PLIOCENE-QUATERNARY GEOLOGY OF NORTHERN NEW JERSEY

60th Annual Reunion of the Northeastern Friends of the Pleistocene

Southern entrance of the Delaware Water Gap.

May 30-June 1, 1997
Ledgewood, NJ

Hosts:
New Jersey Geological Survey
Rutgers University

Guidebook for the 60th Annual Reunion
of the
Northeastern Friends of the Pleistocene

PLIOCENE-QUATERNARY GEOLOGY OF NORTHERN NEW JERSEY

Leaders: Scott D. Stanford, New Jersey Geological Survey
and Rutgers University
Ron W. Witte, New Jersey Geological Survey

Contributors: Gail M. Ashley, Rutgers University
Jack B. Epstein, United States Geological Survey
Richard K. Shaw, Natural Resources Conservation Service
John Wright, National Park Service

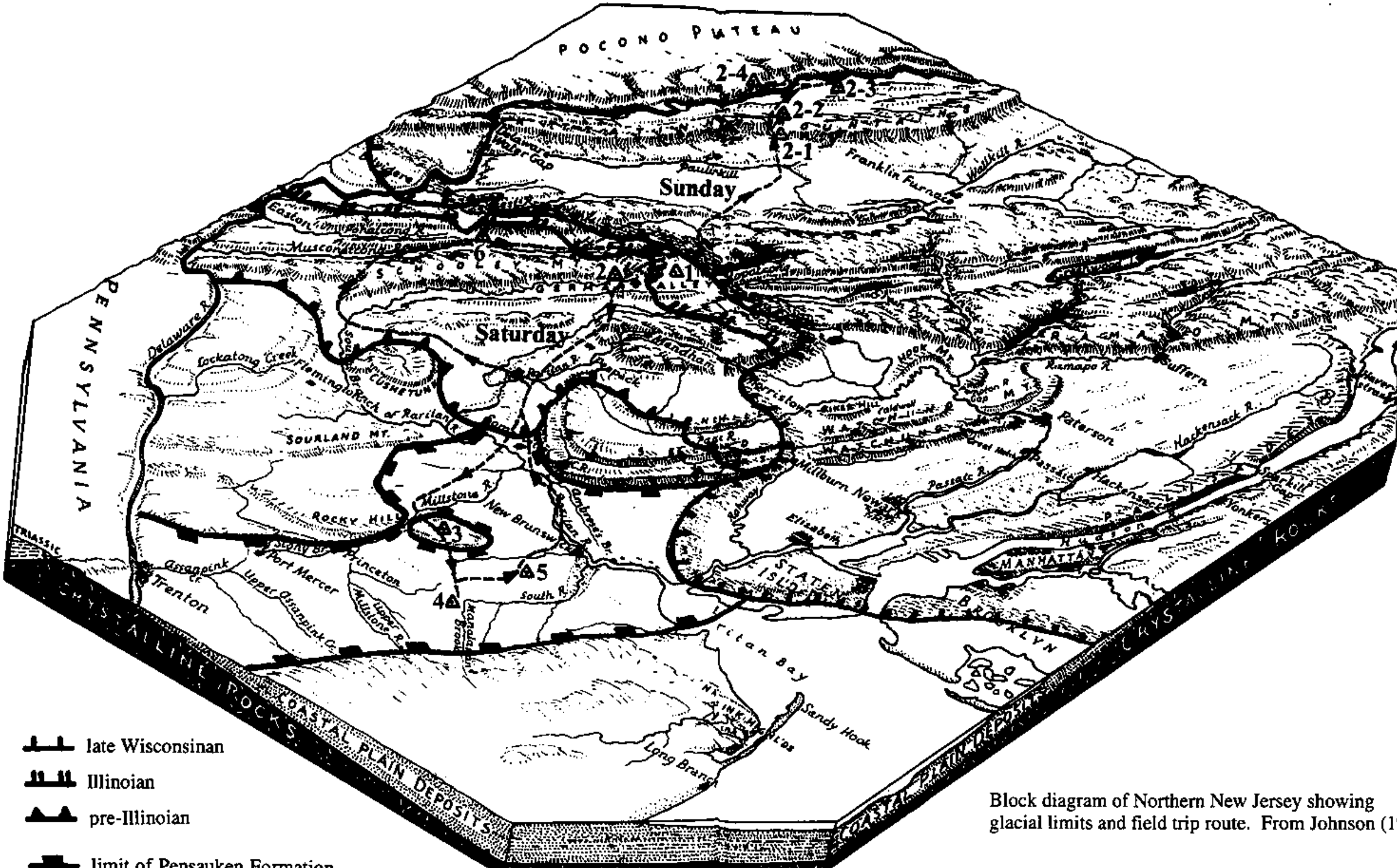
May 30-June 1, 1997

Days Inn
Ledgewood, NJ

Hosts:
New Jersey Geological Survey
Rutgers University

Guidebooks available from: Scott Stanford
New Jersey Geological Survey
CN 427
Trenton, NJ 08625-0427
609-292-2576
e mail: scotts@njgs.dep.state.nj.us
(available for cost of reproduction, while supplies last)

Cover illustration: 1844 engraving of the Delaware Water Gap, by Henry Howe.



Block diagram of Northern New Jersey showing glacial limits and field trip route. From Johnson (1931).

FOREWORD

The Friends have visited New Jersey on three previous occasions. In 1939 the Friends, led by Paul MacClintock, Meredith Johnson, and Henry Herpers, examined the Wisconsinan and "Jerseyan" drifts (see the announcement on the following page). In 1949 the Friends, again led by Paul MacClintock (with John Lucke), examined the Pensauken Formation (copies of the assembled notes from this trip are available on request; the first page is reproduced on page iii). In 1985, Ed Evenson, Jack Ridge, Ron Witte, and Scott Stanford examined the late Wisconsinan deglaciation of the Kittatinny Valley. We will revisit many of these issues on this trip, and perhaps engage some new ones. Over the more than 60 years that the Friends have been meeting much has changed in Pleistocene geology. Marine and ice-core records now document ice-volume and climate history in great detail. Isotopic and magnetic dating techniques provide a chronology. Sedimentologic, geomorphic and glaciologic studies yield an understanding of process not available 60 years ago. But good fieldwork is still the foundation for the appropriate application and interpretation of these advances, and the basic character of good fieldwork hasn't changed much since the 19th century. So the insights and methods of the original Friends, and of their predecessors (extending back to the 1870s in New Jersey, and including G. H. Cook, R. D. Salisbury, H. B. Kummel, G. N. Knapp, C. E. Peet, W. M. Davis, and D. W. Johnson) are much the same as those now, even if some of the vocabulary and data have changed.

One of the persistent issues in the Pleistocene geology of northern New Jersey is long-term landscape evolution. Northern New Jersey is a classic locale for studies of this topic, and was the subject of defining works by William Morris Davis and Douglas Johnson. The Schooley and Somerville "peneplains" have type areas in New Jersey. The combination of these surfaces with multiple glacial, fluvial, and marine deposits provides an opportunity to link traditional geomorphology, with its emphasis on geography and physiography, to surficial geology, with its grounding in process, materials, and stratigraphy. Over the past several decades a greater appreciation for the effects of climate, eustasy, and tectonics on landforms, and a greater understanding of the complexity of geomorphic processes, has led to the abandonment of much of this early geomorphologic work. But with improved chronology of the surficial deposits, and a more detailed reconstruction of global eustasy and climate change, the origin of these erosion surfaces can be reexamined. One of the points we hope to illustrate on this fieldtrip is that the present landscape of northern New Jersey is a mosaic of pieces of different age (possibly extending back to the late Miocene), and that some of the surfaces and events described by Davis and Johnson can be reinterpreted in light of new information on local stratigraphy and global climate and eustasy.

ACKNOWLEDGEMENTS

We thank the following property owners who kindly granted access to the field stops: Dr. Tariq Mahmood, Mr. Ed Herbert, and Mr. John Perrucci. We also thank Mr. William Foley, Superintendent of Stokes State Forest, and Mr. Roger Rector, Superintendent of the Delaware Water Gap National Recreation Area, for their kind assistance. This trip and guidebook are based partly on fieldwork conducted between 1983 and 1992 as part of a cooperative geologic mapping project conducted jointly by the New Jersey Geological Survey and the U. S. Geological Survey. We thank our co-workers on this project, Dave Harper (formerly of the N. J. Geological Survey), Byron Stone, the late Jim Owens, and Wayne Newell (all of the U. S. Geological Survey) for their contributions and discussions. Jane Uptegrove, Laura Nicholson, Don Monteverde, Bill Graff, and Ron Pristas (all of the N. J. Geological Survey) generously assisted with logistics for the trip and guidebook. We also thank Haig Kasabach, State Geologist, for providing us the opportunity to prepare this trip.

RECEIVED
JUL 25 1949
Department of Conservation
& Economic Development
Geologic & Top. Survey

ASSEMBLED NOTES FROM THE "FIENDS"

PENSAUKEN FIELD TRIP - MAY

1949

R. F. Flint

Sedimentary Bodies Present

Four Pleistocene (?) sedimentary bodies have been recognized in eastern New Jersey - (1) the Beacon Hill gravel, (2) the Bridgeton formation, (3) the Pensauken formation, and (4) the Cape May formation. Of these, (1), (3), and (4) are distinctive in internal character, but (2) is much like (3), differing from its chiefly in altitude of occurrence. As all occurrences of all four are in erosion remnants, altitude means little; hence Meredith Johnson's suggestion that (2) and (3) are a single sedimentary body is provisionally accepted.

The Beacon Hill gravel is coarser grained and more deeply composed than the Pensauken formation. The Cape May formation is finer grained than the Pensauken, and is little decomposed. Furthermore the Cape May occupies lowlands apparently cut into the Pensauken. Hence, in order of decreasing age, the sedimentary bodies appear to be Beacon Hill, Pensauken, Cape May.

The Cape May is correlated by H. G. Richards with the Pamlico formation, which is (1) interglacial, (2) not younger than Sangamon because of the improbability of so high a sealevel in Wisconsin time, (3) not older than Sangamon because of its fresh topography and slight weathering.

Pensauken formation

Internal character - The Pensauken formation (together with its Bridgeton equivalent(?)) consists of about 90 percent sand, with some pebble gravel and a few boulders. The gravel is notably siliceous - cherts, quartzites, quartzes, suggesting long-term reworking of material previously selected by stream action and deep weathering. However, it includes a few fresh gneissic rock types. It generally (though not universally) exhibits cut-and-fill stratification, with preferred directions of foresetting toward the south and southwest. Average grainsize diminishes in the same directions. No clear-cut unconformities that might be more important than diastems have been noted; yet variations in grainsize and stratification are so abrupt that significant alternations between filling and cutting are suggested. However, none has been proved. The formation is at least 100 feet thick. Similarity to the Beach Hill gravel and the presence of pebbles of ferruginous Beacon Hill conglomerate, suggest extensive reworking of Beacon Hill material.

The boulders are mostly quartzites and mostly small, but they are numerous (on the surface; rarely are they seen actually imbedded in Pensauken sediments. Many exhibit percussion fractures and a few are said to exhibit striations. Nothing is known about their vertical distribution within the sediments.

The Pensauken contains fossil plants said to imply a warm climate.

CONTENTS

Figures are numbered consecutively within chapters. References are provided at the end of each chapter.

CHAPTER 1

Pliocene-Quaternary Geology of Northern New Jersey: An Overview

Scott D. Stanford
New Jersey Geological Survey
CN 427
Trenton, NJ 08625

INTRODUCTION

Northern New Jersey includes the terminal positions of pre-Illinoian, Illinoian, and late Wisconsinan glaciations, multiple fluvial and colluvial deposits outside the glacial border, and extensive rock-cut erosion surfaces, including the type areas of the Schooley and Somerville peneplains. It also includes the northernmost exposed portion of the unglaciated Atlantic Coastal Plain, with a marine stratigraphy extending into the middle Miocene. The conjunction of glacial termini and unglaciated terrain subject to continuous stream and slope erosion with proximity to the Atlantic offers a unique opportunity to evaluate and date the competing effects of eustatic baselevel change, glaciation, and climate cycling on a passive-margin landscape. The following overview will briefly describe the history of surficial geologic investigations in northern New Jersey; the general lithologic, stratigraphic, and weathering characteristics of the surficial deposits; and their associated erosional landforms.

HISTORY OF INVESTIGATIONS

The earliest work on the surficial geology of northern New Jersey was by Cook (1880) who traced the late Wisconsinan terminal moraine and identified many glacial-lake features. This was soon followed by the systematic mapping and stratigraphic studies of R. D. Salisbury, H. B. Kummel, C. E. Peet, and G. N. Knapp. This work was partially published in map form in various USGS folios (Merrill et al., 1902; Darton et al., 1908; Spencer et al., 1908; Bascom et al., 1909; Bayley et al., 1914) and described narratively in Salisbury (1902) and Salisbury and Knapp (1917). Concurrent with, and following, the mapping work, were the geomorphic studies of Davis and Wood (1889), Davis (1890), Salisbury (1898), Johnson (1931), and Campbell and Bascom (1933). Topical studies building on this framework were conducted sporadically for the next several decades. MacClintock (1940, 1954) studied weathering characteristics of the glacial deposits and MacClintock and Richards (1936) continued stratigraphic studies of glacial and interglacial deposits. Ward (1938), Peltier (1959), and Crowl (1971) mapped and correlated Delaware River terrace deposits. Bowman (1966), Owens and Minard (1979) and Martino (1981) studied the petrography, sedimentology, and stratigraphic correlation of the Pensauken Formation. Minard (1961) and Minard and Rhodehamel (1969) describe some additional glacial features, including moraines on Kittatinny Mountain, and Coastal Plain terrace deposits. Herpers (1961) traced the Ogdensburg-Culvers Gap moraine. Connally and Sirkin (1973) reconstructed late Wisconsinan recessional ice margins as part of a regional study of the Hudson-Champlain lobe. Reimer (1984) and Von Schondorf (1987) studied sedimentologic aspects of the late Wisconsinan meltwater deposits. Wolfe (1953) and Walters (1978) describe periglacial features in the Coastal Plain and in the New Brunswick area. Tedrow (1986) describes the soils of the state.

The late Pleistocene and Holocene pollen record has been studied in a number of cored lakes and swamps at and north of the terminal moraine (Heusser, 1963; Harmon, 1968; Nicholas, 1968; Meyerson, 1971; Sirkin and Minard, 1972; Russell, 1980; Cotter et al., 1986; Peteet et al., 1990), and at a few sites south of the glacial border, including thermokarst ponds and Coastal Plain swamps (Sirkin et al., 1970; Watts, 1979) and an upland swamp dammed by colluvium (Russell, in progress).

A renewed round of detailed mapping in the 1980s used the morphosequence approach of Koteff and Pessl (1981) to interpret the late Wisconsinan meltwater deposits. This led to the delineation of previously unknown recessional ice-margin positions, glacial lake basins and stages, valley-fill stratigraphies, and fluvial and lake drainage histories. Harper (1982) initiated this approach at 1:250,000 scale. Ridge (1983) and Witte (1988) used this approach at 1:24,000 scale to establish the late Wisconsinan deglaciation history from the terminal moraine to the Ogdensburg-Culvers Gap moraine in the Delaware and Kittatinny valleys. Stone et al. (1989), Stanford and Harper (1991), and Stanford (1993a) do the same for the Highlands and Newark Basin to the east. Mapping of colluvial and fluvial deposits south of the late Wisconsinan glacial limit has also led to some conclusions about preglacial and periglacial geomorphology (Ridge et al., 1992; Stanford, 1993b). A cooperative geologic mapping effort by the NJGS and USGS from 1984-1990, initiated by the State of New Jersey to provide new statewide surficial and bedrock geologic maps for groundwater management, has produced nearly complete 1:24,000-scale surficial geologic map coverage for the northern half of the state. These maps include rock-surface contours and subsurface stratigraphy inferred from logs of water wells and test borings. The 1:24,000 work was compiled at 1:100,000 for a formal stratigraphic map (Stone et al., in press), which uses formation nomenclature for tills and meltwater sediments, and for a simplified glacial-sediment map emphasizing hydrogeologic properties and valley-fill aquifer structure (Stanford et al., 1990). Twenty-six out of 79 mapped quadrangles are currently published.

BEDROCK GEOLOGY AND PHYSIOGRAPHY

Northern New Jersey includes parts of four physiographic provinces (fig. 1-1): the Valley and Ridge, formed on tightly folded and faulted Paleozoic sedimentary rocks; the New Jersey Highlands, formed on isoclinally folded Proterozoic granite, gneiss, and marble, with downfaulted inliers of Paleozoic sedimentary rock; the Newark Basin, formed on gently folded and normally faulted Triassic and Jurassic sedimentary and igneous rocks; and the Coastal Plain, formed on very gently dipping unconsolidated sediments of Cretaceous through Miocene age. The Valley and Ridge includes Cambrian and Ordovician dolomite and limestone of the Kittatinny Supergroup and Jacksonburg Formation, which floor valley bottoms within Kittatinny Valley (fig. 1-2); shale, slate, siltstone, and sandstone of the Ordovician Martinsburg Formation, which underlies an intermediate upland within Kittatinny Valley, and quartzite of the Silurian Shawangunk Formation, which, along with siltstone and sandstone of the Bloomsburg Redbeds, underlies the prominent Kittatinny Ridge. Thin, interbedded carbonate, shale, and siltstone of Silurian and Devonian age floor the Delaware Valley (locally termed the Minisink Valley) northwest of Kittatinny Mountain.

The Highlands include plateau-like uplands and ridges underlain by gneiss and granite separated by long, northeast-southwest trending valleys floored by marble and downfaulted Paleozoic carbonate rock and shale equivalent to the rocks of the Valley and Ridge. In the Green Pond Outlier, a belt of Paleozoic rock that nearly bisects the Highlands (fig. 1-2), the sedimentary rocks also include Silurian and Devonian quartzites that form long prominent ridges. Narrow, ravine-like valleys form where streams cut through the gneiss uplands.

The Newark Basin includes shale, siltstone, sandstone, and some conglomerate of the Stockton, Lockatong, Passaic, Feltville, Towaco, and Boonton Formations of Triassic and Jurassic age, with some interbedded Jurassic diabase sills and dikes and basalt flows. The rocks generally dip 5 to 15 degrees to the northwest, although they are locally deformed into broad open folds. Shale and arkosic sandstone underlie broad lowlands, including the Raritan, Hackensack, and Passaic lowlands; mudstones and quartz sandstones underlie uplands, including the Hunterdon Plateau; and diabase, basalt, and adjoining hornfels underlie the prominent ridges of the Watchung Mountains, Palisades, Rocky Hill, and Sourland Mountain (fig. 1-2).

The Newark Basin is onlapped to the southeast by unconsolidated sand, silt, and clay marginal marine and shelf deposits of Cretaceous through Miocene age that are divided into 17 formations. These sediments dip very gently (5-50 feet/mile) to the southeast. Most of the Coastal Plain is an area of low relief with broad valleys and subdued interfluves. Locally, residual hills and uplands such as the Mount Pleasant and Clarksburg Hills, capped by quartz gravel and iron-cemented glauconitic sand, rise 100-200 feet above the lowlands (fig. 1-2).

SURFICIAL GEOLOGY AND LANDFORMS

Figure 1-3 is a schematic southeast-northwest section from the Coastal Plain inland across the Newark Basin and Highlands to a glacial valley fill along the terminal moraine. Age controls on the deposits are indicated. The following sections will describe these deposits and landforms, and associated events, from oldest to youngest.

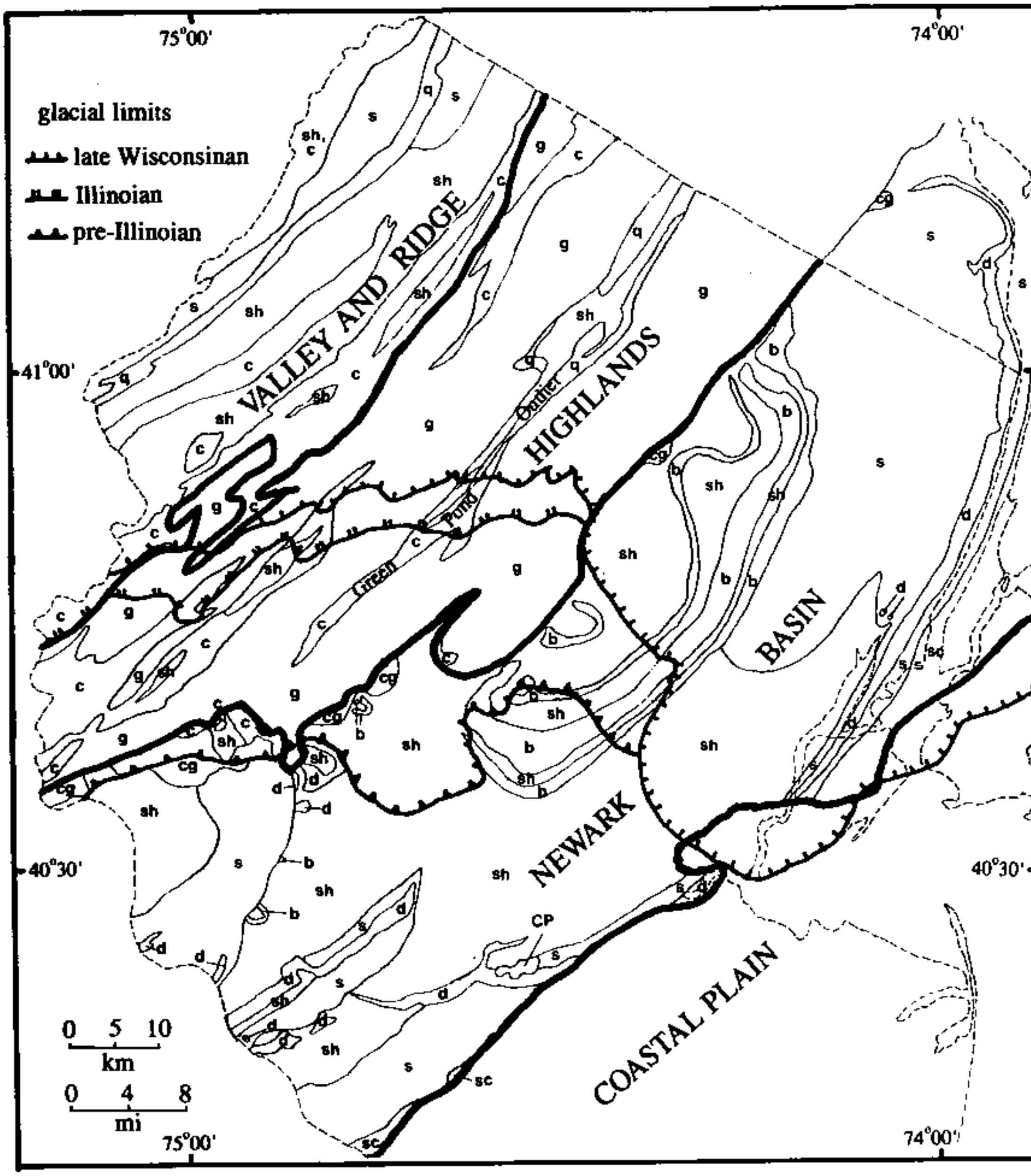


Figure 1-1. Bedrock lithologies of northern New Jersey. Abbreviations are: b=basalt, c=carbonate rock, cg=conglomerate, CP=Coastal Plain sediments, d=diabase, g=gneiss, q=quartzite, s=sandstone and mudstone, sh=shale, sc=schist. Valley and Ridge rocks are Cambrian through Devonian in age; the Highlands 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.

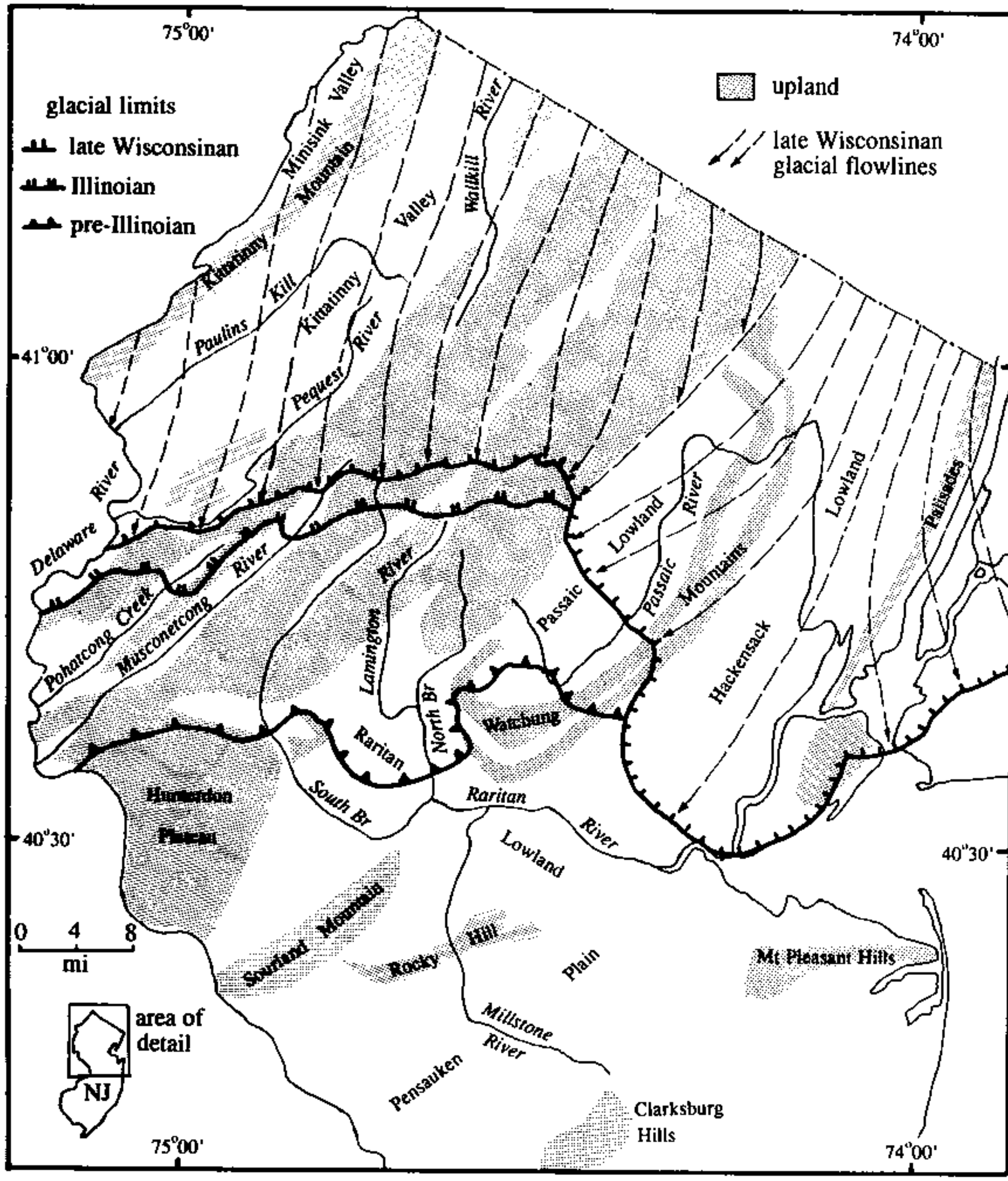


Figure 1-2. Major physiographic features of northern New Jersey, with glacial limits and flowlines of the late Wisconsinan glacial advance. Flowlines are based on drumlin axes, erratic dispersion in till, and several hundred ridgetop and upland striations (Stone et al., in press).

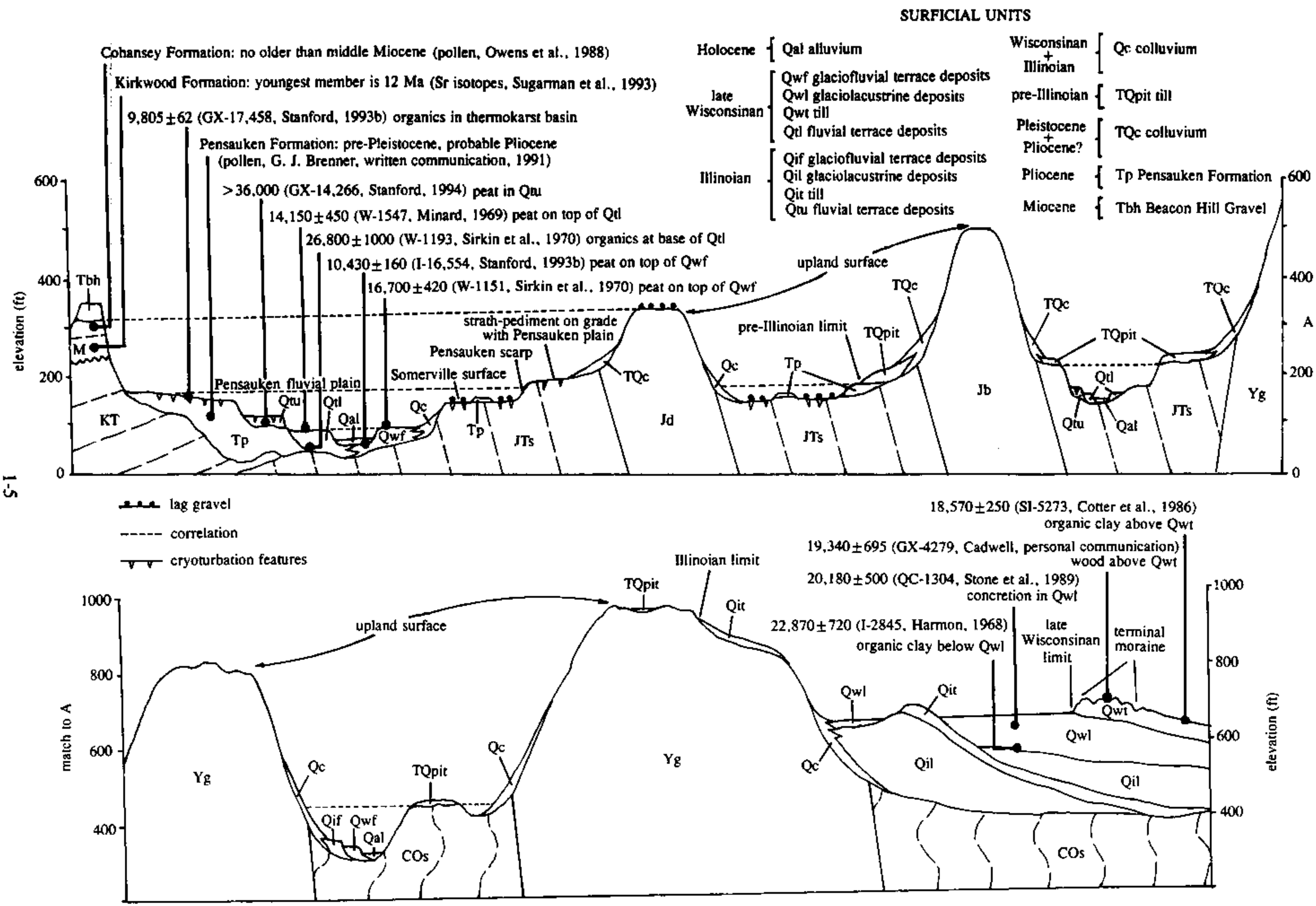


Figure 1-3. Schematic section on a southeast-northwest line from the inner Coastal Plain through the Newark Basin and Highlands to a glacial valley fill along the late Wisconsinan terminal moraine. Age control is shown in the appropriate stratigraphic position. Vertical scale is approximate; no horizontal scale implied. Note scale change at match line. Abbreviations for bedrock units are: M=Miocene Coastal Plain formations, KT=Cretaceous through Eocene Coastal Plain formations, JTs=Triassic-Jurassic shale and sandstone, Jd=Jurassic diabase, Jb=Jurassic basalt, Yg=Proterozoic gneiss, COs=Cambrian and Ordovician dolomite and limestone.

Beacon Hill Gravel and the Upland Erosion Surface

The Beacon Hill Gravel (Salisbury and Knapp, 1917) is the oldest surficial deposit in New Jersey. It is a quartz-quartzite-chert pebble gravel with a quartz sand matrix, generally less than 20 feet thick, that forms mesa-like residual caps on all of the highest hills of the Coastal Plain. Quartz and quartzite clasts are nondecomposed but generally have thin weathering rinds and are easily broken with a hammer due to incipient weathering along crystal boundaries. Many of the chert clasts are fully or partially decomposed to clay. Where soils are preserved, they show evidence of laterization and rubification, suggesting both a long period of soil development and formation in a warm paleoclimate (Trela, 1984).

The base of the Beacon Hill is at 320-340 feet in the Mount Pleasant Hills, and at about 330-340 feet in the Clarksburg Hills. Pebbles and cobbles of quartzite and chert like those in the Beacon Hill sit on an upland flat cut into diabase bedrock at an elevation of 310-330 feet on Rocky Hill near Princeton (fig. 1-4), which is separated from the Clarksburg Hills by the Pensauken Valley of Pliocene age (see next section). The similar elevations of these widely separated locations, the mesa-like rather than fan- or channel-like form of the remnants, and the absence of any intervening or adjoining uplands exceeding these elevations, suggest that they are remnants of a broad, continuous, low-relief plain rather than local inverted fan or channel gravels or let-down lags.

The youngest formation cut by the Beacon Hill is the Cohansey Sand, a marginal marine quartz sand, which is of probable middle Miocene age based on pollen (Owens et al., 1988) and strontium stable isotope ages of the underlying Kirkwood Formation (Sugarman et al., 1993). The Pensauken Formation, of probable Pliocene age (see next section), forms a broad plain inset 200 feet below the Beacon Hill. The unconformable relationship to the middle Miocene Cohansey, and the depth and extent of dissection between the Beacon Hill and the Pensauken, suggests a late Miocene age for the Beacon Hill.

The elevation of the Beacon Hill corresponds approximately to the level defined by the upland erosion surface that truncates the rocks of northern New Jersey. This surface, the Schooley peneplain of Davis and Wood (1889), as revised by Johnson (1931), rises unevenly from about 320 feet on Rocky Hill, where it is capped by the previously-mentioned Beacon Hill-like lag, to about 1000-1100 feet in its type area on Schooleys Mountain (fig. 1-4). Although locally it is quite flat, especially on mudstone on the Hunterdon Plateau, on diabase and gneiss the surface shows up to 100 or 150 feet of gentle relief, with resistant rock holding up low knolls and ridges. If this is indeed a single relict erosion surface then its present elevation may reflect both an original topographic rise and increase in relief to the northwest, and later isostatic warping caused by continental margin sedimentation. If it is contemporaneous with the Beacon Hill Gravel, then it reached its final, pre-dissected form in the late Miocene.

Two sets of aligned wind gaps across Kittatinny Mountain and the Highlands, one from the Delaware Water Gap through wind gaps at Oxford and Glen Gardner, the other from Culvers Gap through a gap from Andover to Ledgewood, may mark southeasterly drainage on this surface (fig. 1-4; see also chapter on Culvers Gap later in guidebook). These gaps are inset several hundred feet below the Kittatinny ridgecrest and the upland level of the Highlands, but are approximately on grade with the upland surface in the Newark Basin. If contemporaneous, these features suggest that gneiss and quartzite stood out with several hundred feet of relief inland from a low-relief plain across the Piedmont and Coastal Plain.

The supermature quartzite-chert lithology of the Beacon Hill Gravel may support the idea of a low-relief inland source area. It indicates derivation from the polycyclic quartz gravel common in the Cohansey Formation, which likely extended further inland before dissection, and from outcropping Shawangunk Formation quartzite and residual chert weathered from Paleozoic carbonate rocks. The complete absence of less-resistant rock types suggests that these units either were not exposed (perhaps they were covered by the Cohansey or inland fluvial correlates of the Cohansey) or were deeply weathered to fine-grained regolith during Beacon Hill time.

The upland erosion surface is best preserved on the Hunterdon Plateau, which, unlike the type area on Schooleys Mountain, was never glaciated. The Plateau (fig. 1-5) is a flat to very gently sloping surface covering about 70 square miles. It is cut into shale, mudstone, and argillite of the Passaic and Lockatong Formations, which dip westward at 5 to 10 degrees. The Passaic rocks include alluvial plain and lacustrine beds composed of quartz (10 to 30%), illite, chlorite, Na and K feldspar (less than 15%), and hematite (Van Houten, 1969). The Lockatong is

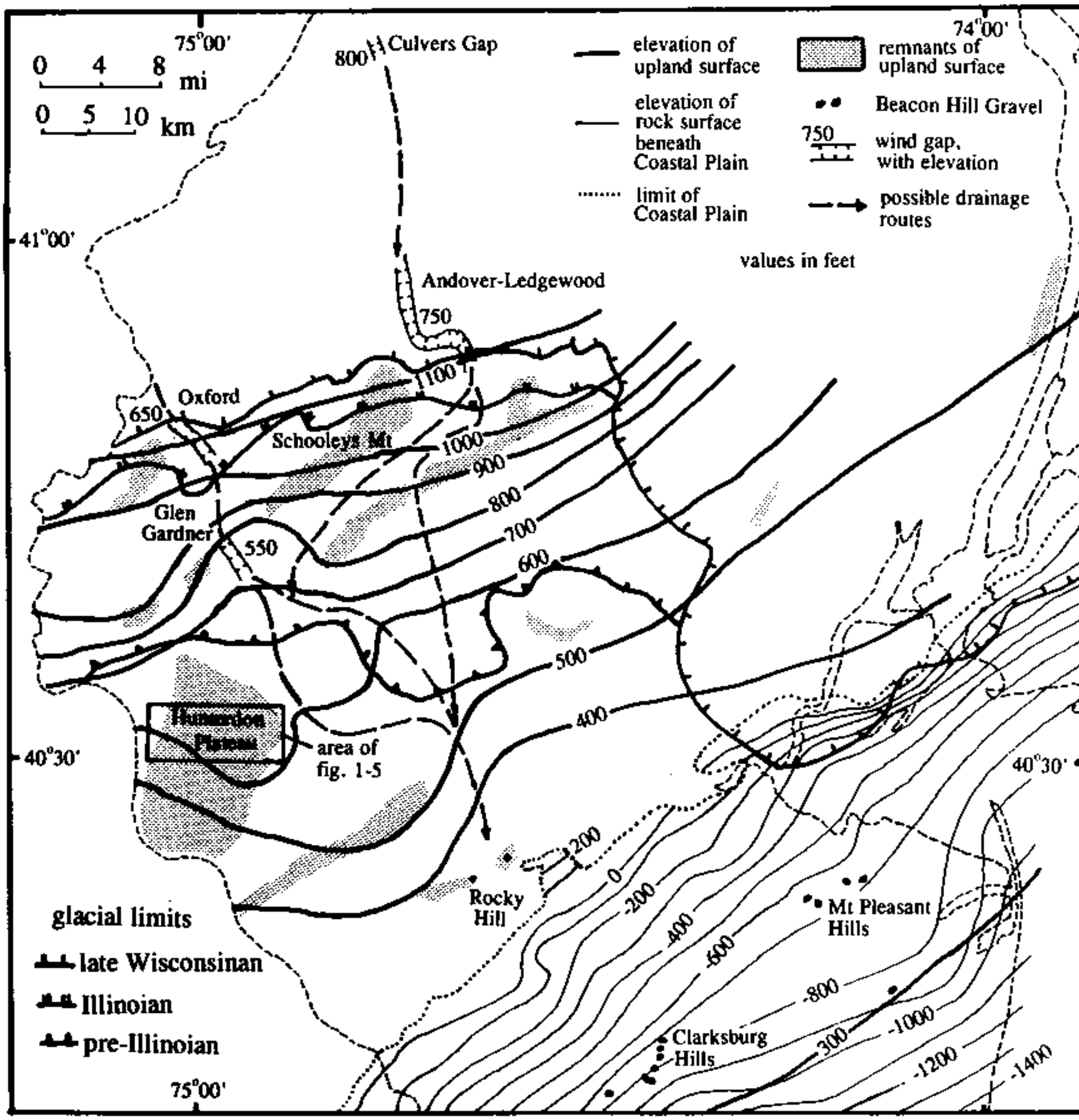
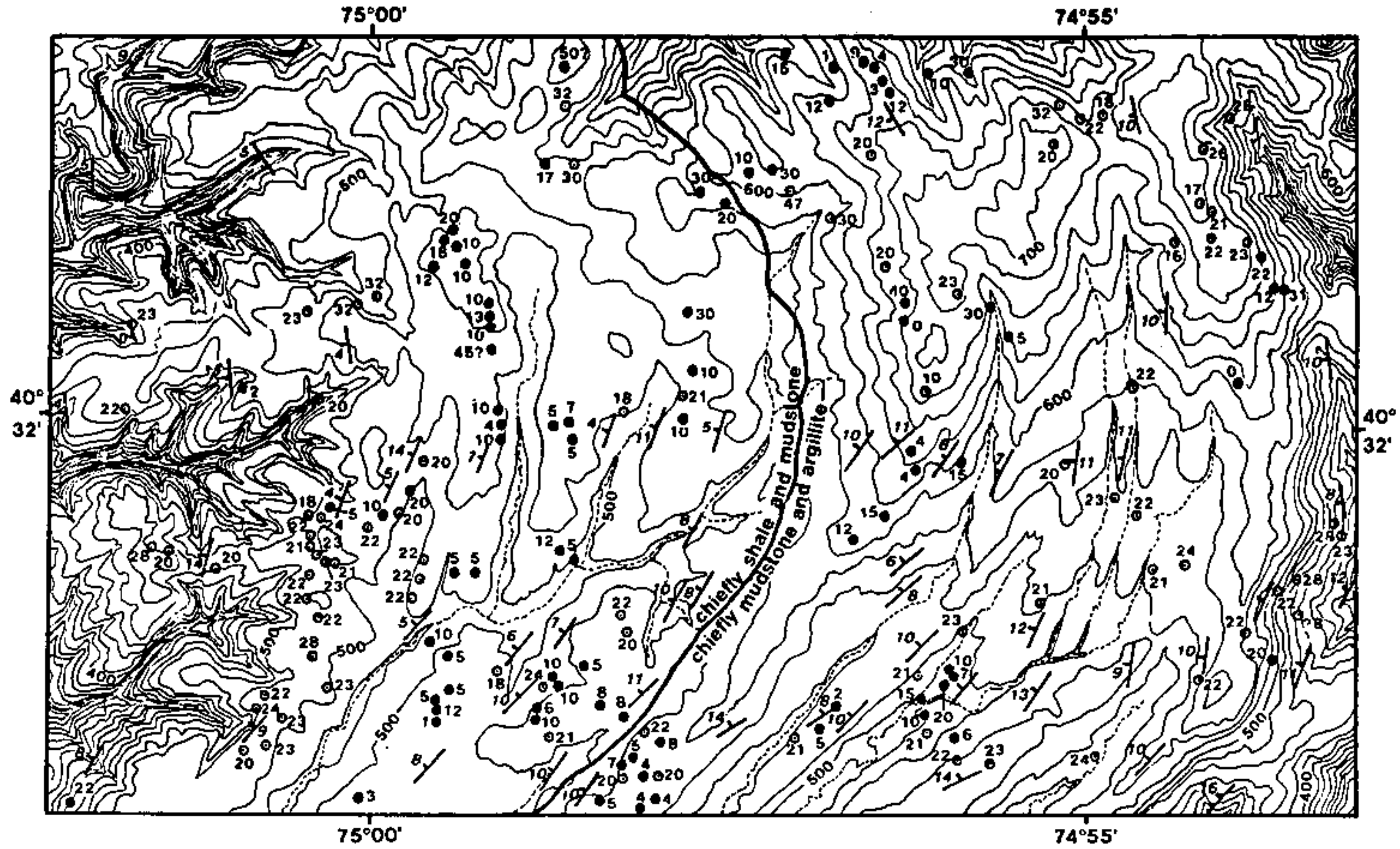


Figure 1-4. Smoothed elevation contours on the upland surface-Beacon Hill Gravel (heavy lines) and elevation of the rock surface beneath the Coastal Plain (light lines, from Gill and Farlekas [1976]). The Coastal Plain basement surface dips more steeply than the upland surface, indicating a significant amount of bedrock denudation after stripping of any Cretaceous sediment, as argued by Johnson (1931) in his revision of Davis (1890). Dashed lines show possible drainage routes marked by wind gap alignments. This drainage may be the same age or slightly younger than the Beacon Hill Gravel.



strike and dip of bedding

bedrock contact

thickness of regolith in water-well

length of well casing, indicating maximum thickness of regolith

0 1
mile

Figure 1-5. Modern topography and regolith thickness on part of the Hunterdon Plateau, an unglaciated part of the upland erosion surface. Location shown on fig. 1-4. Uniformly thin regolith indicates the plateau is a true rock-cut surface without thick colluvial fills or saprolite. Structural data from Herman et al. (1992).

lacustrine and includes detrital beds composed of abundant Na feldspar with some K feldspar, illite, muscovite, chlorite, and calcite, and a little quartz; alternating with evaporite beds composed of as much as 35-40% analcime (a hydrated Na silicate), with albite, dolomite, calcite, illite, and chlorite (Van Houten, 1969). These rocks weather to a yellow to reddish brown clayey silt with varying amounts of weathered chips, flagstones, and blocks of shale and argillite. Erratic clasts are absent, except for occasional archeologic chert. Soils in this weathered-rock regolith include the Croton soil, a Typic Fragiaqualf consisting of quartz, albite, and muscovite (but no analcime) in the silt fraction and kaolinite, illite, and montmorillonite in the clay fraction, that forms chiefly on poorly-drained argillite and shale regolith; and the Quakertown soil, a Typic Hapludult consisting of altered mica clays, kaolinite, vermiculite, and gibbsite in the clay fraction, that forms on better-drained siltstone regolith (Tedrow, 1986).

Exposures and well logs (fig. 1-5) indicate that this material is generally less than 15 feet thick, and rarely exceeds 20 feet. The surface is therefore a true low-relief rock-cut surface, not a flat formed by colluvial filling of valleys or by volume loss and compaction of a thick saprolite. The Plateau today, because of the shallow, clayey soil and low-permeability bedrock, is poorly-drained. Drainage density is low and standing water is common after heavy rains. Planation mechanisms in this current upland setting are obscure. Certainly solifluction, sheetwash, and creep were active on all slopes under periglacial conditions but the absence of colluvial fills indicates that these processes did not redistribute more than a few feet of material, not the tens of feet required to produce the observed planation. It is possible that the gross form of this surface is relict and was largely planed during an extended period of stable baselevel in the middle to late Miocene. Baselevel stability is likely before the sustained middle Miocene eustatic drawdown. With the Coastal Plain undissected at this time, shorter-term or lesser-magnitude drawdowns may not have been sufficient to allow for nickpoint migration through the Coastal Plain, thus maintaining stable baselevel in the Piedmont.

The absence of fluvial clasts suggests that stream erosion was not the primary planation mechanism on the Plateau. Given the low permeability of the rock and regolith perhaps groundwater-related hillslope processes like piping, spring-sapping, and rilling or sheetwash were important after slope angles had declined below the threshold for mass transport.

Given the fragmentary record, projection of the upland surface inland to the narrow quartzite ridges of the Valley and Ridge is speculative, especially given the likely intensity of periglacial erosion on these ridges (Braun, 1989). However, the regional extent of the surface and of the Beacon Hill Gravel, and the global nature of the eustatic event it records, does suggest the possibility of regional correlation, at least in areas bordering the Coastal Plain.

Pensauken Formation and the Somerville Erosion Surface

An extended period of dissection followed deposition of the Beacon Hill Gravel. This dissection was likely initiated by the middle Miocene growth of Antarctic ice (fig. 1-6). It led to the development of a major river system draining southwestward along the Coastal Plain-Newark Basin onlap zone, separated from the Atlantic by a Coastal Plain upland to the southeast. This river system incised broad valleys and lowlands as much as 200 feet below the Beacon Hill level and deposited the Pensauken Formation. The Pensauken is a yellow arkosic sand and pebbly sand, with a basal cobble gravel in places, dominated by tabular, planar cross beds suggesting a shallow, braided channel network (Martino, 1981). It is as much as 140 feet thick in the narrow thalweg of the paleovalley but is more commonly less than 40 feet thick. Gravel composition is, like the Beacon Hill, dominantly quartz, quartzite, and chert; with some mudstone, sandstone, gneiss, diabase, and basalt in the basal gravel. The quartz and chert clasts are iron-stained but generally unweathered, except for some decomposition of chert in poorly-drained locations, but the other clasts are generally fully decomposed as is the feldspar in the sand fraction. Glauconite occurs in the sand fraction along the southeastern edge of the deposit, reflecting contribution from tributaries eroding glauconite-bearing Coastal Plain formations to the southeast. Truncated red Bt horizons as much as 6 feet thick are preserved on scattered erosional remnants of the original aggradation surface of the deposit, generally at altitudes of 150-160 feet in the northern Coastal Plain.

Paleocurrent measurements in the Pensauken show a consistent and strong southwest paleoflow (fig. 1-7). The deposit extends down the Delaware Valley south of Trenton and is the surface deposit over most of the northern Delmarva Peninsula (Owens and Minard, 1979), where it is referred to as the Columbia Formation in Delaware

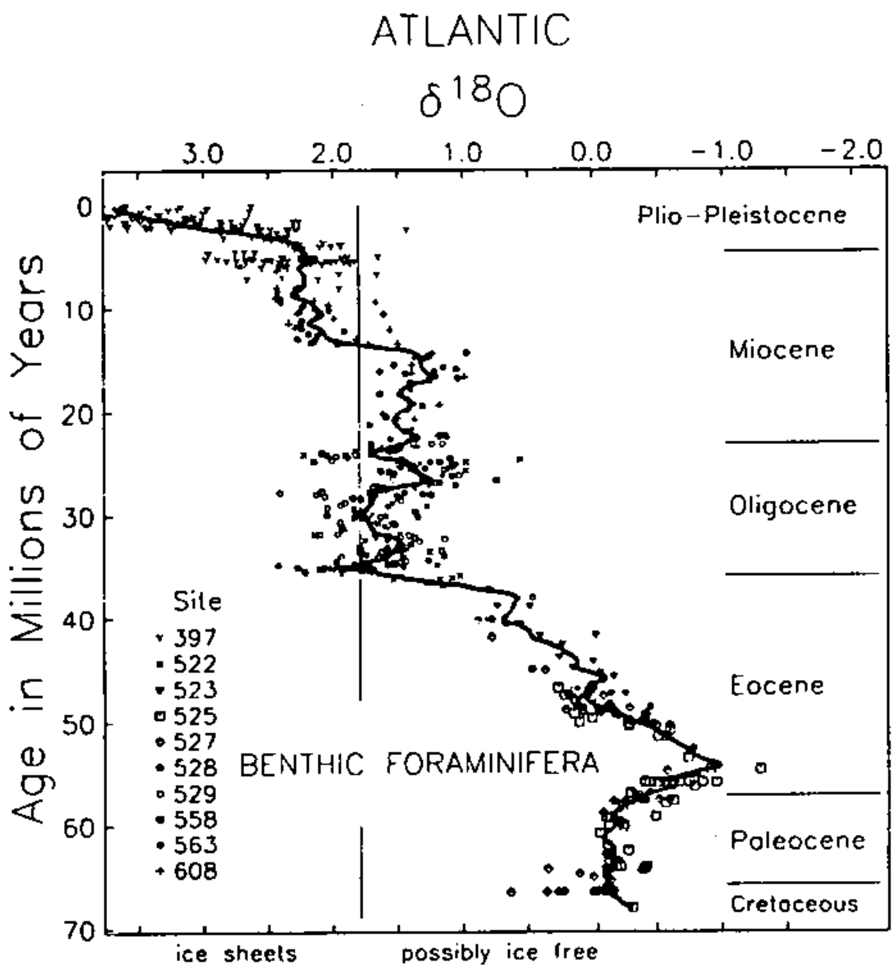


Figure 1-6. Composited and smoothed Tertiary oxygen-isotope record from benthic forams from Atlantic DSDP sites. The sustained enrichments in the middle Miocene and Pliocene-Pleistocene represent growth of Antarctic and northern hemisphere ice sheets, respectively. The resulting long-term eustatic declines may have initiated the two main periods of dissection observed in the northern New Jersey landscape. From Miller et al. (1987).

(Jordan, 1964). North of the late Wisconsinan limit it is glacially eroded but Pensauken gravel clasts are common in till in the Newark Basin north to the New York state line, and remnants of the Pensauken locally underlie the till for several miles north of the limit. The base-of-Pensauken contours, when projected north of the late Wisconsinan limit, intersect the floors of paired wind gaps through the double Watchung basalt ridges at Millburn and Paterson, and the wind gap at Sparkill, New York, through the Palisades diabase ridge (Fig. 1-7) (Stanford, 1993b). The main axis of the Pensauken valley projects across a broad low in the Palisades between Jersey City and uplands on Staten Island. Johnson (1931) interpreted the gaps at Sparkill, Paterson, and Millburn as marking a superposed course of the Hudson River established on the Schooley surface, but he did not tie the gaps into a depositional stratigraphy. The accordance of the gaps with the Pensauken suggests that they mark the route of the Hudson, perhaps superposed from the Beacon Hill fluvial plain, as a tributary to the Pensauken, with the trunk Pensauken river flowing southwesterly through the broad gap in the Palisades from the area of the Long Island Sound lowland. This trunk stream likely included drainage from much of southern New England. The dominance of polycyclic quartz and chert gravel in the Pensauken indicates that it was derived chiefly from earlier Coastal Plain formations that likely extended further inland in southern New England prior to glaciation. These deposits may have included correlates to the quartzose sand and gravel of the Cohansey and Beacon Hill.

Another site showing evidence of superposition is the Rocky Hill Gap (fig. 1-7), where a south-flowing tributary to the Pensauken cut a gap through a diabase ridge. The floor of this gap has a lag of quartz pebbles and falls on the base-of-Pensauken contours. An inner gorge has been cut below the wind gap level by the north-flowing Millstone River, which established its course after diversion of the Pensauken. The upland surface on the diabase ridge above the wind gap has a quartz-chert pebble-to-cobble lag, providing direct evidence of a former Beacon Hill cover over the site of the gap.

The base-of-Pensauken contours also indicate tributaries entering the main valley at Somerville and Jamesburg (fig. 1-7). The deposit around Jamesburg is glauconitic, indicating a local Coastal Plain tributary, and the deposits near Somerville are somewhat enriched in chert, suggesting supply from the chert-bearing carbonate rocks of the Kittatinny Valley or, possibly, the Green Pond Outlier. The wind gap alignments of Culvers Gap-Andover-Ledgewood and Delaware Water Gap-Oxford-Glen Gardner, which may mark former superposed drainages across the Highlands during or slightly after Beacon Hill time, may have brought this material into the Raritan lowland, to be reworked later by the Pensauken tributaries. Alternatively, the chert may have come via southwesterly drainage from carbonate rocks in the Green Pond Outlier.

In the Newark Basin the Pensauken was deposited on a broad strath cut into shale of the Passaic Formation. The northwestern edge of the Pensauken in the basin is marked by a 40- to 60-foot high scarp cut into the shale west of the Somerville area and between Princeton and Trenton (fig. 1-7). Today, erosional remnants of the Pensauken in the Basin occur as small isolated mesaform caps at elevations between 140 and 160 feet, which correspond roughly to the top of the rock-cut scarp to the northwest. These remnants are on low hills rising 20 to 40 feet above a broad flat surface that is cut into shale bedrock and veneered with quartz and chert pebbles from the Pensauken. This broad surface is the type area of the Somerville peneplain of Davis and Wood (1889) but the close association of the surface with the Pensauken Formation, as evidenced by the pebble lags, termination against the bordering scarp, and remnants of Pensauken on all hills rising above the surface, indicate that it is likely the exhumed strath beneath the Pensauken. Once valleys incised below the Pensauken/shale contact, spring sapping by groundwater moving laterally along the contact could have gradually stripped away the sand and gravel, exposing the shale surface and leaving a gravel lag. Correlations from this local strath to regional erosion surfaces are therefore suspect.

A broad lowland cut in shale, now incised 60 to 100 feet by modern streams, also extends up the Raritan lowland north and west of the Pensauken plain. In the narrow Delaware Valley, cut through diabase, mudstone, and argillite north of Trenton, there is a discontinuous set of rock-cut terraces, some with quartzite and chert cobbles, that grade to the Pensauken plain at Trenton. These terraces also grade upvalley to the flat interfluves on carbonate rock and, locally, shale, in the Musconetcong and Pohatcong valleys, which are incised 80 to 200 feet by modern streams. These surfaces may be straths and, locally along the junction with bordering uplands, colluvial pediments formed during the extended period of stable baselevel represented by the Pensauken Formation. They were correlated to the Harrisburg peneplain by Johnson (1931) but, again, regional correlation of surfaces that may instead reflect local baselevel stability is suspect.

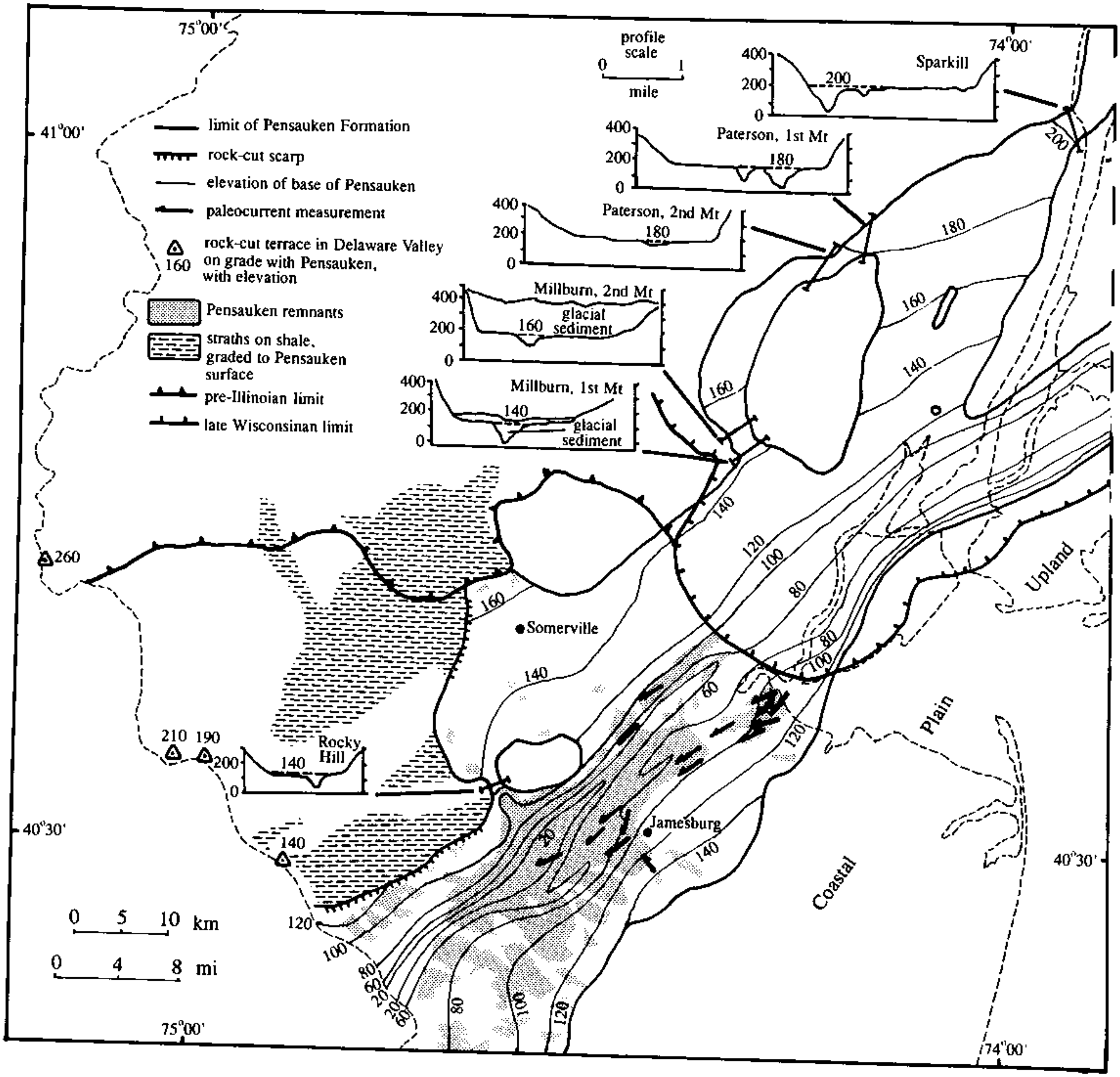


Figure 1-7. Pensauken Formation, straths on grade to the Pensauken surface, and location and profiles of wind gaps on grade to the Pensauken. The Pensauken records a major southwesterly flowing drainage system in the Pliocene that included the Hudson and Delaware as tributaries to a trunk river draining from southern New England. It was likely diverted by the pre-Illinoian glaciation. Note basal contours and paleocurrents indicating tributaries entering the Pensauken valley at Somerville and Jamesburg.

The age of the Pensauken is not firmly established. Owens and Minard (1979) assigned it a late Miocene age based on correlation to subsurface units in the Delmarva Peninsula. Salisbury and Knapp (1917) and Jordan (1964) considered it to be of Quaternary age. Berry and Hawkins (1935) describe warm-temperate plant fossils, including magnolia, sassafras, chestnut, willow, and viburnum leaves from the base of the Pensauken near New Brunswick, that they consider to be early Pleistocene. Pollen from several samples of a black clay bed within the upper part of the Pensauken recovered at two locations near Princeton was analyzed by G. Brenner (written communication, 1991). Pine, oak, and hickory dominate; with hemlock, spruce, fir, and birch dominant in several samples, and the pre-Pleistocene exotics *Engelhardia* and *Pterocarya* present in several samples. The combination of cool temperate climate indicators and exotic pollen suggests a Pliocene age. A Pliocene age is also consistent with the deeply-inset geomorphic relation of the Pensauken with the late Miocene Beacon Hill, and its relation to the pre-Illinoian till, which overlies and erodes the Pensauken north of Somerville. The size and stability of the Pensauken river system, which was likely in place until diversion by valley blockage or capture during the pre-Illinoian glaciation, indicates that the Pensauken may represent a long period of deposition between the late Miocene and late Pliocene or early Pleistocene. The deposit south of Trenton, where the Delaware enters the valley, may have remained active, after the main trunk stream was dislocated from the valley north of there.

Although the Pensauken is largely a braided stream deposit it is probably not of glacial origin, given the temperate flora and pollen it contains, and the abundance of gibbsite and limonite indicative of lateritic weathering (Bowman, 1966; Bowman and Lodding, 1969). Also, the quartz-chert gravel lithology and arkosic quartz sand lithology is distinctly unlike the gneiss-quartzite-sandstone-mudstone gravel and lithic sand dominant in the pre-Illinoian deposits. It may have been deposited primarily during the period of high sea level and warm climate in the early Pliocene, just before major northern hemisphere cooling starting about 2.5 Ma, that is documented by marginal marine deposits such as the Yorktown and Duplin Formations of Virginia and the Carolinas and the Beaverdam Formation of Delaware and southernmost New Jersey (Pazzaglia and Gardner, 1994).

Pre-Illinoian Glaciation and Abandonment of the Pensauken Valley

Pre-Illinoian till, and a few stratified deposits (collectively designated as the Port Murray Formation in Stone et al. [in press]) are preserved in thin (generally less than 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. The distribution of the remnants, which are absent on flat surfaces below the interfluve level within valleys, suggest that the modern valleys were incised anywhere from 80 to as much as 200 feet below the pre-Illinoian valley surface, a feature noted by Salisbury (1902). This is in marked contrast to the Illinoian deposits, which form subdued but recognizable landforms within modern valley bottoms. 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 Newark Basin, 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 lowland (fig. 1-2). 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 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. 1-9) dammed between the pre-Illinoian ice margin and Second Watchung Mountain in the Passaic lowland, with a possible spillway 4 miles to the south at an elevation of about 430-440 feet across a gap in the ridgetop.

The age of this drift is uncertain but it likely correlates to the magnetically reversed drift in central and eastern Pennsylvania (Gardner et al., 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, and the possible association of the pre-Illinoian glaciation to the diversion of the Pensauken River, it may be correlative to the ash-dated late Pliocene pre-Illinoian K till of the midcontinent (fig. 1-8) (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. 1-8), with 40 ka periodicity, which may not be sufficient for extreme Laurentide glaciation, dominant before then back to the initiation of northern hemisphere glaciation at about 2.4 Ma. The exception is a 100 ka periodicity leading up to stages 82 and 78 at 2.03 and 1.94 Ma respectively (Raymo et al., 1989), which corresponds to the 2.01-2.14 Ma range for the 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. 1-9). 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. 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 a finely laminated gray clay which, below 57 feet, is sandy 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 Pensauken valley north of Trenton no longer has a southwesterly flowing stream, but is instead traversed by small westerly to northeasterly-draining tributaries to the Raritan. These streams are inset as much as 100 feet below the Pensauken surface, and they contain Illinoian and late Wisconsinan terrace deposits. This drainage change occurred after the main Pensauken river, including the Hudson and southern New England drainages, was diverted from the valley in the vicinity of New York City. The Illinoian and late Wisconsinan glaciations have eroded and overprinted any deposits or landforms in this area that would record this diversion, but the breach in the Coastal Plain upland between Long Island and New Jersey in the New York Bay-Narrows area, and the refocusing of continental shelf deposition from the Baltimore Canyon area to the Hudson Canyon area between the Pliocene and Pleistocene (Poag and Sevon, 1989) suggest that this was the diversion point. The diversion may have been caused by direct glacial blockage and erosion of the valley, because the pre-Illinoian glacier almost surely covered the area; or by headwater erosion of steepened Atlantic-slope streams through the Coastal Plain upland.

Pleistocene Valley Incision

Following diversion of the Pensauken, local streams established new courses on the now-abandoned Pensauken plain. These include the lower Raritan, which flowed northeasterly across the plain to the New York City area, and the Millstone, which flowed northerly through Rocky Hill Gap to join the lower Raritan. These routes are opposite to the southwesterly paleoflow of the Pensauken. If the Pensauken diversion occurred during the pre-Illinoian glacial maximum then glacioisostatic depression to the north could have levelled or reversed the gradient of the Pensauken plain, which was estimated at .00042 by Martino (1981) on the basis of sedimentologic analysis. Isostatic depressions of .0004 to .0007, sufficient to level or reverse the Pensauken slope, are known from late Wisconsinan glacial lakes (Ridge, 1983; Witte, 1988; Stone et al., 1989; Stanford and Harper, 1991), and similar values are likely for the pre-Illinoian glaciation.

The Delaware entered the Pensauken valley at Trenton and turned southwest to follow that valley to the Delmarva

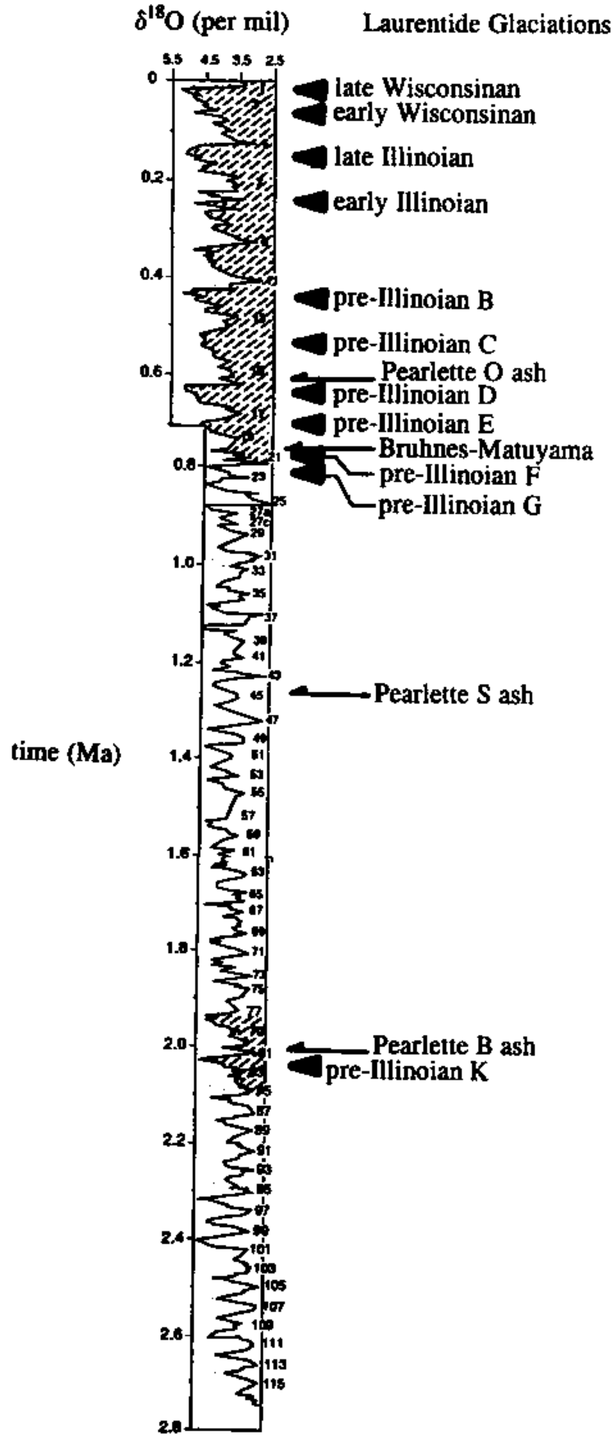


Figure 1-8. Oxygen-isotope record from benthic forams at DSDP site 607 in the North Atlantic and Laurentide glaciations. Ruling indicates 100 ka periodicity possibly marking Laurentide glaciation. Oxygen-isotope curve from Plinth et al. (1992), compiled from Chappell and Shackleton (1986), Ruddiman et al. (1986) and Raymo et al. (1989). Laurentide glaciations and ash stratigraphy from Richmond and Fullerton (1986).

Peninsula. However, abandoned valleys on the Delaware-Millstone divide on the former Pensauken plain southwest of Princeton suggest that the Delaware, too, may have flowed northeastward from Trenton for a time under the influence of pre-Illinoian isostatic depression. This significant added flow would have helped set the lower Raritan and Millstone into their valleys. That such diversion is possible is documented by the late Wisconsinan glaciofluvial terrace deposit in the Delaware Valley, which overtopped this divide and descended the Millstone Valley under the influence of isostatic depression (Stanford, 1993b).

With the new drainage pattern established, streams incised their valleys anywhere from 80 to 200 feet below the Pensauken plain and, northwest of the plain in the Raritan and Delaware basins, below the straths and pediments graded to the plain. Terrace deposits of late Wisconsinan and Illinoian age occur within these valleys, so the incision was accomplished in the early to middle Pleistocene. The Pensauken and pre-Illinoian drift were extensively stripped and dissected, particularly in the Newark Basin, where the Somerville strath surface was exhumed. This incision may reflect, in part, local baselevel lowering due to diversion of the Pensauken on a shorter route to the Atlantic, and sustained eustatic lowering related to the growth of northern hemisphere ice sheets (fig. 1-6).

Illinoian Glaciation

Illinoian till and stratified deposits crop out in a belt in western New Jersey extending about 5 miles south of the late Wisconsinan limit (fig. 1-9). They also occur in the subsurface in valley fills beneath late Wisconsinan deposits south of, beneath, and for short distances north of the late Wisconsinan terminal moraine. Illinoian till also occurs in the cores of some drumlins north to the New York state line. East of the Denville area late Wisconsinan deposits overprint the Illinoian deposits, which continue beneath the late Wisconsinan deposits eastward across the Passaic lowland to the Summit area. Illinoian deposits have not been identified in the subsurface east of the Summit area, although probable Illinoian till (the Bergen Till of Stone et al. [in press]) cores some drumlins in northeastern New Jersey near the New York state line.

In the outcrop belt, Illinoian till derived largely from gneiss (the "Budd Lake Till" of MacClintock [1940]; formally designated as the Flanders Till in Stone et al. [in press]) 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, however, Illinoian fluvial, lacustrine, and moraine deposits form fills up to 150 feet thick, with recognizable delta, moraine, and fluvial-plain landforms. The fluvial and lacustrine deposits are collectively termed the Lamington Formation (Stone et al., in press).

Old soils are stripped or truncated; 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 only a few mm 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 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.

The Illinoian age is assigned based on comparison to the marine oxygen-isotope record, by the local presence of red Sangamon-like soil on the drift, and by correlation 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). However, there is no direct dating of this drift. An early Wisconsinan age is considered untenable based on assumed relations of ice volume and amplitude of the oxygen-isotope record (Ridge et al., 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

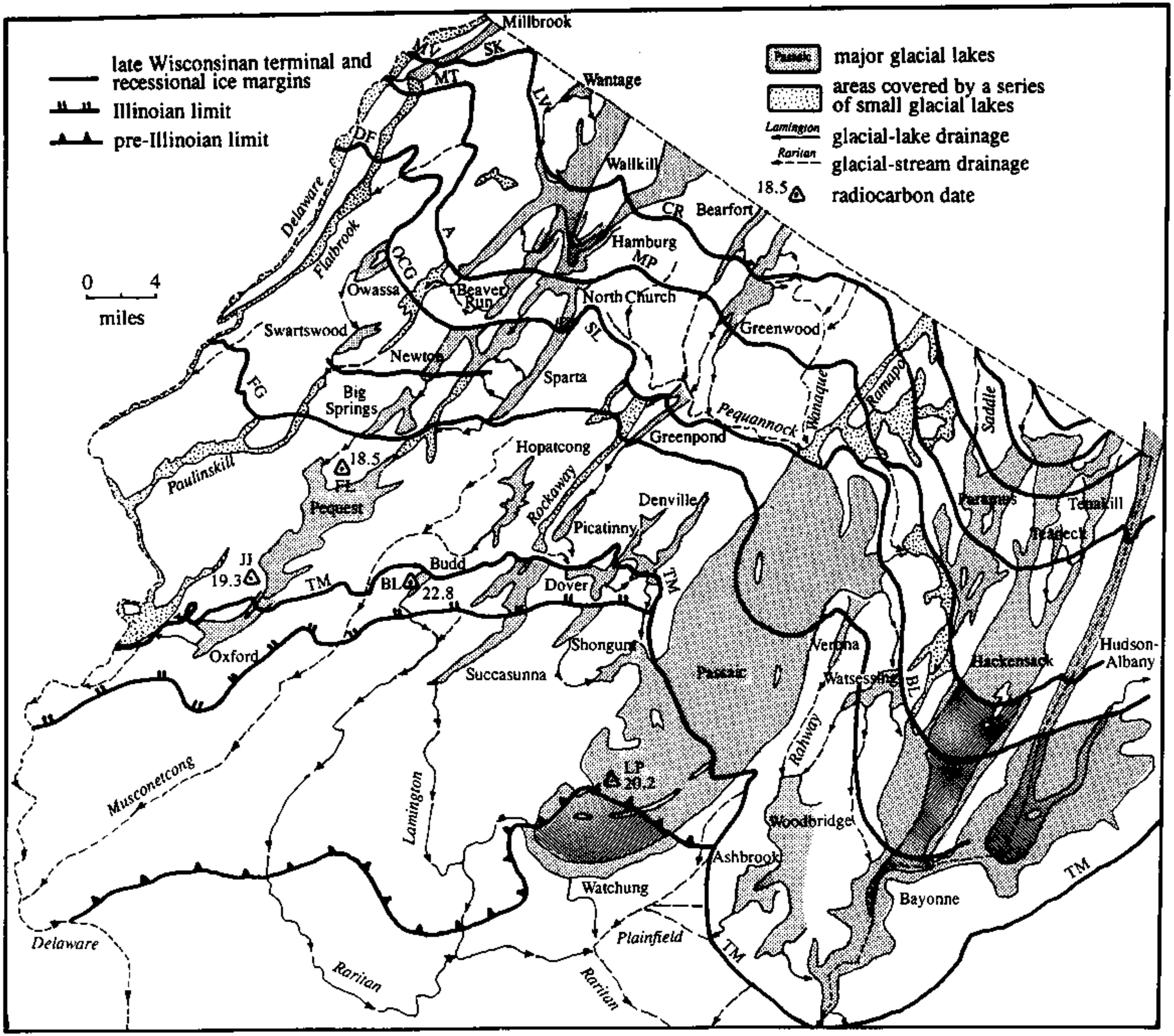


Figure 1-9. Glacial limits, late Wisconsinan recessional ice margins, selected radiocarbon date locations, glacial lakes and meltwater drainage. Numerous minor recessional ice margins and small lakes not shown. Lakes Oxford, Budd, Succasunna, and Passaic likely existed during both the Illinoian and late 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), Stone et al. (1989), Stanford and Harper (1991), Stanford (1993), and Stone et al. (in press).

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.

Outside the Illinoian limit in the Raritan Valley and in the valleys of the northern Coastal Plain there are stream terraces that are higher than the dated late Wisconsinan terraces. These upper terraces are roughly on grade with Illinoian glaciofluvial terraces in the Delaware Valley, show clast weathering similar to the Illinoian drift, support cryoturbation structures (indicating that the terraces were in place and inactive during the last periglacial period) and have yielded dead carbon (fig. 1-3). These features suggest that the terraces are of Illinoian age, and record periglacial alluviation similar to that observed in the late Wisconsinan. On Coastal Plain streams that drain to the Atlantic these terraces seem to grade to the Cape May marine terrace deposits at about 30 feet in elevation, which are of presumed Sangamon age (Ashley et al., 1991; Newell et al., 1995), suggesting that at least the lower, coast-proximal reaches of some of the terraces may be highstand backfills. Such backfilling alluvial plains, however, are not observed today along the coast. Instead, narrow floodplains merge into salt marshes that fill the drowned valleys.

Late Wisconsinan Glaciation

Striations, drumlin axes, and till clast and matrix provenance, indicate that late Wisconsinan ice advanced across northern New Jersey in two main lobes (fig. 1-2): 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 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.

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 rock in the Valley and Ridge also have extensive outcrop where they rise above valley-fill deposits. Locally, however, 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 ramps on slopes facing advancing ice, where it may be as much as 150 feet thick; in drumlins, where it may overlie cores of weathered gneiss (in some drumlins in the Highlands), Illinoian till (statewide), and stratified drift (in some drumlins in the northernmost Newark Basin), and may be as much as 150 feet thick; and as a sheet on the rock surface, which is regionally continuous in the Newark Basin, where it is generally 20 to 40 feet thick, but is thin and patchy elsewhere. In 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.

The late Wisconsinan till is divided into three lithically-distinct units. Light gray to yellowish brown sandy till derived chiefly from gneiss (the Netcong Till) covers most of the Highlands and part of the northwesternmost Newark Basin. Reddish brown sandy-silty till derived chiefly from red shale and sandstone in the Newark Basin (the Rahway Till) 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. Olive brown to grayish brown sandy-silty till derived from Paleozoic sedimentary rock (the Kittatinny Mountain Till) covers the Valley and Ridge and small parts of the westernmost Highlands.

The Budd Lake core provides 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 (OC-1304) (Stone et al., 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) (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 et al., 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, cited in Stone and Borns, 1986) 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. 1-9), indicate the late Wisconsinan ice margin had retreated north of New Jersey by about 18 ka.

Recessional ice margins (fig. 1-9) are marked by four till moraines and by stratified drift forming fluvial and lacustrine morphosequences. The stratified drift is collectively classed as the Rockaway Formation in Stone et al. (in press). The moraines are well developed on Kittatinny Mountain and are discontinuous but readily traceable in the Kittatinny and Minisink Valleys and the western Highlands, but are only locally developed east of there. The fluvial and lacustrine deposits, which are divided into about 100 informal allostratigraphic units in Stone et al. (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. Lakes formed in valleys that drained toward the ice margin, or in valleys where drainage was blocked by previously-deposited drift. Fluvial deposits occur in valleys where drainage was not blocked and also cover lacustrine deposits in some valleys where meltwater continued to discharge after lakes drained or filled completely with sediment. The history of lake levels and drainage, determined from the elevations of deltas and lacustrine fans and 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. Details of the deglaciation are provided by Ridge (1983), Witte (1988), Stone et al. (1989), Stanford and Harper (1991), Stanford (1993a), and Stone et al. (in press), and on quadrangle maps available from the New Jersey Geological Survey.

Deposition of the valley-fill deposits, and of the terminal moraine, caused a number of drainage dislocations. Among the most prominent are the lower Raritan, which was diverted southeastward into Raritan Bay when its northeast-trending preglacial valley was filled. 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. 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.

Periglacial Colluvial and Fluvial Deposits

Colluvium is widespread south of the terminal moraine and forms continuous aprons along the base of long, steep slopes. The aprons are best developed along the escarpment slopes separating main valleys and lowlands from uplands, generally at the contacts of gneiss with carbonate rock or shale, or diabase and basalt with shale. Colluvium in these aprons may be as much as 50 feet thick. It is not well developed north of the moraine, except as talus at the base of cliffs, and as a few small aprons at the base of steep till-covered slopes. Some of these latter aprons have ice-contact scarps at their distal margins, indicating that they were deposited just after melting ice uncovered the sourcing slopes. These observations suggest that colluviation occurred primarily before or during the glacial maximum rather than during deglaciation.

In some exposures the colluvium consists of a lower layered white, yellow, and red clayey to sandy material that can be traced to saprolite in places, which passes upward to a blocky brown diamicton with a strong slope-parallel clast fabric. Water-washed or current-bedded material is rarely observed, and lobate landforms are not observed on the apron surface. These features suggest that creep was the dominant colluvial process, at least on blocky-weathering rocks like gneiss, diabase, and basalt, and that there has been a gradual inversion of the saprolite-over-weathered rock stratigraphy on the feeder slope. In several exposures a lower, weathered colluvium of presumed Illinoian age, in places including a truncated red Bt horizon, was exposed below several feet of brown fresh-clast colluvium of presumed Wisconsinan age. This stratigraphy indicates episodic colluviation and stacking of slope debris in the aprons.

In valleys south of the terminal moraine that contain late Wisconsinan outwash plains, the distal ends of the aprons grade to the plains and locally interfinger with the outwash, indicating that the surface colluvium is, in part, of late Wisconsinan age. Currently the aprons are being slowly eroded by bank cutting at the toe, and by spring-sapping and gullying on the apron surface. Thus, slopes appear to be mobile during periods of maximum cold (although permafrost may not be necessary) and the resulting debris is stored along the slope base until slowly removed by stream erosion during interglacials. Flat to gentle slopes, however, hold old surficial deposits and locally support cryoturbation features, especially on the Pensauken plain and Somerville strath, where substrate material is more deformable and drainage poor. These observations suggest that gentle slopes are not extensively eroded under periglacial conditions but rather are stable or are deformed in-situ.

Valleys south of the terminal moraine that are not connected to glacial meltwater sources contain local stream terraces that are on grade with the colluvial aprons and with late Wisconsinan glaciofluvial terraces downvalley along either the Delaware, Millstone, or Raritan rivers. These relationships, and radiocarbon dates in the Coastal Plain that bracket the lower terrace deposits there to between about 27 and 14 ka (fig. 1-3), indicate that valleys alluviated under periglacial conditions and, like the colluvial aprons, entrenched during interglacials.

Postglacial Events

Following deglaciation, streams adjusted to declining 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 reversed gradients of two such terraces, and a radiocarbon date of 12,870+/-200 yrs BP (QC-297, Averill et al., 1980) on a basal peat on top of the pre-rebound terrace, suggest that the onset of rebound was delayed until about that time (Stanford and Harper, 1991). A date of 10,430+/-160 (I-16554, Stanford, 1993b) on basal peat on top of the late Wisconsinan 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, marked in places by sand-capped straths cut into the glacial deposits, streams established generally narrow floodplains. North-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.

The rising sea level entered the Hudson Valley, which may have been deeply-scoured of its glacial sediment fill by meltwater floods sourced from Great Lake drainages during periods when the Mohawk Valley was ice-free, before 12 ka (Newman et al., 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-21687, Stanford, unpublished) from basal estuarine deposits at Perth Amboy.

Postglacial pollen records show dominantly pine and spruce, with elevated sedge and grass, extending from before the glacial maximum to about 10 ka. This mix of arboreal and nonarboreal pollen suggests a mixed boreal woodland-tundra vegetation. Pine dominates from about 10 to 9 ka, with sedge and grass declining, and oak dominates after 9 ka (Harmon, 1968; Nicholas, 1968; Sirkin and Minard, 1972; Watts, 1979; Cotter et al, 1986; Peteet et al., 1990). Peteet et al. (1990) detected a increase in spruce, fir, larch, birch, and alder between 11 and 10 ka in a core from a bog on the Palisades Ridge that they interpret as indicating a Younger Dryas cooling.

CONCLUSIONS

The landscape and surficial deposits of northern New Jersey are a mosaic of relict features dating to the Pliocene and, possibly, the late Miocene, and more recent features reflecting the glacial-interglacial climate cycles of the late Pleistocene. The gross topographic elements such as erosion surfaces, wind gaps, straths, pediments, and incised valleys record dissection episodes driven by sustained baselevel lowerings, particularly the eustatic declines due to middle Miocene growth of the Antarctic ice sheet, and the late Pliocene growth of northern hemisphere ice sheets, and the glacially induced diversion of the Pensauken river system in the late Pliocene or early Pleistocene. The large-scale landscape features thus take shape over periods of several hundred thousand to several million years, while the climate-driven processes of periglacial colluviation and alluviation, and interglacial channel incision, that form landscape details within the large-scale forms, cycle over periods of several thousand to several tens-of-thousands of years.

The type areas of the Schooley and Somerville peneplains are true rock-cut erosion surfaces, and their close association with the Beacon Hill and Pensauken fluvial deposits provide tentative age control and suggest a relict origin. The Schooley surface, at least in the Newark Basin proximal to the Coastal Plain, may represent a planation surface, perhaps cut largely by colluvial processes, formed during an extended period of stable baselevel prior to the middle Miocene eustatic decline and resulting dissection of the Coastal Plain. Inland correlations are speculative, given the degree of glacial and periglacial modification of Appalachian ridgetops. The Somerville surface, and the supposed correlate to the Harrisburg surface, however, are valley straths tied spatially and temporally to the relatively long period of stable local baselevel represented by the Pensauken Formation, and thus their identification and extension as regional planation surfaces is suspect.

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 the Pensauken Formation, and correlation to till in central 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.

Remaining questions include the age and correlation of the pre-Illinoian and Illinoian? drifts, and of the Pensauken Formation; the age, origin, and degree of modification of the upland surfaces; and the processes involved in colluviation, planation, and moraine and morphosequence deposition. A broader question is: to what extent are glacial advances, retreats, and recessional stillstands responses to global climate change, to regional climate factors, and to non-climatic glaciologic factors? How does the terrestrial record of the timing and extent of the Laurentide ice sheet compare to the marine record of ice volume and ocean circulation, and to the outputs of global climate models? Deeper, older pollen records that may be preserved beneath late Wisconsinan clastic layers in Illinoian lake basins south of the terminal moraine (such as lakes Succasunna, Budd, Oxford, Shongum, Oxford, and Passaic on fig. 1-9) could provide some data that bear on these questions. Improved chronology for the pre-late Wisconsinan deposits would also be of significant value in comparing the terrestrial record of glaciation and climate to the marine record.

REFERENCES

Ashley, G. M., and 3 others, 1991, Clastic sequences developed during late Quaternary glacio-eustatic sea-level fluctuations on a passive margin: example from the inner continental shelf near Barnegat Inlet, New Jersey: Geological Society of America Bulletin, v. 103, p. 1607-1621.

Averill, S. P., and 3 others, 1980, Late Wisconsinan-Holocene history of the lower Hudson region: new evidence from the Hackensack and Hudson River valleys, *in* Manspeizer, W., ed., Field studies of New Jersey geology and guide to field trips: New York State Geological Association, 52nd Annual Meeting, Rutgers University, Newark,

N. J., p. 160-186.

Bascom, F., and 5 others, 1909, Description of the Trenton quadrangle, New Jersey-Pennsylvania: U. S. Geological Survey Geologic Atlas 167, 24 p.

Bayley, W. S., Salisbury, R. D., Kummel, H. B., 1914, Description of the Raritan quadrangle: U. S. Geological Survey Geologic Atlas 191, 32 p.

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

Bowman, J. P., 1966, Petrology of the Pensauken Formation: unpublished Ph. D. dissertation, Rutgers University, New Brunswick, N. J., 155 p.

Bowman, J. P., and Lodding, W., 1969, The Pensauken Formation--a Pleistocene fluvial deposit in New Jersey, *in* Subitzky, S., ed., Geology of selected areas in New Jersey and eastern Pennsylvania and guidebook of excursions: New Brunswick, N. J., Rutgers University Press, p. 3-6.

Braun, D. D., 1989, Glacial and postglacial erosion of the Appalachians: Geomorphology, v. 2, p. 233-256.

Campbell, M. R., and Bascom, F., 1933, Origin and structure of the Pensauken gravel: American Journal of Science, v. 26, no. 153, p. 300-318.

Chappell, J., and Shackleton, N. J., 1986, Oxygen isotopes and sea level: Nature, v. 324, p. 137-140.

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

Cook, G. H., 1880, Surface geology--report of progress: N. J. Geological Survey Annual Report for 1880, p. 16-97.

Cotter, J. F. P., and 5 others, 1986, The Wisconsinan history of the Great Valley, Pennsylvania and New Jersey, and the age of the "Terminal Moraine", *in* Cadwell, D. H., ed., The Wisconsinan stage of the First Geological District, eastern New York: New York State Museum Bulletin 455, p. 22-50.

Crowl, G. H., 1971, Pleistocene geology and unconsolidated deposits of the Delaware Valley, Matamoras to Shawnee on Delaware, Pennsylvania: Pennsylvania Geological Survey General Geologic Report 60, 40 p.

Darton, N. H., and 3 others, 1908, Description of the Passaic quadrangle: U. S. Geological Survey Geologic Atlas 83, 19 p.

Davis, W. M., 1890, The rivers of northern New Jersey, with notes on the classification of rivers in general: National Geographic Magazine, v. 2, p. 81-110.

Davis, W. M., and Wood, J. W., 1889, The geographic development of northern New Jersey: Proceedings of the Boston Society of Natural History, v. 24, p. 365-423.

Gardner, T. W., Sasowsky, I. D., Schmidt, V. A., 1994, Reversed polarity glacial sediments and revised glacial chronology, West Branch Susquehanna River, central Pennsylvania: Quaternary Research, v. 42, p. 131-135.

Gill, H. E., and Farlekas, G. M., 1976, Geohydrologic maps of the Potomac-Raritan-Magothy aquifer system in the New Jersey Coastal Plain: U. S. Geological Survey Hydrologic Investigations Atlas HA-557.

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

Harper, D. P., 1982, Late Wisconsinan glacial geology of New Jersey: Geological Society of America Abstracts with Programs, v. 14, no. 1-2, p. 23.

Herman, G. C., and 3 others, 1992, Bedrock geologic map of the Pittstown and Flemington quadrangles, Hunterdon and Somerset counties, New Jersey: N. J. Geological Survey Open File Map 10, scale 1:24,000.

Herpers, H., 1961, The Ogdensburg-Culvers Gap recessional moraine and glacial stagnation in New Jersey: N. J. Geological Survey Geologic Report Series 6, 16 p.

Heusser, C. J., 1963, Pollen diagrams from three former cedar bogs in the Hackensack tidal marsh, northeastern New Jersey: Bulletin of the Torrey Botanical Club, v. 90, p. 16-28.

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

Jordan, R. R., 1964, Columbia (Pleistocene) sediments of Delaware: Delaware Geological Survey Bulletin 12, 96 p.

Koteff, C., and Pessl, F., 1981, Systmeatic ice retreat in New England: U. S. Geological Survey Professional Paper 1179, 20 p.

MacClintock, P., 1940, Weathering of the Jerseyan till: Geological Society of America Bulletin, v. 51, p. 103-116.

MacClintock, P., 1954, Leaching of Wisconsinan glacial gravels in eastern North America: Geological Society of America, v. 65, p. 369-384.

MacClintock, P., and Richards, H. G., 1936, Correlation of late Pleistocene marine and glacial deposits of New Jersey and New York: Geological Society of America Bulletin, v. 47, p. 289-338.

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

Merrill, F. J. H., and 6 others, 1902, Description of the New York City quadrangle: U. S. Geologic Atlas 83, 19 p.

Meyerson, A. L., 1971, Glacial Lake Passaic; palynological evidence for draining of the Great Swamp stage: N.J. Academy of Science Bulletin, v. 15, no. 1-2, p. 10-12.

Miller, K. G., Fairbanks, R. G., Mountain, G. S., 1987, Tertiary oxygen isotope synthesis, sea level history and continental margin erosion: Paleoceanography, v. 2, p. 1-19.

Minard, J. P., 1961, End moraines on Kittatinny Mountain, Sussex County, New Jersey: U. S. Geological Survey Professional Paper 424C, p. C61-C64.

Minard, J. P., 1969, Geology of the Sandy Hook quadrangle in Monmouth County, New Jersey: U. S. Geological Survey Bulletin 1276, 43 p.

Minard, J. P., and Rhodehamel, E. C., 1969, Quaternary geology of part of northern New Jersey and the Trenton area, *in* Subitzky, S., ed., Geology of selected areas in New Jersey and eastern Pennsylvania and guidebook of excursions: New Brunswick, N. J., Rutgers University Press, p. 279-313.

Newell, W. L., and 3 others, 1995, Surficial geological map of New Jersey: southern sheet: U. S. Geological Survey Open File Map 95-272, scale 1:100,000.

Newman, W. S., and 4 others, 1969, Late Quaternary geology of the Hudson River estuary: a preliminary report: Transactions of the New York Academy of Sciences, v. 31, p. 548-570.

Nicholas, J. T., 1968, Late Pleistocene palynology of southeastern New York and northern New Jersey: unpublished Ph. D. dissertation, New York University.

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

Owens, J. P., and 5 others, 1988, Stratigraphy of the Tertiary sediments in a 945-foot-deep corehole near Mays Landing in the southeastern New Jersey Coastal Plain: U. S. Geological Survey Professional Paper 1484, 39 p.

Pazzaglia, F. J., and Gardner, T. W., 1994, Late Cenozoic flexural deformation of the middle U. S. Atlantic Coastal Plain: Journal of Geophysical Research, v. 99, no. B6, p. 12143-12157.

Peltier, L. C., 1959, Late Pleistocene deposits, *in* Willard, B., ed., Geology and mineral deposits of Bucks County, Pennsylvania: Pennsylvania Geological Survey 4th Series Bulletin C9, p. 163-184.

Peteet, D. M., and 5 others, 1990, Younger Dryas climatic reversal in northeastern USA? AMS ages for an old problem: Quaternary Research, v. 33, p. 219-230.

Plint, A. G., and 3 others, 1992, Control of sea level change, *in* Walker, R. G., and James, N. P., eds., Facies Models: Geological Association of Canada, St. Johns, Newfoundland, p. 15-25.

Poag, C. W., and Sevon, W. D., 1989, A record of Appalachian denudation in postrift Mesozoic and Cenozoic sedimentary deposits of the U. S. Middle Atlantic continental margin: Geomorphology, v. 2, p. 119-157.

Raymo, M. E., and 4 others, 1989, Late Pliocene variation in northern hemisphere ice sheets and north Atlantic deep water circulation: Paleoceanography, v. 4, p. 413-446.

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

Richmond, G. M., and Fullerton, D. S., 1986, Summation of Quaternary glaciations in the United States of America: Quaternary Science Reviews, v. 5, p. 183-196.

Ridge, J. C., 1983, The surficial geology of the Great Valley section of the Ridge and Valley Province in eastern Northampton County, Pennsylvania and Warren County, New Jersey: unpublished M. S. thesis, Lehigh University, Bethlehem, Pa., 276 p.

Ridge, J. C., Evenson, E. B., Sevon, W. D., 1992, A model of late Quaternary landscape development in the Delaware Valley, New Jersey and Pennsyvania: Geomorphology, v. 4, p. 319-345.

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

Ruddiman, W. F., Raymo, M. E., and McIntyre, I., 1986, Matuyama 41,000-year cycles: north Atlantic ocean and northern hemisphere ice sheets: Earth and Planetary Science Letters, v. 80, p. 117-129.

Russell, E. W. B., 1980, Vegetational change in northern New Jersey from precolonization to the present: a palynological interpretation: Bulletin of the Torrey Botanical Club, v. 107, no. 3, p. 432-446.

Salisbury, R. D., 1898, The physical geography of New Jersey: N. J. Geological Survey Final Report, v. 4, 170 p.

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

Salisbury, R. D., and Knapp, G. N., 1917, The Quaternary formations of southern New Jersey: N. J. Geological Survey Final Report, v. 8, 218 p.

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

Spencer, A. C., and 4 others, Description of the Franklin Furnace quadrangle: U. S. Geological Survey Geologic Atlas 161, 27 p.

Sirkin, L. A., and 3 others, 1970, Palynology of some upper Quaternary peat samples from the New Jersey Coastal Plain: U. S. Geological Survey Professional Paper 700D, p. D77-D87.

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

Stanford, S. D., 1993a, Late Wisconsinan glacial geology of the New Jersey Highlands: Northeastern Geology, v. 15, no. 3 and 4, p. 210-223.

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

Stanford, S. D., 1994, Surficial geology of the northernmost New Jersey Coastal Plain: Geological Society of America Abstracts with Programs, v. 26, no. 3, p. 74.

Stanford, S. D., and Harper, D. P., 1991, Glacial lakes of the lower Passaic, Hackensack, and lower Hudson valleys, New Jersey and New York: Northeastern Geology, v. 13, no. 4, p. 271-286.

Stanford, S. D., Witte, R. W., Harper, D. P., 1990, Hydrogeologic character and thickness of the glacial sediment of New Jersey: N. J. Geological Survey Open File Map 3, scale 1:100,000.

Stone, B. D., and Borns, H. W., Jr., 1986, Pleistocene glacial and interglacial stratigraphy of New England, Long Island, and adjacent Georges Bank and Gulf of Maine: Quaternary Science Reviews, v. 5, p. 39-52.

Stone, B. D., Reimer, G. E., Pardi, R. R., 1989, Revised stratigraphy and history of glacial Lake Passaic, New Jersey: Geological Society of America Abstracts with Programs, v. 21, no. 2, p. 69.

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

Sugarman, P. J., and 3 others, 1993, Strontium-isotope and sequence stratigraphy of the Miocene Kirkwood Formation, southern New Jersey: Geological Society of America Bulletin, v. 105, p. 423-436.

Tedrow, J. C. F., 1986, Soils of New Jersey: Malabar, Florida, Robert E. Krieger, 479 p.

Trela, J. J., 1984, Soil formation on Tertiary land surfaces of the New Jersey Coastal Plain: unpublished Ph. D. dissertation, Rutgers University, New Brunswick, N. J., 620 p.

Van Houten, F. B., 1969, Late Triassic Newark Group, north central New Jersey and adjacent Pennsylvania and New York, *in* Subitzky, S., ed., Geology of selected areas in New Jersey and eastern Pennsylvania and guidebook of excursions: New Brunswick, N. J., Rutgers University Press, p. 314-347.

Von Schondorf, A., 1987, Sedimentary facies and paleohydraulics of an ice-contact glacial outwash plain, Germany Flats, New Jersey: unpublished M. S. thesis, Rutgers University, New Brunswick, N. J., 122 p.

Walters, J. C., Polygonal patterned ground in central New Jersey: Quaternary Research, v. 10, p. 42-54.

Ward, F., 1938, Recent geological history of the Delaware Valley below the water gap: Pennsylvania Topographic and Geologic Survey Bulletin G10, 76 p.

Watts, W. A., 1979, Late Quaternary vegetation of central Appalachia and the New Jersey Coastal Plain: Ecological Monographs, v. 49, p. 427-469.

Witte, R. W., 1988, The surficial geology and Woodfordian glaciation of a portion of Kittatinny Valley and the New Jersey Highlands in Sussex County, New Jersey: unpublished M. S. thesis, Lehigh University, Bethlehem, Pa., 276 p.

Wolfe, P. E., 1953, Periglacial freeze-thaw basins in New Jersey: Journal of Geology, v. 61, p. 131-141.

CHAPTER 6
ROAD LOG AND DESCRIPTION OF FIELD STOPS
SATURDAY MAY 31, 1997

Scott D. Stanford
New Jersey Geological Survey
CN 427
Trenton, NJ 08625

0.0 Turn right from hotel parking lot onto US 46 west. This small valley is floored by a patch of non-morainic late Wisconsinan till extending about 0.25 mile beyond the front of the terminal moraine. Distal edge of the till, marking the terminus of late Wisconsinan ice, runs along the base of the hillslope to the south (L), where it feathers out onto weathered gneiss.

0.5 Cross under I-80.

0.7 Enter terminal moraine. Ridge, knoll, and swale moraine topography shows up to 60 feet of relief but is obscured by urbanization. Till is 80 to 120 feet thick in this area.

1.7 Turn left onto US 206 south.

2.1 Cross under I-80.

2.6 Netcong wellfield on right. Several wells between 80 and 120 feet deep pump from sandy till of the terminal moraine or, possibly, sandy gneiss saprolite, just above gneiss bedrock. Yields of wells are as much as 250 gpm.

2.8 Turn right on Gold Mine Road. Industrial park here is built in a former gravel pit, which removed a 40-foot-tall ridge of flowtill marking the front of the terminal moraine.

3.4 Make U turn at Morris County trash transfer station.

3.7 Park along road at end of fence.

STOP 1. Late Wisconsinan and Pre-late Wisconsinan Till (Scott Stanford)

Setting: This exposure is in a small fringe of late Wisconsinan till extending just beyond the front of the terminal moraine (here a part of the "Budd Lake Moraine" of Stone et al. [in press]). A former gravel-pit operation removed the upper 20 to 40 feet of the moraine between here and US 206, and later became a quarry operation after the gravel was exhausted. The original topography is shown on the base map (fig. 6-1). The front edge of the moraine was sharply defined by an asymmetric till ridge with a steep, ice-contact north slope and gentler south slope. A similar ridge is preserved just to the west of here (fig. 6-1), and at several other places in the vicinity. The morphology and composition of these ridges suggest that they are colluvial ramparts formed by deposition of flowtill from the ice front. Based on radiocarbon dates from Budd Lake (2 miles west of here), late Wisconsinan ice arrived no earlier than about 22 ka (see Chapter 1) and, based on top-of-till dates north of the moraine, was in retreat by about 20 ka. The southern edge of the moraine, and of the few small fringes of nonmorainic till that extend beyond the moraine, is a sharp contact with blocky-weathering gneiss or gneiss colluvium on steep to moderate slopes, and patches of older weathered till on flat to gentle slopes where it has been protected from colluviation (fig. 1-4). Gneiss clasts in this older till consistently show at least a thin weathering rind, in contrast to the late Wisconsinan till, where there are always some fresh gneiss clasts. This older till is probably Illinoian,

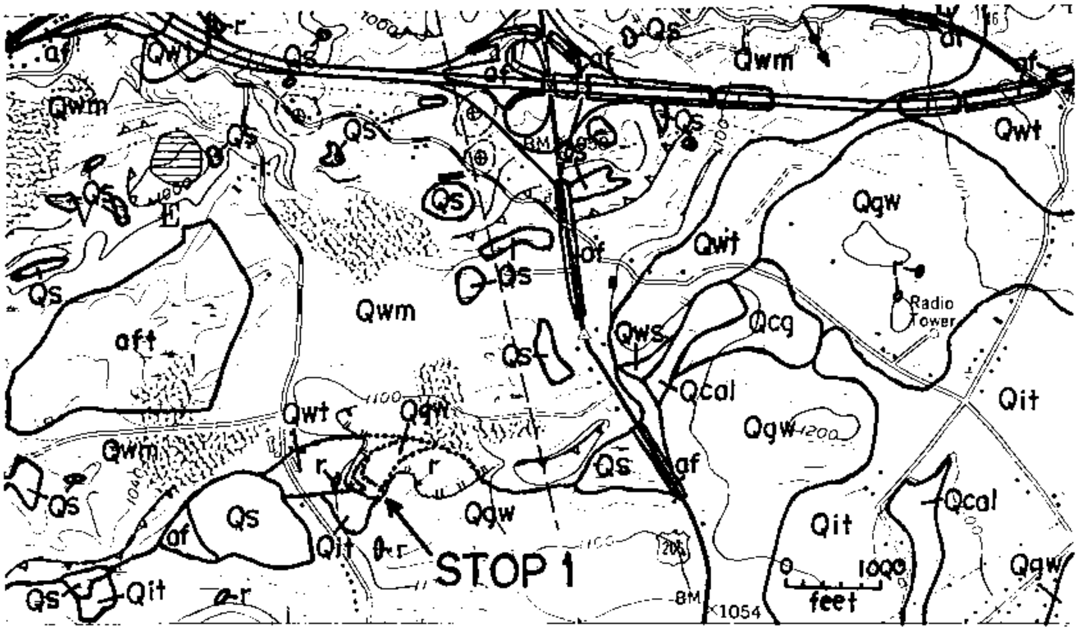


Figure 6-1. Surficial geologic map of area around Stop 1. Units are: af=artificial fill, aft=trash fill, Qs=swamp deposits, Qcal=colluvium and alluvium, undivided; Qcg=gneiss colluvium; Qgw=weathered gneiss and thin, patchy colluvium; Qwm=till of the late Wisconsinan terminal moraine; Qwt=late Wisconsinan till, without morainic landforms; Qit=Illinoian till; r=bedrock outcrop. Symbols are: =striation; =asymmetric moraine ridge, barbs on gentle slopes (includes flowtill aprons where barbs point south and push ridges where barbs point north); =crest of symmetric moraine ridge; =moraine hill, dashed line around base, symbol on summit; =moraine plateau; =edge of excavation, contacts within excavation are dotted. From Stanford et al. (1997b). Base is U. S. Geological Survey Stanhope 7.5 minute quadrangle.

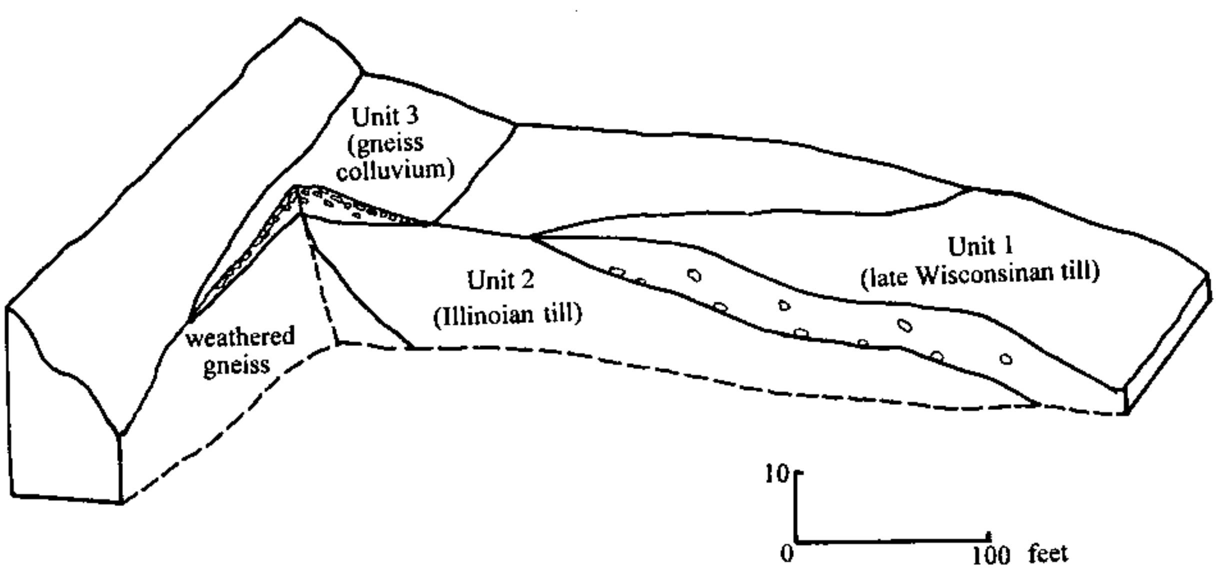


Figure 6-2. Block diagram of Stop 1. Perspective view to south. Dashed lines show edges of pit faces, solid lines show unit contacts. Scale is approximate.

in contrast to the more deeply weathered and eroded pre-Illinoian till further south (refer to Chapter 1).

Materials: Both tills, and some gneiss colluvium, are exposed in this cut (fig. 6-2). Diamicton 1 (fig. 6-2) is interpreted as late Wisconsinan till. It is a yellowish brown (10YR5/6) matrix-supported silty sand containing a mix of fresh, weathered, and fully decomposed gneiss pebbles and cobbles, of subangular to subrounded shape, and a few generally unweathered, subrounded gray and brown mudstone and sandstone pebbles, and a trace of decomposed carbonate rock (as yellow silty ghosts, often smeared or flattened), that were transported southward from Paleozoic rock of the Kittatinny Valley. This mix of clast weathering is typical of late Wisconsinan till in and near the terminal moraine, where a significant amount of weathered material from Illinoian drift and saprolite is incorporated into the younger deposits.The contact with diamicton 2 is sharp and marked visually by a rather sparse boulder line, in part reflecting the decomposition of boulders in the lower unit. Diamicton 2 is interpreted as Illinoian till. It is a brownish yellow (10YR6/6 to 6/8, slightly more yellow than diamicton 1) to light olive brown (2.5Y5/6) where gleyed, silty sand to sandy silt. It is matrix supported, but pebbles and cobbles are much more abundant than in unit 1. Gneiss clasts, including some up to small boulder size, are entirely saprolitized or have thick (2-3 cm) weathering rinds surrounding a partly weathered core. Gneiss boulders are fractured in-situ and also have thick weathering rinds. Yellow to olive silty ghosts of carbonate rock (in places replaced by black MnO) are more abundant than in unit 1, and gray mudstone pebbles are also weathered here, in contrast to unit 1. White to gray quartzite and chert are generally unweathered. Black MnO staining is common and coats clast and parting surfaces. There are a few wavy lenses of clayey sand within the diamicton that likely reflect subglacial, englacial or superglacial water-sorting, although some near the contact with unit 1 may be related to later sheetwash or solifluction. There is little to no indication of a buried B horizon, indicating that the surface of this unit was truncated (probably by Wisconsinan slope erosion) before deposition of unit 1. Diamicton 3 overlies, with sharp, wavy contact, the outcropping part of unit 2 adjacent to the gneiss ridge. It is interpreted as late Wisconsinan colluvium. It is a yellowish brown (10YR5/6) silty sand containing mostly fresh angular gneiss clasts, with a-b planes of tabular clasts generally lying parallel to the slope. Clast abundance increases toward the surface, suggesting sequential stripping of soil followed by rock rubble on the feeder slope. Weathered gneiss, including a mix of intact outcrop, angular-clast rubble, and yellowish brown silty sand saprolite, is exposed on the east wall of the pit.

Discussion: The intensity of weathering in unit 2 is greater than usually observed in Illinoian till, and is similar to that usually seen in pre-Illinoian till. However, this is a poorly drained site where the time of contact of the till with groundwater is likely significantly longer than at better-drained sites, producing faster weathering. This type of variation in site conditions produces some overlap in the weathering features of Illinoian and pre-Illinoian drift. In the absence of a well-defined soil chronosequence, where drainage, topographic position, and parent material are held constant, the topographic position and extent of erosional preservation of the deposits are often more effective than weathering characteristics in distinguishing the two drifts. The continuity of the till on gentle slopes, and the regional association of the till with subdued construction landforms in valleys, both features that are not observed for pre-Illinoian till, indicate an Illinoian age for unit 2. The greater clast content of unit 2, and the abundance of grit and granule-sized gneissic fragments in the matrix, may reflect the greater extent and thickness of weathered gneiss on the landscape during Illinoian advance than during late Wisconsinan advance, producing a more pebbly, gritty till. The greater carbonate content of unit 2 perhaps reflects exposure of carbonate rocks in the Musconetcong Valley several miles to the north (known from borings but not now exposed) during Illinoian advance that were buried by Illinoian lacustrine deposits during late Wisconsinan advance, and so not readily available to the late Wisconsinan glacier. Also, the sharp contact and abrupt outer border of the late Wisconsinan till argue against a "diffuse" margin beyond the moraine over which late Wisconsinan ice reworked earlier materials.

3.7 Proceed east on Gold Mine Road back to 206.

4.0 Turn right onto US 206 south. Cross southern edge of terminal moraine (removed by gravel-pit operations) and a small moraine-dammed wetland.

4.2 Cross gently sloping upland surface on gneiss bedrock. Gentlest slopes are mantled with Illinoian till; moderate and steep slopes are underlain by weathered gneiss and gneiss colluvium. Weathered gneiss, including saprolite and mixed saprolite and rock rubble, is as much as 100 feet thick throughout the Highlands (even north of the terminal moraine in places) but thickness is sharply variable, depending on mineralogic and structural properties of the gneiss. Metasedimentary units generally show thicker saprolites than gneissic granites.

5.0 Begin descent into Long Valley, which is underlain by downfaulted Cambrian and Ordovician carbonate rock (Allentown and Leithsville Formations) of the Green Pond Outlier.

6.4 Approximate location of fault contact of Proterozoic gneiss and Paleozoic carbonate rock. Travel along apron of gneiss colluvium at base of escarpment. Colluvium covers contact.

6.9 Turn right into driveway for Flanders Office and Research Park (230 Route 206). Proceed to cul-de-sac at end.

7.1 Park at cul-de-sac.

STOP 2. Wisconsinan and Pre-Wisconsinan Colluvium (Scott Stanford)

Setting: This exposure is cut into the upper part of an apron of gneiss colluvium (fig. 6-3). This apron is continuous (except where replaced by alluvial fans at the mouths of drainages) along the base of the escarpment separating Schooleys Mountain to the west, underlain by gneiss, and Long Valley to the east, floored by carbonate rock. This site is just outside the Illinoian limit, which is marked by a till moraine and outwash plain in Long Valley (fig. 6-3), and till patches atop Schooleys Mountain. The colluvial aprons typically have a distinct break-in-slope at their tops to a steeper feeder slope in blocky-weathering gneiss, and feather valleyward onto weathered carbonate or shale, locally capped with pre-Illinoian till. Modern streams typically are incised 10-15 feet into the apron surface. Distal parts of the aprons are also gullied and spring-sapped in areas of groundwater discharge. Exposures and well data suggest that the rock surface beneath the aprons slopes less steeply than do the feeder slopes, and is of the same lithology, indicating that the colluviation process can produce a pediment and flatten slopes.

Materials: Five units are exposed here (fig. 6-4). Unit 1 is exposed on the north end of the cut, where the outcrop plane cuts behind the colluvium. It is a white, yellow, and yellowish brown silty sand gneiss saprolite grading to angular pebble-to-cobble gneiss rubble. Intact gneiss crops out in several spots. Unit 2 is a yellowish brown (10YR5/6 to 5/8) clayey silt rottenstone diamicton with saprolitized pebbles and cobbles of gneiss, gray mudstone, and yellow to olive ghosts of carbonate rock. This unit may be either pre-Illinoian till or colluvium derived from stripping of pre-Illinoian till. Unit 3 overlies unit 2 with sharp, wavy contact. It is a dark brown to dark yellowish brown (10YR3/3 to 3/6) silty sand diamicton with clast lithology and weathering as in unit 2, although with more gneiss. MnO coats clasts and imparts the dark matrix color. The similarity in clast weathering and composition to unit 2 suggests that unit 3 is an upper facies of the same colluvial event, representing uncovering of gneiss saprolite underlying the pre-Illinoian till on the feeder slope. However, the sharp contact and distinct texture imply a hiatus between the two units, perhaps indicating that 2 is till and 3 is a later colluvium. Unit 4 overlies unit 3 with a distinct, wavy contact with several feet of relief. It is a yellowish brown (10YR5/6 to 5/8) silty sand diamicton with a mix of saprolitized clasts as in units 2 and 3 and less weathered, larger caliber gneiss clasts and a few unweathered reddish quartzite erratics. The increased size of the gneiss clasts, and the less-intense weathering, suggest that unit 4 represents a later colluvial event that began to mine the fractured-rock rubble beneath the saprolite. The thickness of the weathering rinds on the gneiss cobbles is similar to that in Illinoian till, suggesting that unit 4 is Illinoian colluvium. Unit 5, which is poorly exposed, is the surface

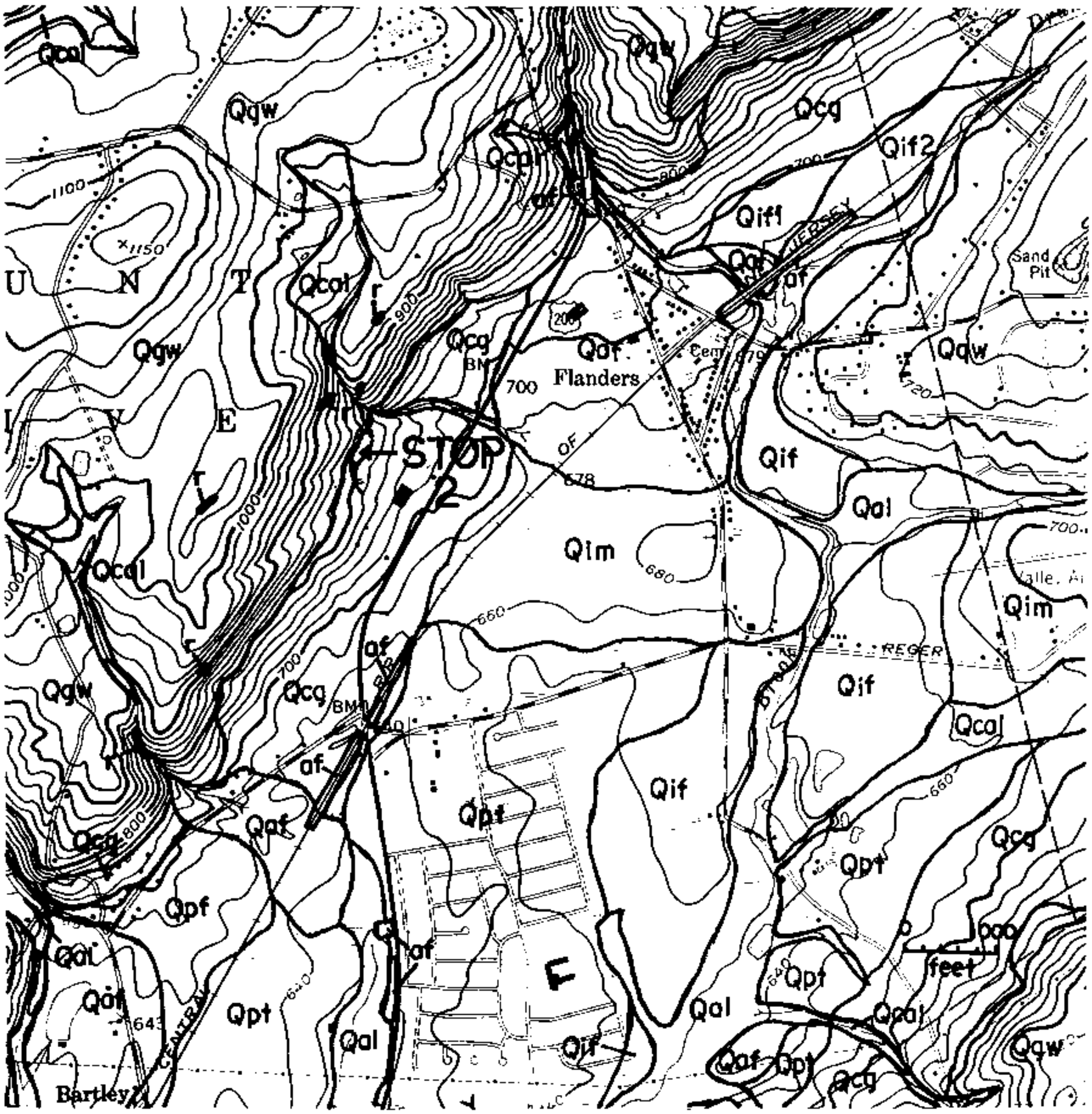


Figure 6-3. Surficial geologic map of area around Stop 2. Units are: af=artificial fill; Qal=alluvium; Qaf=alluvial- fan deposits; Qcal=colluvium and alluvium, undivided; Qcg=gneiss colluvium; Qgw=weathered gneiss and thin, patchy colluvium; Qqw=weathered quartzite and thin, patchy colluvium; Qim=till of the Illinoian terminal moraine; Qif=Illinoian glaciofluvial deposits (suffixes 1 and 2 denote upper and lower sequences, respectively, east of Flanders); Qpt=pre-Illinoian till; Qpf=pre-Illinoian alluvial-fan deposits; r=bedrock outcrop areas. Surficial deposits on the valley bottom are underlain by carbonate-rock residuum. Base is U. S. Geological Survey Chester 7.5 minute quadrangle.

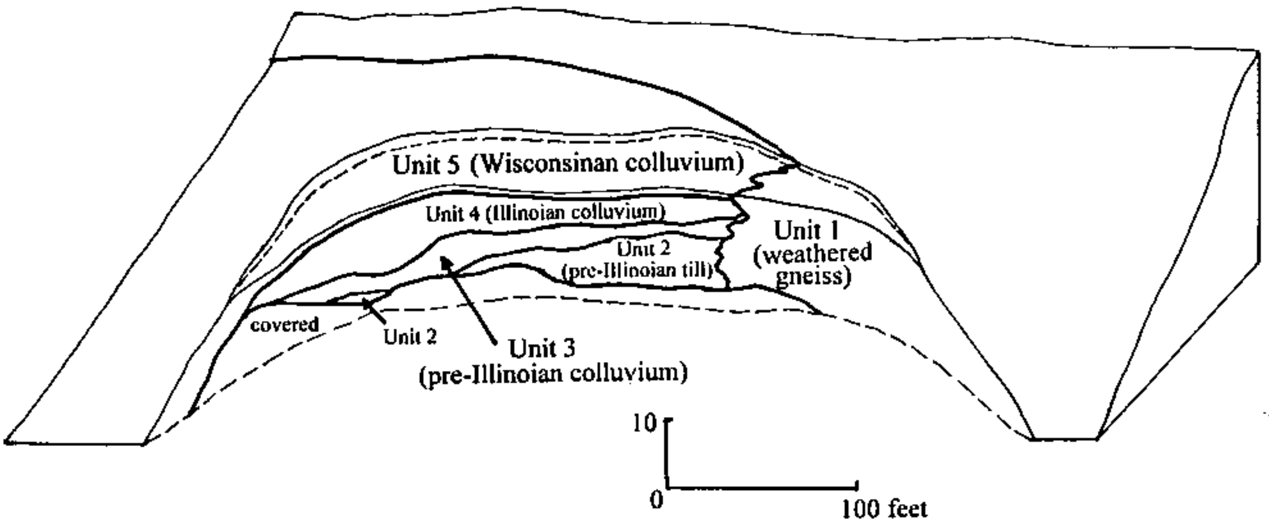


Figure 6-4. Block diagram of Stop 2. Perspective view to west. Dashed lines show bottoms of exposure faces, heavy solid lines show unit contacts. Unit 5 is exposed only in the stripped-back area above the main cut. Scale is approximate.

colluvium. It is a yellowish brown (10YR5/6) silty sand diamicton with many angular, fresh gneiss pebbles and cobbles, similar to unit 3 at stop 1. It is the Wisconsinan colluvium, mined from fractured-rock rubble on the feeder slope.

Discussion: The most straightforward interpretation of these units is: Wisconsinan colluvium over Illinoian colluvium over pre-Illinoian colluvium, possibly over pre-Illinoian till, over gneiss saprolite. Multiple colluviation events separated by hiatuses is indicated by the sharp contacts between the units, and their distinct weathering characteristics. The increase in both size and abundance of gneiss clasts, and decrease in erratic material, upward through the section represents inversion of the till-over-saprolite-over-bedrock stratigraphy on the feeder slope. The stacking of colluvium in these aprons, which, based on available exposure, seems to be common, indicates that much of the slope regolith mobilized under periglacial conditions stays on the slope. Thus, periglacial erosion on steep slopes essentially redistributes regolith on the slope, perhaps forming a pediment in the process. The colluvium is slowly removed by gullying, sapping, and bank erosion under interglacial conditions but the stacking indicates that this removal does not keep pace with the addition of material under periglacial conditions.

7.1 Return to US 206.

7.3 Turn right onto US 206 south. Subdued Illinoian terminal moraine (not visible) traverses valley from east to west ahead across 206 before looping northward about 1 mile to ascend Schooleys Mountain.

7.9 Contact of gneiss colluvium with pre-Illinoian till, which overlies carbonate residuum on much of the valley bottom. Carbonate residuum may be as much as 100 feet thick (or more) but, like weathered gneiss, thickness is sharply variable.

9.2 Cross onto apron of gneiss colluvium on east side of valley.

10.0 Crest of Fox Hill, a narrow ridge of Proterozoic gneiss. Descend into Lamington Valley, another downfaulted sliver of Cambrian and Ordovician carbonate rock within the Green Pond Outlier.

10.4 Cross marsh along the Lamington River, marking the bottom of glacial Lake Succasunna. This lake, which formed during both the Illinoian and late Wisconsinan glaciations, occupied a north-draining preglacial valley that is now filled with as much as 300 feet of Illinoian and late Wisconsinan glaciolacustrine sediment. It was controlled by a spillway across the former drainage divide about 1.5 miles southwest of here. Drainage today is southward, across the former lake spillway. The divide is now about 8 miles to the north, on the late Wisconsinan terminal moraine.

10.7 Cross from lake bottom to bordering apron of gneiss colluvium.

11.1 Cross top of apron onto weathered gneiss.

11.6 Cross NJ 24. Town of Chester to left. Continue on remnant of upland surface, here with about 100 feet of gentle relief and entirely in weathered gneiss.

13.2 Begin descent from upland surface into the Raritan Lowland.

14.7 Fault contact (not exposed) of Proterozoic gneiss and red Jurassic-Triassic conglomerate and sandstone (Passaic Formation). Cross from Highlands into Newark Basin.

17.3 Dissected topography, with 100-150 feet of relief, hereabouts, cut into sandstone and mudstone of the Passaic Formation. Flat interfluves are capped, in places, with a thin pre-Illinoian till. The till is absent on similar flats below the interfluve level, suggesting that dissection occurred after deposition of the till.

19.9 Cross county route 523. West end of Second Watchung Mountain, held up by basalt, in front. This is a 40-mile-long arcuate ridge that contacts the Highland escarpment and so encloses a 40 by 10 mile elliptical lowland in the upper Passaic drainage basin. With preglacial gaps through the ridge at Paterson and Millburn blocked by ice and, later, by moraine deposits (at Millburn), this lowland was filled by glacial Lake Passaic during the late Wisconsinan and, possibly, Illinoian, glaciations.

20.1 Cross from shale to fine-grained upper stream terrace along the North Branch of the Raritan River. This upper terrace, which locally shows cryoturbation features, is prominent throughout the Raritan Basin. It is about 15-20 feet above a lower terrace that grades to late Wisconsinan glaciofluvial terraces. Clast weathering and terrace height in the upper terrace are similar to Illinoian glacial deposits, suggesting an Illinoian age. However, there is little glacial sediment supply to this basin in the Illinoian so the alluviation may be mostly periglacial.

20.6 Drop from upper to lower stream terrace, here a pebble-to-cobble gravel. The lower terrace contains fresh clasts and is likely of Wisconsinan age. Like the upper terrace, it is nonglacial. The spillway for the high stage of Lake Passaic is through a gap in Second Watchung Mountain about 1.5 miles east of here. A dry gorge (Moggy Hollow) about 60 to 80 feet deep, cut in basalt, leads downhill.from the spillway and empties into the main valley just east of here. However, there is little basalt in these terrace gravels, so the eroded basalt is either confined to the downstream end of the gorge, or is beneath the surface gravel, which may have been deposited after the spillway was abandoned.

21.2 Cross North Branch Raritan River.

21.3 Exit right to I-287 south.

21.8 Upland to left (E) capped with pre-Illinoian till, overlapped by basalt colluvium to east along mountain front.

22.4 Cross under I-78. Continue through dissected topography cut below the pre-Illinoian surface into Passaic Formation shale. Pre-Illinoian till caps the higher ground along the base of First Watchung Mountain to the left (E).

24.6 Move left to exit 13.

25.6 Exit left to US 206 south at exit 13.

26.2 Approximate position of pre-Illinoian terminus, as defined by hilltop remnants of till to the east and west (not visible). The till remnant to the east (L) overlaps remnants of the Pensauken Formation, a Pliocene fluvial deposit that is topographically preserved in a similar manner to the pre-Illinoian till.

27.0 Move right.

27.2 Exit right to US 206 south. Stay in left lane on exit ramp. Proceed around traffic circle, take third spoke (US 206 south). At circle cross from weathered shale onto the upper Raritan terrace deposit, here mostly a pebbly sand. This upper terrace is the downstream equivalent of that described at mile 20.1.

27.9 Cross under NJ Transit tracks (formerly Central Railroad of NJ).

28.1 Cross onto Raritan River floodplain, here largely filled.

28.7 Cross from floodplain onto a fragment of the lower Raritan terrace, the downstream equivalent of that described at mile 20.6. Here it is mostly a pebbly sand.

29.0 Cross mainstem of the Raritan River.

29.2 2000-acre estate of Doris Duke, recently deceased tobacco heiress, to right behind stone fence. Much of the estate is on the upper terrace. Inheritence litigation ongoing.

29.4 Ascend scarp onto upper terrace surface. Cryoturbation structures in pebbly sand terrace deposits were exposed in foundation excavations for offices to left.

29.8 Ascend slightly from the terrace surface (visible in fields to right) onto weathered shale. From here to mile 41.9 we will be travelling across the type area of the Somerville erosion surface. This is a flat to very gently sloping surface cut into shale of the Passaic Formation, below which modern streams are incised 40 to 60 feet. Shale bedrock is within 1 to 3 feet of the surface over the entire area, so it is a true rock-cut erosion surface. A few widely separated low hills rise about 40 feet above the surface. Each of these hills is capped, at altitudes above 140 feet, by a thin mesa-form remnant of yellow arkosic sand and quartz-chert pebble gravel identical in lithology and altitude to the more extensive Pensauken deposits just to the southeast, which will be seen at Stop 4. The quartz and chert pebbles also occur as a lag everywhere on the Somerville surface between the Pensauken remnants. The Pensauken remnants, pebble lag, and the Somerville surface all terminate against a 40 to 60 foot scarp cut in shale to the northwest (fig. 1-7), beyond which there are no Pensauken remnants or pebble lags. These characteristics suggest that the Somerville surface is a fluvial strath exhumed from beneath the Pensauken Plain rather than a regional erosion surface. Exhumation may have been accomplished by backsapping along the Pensauken-shale contact after post-Pensauken stream incision. This process is observed today where more extensive Pensauken remnants overlie shale or Cretaceous clay.

30.6 Cross under Conrail tracks (formerly Lehigh Valley, soon to be Norfolk Southern most likely). This is a mainline freight route from the Port of New York-New Jersey to points west.

31.9 View to left across fields illustrates Somerville surface. Shale at 1-3 foot depth throughout.

36.0 Keep to left at fork on US 206 south. Cross over Conrail tracks (formerly Reading Railroad, soon to be CSX most likely). This is another mainline freight route to points south and west. Double-stack container trains must use this route to avoid low tunnel clearance through the Pattenburg Tunnel in the New Jersey Highlands on the former Lehigh Valley line.

39.2 Roadcuts to left may expose Pensauken pebble lag on height-of-land rising above Somerville surface. Here, as at other highs that do not reach the 140-foot altitude, there is a heavy gravel lag that is cryo- and bioturbated into weathered shale to create a pebbly diamicton rather than a true Pensauken remnant.

40.7 Turn left at light onto county route 518 east. A Pensauken remnant caps the upland here, above 140 feet in elevation. A cut behind the shopping mall to the left, just downslope from the basal contact of the Pensauken, exposed a red clayey diamicton with quartz and chert pebbles. This diamicton, like that at 39.2, was formed by mixing of lag pebbles into underlying weathered shale (which is thicker and more clayey beneath the Pensauken remnants than elsewhere). However, it was interpreted as till and named the Rocky Hill Till by Neuman (1980). This unit appears in the Fullerton (1986) compilation but these relationships, and regional mapping of the Pensauken and pre-Illinoian till, indicate it is not a glacial diamicton. Proceed through town of Rocky Hill.

41.8 Cross Millstone River. Just to the south is the Rocky Hill Gap through the Rocky Hill diabase ridge.

This gap consists of an upper broad wind gap with a floor between 120 and 140 feet in altitude which is capped by a lag of Pensauken pebbles on weathered diabase. This upper gap marks the route of a Pensauken tributary, likely superposed from earlier fluvial deposits that formerly covered the area (see Stop 3), flowing southward across the diabase from the Raritan Lowland to the main Pensauken trunk valley to the south. A narrow inner gap is inset about 60 feet into the upper gap and the Millstone River flows northward through this inner gorge. The Millstone established northerly drainage on the abandoned Pensauken Plain after the main Pensauken trunk drainage was diverted from the valley by the pre-Illinoian glacier in the New York City area. The northerly drainage may may have been established in response to isostatic depression during that glaciation, which would have reversed the former southwesterly gradient of the Pensauken Plain (see Chapter 1).

41.9 Cross Delaware and Raritan canal. This canal, active from 1833 to 1933, links the Delaware River north of Trenton to the Raritan River at head-of-tide at New Brunswick. It gravity-feeds from the Delaware valley across a low divide in an inset abandoned valley on the Pensauken Plain, and then down the Millstone to the Raritan. The canal parallels and is at the same altitude as a Delaware-sourced late Wisconsinan glaciofluvial terrace that also crosses the divide and descends the Millstone. Partial diversion of Delaware drainage across this low divide occurred in response to isostatic depression during the late Wisconsinan glacial maximum, similar to that proposed for the Millstone in the pre-Illinoian. The canal today serves as a water supply for New Brunswick.

42.3 Diabase colluvial apron on left; large quarry in diabase on right which has removed the east half of the Rocky Hill Gap.

42.9 Rockingham on left. This house served as a headquarters for George Washington from August to November 1783 while the Continental Congress met in Princeton to draft terms of peace with England. He drafted his farewell address to his army here.

43.3 Turn left on Carroll Place.

43.5 Turn right on Old Georgetown Road.

43.7 Park on left shoulder.

STOP 3. Upland Surface and Beacon Hill Lag (Scott Stanford)

Setting: This flat surface (fig. 6-5), at an elevation of about 320 feet, is cut into diabase of the Rocky Hill diabase sheet, the southern extension of the Palisades sill where it emerges from beneath the Cretaceous onlap. The sill here dips about 15 degrees to the northwest (Parker and Houghton, 1990), so this surface is a true rock-cut surface, not a structural caprock. The surface is mantled with a regolith having a matrix that ranges from a yellow (10YR7/6) to white (10YR8/2) to light gray (10YR7/2) silty clay to a yellowish brown (10YR5/6) to dark yellowish brown (10YR4/4) silty clayey fine sand. The regolith contains a mix of angular weathered to fully saprolitized diabase pebbles and cobbles and rounded white to gray quartz and quartzite pebbles, and a few black chert pebbles and, in places, reddish brown ironstone pebbles. Many of the quartz and quartzite clasts at and near the surface have a reddish brown weathering rind, and are pitted. Well logs (fig. 6-5) indicate that this regolith is between 5 and 25 feet thick. The lag gravel is similar to the Beacon Hill Gravel, a quartzite-chert fluvial gravel (locally iron-cemented) that forms mesa-like caps on the highest hills of the Coastal Plain, with base elevation ranging from 320-340 feet (see fig. 1-4). We are separated from these uplands by a 15-mile-wide Pliocene fluvial plain inset 150-200 feet below the Beacon Hill surface (see fig. 1-7 and discussion at Stop 4). To the northwest this surface corresponds to the low-relief upland erosion surface (Schooley peneplain of Davis and Wood [1889] and Johnson [1931]) that truncates the rocks (including diabase, basalt, mudstone, argillite, sandstone, shale, and gneiss) of northern New Jersey. This upland surface, which again is a true rock-cut surface (see fig. 1-5), rises

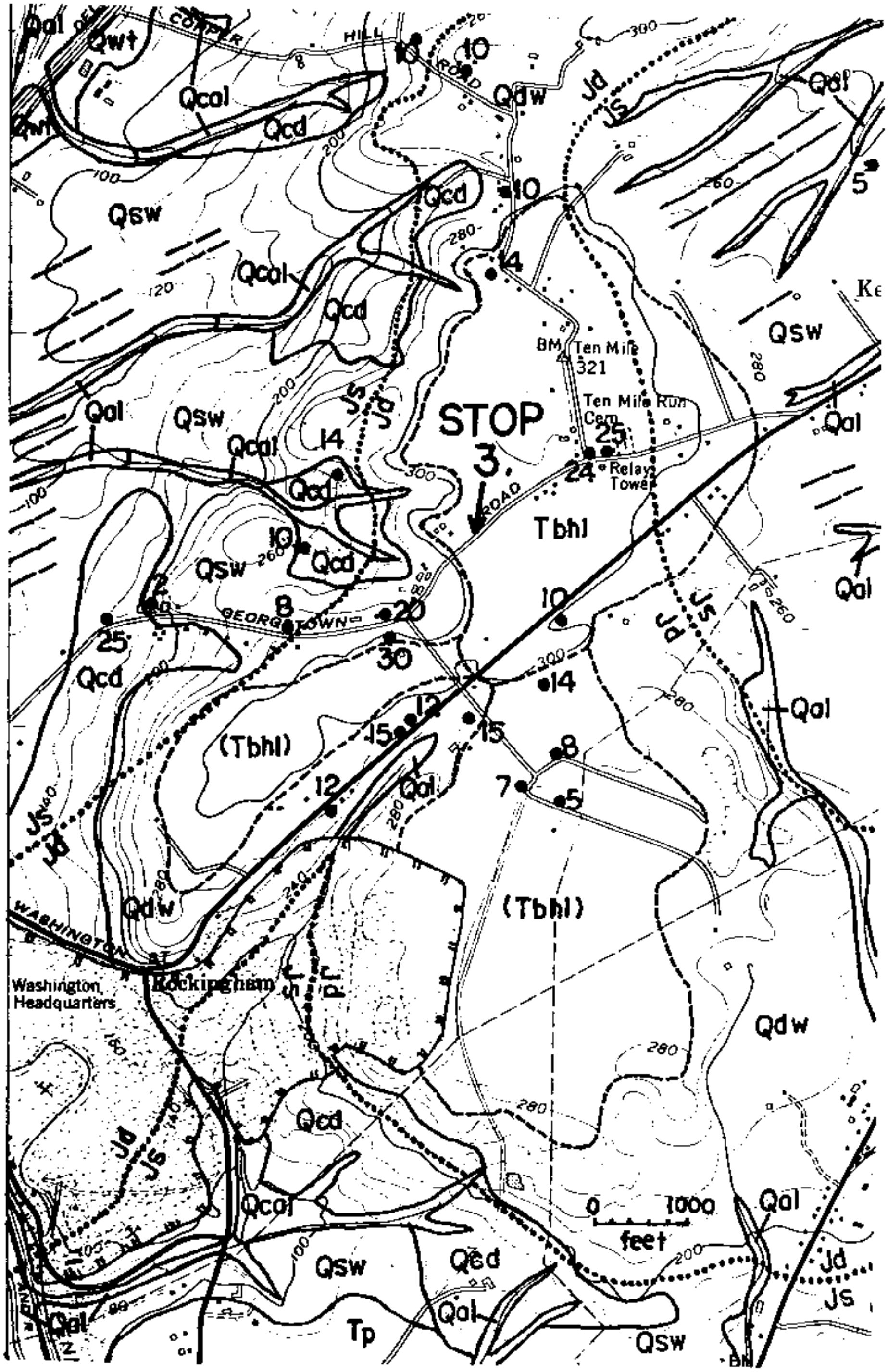


Figure 6-5. Surficial geologic map of area around Stop 3. Solid lines are contacts of surficial units. Dotted lines are contacts of bedrock units (from Parker and Houghton, 1990). Long-dashed lines are crests of low rock-controlled ridges parallel to the strike of beds in sedimentary bedrock. Short-dashed lines are the approximate limit of lag gravels. Hachured line is edge of quarry. Numbered dots show thickness of regolith reported in water-well or test-boring record. Map units are: Qal=alluvium, Qcal=colluvium and alluvium (undivided), Qcd=diabase colluvium, Qdw=weathered diabase, Qsw=weathered sedimentary rock (shale, siltstone, hornfels), Qwt=Wisconsinan stream terrace deposits, Tp=Pensauken Formation, Tbhl=Beacon Hill gravel clasts mixed into weathered diabase, (Tbhl)=scattered gravel clasts mixed into weathered diabase, Jd=diabase, Js=sedimentary rock. Base is U. S. Geological Survey Monmouth Junction 7.5 minute quadrangle.

gradually from about 320 feet here to about 1000 feet on Schooleys Mountain (fig. 1-4), and shows about 100 to 150 feet of gentle local relief.

Materials: The regolith on this surface is likely a mix of diabase saprolite, weathered diabase rock, and sandy matrix material associated with the gravel lag. A pocket of diabase saprolite exposed in the quarry just to the west was a white and yellow clayey coarse sand, similar to some of the regolith here. The diabase consists mostly of plagioclase, pyroxene, and magnetite-ilmenite; with apatite, quartz, alkali feldspar, hornblende, sphene, and zircon as accessories (Owens et al., 1996). The soil mapped here is Keyport (Kirkham, 1976), which is developed on outcropping Cretaceous clay, but is more likely a mix of Neshaminy and Mount Lucas soils in well-drained and poorly-drained sites, respectively. These soils are developed on diabase residuum (Tedrow, 1986). Clay minerals in these soils are predominantly kaolinite, with minor vermiculite and chlorotized intermediate, and a trace of gibbsite and hydrous mica (Tedrow, 1986). Thin-section study of weathering rinds, sand-fraction mineralogy of soil horizons, and profile chemistry, indicate that plagioclase alters to sericite, then to clay, and pyroxene alters to both chlorite and iron oxide, with concentration of quartz and resistant accessory heavy minerals in the A and B horizons (Tedrow, 1986).

Discussion: Upland rock-cut flats like this have been interpreted either as 1) remnants of relict low-relief erosion surfaces (Davis and Wood, 1889; Johnson, 1931; Cleaves, 1989), 2) cryoplanation surfaces (Clark and Hedges, 1992), or 3) surfaces formed by volume loss when saprolite is converted to soil (Pavich, 1989). Favoring some version of hypothesis 1 are the gravel lag, which indicates former fluvial activity; the absence of thick colluvium (which would be expected if cryoplanation had planed off interfluves and filled adjoining valleys) on this and similar, more extensive parts of the upland surface (see fig. 1-5); the regional continuity of the surface over various rock types with different rates and mechanisms of saprolite development; and the temporal association, via the link to the Beacon Hill Gravel, to a sustained, global eustatic event that initiated extensive post-middle Miocene dissection (Chapter 1, fig. 1-6), leading to incision and partial preservation of the surface. Cryoplains described in modern periglacial environments (Priesnitz, 1988) and in the Appalachian Highlands (Clark and Hedges, 1992) are developed on blocky-weathering bedrock like quartzite, show tread-riser-tor morphology, which is not observed here, and are an order of magnitude or more smaller in area than the upland flats in New Jersey. The saprolite compaction model of Pavich (1989) was described for thick (>10 m) quartz-framework saprolites on narrow interfluves in the deeply-dissected Virginia Piedmont. These conditions, which produce significant groundwater drainage, are very unlike the thin clayey regolith on broad, poorly-drained upland flats here, where groundwater drainage is minimal (the large quarry to the east, which is dug below the level of the Millstone River, is dry). While periglacial and rock-weathering processes have no doubt modified these upland flats, as have pedogenic and biologic activity, is it reasonable to think they created them? The lag gravel here suggests that this particular surface may be associated with fluvial erosion but the other unglaciated upland surfaces do not have gravel lags. They may have been formed primarily by slope processes like spring sapping and piping working in concert with rock weathering over an extended period of stable baselevel. These groundwater-related processes would be especially effective on the low permeability regolith and bedrock typical of the upland flats in the Newark Basin. An extended period of stable baselevel might be expected in this part of the Piedmont before the late Miocene, when the undissected expanse of the Coastal Plain would have buffered Piedmont streams from shorter-term sea-level-induced baselevel changes.

Continue east on Old Georgetown Road.

44.4 Turn left onto county route 518 east.

44.9 Turn left at light onto NJ 27 north. On weathered shale of the Passaic Formation here, above the Pensauken Plain level.

45.8 Turn right onto New Road at light.

47.0 Rise slightly from weathered shale to an outlier (the Sand Hills outlier) of Cretaceous sand. This is probably the Farrington Sand member of the Raritan Formation, a fluvial quartz coarse sand to granule gravel of Cenomanian age (97-91 Ma).

47.3 Turn left into Woodlot Park. Lunch. Farrington Sand exposed in cuts behind ballfield at north end of park.

47.3 Turn left onto New Road from park exit.

47.8 Descend slightly from Cretaceous sand onto weathered diabase.

48.0 Stay in middle lane. Cross US 1 at light. After US 1, descend slope of weathered diabase (note corestones in yards and woodlots) that is an exhumed part of the rock surface beneath Cretaceous deposits. The weathered-rock regolith on diabase, shale, and sandstone beneath or exhumed from beneath the Cretaceous deposits is significantly thicker (up to 50 feet thick) and more intensely weathered than that developed on other, younger surfaces. Because this regolith underlies unweathered Cretaceous clays, it is likely of Cretaceous age. This is the Fall Zone surface of Johnson (1931) and, as Johnson claimed, in this area it dips more steeply than the upland (or Schooley surface) (see figure 1-4). In addition to the regional dip of between 50-100 feet/mile there is about 100 feet of gentle relief on the surface (Sandberg et al., 1996), with diabase forming highs. Cretaceous fluvial sands fill the lows around the diabase highs; overlying marginal marine clay (the Woodbridge Clay member of the Raritan Formation) cover both.

48.7 Turn left at light on county route 522 east.

49.2 Cross from weathered diabase onto the Pensauken Formation, which here covers the Cretaceous onlap. Pleistocene valleys incised below the Pensauken Plain in this area show barbed drainage patterns and abandoned valley segments that mark capture of former west-draining Millstone tributaries by northeast-draining Raritan tributaries. The Raritan tributaries eroded headward in response to lowered baselevel established when the lower Raritan was relocated to a more southerly valley at New Brunswick by glacial blockage of its preglacial valley further north during the late Wisconsinan, or, possibly, Illinoian glaciation.

49.6 Cross Amtrak tracks (formerly Pennsylvania Railroad). This is the Northeast Corridor mainline, used exclusively for passenger service.

50.0 Turn right at light onto Ridge Road (still 522 east). We are now on the Pensauken Plain. About 30 to 40 feet of Pensauken have been stripped from the surface here during Pleistocene erosion, based on the elevation of flat-topped Pensauken remnants supporting deep red soils in the area. The Pensauken is about 80 feet thick here, in the rather narrow thalweg of the Pensauken Valley, with a basal elevation between 0-20 feet. The Pensauken includes a basal gravel which is a locally productive domestic aquifer where it overlies Cretaceous clay or clayey saprolite. It is too shallow for public-supply wells but was formerly used for irrigation before urbanization.

51.2 Proceed straight through light in Dayton (still 522 east). Pensauken about 50 feet thick hereabouts.

51.7 Cross under US 130.

53.7 Cross NJ Turnpike. Water-filled pit to left before Turnpike mines Cretaceous sand by dragline from the Magothy Formation (a marginal marine unit of Turonian to Santonian age [about 90-85 Ma]) beneath about 10 to 15 feet of Pensauken .

54.1 Cross county route 535 at traffic light.

54.3 View to left and right across Pensauken Plain, here about 20 feet thick over Cretaceous sand of the Magothy Formation. Thermokarst basin development is poor here compared to the area around stop 5 (see below) due to good subsurface drainage through the Magothy.

54.7 Descend into Manalapan Brook valley, another incised tributary eroding headward from the glacially relocated lower Raritan, capturing former Millstone drainage.

54.9 Cross contact of Pensauken on the Magothy Formation. Base of Pensauken here at about 85 feet.

55.3 Turn right onto Bordentown Turnpike, the small side street just before stop sign, then immediately turn right again on Dayton Road.

55.7 Turn left on county road 619 south (Possum Hollow Road). Exposures of Pensauken to left.

56.0 Turn left into entrance to Herbert Sand and Gravel Co. Park.

STOP 4. Pensauken Formation (Scott Stanford)

Setting: This pit is dug into the basal 20 to 40 feet of the Pensauken Formation (fig. 6-6).The Pensauken is a Pliocene fluvial unit deposited by a large river flowing southwesterly from the Long Island Sound lowland area to the Delmarva Peninsula (see Chapter 1). It likely included Hudson Valley drainage as well as drainage from southern New England, including, possibly, the Connecticut Valley. The Pensauken drainage was established as the Beacon Hill Gravel-upland surface was dissected. Possible superposition of the Hudson on the Palisades diabase and Watchung basalts from the Beacon Hill is indicated by wind gaps at Sparkill (through the Palisades) and at Paterson and Millburn (through the Watchungs) that are on grade with the Pensauken (see fig. 1-7). A broad low in the Palisades north of Staten Island may mark the route of the trunk drainage from the Long Island Sound lowland. In the New York-to-Philadephia segment of the drainage system, the thalweg of the valley, which locally exends below modern sea level, is keyed into the onlap zone of Cretaceous deposits on Newark Basin or Piedmont crystalline rocks. After incision of this valley the river system aggraded and deposited a broad braidplain, which is as much as 140 feet thick over the thalweg between Perth Amboy and Trenton, where it has not been deeply eroded as it has south of Trenton (by the Delaware River) and north of Perth Amboy (by glaciation). The plain extends down the present Delaware Valley and broadens to cover much of the northern Delmarva Peninsula (Owens and Minard, 1979), where it includes the Columbia Formation of Delaware (Jordan, 1964). This period of aggradation may have occured during the early Pliocene eustatic highstand marked by the marine Yorktown and Beaverdam Formations in Delmarva. The Pensauken River was diverted from its valley in the New York City area during the pre-Illinoian glaciation, which may have been in the late Pliocene (see Chapter 1). The diversion may have been by direct glacial blockage and erosion of the valley, or by headward erosion of Atlantic-slope streams through the Coastal Plain upland in the area of the present New York Bight-Hudson Shelf Valley. At the time of diversion, the segment of the valley between New York and Trenton was abandoned and local streams established new drainage patterns on the Pensauken Plain, including the northerly flowing lower Millstone and Raritan rivers. South of Trenton, the Delaware continued to drain southeasterly down the Pensauken Valley, perhaps keeping the Pensauken deposits active in that segment of the valley later than to the north. We are now on the southeastern edge of the main Pensauken Valley, at the junction of the main valley (which we have traversed since mile 49.2) with a northwesterly flowing tributary that drained the Coastal Plain upland to the southeast. The elevation of remnants of the original Pensauken surface to the north and south of us indicates that about 30 to 40 feet of Pensauken material has been removed here during Pleistocene erosion, in addition to another 10 to 20 feet removed by pit excavation above the level of the current working face, so the current exposure is far below any original weathering zone.

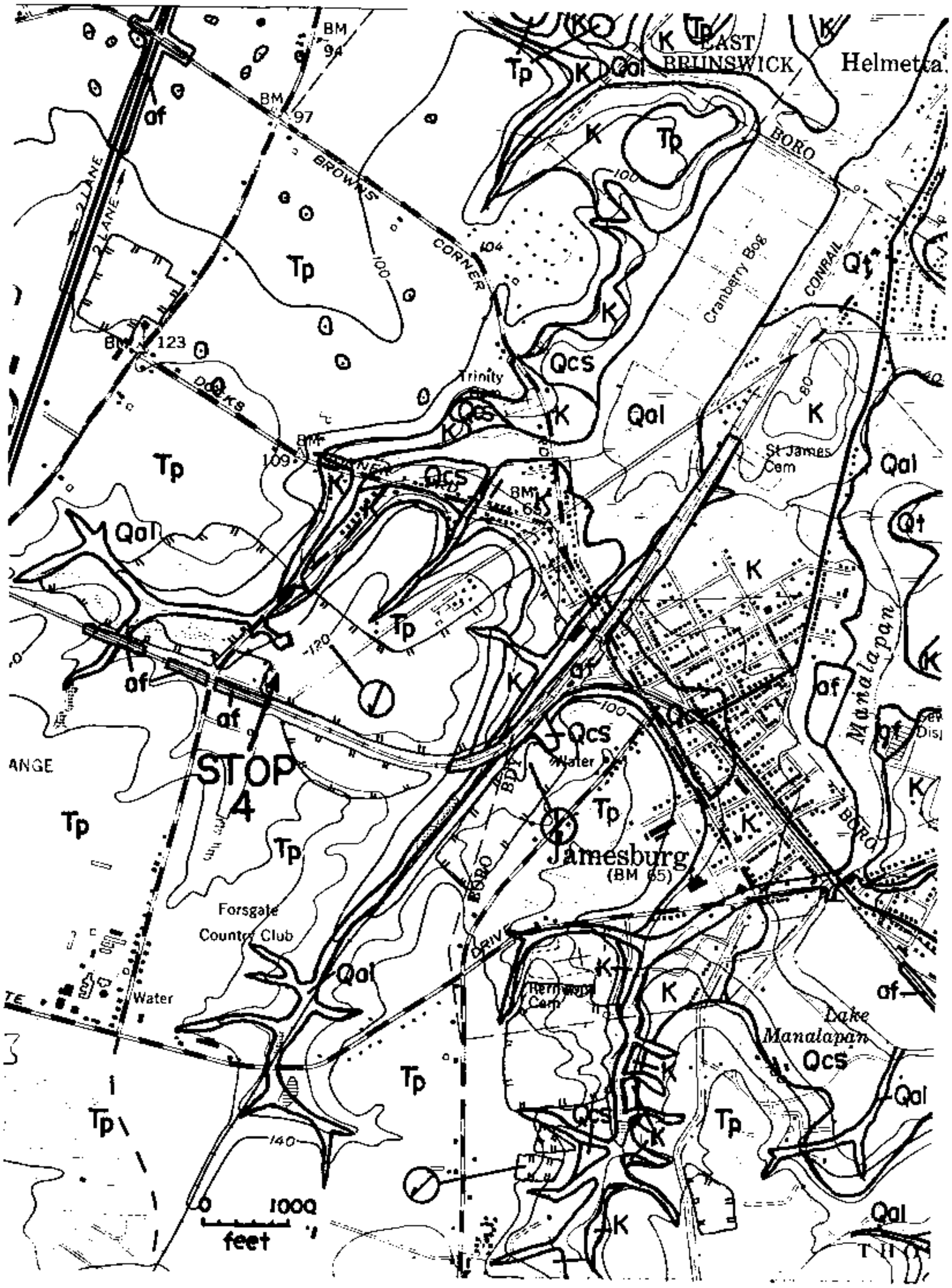


Figure 6-6. Surficial geologic map of area around Stop 4. Solid lines are contacts of surficial units, hachured lines are edges of gravel pits. Contacts are drawn to the base-map topography, which has been substantially altered by excavation in places. Circled arrows show paleoflow directions measured on planar, tabular cross beds in the Pensauken Formation. Circled dots show thermokarst basins. Map units are: af=artificial fill, Qal=alluvium (including some colluvium in heads of drainages), Qcs=sand and gravel colluvium, Qt=fluvial terrace deposits, Tp=Pensauken Formation, K=outcropping Cretaceous sand of the Magothy Formation, overlain by thin, patchy sand and gravel colluvium. Base is U. S. Geological Survey Jamesburg 7.5 minute quadrangle.

Materials: The Pensauken here is a planar-trough to trough-trough cross-bedded reddish yellow glauconitic quartz sand (with a trace of feldspar) and pebble gravel. Gravel clasts are predominantly rounded quartz and quartzite pebbles, with a few dark reddish brown ironstone pebbles and bluish-black cherts. These clasts are iron-stained, giving the characteristic yellow color of the Pensauken, but are otherwise unweathered. There are also a few white decomposed to saprolitized chert, crystalline (probably gneiss and diabase) and sedimentary-rock (sandstone and mudstone) pebbles. Depth of exposed channels is as much as 6 feet but more commonly is less than 2 feet. Scattered channel-bound faults and rotated blocks mark syndepositional deformation, perhaps reflecting bank erosion and scour undercutting. Paleoflow is variable, but the overall sense is southwesterly, as is the case throughout the deposit north of the Delmarva Peninsula (Owens and Minard, 1979; Martino, 1981). Southwesterly paleoflow is particularly strong in the main valley. The variable paleoflow and abundant glauconite in this pit reflect the contribution of the Coastal Plain tributary entering the main valley here. In the base of the pit, on the lower working face, the Pensauken is in sharp contact with the underlying Magothy Formation, a white to reddish purple fine micaceous quartz sand. The Magothy is a marginal marine deposit of Late Cretaceous age. Flaser bedding in the Magothy here suggests a tidal flat origin. The contact is usually marked by an iron-cemented cobble-gravel bed at the base of the Pensauken, with abundant crystalline and sedimentary rock clasts, although only a scattering of cobbles occur on the contact in the present exposure. The paucity of these northerly derived rock types here again likely reflects the dominantly Coastal Plain provenance at this location.

Discussion: The principal accessory mineral in the sand fraction of the Pensauken over most the main valley is feldspar (Bowman and Lodding, 1969; Owens and Minard, 1979), which is commonly weathered to clay; the glauconite here reflects input from the tributaries draining glauconite-rich Cretaceous and Paleogene units to the southeast in the Coastal Plain. The dominance of rounded quartz, quartzite and chert, and the presence of ironstone, in the gravel reflects derivation from polycyclic Coastal Plain gravels, for example, the Beacon Hill and its northerly equivalents in the Long Island-southern New England area (now removed by Pleistocene fluvial and glacial erosion) that covered much of the source area during deposition of the Pensauken. Braided streams are most common today in glacial or arid environments, but the pollen and leaf-fossil data from the Pensauken (Chapter 1) indicate a temperate climate, perhaps slightly warmer than today (although the pollen suggest some cooler intervals toward the top), and the pre-Illinoian till overlaps and erodes the Pensauken, and has a different lithology (a much higher gneiss and sedimentary-rock content), indicating that it postdates the Pensauken. Can a large river system in a temperate, forested setting deposit a braidplain? Does it require backfilling from a eustatic highstand (like that in the early Pliocene) or does reworking of older, extensive quartz sand and gravel deposits in a low-relief landscape lead to a braided channel system? The drowned estuaries that characterize most temperate coastal rivers today don't provide a clear analog to conditions in the Pliocene, before glacial lowstand-interglacial highstand cycles in the Pleistocene greatly modified coastal river landscapes.

56.0 Turn right from pit exit onto Possum Hollow Road.

56.5 Turn right onto Dayton Road.

56.8 Turn left onto Bordentown Turnpike, then immediately left again onto county road 522 west (Browns Corner Road).

57.5 Reascend to Pensauken Plain.

58.0 Turn right onto county road 535 north (Cranbury Road).

58.5 Turn left on Deans-Rhode Hall Road.

58.8 Rise onto a higher remnant of the Pensauken. The highest Pensauken remnants, and deepest, reddest soils, are on subcrop of Cretaceous sand, where there is less surface runoff and spring-sapping than on clay or shale.

58.9 Cross NJ Turnpike. Good thermokarst basins in fields to right after Turnpike.

59.4 Large thermokarst basin to right.

59.6 Turn right into gravel pullout along road.

STOP 5. Thermokarst Basins on Pensauken Formation (Scott Stanford)

Setting: Pensauken sand and pebble gravel is between 15-25 feet thick over a subcrop consisting of a thin feather edge of Magothy Formation sand on the Woodbridge Clay member of the Raritan Formation, which is about 100 feet thick in this area (fig. 6-7) (Stanford et al., 1997a). The Woodbridge Clay is a significant confining unit and impedes drainage in the overlying Pensauken. The belt where the Woodbridge underlies thin (<40 feet thick) Pensauken marks the best-developed thermokarst features in this area. The basins in this belt are numerous, up to several hundred feet in diameter, with over 15 feet of closure in places, and with distinct gravelly rims. Most are ellipsoidal in plan, with the long axis oriented northwest-southeast. Basins similar to those here are common in poorly to moderately drained terrain throughout the Coastal Plain, but are generally more shallow, less numerous, and without well-marked rims. Small, shallow, generally rimless basins of probable thermokarst origin also occur north of the terminal moraine where thin stream terrace sand overlies glacial-lake clay (for example, on terraces deposited on the floors of lakes Hackensack, Passaic, and Pequest, see fig. 1-9).

Materials: The rims and floors of the basins are composed of sand and pebble gravel of the Pensauken Formation. Some of the larger basins have a fill of peat and organic silt and clay. This fill may be as much as 10 feet thick and has yielded a pollen record extending well back into the spruce zone, with minimum dates of 11,950±100 (QL-965) and 11,400±150 (QL-964) (Szabo Pond site of Watts, 1979) and, at another basin, 9,805±62 (GX-17,456) (Stanford et al., 1997a) (fig. 6-7). Although none of the basins has been exposed by excavation, other excavations throughout the area routinely show cryoturbation structures in the upper 8 feet or so of the Pensauken Formation, including ice wedges and, most commonly, involutions.

Discussion: The gravelly substrate, and absence of any windblown sediment in the area, indicate that these basins are not of eolian origin. The radiocarbon dates and pollen record indicate that they predate the Holocene. The absence of organic sediment as a blanket over the area or in the basin substrate would seem to rule out a palsa origin, since palsas (mounds cored by thin ice lenses within permafrost) generally form in bogs (Washburn, 1973). The absence of any wetland or lacustrine sediments in the basin substrate eliminates a closed-system pingo origin, which requires a lake overlying the pingo site prior to pingo growth (Washburn, 1973). Also, the ubiquity of these basins in a variety of landscape positions on interfluves, slopes, and valley bottoms argues against a closed-system pingo origin, since these type of pingos typically occur singly only on flat surfaces. The gravelly substrate, poor groundwater drainage, and ubiquity of the basins across the landscape all argue for an open-system pingo origin (at least for the rimmed basins), where ice cores accrete from freezing of groundwater discharging to the surface under artesian conditions produced by the overlying impermeable permafrost (Washburn, 1973). The northwest-southeast elongation of many of the basins parallels the surface drainage and the expected shallow groundwater flow direction from the local recharge area on the upland to the southeast to the discharge area along Lawrence Brook to the northwest. The basin elongation, and the occasional tendency for the basins to string together in the same direction as the elongation, may therefore reflect ice-lense accretion along groundwater flow paths. The elongation may also reflect preferential wind-driven erosion of thaw-lake shorelines, as suggested by Wolfe (1953), although most of the basins seem too small to show this effect. The many rimless basins in the area, and throughout New Jersey, may reflect differential thawing of permafrost without significant ice-core growth (Wolfe, 1953).

The preservation of these basins, and of involution structures, on flat to gentle slopes indicates that these slopes were not severely eroded under periglacial conditions, but instead were deformed in situ. This stability contrasts with the periglacial mobility of regolith on steep slopes as evidenced by the extensive colluvial deposits at their

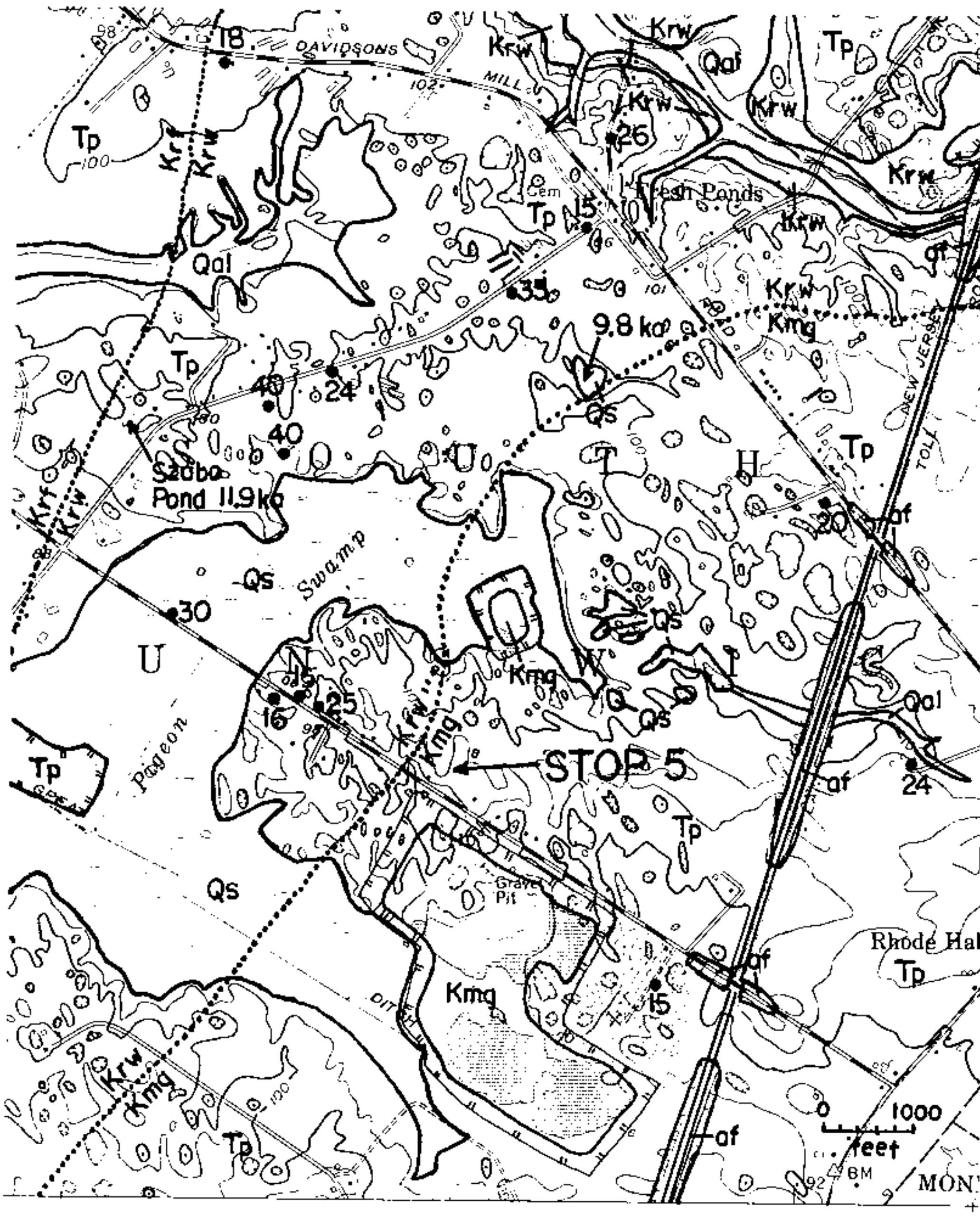


Figure 6-7. Surficial geologic map of area around Stop 5. Solid lines are contacts of surficial units, dotted lines are contacts of Cretaceous units beneath Pensauken Formation. Hachured lines are edges of sand pits. Contacts within pits are drawn as they were in 1991. Numbered solid circles show depth to underlying Cretaceous unit as reported in water-well or test-boring record. Circled dots show thermokarst basins visible on air photos that are not picked up by the base-map topography (10-foot contour interval). Many small basins, particularly in wooded areas, are not shown. A few of the larger basins contain mappable marsh deposits (unit Qs). Units are: af=artificial fill, Qal=alluvium, Qs=swamp and marsh deposits, Tp=Pensauken Formation, Kmg=Magothy Formation (chiefly sand), Krw=Woodbridge Clay member of the Raritan Formation, Krf=Farrington Sand member of the Raritan Formation. From Stanford et al. (1997a). Base is U. S. Geological Survey New Brunswick 7.5 minute quadrangle.

bases (Stop 2). These observations suggest that significant periglacial erosion is largely restricted to steep slopes. On the other hand, the original elevation of the Pensauken surface here, judging from remnants to the east supporting old soils, was between 150-160 feet, indicating 50 to 60 feet of surface lowering since abandonment of the Pensauken Plain in the Pliocene. Did this erosion occur under periglacial conditions or was it accomplished by stream erosion and spring-sapping under more temperate conditions in the late Pliocene and early Pleistocene?

59.6 Continue east (right) on Deans-Rhode Hall Road.

60.6 Crossing Pigeon Swamp. Cretaceous clay (Woodbridge Clay Member of the Raritan Formation) under 15-20 feet of Pensauken here. Pigeon Swamp occupies a former westward drainage network beheaded by headward erosion of Manalapan Brook, as described at 54.7. Drainage today is impeded by the underlying Cretaceous clay and by outcropping diabase at the outlet of the swamp 2 miles to the west of here, where it drains via a barbed pattern into Lawrence Brook, a Raritan tributary.

61.3 Recross thalweg of Pensauken Valley. Pensauken about 60 feet thick here; base at an elevation of 30 to 40 feet.

61.8 Turn right onto US 130 north.

62.2 Cross Lawrence Brook, which drains north to the Raritan. Weathered diabase exposed beneath Pensauken hereabouts. Base of Pensauken at an elevation of about 80 feet. From here to mile 70.5 we will traverse a dissected part of the Pensauken Plain. Pensauken remnants cap interstream flats, with a basal elevation of about 80 to 90 feet, dropping somewhat sharply to the east to 30-40 feet where the valley thalweg is keyed into the onlap of Cretaceous sediments on Newark Basin rocks. Lockatong and Passaic Formation rocks (red and gray shale and hornfels in the Lockatong adjacent to the diabase) crop out in Pleistocene valleys incised below the Pensauken.

67.0 Exit right on ramp onto US 1 north.

68.2 Rutgers University experimental farm on right. Pensauken 10 to 20 feet thick over Passaic shale.

69.5 Exit right under bridge onto NJ 18 north.

70.5 View to right of Raritan River, here set in a narrow gorge cut 50 to 100 feet into shale. This gorge, which extends from the Bound Brook area (about 8 miles upstream) to about 2 miles downstream from here (where the valley broadens after crossing onto Cretaceous deposits), was cut when the Raritan was diverted from its preglacial valley, which ran northeasterly from Bound Brook to the Newark Bay area. The preglacial valley was filled with moraine and meltwater deposits during the late Wisconsinan and, possibly, Illinoian, glaciations. The lower Raritan was rerouted across a low shale divide (the divide was probably about here, judging from the height of the shale upland in this area) and now drains into Raritan Bay just beyond the southernmost position of the terminal moraine at Perth Amboy. The upland flats on either side of the gorge are part of the Somerville strath surface, as described at mile 29.8, with isolated Pensauken remnants present at elevations above 140 feet.

70.8 Outcrop of Passaic Formation on left. This is part of the type section of the Brunswick Formation, now elevated to group status.

71.8 Cross under Amtrak bridge. City of New Brunswick to left. A deposit of quartz sand and gravel, derived from Pensauken material, caps a rock-cut terrace at 40 to 50 feet in elevation in downtown New Brunswick and shows northerly paleoflow. It likely marks local drainage predating the Raritan diversion.

72.6 Move into left lane.

72.8 Cross Raritan River.

73.1 Enter left lane and turn left onto NJ 18 north.

73.4 Parkland to left is on the Raritan floodplain. The well-developed lower terrace deposits in the Raritan Valley upstream from Bound Brook are only sparsely present in the rerouted valley downstream from Bound Brook (in places, as just a patchy lag on a narrow rock-cut terrace above the floodplain). The upper terrace deposits are absent altogether in this reach.

75.7 Lower terrace to left of road, here a silty pebbly sand.

76.7 Exit right onto I-287 north after overpass.

77.2 Cross Raritan River.

78.9 First Watchung Mountain in front, formed on Jurassic basalt (the Orange Mountain Basalt). This marks the northern limit of the Pensauken Plain and Somerville strath.

79.4 Recross Raritan River, here upstream from diversion point. Note the wide floodplain and bordering lower terrace here, in contrast to the narrow valley downstream from Bound Brook.

80.0 View to right of Chimney Rock Gap cut through First Watchung Mountain. This gap drains the intermontane valley between First and Second Watchung Mountains and may have conducted lake drainage from pre-Illinoian glacial Lake Watchung (see figure 1-9). Wide basalt-rich upper and lower terrace deposits occur downstream from the gap and grade to their counterparts in the main valley.

81.1 Cross onto apron of basalt colluvium. Colluvium is continuous and as much as 50 feet thick along the base of First Mountain.

81.3 Cross US 22.

82.3 Cross back from colluvium onto the Somerville strath. Several Pensauken remnants to the north of here have bases at about 150 feet and tops at 160-180 feet. The Pensauken here is overlapped by the terminal edge of the pre-Illinoian till, and both the till and Pensauken are overlapped by basalt colluvium of probable late Pleistocene age.

87.5 Exit left onto I-78 west at exit 21B.

89.3 Cross onto upper terrace along North Branch of the Raritan River.

90.1 Cross North Branch Raritan River.

90.7 Roadcut through shale hill capped with thin pre-Illinoian till. All topography below these hilltop caps is free of till and was presumably cut by dissection after the pre-Illinoian glaciation.

92.2 Cross Lamington River.

93.8 View to left of Cushetunk Mountain, a circular ridge formed on a combination sill-dike intrusion of Jurassic diabase. The pre-Illinoian margin, as defined by till patches and erratic distribution, wraps around the north end of the mountain and extends several miles down the shale lowlands on either side, as befits a lobate glacial margin. From here to mile 98.5 pre-Illinoian till caps shale hilltops and interfluves. These remnants define a gently sloping surface with about 80 feet of relief that is incised by the modern drainage, which has up to 150 feet of relief.

95.5 View to right of Highland Front, where Proterozoic gneiss is in fault contact with Newark Basin rocks.

97.7 View to left of one of the dams for Round Valley Reservoir, which fills the valley enclosed by Cushetunk Mountain.

98.5 Cross fault contact of Passaic Formation and Proterozoic gneiss.

99.9 Cross sedimentary contact of Proterozoic gneiss and Paleozoic sedimentary rocks.

100.9 Exit right onto NJ 31 north at exit 17.

101.9 Outcrops of Cambrian and Ordovician carbonate rock (Allentown Formation) in roadcut on left. Pre-Illinoian till continues to cap flat interfluves through here. Musconetcong Mountain in front. The Pattenburg tunnel on the fomer Lehigh Valley Railroad (mentioned at mile 36.0) bores through this mountain near the prominent I-78 roadcut visible ahead.

102.2 Cross South Branch of the Raritan River.

102.5 Dike for Spruce Run Reservoir to left.

103.6 Spruce Run Reservoir to left. Cross sedimentary contact from Paleozoic rocks to Proterozoic gneiss. Pre-Illinoian striations at S33W on Cambrian quartzite were described in this cut by H. B. Kummel (circa 1900) but are no longer exposed.

104.3 Light at intersection with Van Syckels Road, on floodplain gravel of Spruce Run.

104.5 Enter Glen Gardner. This is a 3-mile-long gap cut through Musconetcong Mountain, partly occupied and deepened by Spruce Run, a local drainage. The glen heads at a wind gap at 550 feet in elevation, which also roughly corresponds to a break in slope to a more open valley within the lower glen, about 150 to 200 feet above Spruce Run. Glen Gardner lines up with wind gaps to the north near Oxford, and with the Delaware Water Gap (see figure 1-4), suggesting a pre-Pliocene (late Miocene?) river alignment before strike-valley capture and assembly of the present Delaware drainage, which has likely been in place since the late Pliocene (see Chapters 2 and 3).

107.4 Summit of wind gap. Cross former Central Railroad of New Jersey tracks (now abandoned). Descend into Musconetcong Valley, crossing thrust fault contact of Proterozoic gneiss and Paleozoic carbonate rock about halfway downslope.

108.5 Cross Musconetcong River, bordered here by a narrow late Wisconsinan terrace, in part consisting of glaciofluvial sediment from the terminal moraine upstream at Hackettstown.

109.2 Turn right at light onto county route 632 north. Travel on valley strath surface cut into carbonate, and, locally, shale bedrock and capped in places by pre-Illinoian till. Views to right into Musconetcong inner valley show depth of Pleistocene incision below this surface.

110.1 View to left of Pohatcong Mountain, which forms the west side of the Musconetcong Valley here. The Illinoian terminus was pinned on the west (far) side of Pohatcong Mountain in this area.

110.8 Erratic boulders of quartzite and gneiss along road, from pre-Illinoian till. Same at 112.1 and 113.1.

113.7 Turn right at stop sign onto NJ 57 east.

114.0 Turn left at light onto county road 629 west.

114.6 Flat interfluve cut into Ordovician shale (Martinsburg Formation), capped by pre-Illinoian till.

114.8 Turn left onto Brickyard Road.

115.3 Bridge across Conrail track.

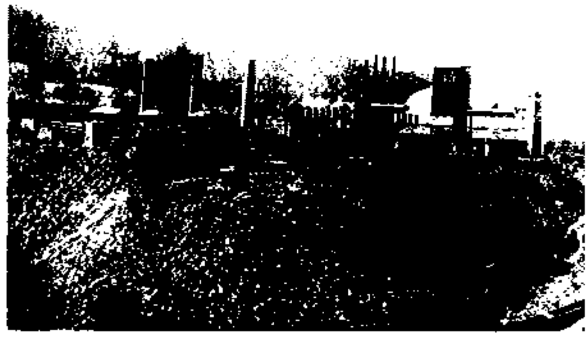
115.4 Park in gravel turnaround.

STOP 6. Pre-Illinoian Till and Weathered Slate

Ron W. Witte
(N.J. Geological Survey)
and
Richard K. Shaw
(Natural Resources Conservation Service)

Setting: Stop 6 is at an abandoned "clay" pit near Port Murray, Warren County, New Jersey (fig. 6.8). It lies on the western side of Musconetcong Valley , in the Washington, N.J., 7 ½ minute quadrangle. The physiography of the area is largely the result of differential weathering on chiefly southwest-trending fold and thrust belts of dolostone, slate and siltstone, and gneiss and granite. The uplands rise as much as 700 feet above the floor of the valley, and they are underlain by gneiss and granite. Topography is rugged and the landscape is deeply dissected. 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. North of the late Wisconsinan glacial limit, rock outcrops are very abundant because in many places the preglacial cover of loose rock and soil has been removed by the erosive action of the last ice sheet. South of the limit, rock outcrops are fewer and the topography is not as rugged because in many places the rock surface is covered by saprolite and colluvium. The valley is underlain by slate, siltstone, dolostone, and minor limestone. The higher areas are held up by slate and siltstone, and the landscape consists of gently to moderately rolling hills and swales. In places where the bedrock is covered by a thin mantle of Pre-Illinoian glacial drift, it is weathered to depths exceeding 20 feet (fig.6.9). In places where the drift has been stripped by erosion, the surface is typically underlain by thin slate-chip colluvium. Areas underlain by carbonate rock are typically 100 feet lower, have less relief, and in many places are also capped by the older glacial drift. Records of nearby wells (Witte and Stanford, 1995) show weathered carbonate rock to depths exceeding 50 feet. The modern river valley is cut down in rock and colluvium and it is very narrow. It chiefly follows the strike of the carbonate rock and it lies 100 to 200 feet below the main valley floor.

The Port Murray pit was formerly worked by the National Fireproofing Co. (see unnumbered figure below) in the late 19th and first half of the 20th century. Based on notes by Meredith Johnson (1949, on file at the N.J. Geological Survey, Trenton, New Jersey) common and face brick was made from a screened mixture of weathered slate and older till. The material was largely dug with a pick and shovel, loaded on a tram car, and taken to the processing plant where it was ground by a Huntingdon mill. The ground material was mixed with water and fed to a die machine that compressed the clayey mixture into a large bar. The compressed ingot was cut into 18 bricks by a mechanical cutter and then loaded on a tram car and sent off to the driers. The average production per day was about 42,000 bricks.



National Fireproofing Co. Port Murray Plant, 1901.

Materials: Surficial geologic materials in the Port Murray area include; weathered bedrock, Pre-Illinoian till, colluvium, and alluvium (fig. 6-8) It was reported by Meredith Johnson (1949, permanent notes on file at the New Jersey Geological Survey, Trenton, New Jersey) that there was as much as 25 feet of old glacial drift overlying weathered slate of the Martinsburg Formation at the Port Murray pit. Based on the degree of weathering exhibited by the clayey drift it was thought to have been laid down during the oldest glaciation. Just northwest of the pit lies a large colluvial apron that consists of crystalline material shed off Pohatcong Mountain. In Musconetcong Valley colluvium of Wisconsinan age overlies weathered colluvium of presumably pre-Wisconsinan age, and a truncated red soil marks the contact between the two (fig. 6.9). Bedrock exposed in the pit is the Bushkill Member of the Martinsburg Formation. Where it is not weathered, it is a dark- to medium-gray, thinly-bedded slate containing thin beds of graywacke and graywacke siltstone. The attitude of bedding is N 10^0 E 75^0 SE (bedrock geology from Drake and others, 1994).

Section 1. - - Port Murray Formation, approximately .8 km southwest of Port Murray on the west side of the Erie-Lackawanna railroad in the northwest corner of an abandoned clay pit, Washington 7 1/2 minute quadrangle, SE 1/4.

0.0 to 3.0 meters - Till, 5YR 5/6-8 sandy silt with some clay, high dry strength, high compaction. Contains approximately 2 to 3 percent gravel clasts by volume. Gravel clasts are highly weathered, consist of a mix of subangular to subrounded chert, gneiss, shale, sandstone, quartzite, and conglomerate, and all exhibit extensive rubification. Gneiss and shale clasts are throughly decomposed or have very thick weathering rinds. The more resistant lithologies exhibit thin weathering rinds, ferro-manganese staining is pervasive on clasts and the surface of subvertical joints. In places color is variegated and alternates between reddish yellow and yellowish brown (10YR 5/8).

3.0 - 5.9 meters - Till, 10YR 5/6 - 5/8 sandy silt with some clay, high dry strength, high compaction. Contains approximately 5 percent gravel clasts by volume. Similar characteristics as overlying unit.

5.9 - 6.8 meters - Till, 10YR 5/6 - 5/8 sandy silt with some clay, high dry strength, high compaction. Contains approximately 10 percent gravel clasts by volume (> 60 percent shale). Rubification is minor.

6.8 - 7.9 meters - Weathered shale, 2.5YR 5/8 clayey silt with many highly weathered, and ferro-manganese-stained clasts of shale.

Below 7.9 meters - Shale, highly cleaved, and fractured regolith. Pervasive reddish-colored (2.5YR 5/8) stain on joint and cleavage surfaces.

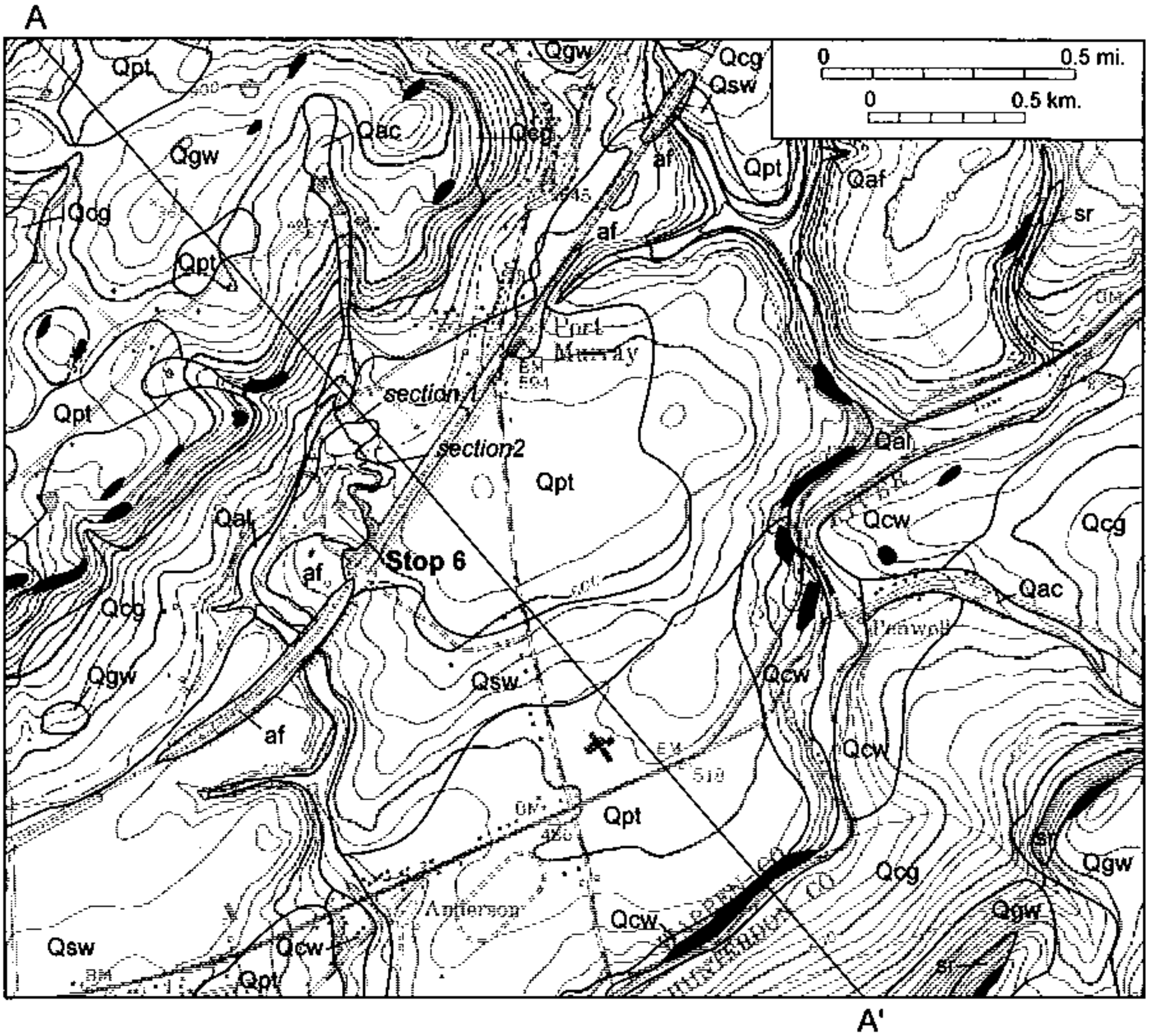


Figure 6.8. Surficial geologic map of part of the Washington, N.J. 7 1/2 minute quadrangle near Port Murray, New Jersey. List of map units: af - artificial fill, Qal - alluvium, Qac - undifferentiated alluvium and colluvium, Qaf - alluvial fan, Qpt - Pre-Illinoian till, Qcg - granitic and gneissic colluvium, Qgw - weathered gneiss and granite, Qsw - weathered slate, Qcw - weathered dolostone (weathered bedrock units include thin colluvium), sr - shallow rock, chiefly rock waste on steep slopes. Shaded areas represent extensive bedrock outcrop. Section A - A' shown on figure 6.9.

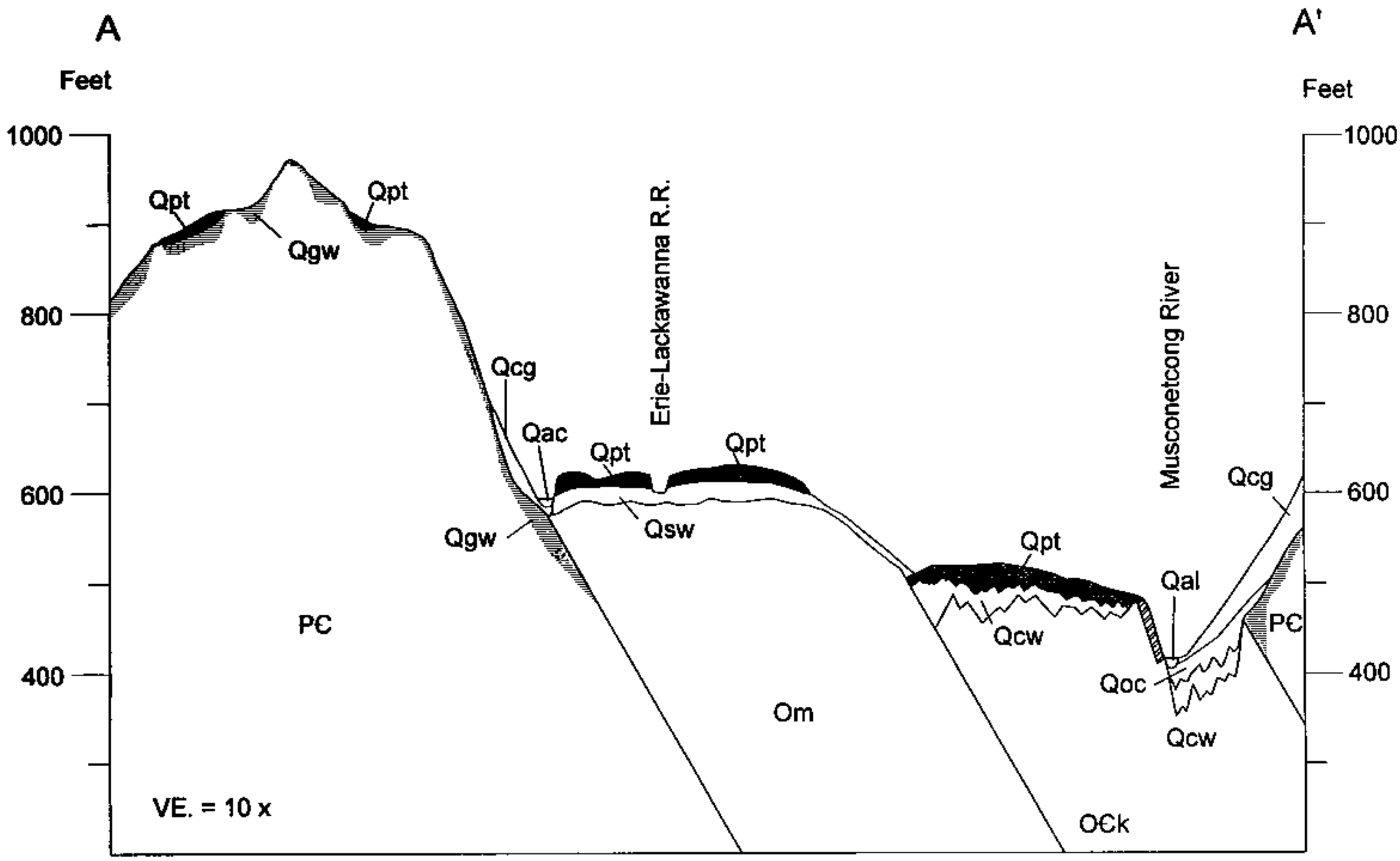


Figure 6.9. Geologic cross-section A-A' from figure 6.8. List of map units: Qal - alluvium, Qac - undifferentiated alluvium and colluvium, Qpt - Pre-Illinoian till, Qcg - granitic and gneissic colluvium, Qgw - weathered gneiss and granite, Qsw - weathered slate, Qcw - weathered dolostone (weathered bedrock units include thin colluvium), PЄ - chiefly gneiss and granite, OЄk - dolostone, and Om -slate and graywacke siltstone. Scale 1;24,000.

Section 2. - - Port Murray Formation, approximately .8 km southwest of Port Murray on the west side of the Erie-Lackawanna railroad in the northeast corner of an abandoned clay pit, Washington 7 1/2 minute quadrangle, SE 1/4.

0.00 - 0.79 meters - Till, 7.5YR 5/6-8 sandy silt with some clay, high dry strength, high compaction. In places minor mottling (7.5YR 6/4) noted. Contains approximately 2 to 3 percent gravel clasts by volume. Gravel clasts are highly weathered, and consist of a mix of subangular to subrounded chert, gneiss, shale, sandstone, quartzite, and conglomerate, and all exhibit extensive rubification. Gneiss and shale clasts are throughly decomposed or have very thick weathering rinds. The more resistant lithologies exhibit thin weathering rinds, Ferro-manganese staining is pervasive on clasts and the surface of subvertical joints. In places there are small clots (< 3cm) of ferro-manganese oxide.

0.79 - 1.20 meters - Near base of unit, there are thin lenses (< 3 inches thick, and < .6 m wide) of clayey silt with shale fragments. These beds are moderately deformed and appear to have been sheared.

1.20 - 1.28 meters - Clayey silt with many weathered shale clasts. Shale clasts are chiefly oriented on their sides in a horizontal position.

1.28 - 5.49 meters - Weathered shale, buff to pink to light yellow clayey silt, high dry strength, high compaction. Contains many fragments of weathered shale, some quartz, and many small clots of ferro-manganese oxide. Shale clasts do not appear to have a preferred orientation. Material has a subvertical platy structure with minor clay films and Fe-Mn staining on plate surfaces. A few large vertical joints extend downward from the overlying till into the residuum. They are as much as 3 cm wide and contain sandy silt derived from the overlying till. Shale clasts increase in percentage toward the base of the outcrop.

Miscellaneous comments. Near section 2, a large (15' x 12' x 10') subrounded, striated boulder of gneiss lies on the pit floor. The surface of the shale regolith is broadly curvilinear and it appears to rise in a southerly direction.

Description of soil at section 1 (by R.K. Shaw)

0-4 inches	A	Dark yellowish brown (10YR 4/4) gravelly loam; fine to medium granular structure; friable; 12 % (volume) shale channers; 5 % subrounded gneiss gravel; gradual irregular boundary.
4-18 inches	2Bt1	Strong brown (7.5YR 4/8) gravelly loam; moderate fine to medium subangular blocky structure; friable; distinct clay films in pores and on coarse fragments; few black manganese coatings on coarse fragments and throughout matrix; 15 % shale channers; 4 % subrounded gneiss gravel; clear smooth boundary.
18-46 inches	2Bt2	Yellowish red (5YR5/8) gravelly, sandy clay loam with patches of red (2.5YR 4/8) and strong brown (7.5YR 4/8); moderate medium to coarse subangular blocky structure; firm; distinct clay films in pores and on coarse fragments; black manganese coatings common on coarse fragments and throughout matrix; 18 % subrounded gneiss gravel; clear smooth boundary.
46-80 inches	BC	Strong brown (7.5YR 4/8) to reddish yellow (7.5YR 6/8) sandy clay loam; moderate medium to thick platy structure, parting to medium subangular blocky; firm; clay films in pores and on coarse fragments and throughout matrix; local areas with mica flakes common; 8 % subrounded gneiss gravel.

Particle Size Analysis (hydrometer method)					sieve percent by weight
percent of < 2mm fraction					
Horizon	**sand**	silt	clay	texture	> 2mm
A	48.8	30.7	20.5	loam	31.6
2Bt1	46.9	35.0	18.1	loam	33.6
2Bt2	52.7	15.8	31.5	sandy clay loam	31.1
2BC	54.4	20.3	25.3	sandy clay loam	11.1

Mineralogy (sample collected at section 1, depth = 4.2 meters; analyzes provided by F.L. Muller, New Jersey Geological Survey)

Texture - gravel (10 %), sand (36 %), clay/silt (54 %).

Mineralogy (based on visual estimates)

granules - 33 % goethitic ironstone, 33 % white to gray/white shale, 34 % quartz.

sand (light fraction, < 2.89 SD) - 10 % fine-grained sandstone w/ iron (goethitic cement) and/or stained shale, 25 % goethitic ironstone, 1 % hematite, 1 % chert, 35 % feldspar (XRD shows both microcline and orthoclase), 2 % biotite, 2 % granitic fragments, 1 % white bleached siltstone, 1 % red shale, 20 % quartz, 3 % unidentified.

sand (heavy fraction, >2.89 SD) - 40 % goethitic ironstone, 2 % muscovite, 15 % ilmenite, 3 % biotite, 20 % magnetite, 2 % aluminosilicates, 1 % pyrite, 1 % zircon, 5 % quartz, 10 % unidentified, trace: tourmaline, rutile, monazite, pyroboles (very rare).

Discussion: The mineralogy, texture, clast shape, and the occurrence of faceted and striated boulders show that the drift at the Port Murray "clay" pit is till. Furthermore, its position on the landscape and on deeply-weathered bedrock, and its degree of weathering compared to other till of similar parent material, indicates the drift is Pre-Illinoian age. The age of this glaciation is uncertain; it may represent single or multiple glaciations during the early to middle Pleistocene (2.0 - 0.8 ma). Recent work on glacial lake-bottom deposits in the West Branch Susquehanna River valley by Gardner and others (1994) showed that these deposits were laid down during a period of reversed magnetic polarity. This places the age of the deposits at older than 788 ka, and based on the oxygen isotope record and the position of the deposits they suggested that the deposits were laid down during the Pre-Illinoian F or G glaciation.

Incision of the "Port Murray surface" is also an indication of the drift's antiquity. The present position of the older till (figs. 6.8 and 6.9) lies well above the modern drainage in areas protected from alluvial and hillslope erosion. The position of Illinoian outwash in the modern Delaware River and Musconetcong River valleys (Witte and Stanford, 1995) suggests that the landscape had been lowered or nearly lowered to modern levels by the Illinoian glaciation. The difference in the distribution of Illinoian and Port Murrayan glacial drift suggests that uplift and/or a drop in sea level in post-Port Murrayan time had lowered base level. The Delaware River and its tributaries adjusted by downcutting as much as 100 feet. 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 narrow, rock-walled valleys through which the Delaware and Musconetcong Rivers flow seem to support the above hypothesis.

The presence of glacial till of varying ages and similar parent material allows for the examination of the time factor in soil development. Several reports have detailed the degree of weathering in till in New Jersey (MacClintock, 1940;

Tedrow, 1954). Tedrow (1986) depicted an idealized depth of weathering for the late Wisconsinan (2.5 feet), Illinoian (7 to 8 feet), and Pre-Illinoian (15 to 16 feet) till. Other soil characteristics correlated with increasing age that may apply to the Port Murray drift include: 1) an increase in the extent of surface alteration of sand-sized quartz grains (Douglass and Platt, 1977), 2) a decrease in the amount of clay-sized quartz in till composed of Pocono sandstone (Tedrow, 1954), 3) an increase in the amount of pedogenic chloritization of 2:1 clays, in till composed of the Martinsburg Shale (Novak and others, 1971), and 4) an increase in iron oxide formation, increase in total clay and fine clay to total clay ratio, and an increase in the depth and thickness of the argillic horizon in some Pennsylvania till soils (Levine and Ciolkosz, 1983).

115.4 Retrace route on Brickyard Road.

116.0 Turn right onto county road 629 east.

116.8 Turn left at light onto NJ 57 east.

117.5 Descend to Musconetcong River.

118.0 Shale of the Martinsburg Formation crops out on left (to 118.5); narrow Musconetcong floodplain to right.

119.6 On late Wisconsinan terrace here (to 120.1).

120.6 Ascend from incised valley onto carbonate upland.

121.5 Illinoian terminus crosses the valley hereabouts but, although there is some Illinoian till along the valley walls in swales just beyond the colluvial aprons, the bedrock upland in the center of the valley (mostly shale here) has no mappable till, and, unlike carbonate-floored valleys, there is no Illinoian moraine here. Perhaps lower permeability of the shale has led to extensive surface runoff that stripped the till.

122.6 Descend to the late Wisconsinan glaciofluvial terrace.

123.5 Turn left at light onto NJ 182. Continue on terrace, here a cobble gravel.

124.4 Rise from terrace onto carbonate upland.

124.5 Turn right at light onto US 46 east.

124.6 Cross Musconetcong River and return to terrace. Terrace gravel about 30-50 feet thick here, over carbonate rock and carbonate residuum. The terminal moraine, from which the terrace was deposited, is about 1 mile upstream (to the left) from here.

125.2 View over field to right shows gneiss colluvium and an alluvial fan grading to the terrace surface, suggesting contemporaneity.

125.6 Cross fault contact of Paleozoic carbonate rock and Proterozoic gneiss and begin ascent up Schooleys Mountain.

127.2 Top of Schooleys Mountain. Gently sloping terrain here is veneered with Illinoian till and a few patches of stratified drift. The Illinoian terminus, as defined by the till patches, is just to the south of US 46. On the broader, less dissected part of Schooleys Mountain further south, there are patches of pre-Illinoian till on the flattest areas.

128.6 Cross outlet channel for glacial Lake Budd, now the headwater of the South Branch of the Raritan River.

129.2 Budd Lake on left. Budd Lake occupies the headwater area of a north-draining preglacial valley that was dammed by the late Wisconsinan terminal moraine, and by underlying late Wisconsinan and Illinoian lacustrine valley-fill sediments. A 60-foot pollen core taken by Harmon (1968) on the far shore of the lake, along the edge of a floating bog, yielded a pre-advance date of 22,870±720 (I-2845) at 37 feet and yielded pre-Pleistocene exotic pollen taxa from 46-60 feet, possibly from pre-Illinoian glacial-lake clay (see Chapter 1). If not reworked, these pollen taxa suggests a Pliocene age for the pre-Illinoian glaciation.

129.8 North end of Budd Lake. A small glacial delta fronts the terminal moraine to the left here. Most of the delta was overrun during ice advance and is covered by till of the moraine.

131.5 Cross onto terminal moraine. The filled preglacial valley mentioned at 129.2 parallels US 46 just to the west, and is about 0.5 mile wide and 2 miles long. It is filled with a top-to-bottom sequence of late Wisconsinan till of the moraine over late Wisconsinan deltaic sand over late Wisconsinan and Illinoian lake-bottom clay over thick Illinoian and pre-Illinoian? lacustrine-fan? sand and gravel (Stanford et al., 1997b). Total thickness of fill is as much as 250 feet. The basal sand and gravel here, as in other valley fills along the terminal moraine, is a productive confined aquifer.

132.2 Cross under I-80. Stay in right lane on US 46 east. Continue through terminal moraine, which overlaps a bedrock ridge to the right.

133.0 Junction with US 206. Stay on US 46 east at circle. Still in moraine.

134.2 Cross under I-80.

134.8 Turn left into hotel parking lot. End of Day 1.

REFERENCES

Bowman, J. P. and Lodding, W., 1969, The Pensauken Formation--a Pleistocene fluvial deposit in New Jersey, *in* Subitzky, S., ed., Geology of selected areas in New Jersey and eastern Pennsylvania and guidebook of excursions: New Brunswick, N. J., Rutgers University Press, p. 3-6.

Clark, G. M. and Hedges, J., 1992, Origin of certain high-elevation local broad uplands in the central Appalachians south of the glacial border, U. S. A.--a paleoperiglacial hypothesis, *in* Dixon, J. C., and Abrahams, A. D., eds., Periglacial geomorphology: New York, John Wiley, p. 31-61.

Cleaves, E. T., 1989, Appalachian Piedmont landscapes from the Permian to the Holocene: Geomorphology, v. 2, p. 159-179.

Davis, W. M., and Wood, J. W., 1889, The geographic development of northern New Jersey: Proceedings of the Boston Society of Natural History, v. 24, p. 365-423.

Douglas, L. A., and Platt, D. W., 1977, Surface morphology of quartz and age of soils. Soil Sci. Soc. Am. J. 41 : 641-645.

Fullerton, D. S., Stratigraphy and correlation of glacial deposits from Indiana to New York and New Jersey: Quaternary Science Reviews, v. 5, p. 23-37.

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

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

Jordan, R. R., 1964, Columbia (Pleistocene) sediments of Delaware: Delaware Geological Survey Bulletin 12, 96 p.

Kirkham, W. C., 1976, Soil survey of Somerset County, New Jersey: U. S. Government Printing Office, 114 p.

Levine, E. R., and Ciolkosz, E. J., 1983, Soil development in till of various ages in northeastern Pennsylvania, Quaternary Research, 19 : 85 - 99.

MacClintock, Paul, 1940, Weathering of the Jerseyan till. Bull. Geol. Soc. Amer. 51 : 103 - 116.

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

Neuman, R. P., 1980, Evidence for pre-Wisconsinan (Jerseyan?) glacial deposits in the Rocky Hill-Kingston area, New Jersey: Bulletin of the New Jersey Academy of Sciences, v. 25, no. 1, p. 12-16.

Novak, R. J., Motto, H. L., and Douglas, L. A., 1971. The effect of time and particle size on mineral alteration in several Quaternary soils in New Jersey and Pennsylvania, *In* Paleopedology: Origin, Nature and Dating of Paleosols, D.H. Yaalon, ed. Hebrew University Press, Jerusalem, pp. 211 - 224.

Owens, J. P., and Minard, J. P., 1979, Upper Cenozoic sediments of the lower Delaware Valley and northern Delmarva Peninsula, New Jersey, Pennsylvania, Delaware, and Maryland: U. S. Geological Survey Professional Paper 1484, 39 p.

Owens, J. P., and 7 others, 1995, Geologic map of New Jersey: central sheet: U. S. Geological Survey Open File Report 95-253, scale 1:100,000.

Parker, R. A., and Houghton, H. F., 1990, Bedrock geologic map of the Monmouth Junction quadrangle, New Jersey: U. S. Geological Survey Open File Report 90-219, scale 1:24,000.

Pavich, M. J., 1989, Regolith residence time and the concept of surface age of the Piedmont "peneplain": Geomorphology, v. 2, p. 181-196.

Priesnitz, K., 1988, Cryoplanation, *in* Clark, M. J., ed., Advances in periglacial geomorphology: New York, J. W. Wiley, p. 49-67.

Sandberg, S. K., and 4 others, 1996, Geophysical investigation of the Potomac-Raritan-Magothy aquifer system and underlying bedrock in parts of Middlesex and Mercer counties, New Jersey: N. J. Geological Survey Report GSR 37, 33 p.

Stanford, S. D., and 4 others, 1997a, Geologic map of the New Brunswick quadrangle, Middlesex and Somerset counties, New Jersey: N. J. Geological Survey Open File Map 23, scale 1:24,000.

Stanford, S. D., Stone, B. D., and Witte, R. W., 1997b, Surficial geology of the Stanhope quadrangle, Morris and Sussex counties, New Jersey: N. J. Geological Survey Open File Map 22, scale 1:24,000.

Stone, B. D., Stanford, S. D., Witte, R. W., in press, Surficial geologic map of northern New Jersey: U. S. Geological Survey Miscellaneous Investigations Series Map, scale 1:100,000.

Tedrow, J.C.F., 1954, Clay minerals in three podzol profiles, Soil Sci. Soc. Am. Proc. 18 : 479 - 481.

Tedrow, J. F., 1986, Soils of New Jersey: Malabar, Florida, Krieger, 479 p.

Washburn, A. L., 1973, Periglacial processes and environments: London, Edward Arnold, 320 p.

Watts, W. A., 1979, Late Quaternary vegetation of central Appalachia and the New Jerey Coastal Plain: Ecological Monographs, v. 49, p. 427-469.

Wolfe, P. E., 1953, Periglacial freeze-thaw basins in New Jersey: Journal of Geology, v. 61, p. 131-141.

CHAPTER 7

ROAD LOG AND DESCRIPTION OF FIELD STOPS

SUNDAY- JUNE 1, 1997

Ron W. Witte
New Jersey Geological Survey
CN 427
Trenton, New Jersey

We will assemble in the Days Inn parking lot near the exit onto US Route 46 at 8:00 a.m.. Travel today will be by private vehicles, and we also have the use of a Rutgers University passenger van. The final stop will be near Milford, Pennsylvania, approximately 9 miles south of Interstate Route 84, and 28 miles north of Interstate Route 80. If you plan to travel homeward through the starting area, please use the van or car pool to keep the number of vehicles to a minimum. These vehicles will return to the Days Inn at Ledgewood.

Today's route (fig. 1) will take us along a northwest traverse of the New Jersey Highlands, Kittatinny Valley, Kittatinny Mountain, and Minisink Valley; an area covered by the late Wisconsinan glaciation. The main topics of the field trip will include: 1) late Wisconsinan recessional history of the Kittatinny and Minisink Valley ice lobes, 2) origin of the Ogdensburg-Culvers Gap moraine, 3) late history of the Culvers Gap River and abandonment of Culvers Gap, 4) postglacial alluvial history of Minisink Valley, and 5) archaeology of Minisink Valley.

The traveling distance to Stop 2-1 at Culvers Gap is 26 miles. If the caravan is small, I will try to keep everyone together. If you become disconnected from the main group stay on Route 206 and head north to Culvers Gap. Caffeinated beverages and foodstuffs loaded with carbohydrates will be served at stop 2-2 (Park Headquarters, Stokes State Forest).

Mileage **Route Description**

0.0 Turn right onto US 46 west from hotel parking lot (the road log up to mile 8.7 was written by S.D. Stanford)

1.7 Exit right onto US 206 (NJ 183) north at circle.

1.8 Cross NJ Transit (formerly Lackawanna Railroad) tracks. 50-foot railroad cut to right is entirely in late Wisconsinan till. We are on the axis of a filled preglacial valley, here about 150-200 feet thick. Well logs indicate till as much as 80 feet thick overlies lacustrine deposits that were probably deposited in an ice-dammed lake during the late Wisconsinan advance. The buried lacustrine deposits are a productive aquifer in places.

2.2 Cross the Musconetcong River and former route of the Morris Canal, now obscured. Lake Musconetcong, dammed in part by the terminal moraine, to the right. The Morris Canal, built in the 1820s to carry coal from Pennsylvania to New York City, ran from Phillipsburg on the Delaware River to Jersey City on the Hudson, using a unique series of inclined planes rather than locks. The level of Lake Musconetcong was artificially raised to provide part of the canal route. After crossing the river, leave the terminal moraine and ascend a bedrock hill.

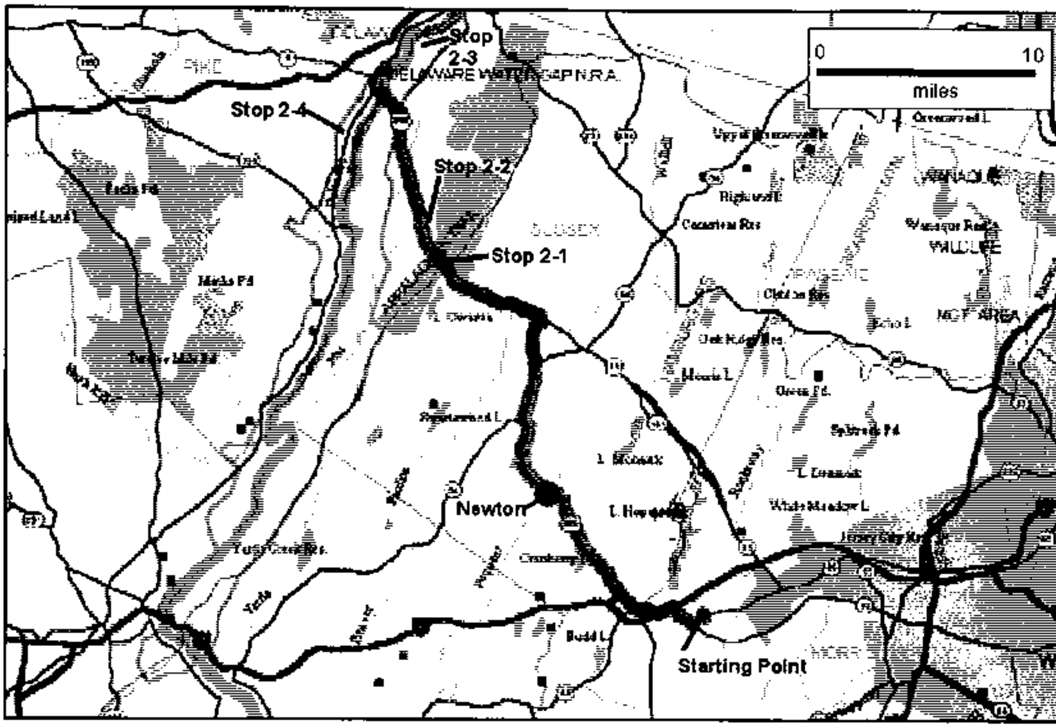


Figure 1. Route for field trip on Sunday, June 1, 1997 and location of stops.

3.3 Merge into US 206 bypass entering from left.

4.1 Cross Lubbers Run, here inset into a low till-over-rock upland but elsewhere bordered by glaciofluvial terrace deposits.

5.2 Quarry in gneiss on right. We are ascending to the Andover-Ledgewood wind gap, here greatly modified by glacial erosion.

5.6 Top out on floor of gap.

5.7 Old pit on right (now overgrown) is dug in gneiss saprolite. Several exposures in this area show late Wisconsinan till containing fresh gneiss clasts overlying gneiss saprolite, indicating incomplete glacial erosion of the saprolite. Exposures and well logs elsewhere in the Highlands, up to the New York border, show pockets of weathered gneiss (generally a grus) as much as 100 feet thick under unweathered till. In places the weathered gneiss forms drumlin cores. Descend slightly to the surface of a glacial delta.

6.4 Cranberry Lake to left. This lake occupies a plucked rock basin, as do many of the small upland lakes and swamps north of the terminal moraine.

6.7 Plucked cliff to left is in a resistant pyroxene granite unit. Striations show ice flow about due south over this ridge, while the cliff strikes about S40W. This divergence produced "obliquely plucked" erosional rock forms on this and on other similarly positioned slopes, where plucked faces are at an angle to the slope and alternate with abraded ledges. Flat terrace on the valley floor is another glacial delta surface.

7.3 Roadcut to left in another glacial delta, one of a series of step-down deltas in the north-draining valley from here to Andover. The ice-dammed lakes discharged westward into the Pequest Valley across local divides.

7.9 Abandoned gravel pit to right, in another glacial delta, a step lower than that at mile 7.3.

8.2 Tunnel under the Lackawanna Cutoff, a 28-mile-long cut-and-fill built by the Lackawanna Railroad between 1909-1911 to shorten rail passenger travel time between New York and points west, so as better to compete with the Erie and New York Central railroads. The fill, as much as 110 feet tall, is the world's largest railroad embankment. It was abandoned in 1979. A proposal in the late 1980s to mine it for aggregate led to state acquisition for possible future extension of a commuter rail line.

8.7 Intersection of U.S. Route 206 with County Route 517 in Andover New Jersey. Descend into Pequest River lowland in Kittatinny Valley. The ridge on your left (west) is a rock-cored drumlin. The low ridge to the right (east), upon which rests a small cemetery, is part of small upland delta that marks the Franklin Grove ice-margin. The elevation of the ridge is similar to the delta plain that lies just to the south. Apparently this part of the deposit was laid down in an ice-crevasse or collapsed tunnel.

9.5 Cross Lehigh-Hudson Railroad. The Pequest River lowland is a broad, strike-controlled, carbonate-floored area that lies between the New Jersey Highlands (to the east) and slate uplands in Kittatinny Valley (to the west). The landscape here is a topographic mosaic of rock-rimmed depressions, solution valleys, low relief ridges, and rock pinnacles; all products of glacial erosion in karst. In many places the lowest parts of the lowland are underlain by glacial outwash; chiefly sediment laid down in small sediment-dammed proglacial lakes that bordered on the margin of the Kittatinny Valley lobe. The low flat plains on your right (east) are part of the upper Pequest Valley deposits (Witte, 1988) laid down from ice-retreatal positions one to two miles upstream.

10.1 Whites Pond, on your right (east) side. The pond is a small kettle formed in the upper Pequest Valley deposits. The name is common to this area and is apparently derived from the thin deposits of marl that cover the pond's floor.

11.1 Large outcrop of the Allentown Dolomite on the right (east). The formation is of Upper Cambrian and Lower Ordovician age and consists of fine- to medium-bedded crystalline, thin- to medium-bedded, massive to laminated, rhythmically bedded dolomite. Unit is characterized by oolites, and stromatolites, and nodular and bedded chert and orthoquartzite are common (unit description modified from Drake, 1992).

11.3 Cross the Pequest River. The route here crosses through the headwaters of the Pequest River, which largely flows from the many springs south of the village of Newton, New Jersey. The many outcrops along this part of the route are Allentown Dolomite. Just to the north lies the remnant of a valley-outwash delta, that at one time covered the floor of the river valley. The low terraces on your right (east) are meltwater-terrace deposits that were largely formed by the drainage of Lake Newton (Witte, Chapter 4 in this guidebook).

11.7 Cross the Pequest River, near the village of Springdale. The broad terrace just ahead is another valley-fill delta remnant. This material appears to have been laid down from the Sparta margin (Witte, Chapter 4 in this guidebook), an ice-retreatal position one mile upstream near Newton. Based on morphosequence mapping, Ridge (1983) and Witte (1988) have identified 12 ice-retreatal positions in the Pequest lowland.

12.1 Cross meltwater channel cut by drainage from Lake Newton.

12.5 Pass St. Pauls Abbey on your left (west). Broad outwash plain on your right (east). Several rock pavements and small ridges of Allentown Dolomite poke through the outwash plain. Striae measured nearby show that the ice moved S 20^0 W, a direction the crosscuts the more southwesterly trend of Kittatinny Valley.

12.9 On your left (west) there are several low ridges of Allentown Dolomite, and a few large depressions in the outwash plain. The irregular shapes of some depressions suggest they are kettles. However, some are aligned with the strike of the country rock and have a more elliptical shape. These may be sinkholes. A common problem in parts of the Pequest and Paulins Kill lowlands in areas of karst is determining the origin of these landforms. The rolling upland on your left (west) is held up by the Martinsburg Formation (slate and graywacke siltstone, Upper Ordovician). Here it is capped by thin till. In this area the Martinsburg Fm. consists of three members, the Bushkill, Ramseyburg, and High Point. The lower member consists of slate with minor graywacke siltstone. It is quarried extensively in Pennsylvania where it is largely used as roofing. In this area there are several small and abandoned quarries.

13.3 Enter a small solution valley formed in the Allentown Dolomite. The valley here is underlain by outwash and it served as a conduit for meltwater emanating from the Kittatinny Valley lobe. The massive, deeply-etched outcrops here are Allentown Dolomite. In several places small aprons of talus (chiefly joint blocks) line the base of the outcrops.

13.5 Enter the village of Newton.

13.7 Ascend out of valley and climb approximately 50 feet onto the edge of the slate upland. At the crest of hill cross the drainage divide between the Pequest River and Paulins Kill. The gentle hillslope on the north side of the divide is underlain by thick till, which in many places forms aprons on the north-facing sides of hills and ridges.

14.6 Traffic light, intersection of Route 206 with Park Place, and the town square. Continue straight ahead.

14.6 Traffic light, intersection of Route 206 with Spring Street. Make a left-hand turn and then bear right down hill following signs for Route 206 (north). At bend in road, Route 206 joins Route 94.

14.8 Cross Paulins Kill, which drains beneath Route 206, northward toward a large swamp northeast of Newton. The swamp, formerly the site of Lake Newton (Witte, Chapter 4 in this guidebook), occupies a broad carbonate-floored valley. Route 206 here lies approximately 250 feet west of the carbonate - slate contact.

14.9 Traffic light, intersection Route 206-94 north and E. Clinton Street.

15.1 Pass large outcrop of the Martinsburg Formation (Bushkill Member) on your left (west). Bedding as shown by a few graywacke siltstone beds dips northwest. The upper five feet of the outcrop has been reworked by cryoturbation forming a loose mantle of slate chips. Till, if it was present, has been stripped off the hillslope by mass weathering. The Bushkill is the more slaty member of the formation and it crops out along a narrow belt that closely follows Rt. 206. The rolling hills that lie further west and at higher elevations are underlain by the Ramseyburg Member (graywacke, and graywacke siltstone). Bedrock geology from Drake and Volkert (1993).

15.3 Traffic light, intersection Route 206-94 and S. Park Drive.

15.6 Pass Sussex Shopping Plaza on the left (west). The plaza sits in a narrow valley that in places is underlain by laminated silt and clay. These materials were laid down in a small embayment of Lake Newton that projected into the slate uplands on the west side of the main lake basin. The floor of the valley is approximately 600 feet above sea level, and the lakes spillway one mile to the southeast, lies at an elevation of 595 feet above sea level (Witte, 1988). The slate uplands on either side of the valley exhibit a streamlined topographic form, a product of glacial erosion. Outcrops here are common, although thin till covers most of the rock surface.

15.8 The small hill on your left (west) behind the strip mall is a klippe of Allentown Dolomite (Drake and Volkert, 1993).

17.0 Traffic light, intersection of Routes 206 and 94. The small valley on your right (east) is underlain by lake-bottom deposits, and its setting is similar to the area beneath the shopping plaza (narrow embayment of Lake Newton). The low rock ridges on the right (east) are held up by the Bushkill Member of the Martinsburg Formation. The high ridge to the left (west) is a rock-cored drumlin. Borings (unpublished data on file at the N.J. Geological Survey, Trenton, New Jersey) show the till is as much as 52 feet thick.

17.5 Intersection of Route 206 and County Route 626. From 17.0 to 17.5 miles there is a slight rise in elevation and you have crossed onto the outwash plain of a small delta laid down in Lake Newton. The low slightly hummocky area immediately north is the Ogdensburg-Culvers Gap moraine, and it forms a low relief cross-valley ridge that terminates against the till-covered bedrock ridge that lies on the right (east) side of the road. Many shovel holes in the morainal area show that it consists of stony till. Its low relief is because it has nearly been drowned by deltaic outwash that lies against its outer and inner margins.

17.7 Pass over the moraine. If you look closely to your left (west), faint moraine-parallel ridges lie along the moraine's outer margin. Further east well records show that the moraine overlies deltaic outwash, which suggests it was laid down following a readvance (Witte, Chapter 4 in this guidebook). The segmented nature of the moraine is more common in this part of Kittatinny Valley. Further west in Kittatinny Valley and on Kittatinny Mountain, its course becomes more continuous.

18.0 Small outcrop of the Martinsburg Formation (Ramseyburg Member). To the left (west) side of the road, there is a large area of collapsed outwash immediately behind the moraine. These deposits are ice-contact deltas and lacustrine-fan deposits laid down in a small proglacial lake formed between the moraine and the margin of the Kittatinny Valley lobe. The spillway for the lake lies 0.5 miles west on the moraine at a similar elevation as the outwash. Proglacial lake deposits behind the moraine are common in Kittatinny Valley (Witte, Chapter 4 in this guidebook). These features mark minor recessional positions of the ice lobe.

18.3 Pass small sand and gravel pit in lacustrine-fan deposit. Descend rapidly (140 feet) to the floor of the Paulins Kill lowland. The Paulins Kill from its headwaters near Newton drains northward to the village of Augusta for a distance of approximately four miles. Retreat of the Kittatinny Valley lobe in this part of the basin resulted in the formation of several proglacial lakes. The largest and highest of these was Lake Newton. During retreat lower spillways were uncovered across local bedrock-floored drainage divides. Near Augusta the river drains southwest toward the Delaware River. Outwash in the Augusta area is also deltaic, although it was laid down in sediment-dammed lakes rather than lakes formed by topographic enclosure. Looking northward will afford an excellent view of the rolling topography formed on the Martinsburg Formation, and off in the distance to the northwest lies the even skyline of Kittatinny Mountain.

18.8 Cross small unnamed tributary of the Paulins Kill. Material in the lowland here consists of alluvium and meltwater-terrace deposits that overlie lake-bottom deposits. The low hill on your left (west) is a lacustrine-fan deposit.

19.1 Ascend onto the flat surface of a delta plain (ice-contact). Elevation of this deposit is approximately 525 feet above sea level. A low divide at 515 feet, about one mile west, may have been the spillway for the small unnamed lake that the delta was laid in.

19.3 Area of extensive outcrop on your left (west) is floored by the Ramseyburg Member of the Martinsburg Formation. The former topography here consisted of several low-relief (20 feet) strike ridges that were presumably leveled to make room for yet another strip mall!

19.5 Cross over Paulins Kill and ascend (20 feet) onto the plain of a large ice-contact delta.

19.6 Traffic light, intersection of US Routes 206 and 15, and County Route 565. A large bedrock ridge (Martinsburg Fm., Ramseyburg Mbr.) lies straight ahead. Turn left at light and follow Rt.206 north toward Culvers Gap via Branchville. Route 206 here follows the distal edge of a delta plain (fig. 2). The large delta here was laid down from the Augusta margin (Witte, Chapter 4 in this guidebook) located about one mile to the north. Typical of the many small and/or narrow glacial lake basins in this part of Kittatinny Valley, outwash has filled in the basin from valley wall to valley wall. Well records show that the outwash is as much as 150 feet thick. On the horizon is Kittatinny Mountain and the low area along its crest is Culvers Gap. In the midground are strike-controlled ridges held up by the Martinsburg Formation.

20.6 Cross Lehigh and New England Railroad (abandoned). The low channel on your right (north) is the outlet for Lake Wallkill (fig. 2) cut down in a large valley-fill delta that slightly predates the Augusta moraine (Witte, Chapter 4 in this guidebook). The floor of the spillway (estimated at 495 feet above sea level) probably stabilized on bedrock given the location of nearby Martinsburg outcrops. However, several probes along the outlet channel have not revealed the true location of the rock threshold because it is covered by swamp deposits. On your left (south) the Paulins Kill turns southwest and begins its descent down the Paulins Kill lowland. I suspect that prior to the late Wisconsinan glaciation the north-draining part of the Paulins Kill may have been part of the Wallkill River drainage basin.

20.6 Traffic light, intersection of Route 206 and Plains Road.

20.8 Cross over the crest of a small ridge and descend into the Branchville lowland. The road cut on your right (north) consists of thick till suggesting the ridge may be drumlin. The west side of the ridge is also the approximate location of the Portland Thrust Fault (Drake and Monteverde, 1992), which lies along the contact between the Martinsburg Fm. and the older Kittatinny Limestone. The Branchville lowland directly ahead is marked by the dissected surface of ice-contact and non ice-contact deltas that had formerly coalesced and filled in the lowland. Later, meltwater drainage accompanied by erosion and lowering of base-level downvalley, led to incision of the delta plain. Meltwater channels and meltwater terraces mark this event. The lowest terraces that lie in the valley on your left (south) are postglacial alluvial deposits laid down by the Paulins Kill and its tributary Dry Brook.

21.1 Cross over a small meltwater channel.

21.4 On your left (south) is an excellent view of the valley floor with its several glacial outwash and alluvial terraces.

21.8 Cross over Dry Brook.

22.0 Traffic light, intersection of County Route 519 and Route 206. Approximately 600 feet past the intersection is the contact between the Kittatinny Limestone, and the Martinsburg Formation (Drake and Monteverde, 1992).

22.60 Martinsburg (Ramseyburg Member) outcrop to the left (south). Ascend through small valley (Culvers Creek) cut by meltwater and drainage from Lake Owassa (Witte, 1997a).

24.00 Pass over the crest of a drumlin (summit - 905 feet), this location affords a good view of Culvers Gap and the crest of Kittatinny Mountain. Nearby well records (Witte, 1997a) indicate the drift here is as much as 294 feet thick.

24.0 Outlet of Culvers Lake.

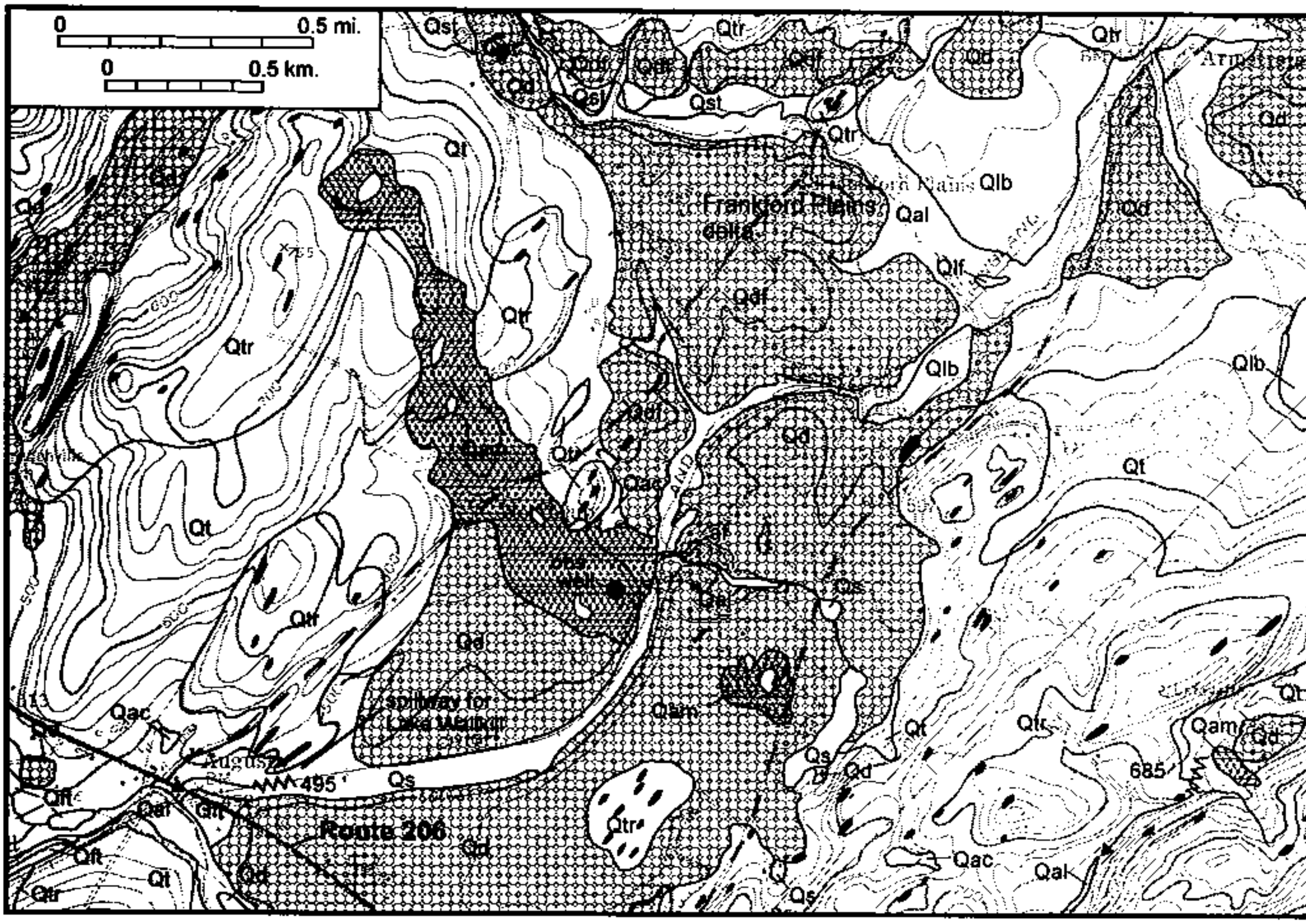


Figure 2. Surficial geologic map of part of the Branchville 7 1/2 minute topographic quadrangle. The area shown is located at Augusta, New Jersey in the SW 1/4 of the quadrangle. List of map units: af - artificial fill, Qal - alluvium, Qst - stream-terrace deposit, Qs - swamp and bog deposits, Qta - talus, Qac - alluvium and colluvium, undifferentiated, Qt - thick till, Qtr - thin till, Qam - Augusta moraine, Qd - ice-contact delta, Qdf - non ice-contact delta, Qlf - lacustrine-fan deposit, and Qft - meltwater-terrace deposit. Geologic log of the observation well located on the Augusta moraine: 0-23 feet, bouldery till with minor interbeds of sand and gravel, 23-35 feet, cobble and cobble-pebble gravel with some sand, 35-51 feet, pebble gravel, pebbly sand, and sand with minor interbeds of cobble-pebble gravel, 51-89 feet, pebble gravel and pebbly sand, gradational to sand and clayey silt below 75 feet, 89-112 feet, gray laminated clayey silt, and clay, and below 112 feet, dark gray siltstone (bedrock).

24.4 Pass over the crest of a drumlin (summit - 930 feet). Nearby wells show the drift here is greater than 300 feet thick. The drumlins near Culvers Lake are part of an anomalously thick belt of drift that extends five miles southwestward down Kittatinny Valley. Most of the well records in this area show till (presumably all of late Wisconsinan age) underlain by a varied mix of material interpreted as sand and gravel, fines, and till. In places this lower material may be as much as 100 feet thick, and because it lies deeply buried its age and genesis are unknown. In the realm of speculation, some of the stratified sediment and fines may have been laid down in a proglacial lake formed in front of the advancing Kittatinny Valley lobe. Alternatively, this material may be Illinoian age.

25.2 Intersection of County Route 521 and Route 206, on the right (north) is McKeown's (restaurant) and beyond is Culvers Lake. Just ahead lies the base of Kittatinny Mountain and the entrance to Culvers Gap. The mountain is held up by the Shawangunk Formation, a extremely resistant quartzite, and quartz-pebble conglomerate. Its lower slope is covered by till that has been mapped here as the Ogdensburg-Culvers Gap moraine, and talus. Depending on the size of our caravan, we may pull off the road into the restaurant's parking lot and take a few moments to view Culvers Gap, Culvers Lake, and the crest of Kittatinny Mountain, or as some will argue the glacially scoured surface of the Schooley peneplain. Please note that there will be no presentation here. The Culvers Gap and Culvers Gap River discussion will be held at stop 2-2 (Park Headquarters, Stokes State Forest, one mile northwest of the gap).

25.7 Culvers Gap, intersection of Lower Shore Road and Route 206 on your right (north), and the Ogdensburg Culvers Gap moraine.

25.8 The parking area for stop 2-1 is on your right (north), just past Lower Shore Road on State property (Stokes State Forest). Additional parking is available off Lower Shore Road. We'll assemble near the edge of a small pit located just northeast of the parking lot (fig. 3). Stop 2-1 will consist of 2 parts. At 1a we'll examine the materials that make up the moraine, and discuss debris sources and debris transport at the margin of the Kittatinny Valley lobe. Stop lb will consist of a short walk (5 minutes) to the moraine's outer margin where we'll reassemble on the crest of the moraine's frontal ridge. Here we'll discuss the moraine's morphology, and its construction (push moraine, dump moraine, stagnation moraine ?).

STOP 2-1 (a and b)

Ogdensburg-Culvers Gap Moraine

Ron Witte (N.J. Geological Survey)

Location - Culvers Gap, N.J. - PA. 7 ½ minute topographic quadrangle in Stokes State Forest, Sussex County, New Jersey (fig.3).

Geologic Setting - Culvers Gap is a preglacial wind gap that lies in the glaciated section of the Valley and Ridge Physiographic Province. It forms a prominent pass through Kittatinny Mountain and links Kittatinny Valley (east) with Minisink Valley (west). The mountain is held up by the Shawangunk Formation, a tough and very resistant quartzite and quartz-pebble conglomerate of Silurian age. The floor of the gap is covered by the Ogdensburg-Culvers Gap moraine, a recessional moraine of late Wisconsinan age laid down at the margin of the Kittatinny Valley lobe (Witte, Chapter 4 in this guidebook). Based on nearby well records

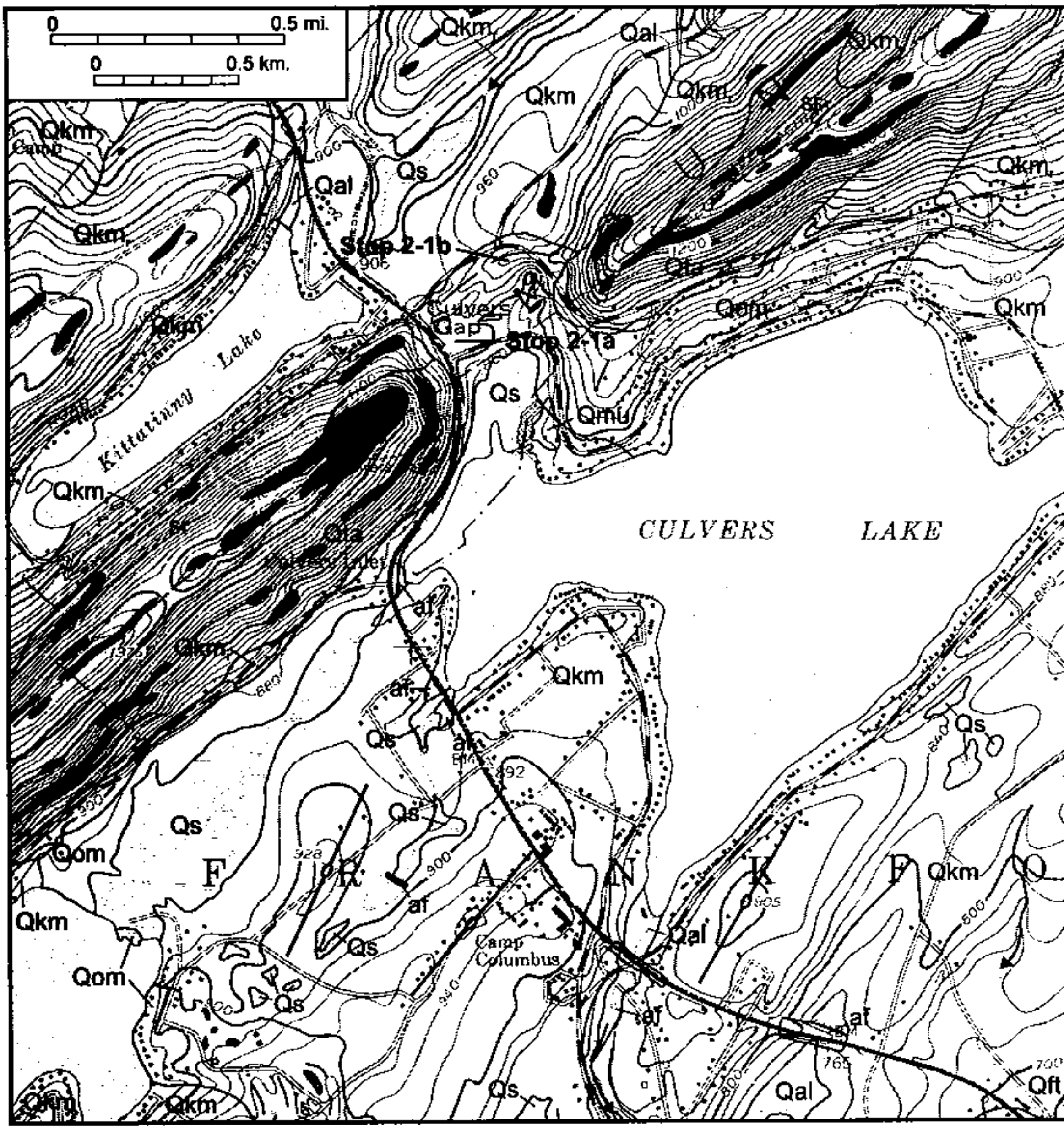


Figure 3. Surficial geology of a portion of the Culvers Gap quadrangle, New Jersey, in the vicinity of Culvers Gap. List of map units: af - artificial fill, Qs - swamp and bog deposits, Qal - alluvium, Qta - talus, Qkm - thick till, Qkmr - thin till, Qom - Ogdensburg-Culvers Gap moraine, Qmu - small undifferentiated meltwater deposits, Qft - meltwater-terrace deposits, sr - regolith, chiefly rock waste on steep hillslopes and ridge crests with minor talus, scattered erratics, and a few rock outcrops. Shaded areas represent extensive rock outcrop. Symbology on Qom: arcuate lines are drawn along the crest of large morainal ridges, and small polygons represent kettles.

(Witte, 1997a), the gap's bedrock floor lies between 750 and 800 feet above m.s.l. Salisbury (1902, p. 350) appears to have been the first to use the term Ogdensburg-Culvers Gap moraine in describing a discontinuous morainal belt that traced a course westward through Kittatinny Valley from Ogdensburg to Culvers Gap. Salisbury also noted that morainal topography in Minisink Valley near Fisher School House and in Wallpack Valley near Layton may have been contemporaneous to the feature at Culvers Gap. Minard (1961) traced the moraine over Kittatinny Mountain and Witte (Chapter 4 in this guidebook) further refined its course and placed it within the morphostratigraphic framework that he developed for the Kittatinny and Minisink Valley ice lobes.

Discussion - Recessional moraines in northwestern New Jersey form conspicuous, arcuate, cross-valley ridges that mark former, stable, ice-marginal positions of the Laurentide ice sheet. These features are late Wisconsinan age, consist chiefly of till, have a varied morphology, and follow looping courses through the Kittatinny and Minisink Valleys. They include the Franklin Grove, Dingmans Ferry-Ogdensburg/Culvers Gap, Montague-Augusta, Fairview Lake, Libertyville, Millville, and Steeny Kill Lake moraines, and they delineate recessional positions of the Kittatinny and Minisink Valley ice lobes (Witte, Chapter 4 in this guidebook). 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 stresses 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.

The recessional moraines consist of non compact, stony, silty-sandy till (diamicton) 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 it crosses river valleys (former glacial lake basins). Their course shows the margin of the Kittatinny and Minisink Valley lobes was distinctly lobate at both a regional and local scale. Topography varies between ridge-and-swale and knob-and-kettle, with the former more prevalent along the outer morainal margin (for additional information on the recessional moraines see Witte, Chapter 4 in this guidebook).

At stop 2-1a (fig. 3) we will examine the materials that make up the moraine. The outcrop in the west wall of the pit exposes about 7 feet of till. Below the B horizon (measured at 42 inches below the top of the A), the parent material consists of poorly sorted, 10 YR 4/4 (dark yellowish brown), slightly to moderately compact, silty sand. Gravel content for clasts greater than 1 inch (7-10 percent), and less than 1 inch (25-30 percent). Matrix has a granular to slightly prismatic structure, and contains indistinct layers and lenses (less than 2 inches thick) of coarse sand and very fine gravel. Most of the clasts are subrounded, although subangular and rounded shapes are present. In places, very faint horizontal layering can be observed that is marked by a concentration of larger clasts at the layer's base.

The mineralogy of the morainal material (listed below) and the course of the moraine (fig. 3) clearly shows that the debris here was derived from sources in Kittatinny Valley. Its gravelly and granular texture, indistinct layering, and sandy lenses suggest that this material may have had a history of sorting; possibly derived from melting glacial ice, post-meltout sorting on the glacier surface, and emplacement as debris flows.

Mineralogy of Ogdensburg-Culvers Gap moraine in Culvers Gap
(sample depth = 77 inches)

Texture - 20 % gravel, 35 % sand, 45% silt and clay.

Mineralogy (granules and sand mineralogy based on visual estimates provided by F.L. Muller, New Jersey Geological Survey)

pebbles (1-3 inches) - 3 % dolostone (Kittatinny Super Group), 41 % slate and graywacke (Martinsburg Fm.), 25 % lithic, quartz sandstone (Martinsburg Fm.), 29 % quartzite and quartz-pebble conglomerate (Shawangunk Fm.), 2% red sandstone (Bloomsburg Red Beds), 1 % vein quartz.

granules - 20 % red sandstone, 35 % gray sandstone - quartzite, 45 % gray siltstone and shale.

sand (light fraction) - 5 % red sandstone, 15 % gray siltstone and shale, 10 % medium- to fine-grained yellow-stained sandstone-quartzite, 15 % light gray to white shale, 40 % quartz, 15 % clay and silt aggregates.

Sand (heavies) - 5 % goethitic ironstone, 5% muscovite, 1 % ilmenite, 1 % garnet, 2 % magnetite, 5 % aluminosilicates, 1 % rutile, 7 % zircon, 1 % pyroboles, 60 % quartz, 5% clay and silt aggregates.

Questions - If this material were largely deposited as debris flows, is it really till? What is the source of the morainal sediment and does there exist a correlation between the thickness of till immediately north (up-ice) from the moraines and their relative size? What is the role of active ice as to sediment transport to the glacier margin?

Head to stop 2-1b (fig. 3) by following the trail marked by the blaze-orange ribbons (northwest direction). This short walk should take about five to ten minutes and we will assemble at the crest of a large morainal ridge for additional discussion. The traverse will take us from the moraine's inner margin to its outer margin, a climb of about 80 feet. The main purpose here is to observe the moraine's morphology and note how it changes from knob-and-kettle to ridge-and-swale. Unfortunately, due to the time of year foliation will mask some morainal landforms (you should've been here a couple of months ago).

The first part of the traverse will take us through a stand of hemlocks across the inner morainal area. Here knob-and-kettle topography is well formed and many kettles contain year-round water. The overall appearance of the topography here is one of collapse. The second part will take us through the outer morainal area (starts approximately where we begin to climb onto the moraine's frontal ridge, and the arboreal cover starts to change to white, red and chestnut oak, hickory, and silver birch. Here knob-and-kettle gives way to ridge-and-swale topography. Ridge crests in this area follow arcuate traces that parallel the course of the moraine. Swales here are typically elongated and typically parallel the adjacent ridges. Continue along the trail and assemble on the crest of a large ridge at stop 2-1b.

The topography observed along the trail is typical of the moraines that lie along the western side of Kittatinny Valley and on Kittatinny Mountain (see Chapter 4, figure 4 in this guidebook). Elsewhere, it is absent or very faintly developed. The recessional moraines are also larger, more continuous, and have better formed ridge-and-swale topography in areas where they lie against thick till.

Questions - How are moraine-parallel ridges formed, are they push ridges or colluvial ramparts. Why are the moraines larger and more continuous in the uplands? Are the moraines end moraines, and do they represent places where the glacier margin remained in a stable position for a significant length of time?

At the conclusion of stop 2-1b head back to your vehicles by retracing our route to stop 2-1a, or you may head down the frontal slope of the moraine to Upper Shore Road, turn left (south) and head back to Rt. 206.

25.8 Continue north on Route 206 and cross over the Appalachian Trail, which is at the intersection of Upper Shore Road and Route 206, and exit Culvers Gap. The topographic divide in the gap marks the drainage divide between Paulins Kill and Flat Brook.

26.0 Pass Kittatinny Lake on your left (south). Cross over the contact between the Shawangunk Formation and the Bloomsburg Red Beds (Monteverde, 1992), which is covered here by thick till.

26.5 Turn right (north) into entrance to Stokes State Forest.

26.6 Turn left into parking lot of Park Headquarters. We will assemble at the south end of the lot following a ten minute coffee break.

STOP 2-2

Culvers Gap River and Abandonment of Culvers Gap

Ron Witte (N.J. Geological Survey)
and
Jack Epstein (U.S. Geological Survey)

Location - Culvers Gap, N.J. - PA. 7 ½ minute topographic quadrangle at the Stokes State Forest Headquarters, Sussex County, New Jersey (fig. 3).

Geologic Setting - Piedmont area of Kittatinny Mountain, 500 feet below the crest of Kittatinny Mountain. The local bedrock is the Bloomsburg Red Beds (red shale and sandstone of Silurian age), and the overall dip of the rock is northwest. However, many secondary folds form small strike ridges (bedrock geology from Monteverde, 1992). In places drumlins, most of which may be rock-cored, lie among the rocky ridges in this shale and sandstone terrane. Further north is an extensive cover of thick till that covers a large part of the Kittatinny Mountain piedmont. The drift here is as much as 175 feet thick (Witte, 1997a); the constructional topography characterized by drumlins, recessional moraine, and ground moraine. The drumlins show that ice flowed in a southerly direction across Kittatinny Mountain whereas the moraines show that ice flow was strongly influenced by the southwesterly topographic grain. On the southeast side of the parking lot in wooded area is a small patch of low-relief hummocky ground. This material is not part of the moraine. It may be ablation till and due to its scant distribution it has not been mapped separately.

Discussion -The original intent was to assemble on the summit of Kittatinny Mountain (on Sunrise Mountain in Stokes State Forest). However, given our shortened day for a field trip, and the theological overtones inherent in assembling on Sunrise Mountain Sunday to discuss abandoned water gaps, I have decided to meet here, well aware that we can't see the gap! The discussion will consist of two parts. First, Jack Epstein will lead a review of the evidence for structural control of wind and water gaps along Kittatinny Mountain. Secondly, I will discuss the late history of the Culvers Gap River and lay out several scenarios by which the river was captured by the Delaware River.

26.7 Exit parking lot on west side and turn left (south) toward Route 206. Turn right (northwest) on Route 206 (mileage measured at intersection of park road and highway) and descend into the Big Flat Brook valley.

27.8 Tuttles Corner. Large meltwater channel on your left (south).

28.4 Cross Big Flat Brook. On the west side of the river is the remnant of a valley-train laid down from the Culvers Gap margin (located about one mile upstream). Most of this material has been eroded by later meltwater (from the Augusta margin), and the postglacial Big Flat Brook. Ascend out of the Big Flat Brook valley. The upland topography is formed by thick to thin till overlying large strike-controlled ridges held up by the Bloomsburg Red Beds.

30.3 Cross Dingmans Ferry moraine and descend into Little Flat Brook valley (mileage measured at the crest of moraine).

31.0 Cross Little Flat Brook. The lowland here is a strike valley that principally follows the very weak Poxono Island Formation, which is a finely crystalline to aphanitic dolomite (Monteverde, 1992). The valley floor is covered by collapsed and dissected outwash laid in a small proglacial lake dammed by the Dingmans Ferry moraine (Witte, 1997a). Some poorly drained areas appear to outline former stagnant ice blocks, an indication of stagnant ice at the margin of the Minisink Valley lobe. The low upland to your left (west) is Wallpack Ridge held up by the Esopus, Schoharie and Buttermilk Falls formations. These Middle Devonian units (listed first to last) consist of arenaceous siltstone, silty to shaly, locally dolomitic limestone, and cherty, clayey to silty limestone. Bedrock geology from Monteverde (1992).

31.4 On your right (east) is the broad alluvial plain of Little Flat Brook. The low terraces that rise above the flood plain are stream-terrace deposits presumably laid down before and during the early stages of delayed rebound.

32.0 Pass Hainesville General Store on the left (west).

32.2 Start ascent out of Little Flat Brook valley, and pass a small sand and gravel pit on your right (east).

32.2 Cross over a drainage divide between Little Flat Brook and White Brook. The surrounding uplands chiefly consist of undulating strike-controlled bedrock ridges covered with thin till.

34.4 Intersection of Route 206 and Clove Road (bear left at Y intersection) and pass into Delaware Water Gap National Recreation Area (DWGNRA). The level plain here is underlain by outwash laid down at the margin of the Minisink Valley lobe (Witte, 1997b). Its position here in a north-draining valley, and similar elevation with a col through Wallpack Ridge (located 2500 feet to the west, elevation 685 feet) suggests the outwash is presumably a delta. The low relief ridge directly ahead (north) is the Montague moraine.

34.7 Descend into Minisink Valley (upper Delaware River) and cross over the Montague moraine.

35.2 Intersection of Route 206 and County Route 521 at bottom of the hill. Turn right onto Rt. 521 and head north up the Minisink Valley (this is the last exit in New Jersey before the toll bridge).

36.5 Cross Shimers Brook. The deep ravine occupied by the brook is a meltwater channel that drained a series of small ice-contact lakes in the uplands to the east. The large outcrop on your right (east) before the bridge is the Esopus Formation.

36.9 Pass an Esopus outcrop on your right (east). The low cross-valley ridge directly ahead is the Millville moraine, and the skyline on your right is formed by the Pocono Plateau.

37.0 Millville Moraine. At the crest of the moraine turn left onto DWGNRA access road and head west toward Delaware River. The road is in very good shape, although in places two vehicles cannot pass each other. Descend into the valley, passing over the moraine.

37.4 Stop 2-3. Turn left (south) into gravel parking lot, and park vehicles. We will assemble on the south side of the lot.

STOP 2-3 (a & b)

Deglaciation and postglacial fluvial history of Minisink Valley

Ron Witte (New Jersey Geological Survey)

Location - Milford, Pa.- N.J. 7 ½ minute quadrangle, New Jersey side of Minisink Valley (upper Delaware Valley) in Sussex County, northern part of the DWGNRA (fig. 4).

Geologic Setting - Glaciated section of the Valley and Ridge Physiographic Province. Minisink Valley is a deep glacially scoured river valley underlain by a complex stratigraphy that consists of glaciolacustrine, glaciofluvial, and alluvial sediment that rests upon northwest dipping Buttermilk Falls Limestone and Marcellus Shale (Middle Devonian age). The upland to the east is a continuation of Wallpack Ridge, and the upland to the west is the western edge of the Valley and Ridge with the Pocono Plateau forming the higher hills in the distance. The large cliff that forms the west valley wall is held up by the Mahantango Shale (Upper Devonian age).

The low broad plain that forms the valley floor is an abandoned flood plain of Holocene age (Qst2 on fig.4). The terrace is underlain by overbank deposits of fine sand and silt (as much as 20 feet thick) that overlie sand and gravel of a large point bar; presumably deposited by the Delaware River as it migrated west across the valley floor. The geometry of channel scrolls on the terrace (fig. 4) preserves this history of lateral migration. To your left (east) is the Millville moraine and just above the parking area is a small bench of sand, and minor pebbly sand (interpreted as a remnant of an abandoned Pleistocene flood plain, Qst3). The high terrace downvalley is outwash laid down from the Millville moraine position. Down the valley to our right (west) in Pennsylvania is a high, conifer-covered terrace that consists of outwash laid down from the Montague moraine position. It lies about 20 feet higher than the Millville outwash.

Discussion - Stop 2-3 will consist of two parts. At 2-3a (parking lot) we will focus the discussion on the style of the late Wisconsinan deglaciation in Minisink Valley (reviewed in Witte, Chapter 5 in this guidebook). Secondly we will take a 15 minute hike out to the Delaware River and look at an outcrop of the upper abandoned flood plain (Qst3).

Several investigations on glaciofluvial terraces in Minisink Valley have suggested that the late Wisconsinan ice sheet disappeared from this area either by regional stagnation or marginal retreat. This controversy is a recurring argument, and it is not unique to the Minisink Valley area. Earlier research by White (1882) and Salisbury (1902) favored a marginal retreat model. Their interpretations were largely based on the identification of recessional moraines, and the ice-contact heads of valley-train deposits that represented positions where the retreating glacier margin had halted. Later work by Happ (1938) and Crowl (1971) favored a stagnation model where the uplands were deglaciated first with residual masses of ice left in the valleys. Large areas of collapsed topography in kames and kame terraces, many kettles, ice-contact slopes, and unpaired terraces were cited as evidence for stagnation. Epstein (1969) and Epstein and Koteff (in press) near Stroudsburg, Pennsylvania, and Ridge (1983), Witte, (1988, Chapter 4 and 5 in this guidebook), and Stone and others (in press) in northwestern New Jersey have returned to the marginal retreat model. Based largely on the morphosequence model of Koteff and Pessl (1981), these investigations have shown that deglaciation took place largely by the systematic melting back of the margins of the Kittatinny and Minisink Valley ice lobes.

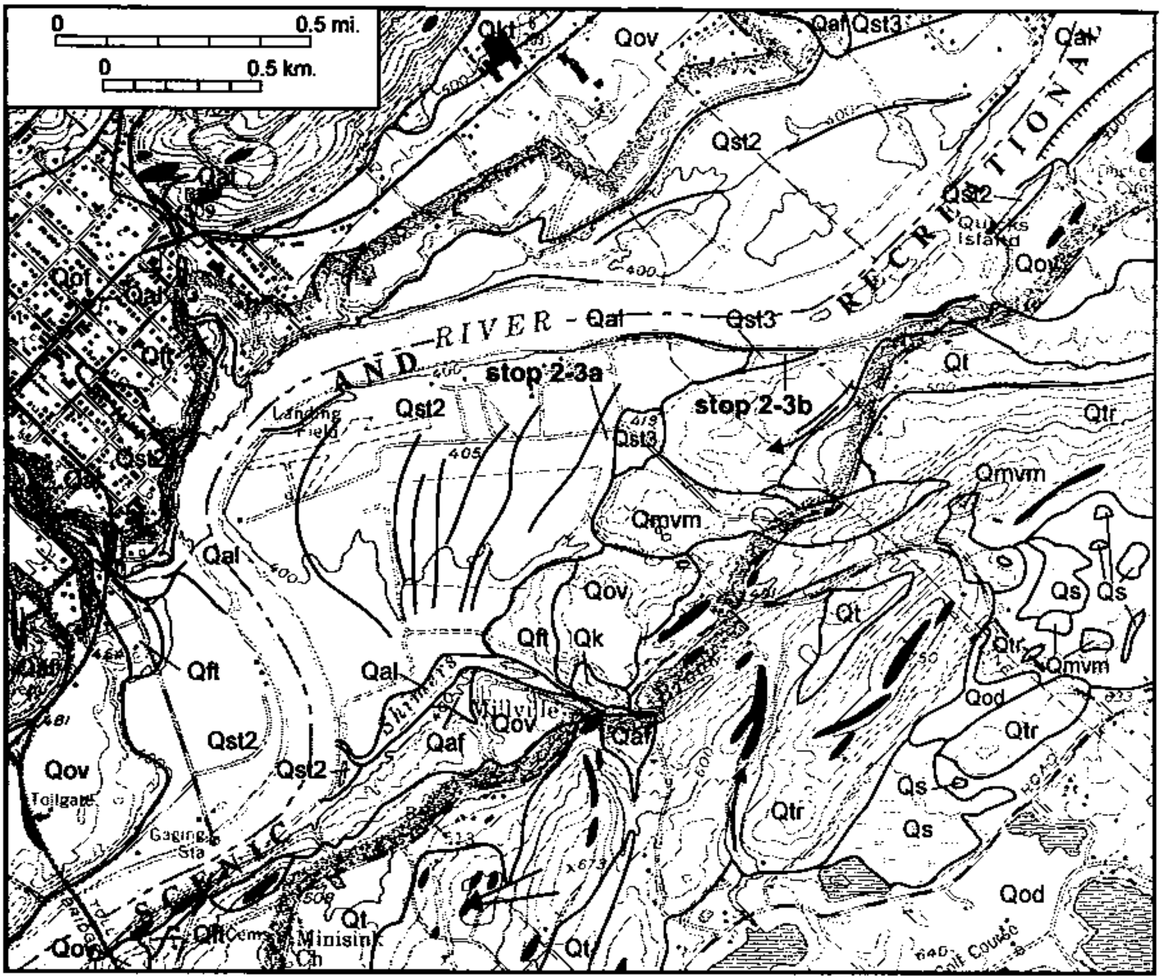


Figure 4. Surficial geologic map of part of the Milford, Pa-N.J. 7 1/2 minute quadrangle near Millville, New Jersey. List of map units: af - artificial fill, Qal - alluvium, Qaf - alluvial fan, Qs - swamp and bog deposit, Qst2 - abandoned flood plain (Holocene), Qst3 abandoned flood plain (Pleistocene), Qft - meltwater-terrace deposit, Qov - valley-train deposit, Qod - ice-contact delta, Qof - outwash-fan deposit, Qkt - kame terrace, Qk - kame, Qmvm - Millville moraine, Qt - thick till, and Qtr - thin till. Shaded areas represent extensive bedrock outcrop and the curved lines on the stream-terrace deposits represent channel scrolls. Surficial data from Witte (1997b).

Directions to stop 2-3b (use fig. 4) - Turn right (east) out of parking lot and follow DWGNRA access road about 120 feet to a trail on your left (north) marked by the stone walls of an old gate. The high ridge on the south side of the access road is the Millville moraine. At the base of the moraine is a low, crudely formed topographic bench underlain by very fine sand and silt with some pebbles. This material may be a remnant of Qst3 (upper abandoned flood plain). Near the road this material is less than three feet thick and it overlies the moraine. Pass through gate walls and head north (900 feet) toward the Delaware along an overgrown road. Along this part of the traverse you will descend onto the T2 terrace and cross several small channels (< five feet deep) before coming to the left bank of the river. The terrace is about 28 feet above the river here and it is underlain by very fine sand and silt. The low terraces across the river in Pennsylvania consist of T2 deposits and the modern flood plain. In places the T2 deposits comprise several levels that represent flood-scoured areas or lower and younger T2 subsets. These intermediate terraces are found throughout the valley. However, based on the 20-foot contour interval of the base maps and the lack of extensive outcrops, I have decided to lump them all as part of the T2 terrace. Turn right and follow the trail along the river bank upstream. In many places the upper part of the terrace scarp is marked by small slumps. This is the principal process by which flood plain material is recycled. Further up stream some slumps are larger and in places gullies have formed at their heads, introducing a second process of erosion. Cross Tennessee Gas Pipeline. The height of the T2 surface here is about 4 feet greater above the river than it was downstream. Along the terrace scarp a very faint levee and its backslope are preserved. Presumably the more proximal part of the levee has been eroded by the Delaware. Overbank materials are only about 5 feet thick here, overlying coarse gravel and sand. Wagner (1994) using GPR and auger borings at the Depew Recreation Site down the valley showed that abandoned flood plain there was underlain by several, stepping-downwards, gravel cut terraces. I believe these features underlie many of the abandoned, flood-plain terraces in the valley, and because they represent the former location of the Delaware River, they preserve its history of incision. Is this material glacial outwash (meltwater-terrace deposits laid down from distant ice margins) or is it outwash reworked by postglacial streams? The low narrow terrace across the river is the modern flood plain, which lies next to the higher T2 terraces. The highest terrace in Pennsylvania (70 feet above the T2 surface) is part of the valley train laid down from the Tristates ice margin, placed about 4.5 miles upstream. Continue traverse upstream, along a narrow foot path. Cross two small gullies. Stop 3b is about 250 feet further in a stand of hemlock, white pine, and rhododendron. The terrace has narrowed considerably and is about 48 feet above the river. The hill slope to the east is part of the Tristates valley-train deposit. Note that the transition from the T2 to T3 terrace is not marked by an erosional scarp, but a gently inclined slope.

The outcrop at stop 3b is at the top of the escarpment overlooking the Delaware River at the headwall of a small slump. Measured from the surface: 0-30 inches, very fine sand and silt that contain a few small pebbles and charcoal fragments, -32 inches, mottled reddish brown horizon that may be a truncated B horizon. This interval also contains the highest amounts of charcoal in the exposed section, - 35 inches, light brown mottled very fine sand, - 44 inches , thinly-bedded pebbly sand with pebbles as large as 1.5 inches, -50 inches, small boulder and cobble gravel lag (stepping stone orientation), below 50 inches is an undetermined thickness of thinly-bedded fine- to coarse-pebbly sand and minor fine gravel, minor cross-bedding in some of the sand beds.

The T3 terrace represents the highest flood plain in Minisink Valley. They have been observed as far downstream as Zion Church (distance of about 25 miles). Above this lies glacial outwash, and below the extensive T2 terrace and the modern flood plain.

Questions - Were the T3 materials laid down by a postglacial river or might they represent the very late stages of deposition by glacial meltwater? If the T3 terrace represents the oldest postglacial flood plain in Minisink Valley, what factors caused the Delaware River to cut down to its present level (rebound, reduction in the overall amount of sediment eroded in the drainage basin)? If the T3 deposits were laid down by a postglacial river than how is it possible that the lower gravels beneath T2 terraces and the modern flood plain have been interpreted as glacial outwash.

Retrace route back to parking area as quickly as possible so that we may depart to our final stop.

39.6 Retrace route to Route 206 and Route 521 intersection (mileage measured at intersection) and turn right (west) on Rt. 206 toward Pennsylvania.

40.0 Milford Toll Bridge. Cross Delaware River and enter Penn's Woods. The floor of Minisink Valley is covered by the head-of-outwash of a valley-train deposit laid down from the Augusta margin (Montague moraine). Except for a few small segments near the edge of the valley walls, the moraine has presumably been eroded by meltwater here. The Delaware lies about 100 feet below the outwash plain and flows along a deeply incised straight and narrow channel. These deposits had been previously mapped by Crowl (1971) as kame terraces.

40.3 Toll booth ($1.00 for passenger vehicles). Note it only costs money to **leave** New Jersey.

40.5 Intersection of US Routes 206 and 209. Turn left (south) following signs for Dingmans Ferry, Bushkill, and Stroudsburg. The highway here follows the Montague valley-train terrace.

40.6 Enter the Delaware Water Gap Recreation Area (technically we really haven't left it) and begin descent onto the eroded surface of the valley train. Meltwater-terrace deposits in the valley generally lie more than 50 feet above the Delaware. Below this, you will find the postglacial alluvial terraces and the modern flood plain. In a few places the upper alluvial terrace (T3) has been eroded revealing meltwater-terrace deposits.

41.7 Shale pit on your right (west). The cliff face is held up by the Mahantango Formation, which in places rises more than 400 feet above the valley floor. The base of the cliff is covered by a thick apron of shale-chip (sharp stone) colluvium. In many places this material has been mined exposing the rock face at road level. Material chiefly used as subbase.

42.2 Another shale pit on your right (west). The valley floor to your left (east) is covered by postglacial alluvial terraces. The higher terrace is an eroded T3. The lower terraces and channels are part of the main Holocene terrace (T2) that covers large parts of the valley's floor. The low channel near the road is a common feature that forms on the landward side of the terrace, pinned against the valley wall and the natural backslope of the terrace, and further eroded during large floods. Over time the Delaware may erode westward and deposit more proximal overbank sediment (levee facies) over the back channel sediment. This stratigraphy was recently revealed in a large excavation for restroom facilities downstream near Bush Kill, where 12 feet of overbank sand and silt was observed to underly 9 feet of orgainic-rich silt and clay.

42.9 Cross Raymonds Kill Creek. Approximately 200 feet past the bridge, turn right (west) onto the driveway of an old homestead and park vehicles. This is the parking area for stop 2-4, and here ends the road log for this odyssey. Stop 2-4 consists of two parts. First and the main focus of this part of the trip is a large outcrop in the T2 terrace on a bluff overlooking Raymondskill Creek (2-4a). Secondly, John Wright of the National Park Service will discuss the archaeology of this area. Please remember that digging and collecting in the park are prohibited under Federal law. The second part of stop 2-4b is optional and will consist of climbing to the top of the Raymonds Kill Creek outwash fan and looking at materials exposed in a recent slump. The site also affords an excellent view of the alluvial and glacial terraces in the valley. Before proceeding to the Raymondskill section a short overview on the geology and archaeology will be given in the parking area.

Stop 2-4 (a & b)

Archaeology and the glacial and postglacial stratigraphy at the Raymondskill Creek Site, Minisink Valley, Delaware Water Gap National Recreation Area

Ron Witte (New Jersey Geological Survey)
and
John Wright (National Park Service)

Location - Milford, Pa. N.J. 71/2 minute topographic quadrangle, Minisink Valley, at the mouth of Raymondskill Creek, upper part of DWGNRA (fig.5). Reportedly a trail, running westward from this area to the Wyoming Valley, was used by the Delaware Indians and early settlers.

Geologic Setting - We are on the western side of Minisink Valley in a reentrant formed by Raymondskill Creek. The T2 terrace forms the low area to the east above the creek. To the south rises a steep bluff (70 feet) held up by glacial outwash (outwash-fan deposit). Most of this material appears to have been laid down by a meltwater stream draining the Raymondskill Creek valley, and at one time it filled the reentrant. Later it was eroded down to its present level by postglacial Raymondskill Creek. The high cliffs to the west and the falls are underlain by the Mahantango Shale. The present location of the creek at its mouth is south of its position shown on the quadrangle map. The creek shifted its course about 600 feet sometime during the last quarter century, presumably during one of the larger flood events.

Directions to 4a - Please cross Route 209 carefully and follow a fishing-access road across the T2 terrace toward the river. At large right-hand bend (about 450 feet from Rt. 209) turn left (north) and follow the foot path through wooded area to the bluff overlooking creek. Please note that the bluff face is extremely unstable. Stay back at least five feet from its edge. At stop 4a the T2 terrace consists of about 15 feet of overbank material, replete with buried A and B soil horizons that preserve a history of episodic alluviation, vertical flood plain accretion, and prolonged periods of land stability marked by soil formation. Beneath this material is coarse gravel and sand (interpreted as an alluvial fan). Based on its elevation, Minisink Island (fig. 5) has also been mapped as a T2 terrace. The lower terraces at the mouth of the creek (modern flood plain) are no higher than 12 feet above the river, and they consist of massively to thickly-bedded fine sand, very fine sand, and silt. Soil development here is immature and it appears that depositional rates here are very high, possibly the result of hydraulic ponding at the mouth of the creek during floods.

See handout for a detailed description of the outcrop.

Discussion (stop 2-4a) - In Minisink Valley, the late Pleistocene is marked by the transition of the Delaware River from a braided glacial meltwater stream to a postglacial non meandering stream. The late glacial Delaware River is assumed to be a braided stream, given the large volume of meltwater that flowed through the valley and the readily available source of sediment.

In contrast to the late glacial river, the modern Delaware is a non meandering stream of low sinuosity flanked by two abandoned flood plains (T3 and T2 terraces). The T2 alluvial sequences at the Shawnee-Minisink (McNett and others, 1985), and Upper Shawnee Island (Stewart, 1989) sites show that the form of the Delaware River at the end of the Pleistocene was also a non braided one with a well-established flood plain. The river was also at or slightly above its present elevation.

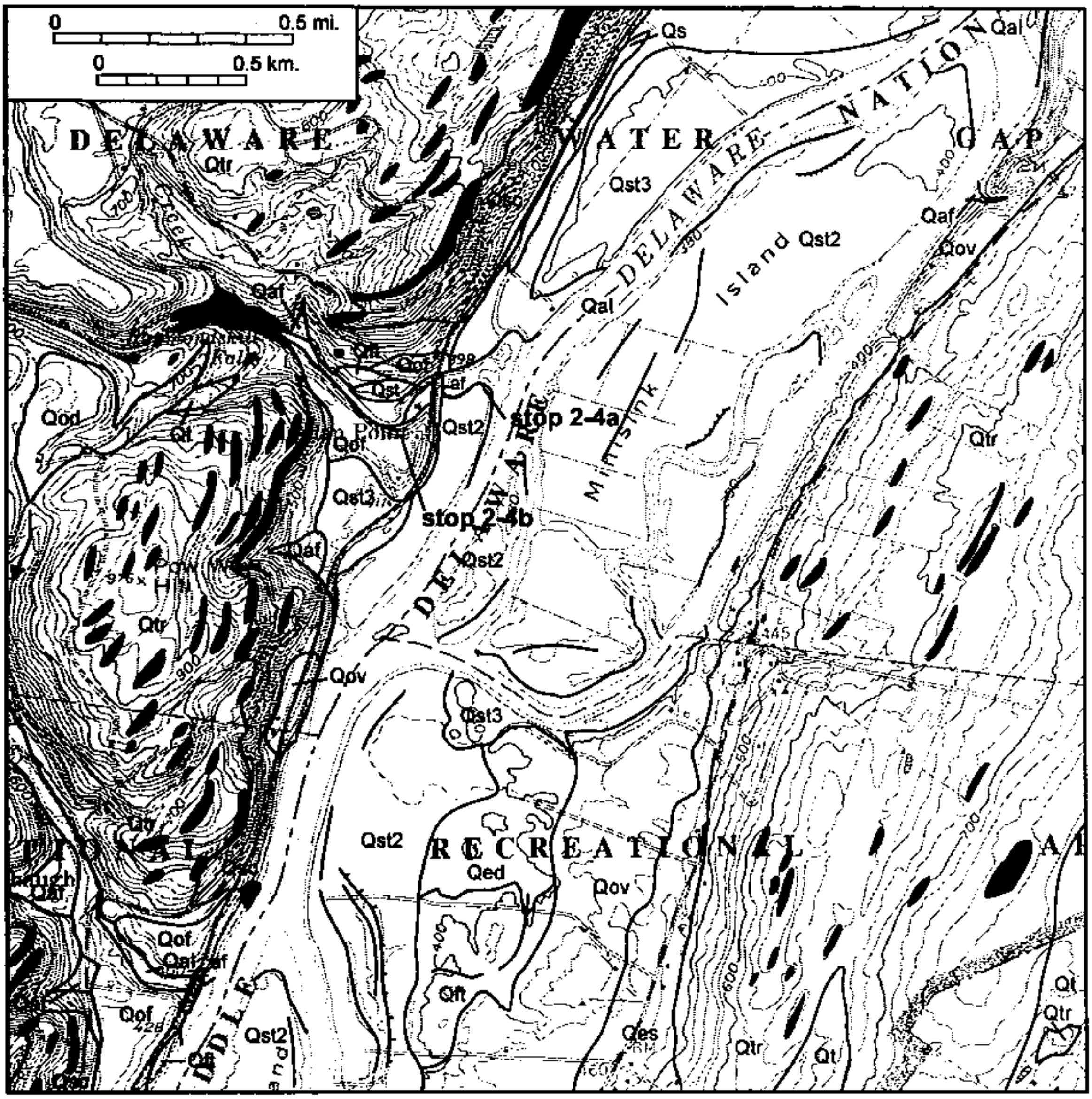


Figure 5. Surficial geologic map of part of the Milford, Pa-N.J. 7 1/2 minute quadrangle near Raymondskill Creek, Pennsylvania. List of map units: af - artificial fill, Qal - alluvium, Qaf - alluvial fan, Qs - swamp and bog deposit, Qsc - shale-chip colluvium, Qes - sand dunes, Qes - thin sheet of wind-blown sand, Qst2 - abandoned flood plain (Holocene), Qst3 abandoned flood plain (Pleistocene), Qft - meltwater-terrace deposit, Qov - valley-train deposit, Qod - ice-contact delta, Qof - outwash-fan deposit, Qt - thick till, and Qtr - thin till. Shaded areas represent extensive bedrock outcrop and the curved lines on the stream-terrace deposits represent channel scrolls. Surficial data from Witte (1997b).

The transition, from a braided glacial stream with a very distant meltwater source to a non meandering stream, represents significant hydraulic changes during the close of the Pleistocene. Most obvious was a substantial decrease in discharge due to the retreat of the Laurentide ice sheet from the Delaware River drainage basin. The minimum date for this event is estimated at 14,000 yrs B.P., based on the mapping and correlation of ice-marginal positions by Ozvath and Coates (1986) in the Western Catskill Mountains, and Fleisher (1986) in the upper part of the Susquehanna drainage basin. The dramatic decrease in discharge was accompanied by a change in channel form from braided to a non meandering channel of low sinuosity. This transition may have already been underway during the later stages of deglaciation of the drainage basin when meltwater found new flow paths into the Susquehanna River and to a lesser extent the Hudson River drainage basins.

At some point in time during the latter part of the late Wisconsinan the Delaware River underwent a period of incision to a level at or near its present elevation. The timing and possible causes of this event will be examined. Previous investigations by Crowl (1971) and Dent (1985) suggested the coarse gravel beneath the T2 terrace is glacial outwash, laid down by meltwater during the latter stages of deglaciation of the Delaware River drainage basin. Based on the oldest dates at the Shawnee-Minisink and Upper Shawnee Island sites (Chapter 5, table 2), the basal gravel is older than 11,000 yrs B.P. The period represented by the sequence of sediments below the older dates and above the coarse gravel is unknown. Because the rate of sedimentation for the late Pleistocene alluvium has not been constrained by radiocarbon dating and is too variable throughout the valley, an accurate estimate of its age cannot be determined. However, ancillary evidence (chiefly stratigraphic) suggests the basal gravel beneath the T2 flood-plain deposits is not glacial outwash, but outwash reworked and incised by the postglacial Delaware River. The position and stratigraphy of the T3 terrace (fig. 3) show that it is older than the T2 terrace and it was laid down by a river that was much higher than the T2 river. The T3 deposits represent the oldest flood-plain deposits in Minisink Valley that were probably laid down by a non meltwater fed stream that had a non meandering channel form. This river, apparently operating under a condition of equilibrium, deposited a thin flood plain, and because the T3 deposits appear to have been laid down by a non meltwater or largely non meltwater-sourced stream, they may date to a time about 14,000 to 15,000 years ago. Incision to the T2 level appears to have been initiated by the onset of delayed isostatic rebound, and possibly a reduction in sediment supply due to the transition form a tundra to a closed boreal forest. The 14,000 yr B.P. maximum date for the start of rebound (Koteff, 1989) and the 14,250 yr B.P. date marking the transition from herb to spruce pollen zones (Cotter, 1983), may be in accordance with the estimated age of the T3 terrace. The T2 terrace represents episodic periods of alluviation throughout the Holocene. Leopold and others (1964, p. 326) noted that the progressive lateral migration of the river channel removes portions of the flood plain and so limits the elevation of its surface. Due to the narrow width of Minisink Valley and low sinuosity of the Delaware River channel, the T2 flood plain outgrew its fluvial setting and eventually became abandoned, receiving sediment only during the greatest of floods.

Retrace steps back to parking area. The trip will officially end here. Please have a safe return trip. Your participation is greatly appreciated.

Time permitting, I will lead the remaining group up to the top of the bluff at stop 2-4b.

Discussion (stop 2-4b) - A recent slump exposed a 15-foot section at the top of the bluff on the south side of the parking area. The most direct route to the outcrop is to climb the steep bluff face (60 feet). The top of the bluff affords an excellent view (up the valley) of the Montague valley train (high terrace in mid ground) and the lower meltwater and postglacial terraces in Minisink Valley. The materials exposed here are part of an outwash-fan deposit (fig. 5) laid down at the mouth of Raymondskill Creek. Postglacial stream erosion has removed about 60 feet of sediment along the course of the modern creek. The south side of the fan is marked by an ice-contact slope, although the fan appears to be graded to the surface of the Montague valley train (see figure 4c in Chapter 5). Presumably the ice was only of local extent.

The outcrop consists of four feet of planar-bedded cobble-pebble gravel overlying, pebble gravel, and pebbly sand with minor cross-bedded sand. The direction of paleoflow is east to southeast, based on the orientation of elongated clasts and cross-bedding. This shows that the outwash here was laid down by meltwater draining the Raymondskill Creek drainage basin, rather than the main trunk valley. The dominant lithologic suite of siltstone and sandstone clasts also shows that most of the material here was largely derived from local upland sources. A small exposure dug out about 2/3 of the way down the slope revealed similar sediment. Preliminary work in this area shows that the fan is a fluvial deposit, rather than deltaic.

Questions - Does the distribution of valley-train and outwash-fan deposits in Minisink Valley indicate deposition against large blocks of stagnant ice, or has most of this material been eroded by meltwater and postglacial streams? Did ice-contact lakes exist in Minisink Valley during deglaciation?

REFERENCES

Cotter, J. F. P., 1983, The timing of the deglaciation of northeastern Pennsylvania and northwestern New Jersey: Doctoral Dissertation, Lehigh University, Bethlehem, Pa., 159 p.

Crowl, G.H., 1971, Pleistocene geology and unconsolidated deposits of the Delaware Valley, Matamoras to Shawnee on Delaware, Pennsylvania, Pennsylvania Geological Survey, 4th., ser., General Geology Report 71, 68 p.

Crowl, G.H., and Sevon, W.D., 1980, Glacial border deposits of late Wisconsinan age in northeastern Pennsylvania, Pennsylvania Geological Survey, 4th ser., General Geology Report 71, 68 p.

Dent, R. J., 1991, Archaeology in the upper Delaware Valley: the earliest populations: in Orr, D. G., and Campana, D. V. (Eds.) The People of Minisink, Papers from the 1989 Delaware Water Gap Symposium: National Park Service, U.S. Department of the Interior, p. 117-143.

Drake, A.A. (1992), Bedrock geologic map of the Newton West quadrangle, Sussex and Warren Counties, New Jersey: U.S. Geological Survey Geologic Quadrangle Map GQ-1703, scale 1:24,000.

Drake, A.A., and Monteverde, D.H. (1992), Bedrock geologic map of the Branchville quadrangle, Sussex County, New Jersey: U.S. Geological Survey Geologic Quadrangle Map GQ-1700, scale 1:24,000.

Drake, A.A., and Volkert, R.A., (1993), Bedrock geologic map of the Newton East quadrangle, Sussex County, New Jersey: U.S. Geological Survey Geologic Quadrangle Map GQ-1707, scale 1:24,000.

Epstein, J. B., 1969, Surficial Geology of the Stroudsburg Quadrangle, Pennsylvania-New Jersey: Pennsylvania Geological Survey, 4th series, Bulletin G57, 67p., scale 1:24,000

Epstein, J.B., and Koteff, Carl, in press, Surficial Geology of the Saylorsburg Quadrangle, Pennsylvania: U.S. Geological Open-file Report.

Fleisher, P. J., 1986, Glacial geology and late Wisconsinan stratigraphy, upper Susquehanna drainage basin, New York: in Cadwell, D.H., (ed.), The Wisconsinan Stage of the First Geological District, Eastern New York: New York State Museum, Bull. no. 455, p. 121-142.

Flint, R. F., 1971, *Glacial and Quaternary Geology*, New York: John Wiley and Sons Inc.

Happ, S. C., 1938, Significance of Pleistocene deltas in the Minisink Valley: Amer. Jour. Sci. Series V, v. XXXVI, p. 417-439.

Koteff, Carl, and Pessl, Fred, Jr., 1981, Systematic ice retreat in New England: U.S. Geological Survey Professional Paper 1179, 20 p.

Koteff, Carl, and Larsen, F. D., 1989, Postglacial uplift in western New England: Geologic evidence for delayed rebound in Gregersen, S., and Basham, P. W., (eds.), Earthquakes at North Atlantic passive margins: Neotectonics and Postglacial Rebound, p. 105-123.

Leopold, L. B., Wolman, M. G., and Miller, J. P., 1964, Fluvial Processes in Geomorphology: W. H. Freeman and Company, San Francisco, 522 p.

McNett, C. W. Jr., McMillian, B. W., and Marshall, S. B., 1977, The Shawnee-Minisink site in Newman, W. S., and Salwen, Bert (eds.) Amerinds and Their Paleoenvironments in Northeastern North America, Annals of the New York Academy of Sciences, v. 288, p. 282-298.

Minard, J. P., 1961, End moraines on Kittatinny Mountain, Sussex County, New Jersey: U.S. Geological Survey Professional Paper 424-C, p. 61-64.

Monteverde, D.H., 1992, Bedrock geologic map of the Sussex County, New Jersey Portions of the Culvers Gap and Lake Maskenozha quadrangles: N. J. Geological Survey Geologic Map Series 92-1, scale 1:24,000

Ozvath, D.L., and Coates, D.R., 1986, Woodfordian stratigraphy in the western Catskill Mountains, in Cadwell, D.H., (ed.), The Wisconsinan Stage of the First Geological District, Eastern New York: New York State Museum, Bull. no. 455, p. 109-120.

Ridge, J.C., 1983, The surficial geology of the Great Valley section of the Valley and Ridge province in eastern Northampton County, Pennsylvania, and Warren County., New Jersey: Unpublished M.S. thesis, Lehigh University, Bethlehem, Pa., 234 p.

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

Sevon, W.D., Berg, T.M., Schultz, L.D., and Crowl, G.H., 1989, Geology and mineral resources of Pike County, Pennsylvania: Pennsylvania Geological Survey, County Report 52, 141 p. 2 plates, scale, 1:50,000.

Stewart, Michael, 1991, Archaeology and environment in the upper Delaware: in Orr, D. G., and Campana, D. V. (Eds.) The People of Minisink, Papers from the 1989 Delaware Water Gap Symposium: National Park Service, U.S. Department of the Interior, p. 79-115.

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

Wagner, D. P., 1994, Pedology and geomorphology of the Depew Recreation Area, in Inashima, P. Y.,Gemorphology, Remote Sensing, and Archealogical Monitoring at Depew Recreation Area, Dept of the Interior, National Park Service, p. 1-36.

White, I.C., 1882, The geology of Pike and Monroe Counties: Pennsylvania Geological Survey, 2d, Report G-6, 333 p.

Witte, R.W., 1988, The surficial geology and Woodfordian glaciation of a portion of the Kittatinny Valley and the New Jersey Highlands in Sussex County, New Jersey: Unpublished M.S. thesis, Lehigh University, Bethlehem, Pa., 276 p.

_____, 1997a, Surficial geology of the New Jersey portion of the Culvers Gap and Lake Maskenozha quadrangles: N.J. Geological Survey Open-file Map, scale 1:24,000.

_____, 1997b, Surficial geology of the New Jersey portion of the Milford and Port Jervis South quadrangles: N.J. Geological Survey Open-file Map, scale 1:24,000.