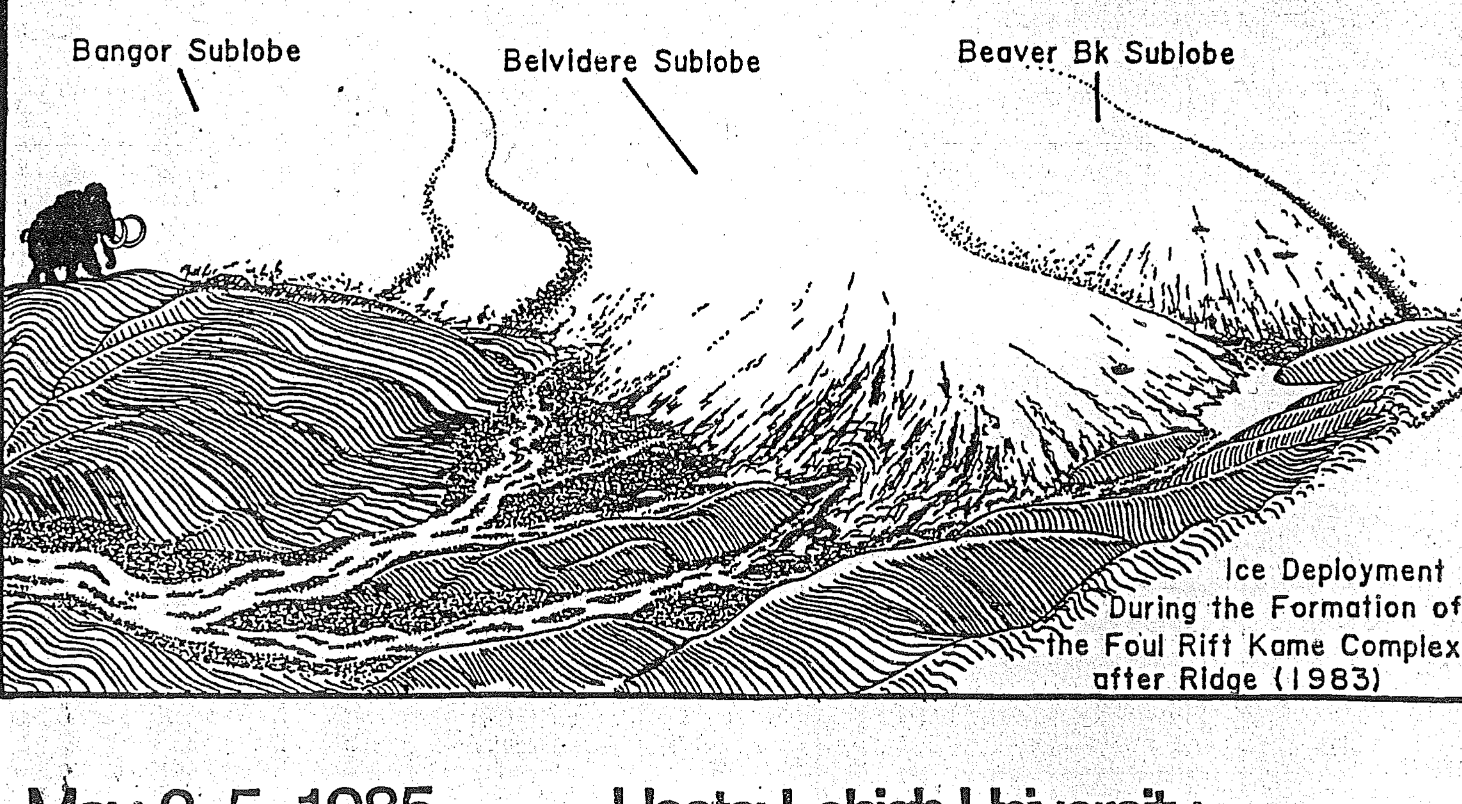
Guidebook

48TH FRIENDS OF THE PLEISTOCENE FIELD CONFERENCE

WOODFORDIAN DEGLACIATION OF THE GREAT VALLEY, NEW JERSEY



Ice Deployment During the Formation of the Foul Rift Kame Complex after Ridge (1983)

May 3–5, 1985
McAfee, N.J.

Hosts: Lehigh University
New Jersey Geol. Survey

GUIDEBOOK for the

48th FRIENDS OF THE PLEISTOCENE FIELD CONFERENCE

WOODFORDIAN DEGLACIATION OF THE GREAT VALLEY, NEW JERSEY

Organizer: Ed Evenson, Lehigh University

Leaders:
Jim Cotter, University Minnesota - Morris
Ed Evenson, Lehigh University
Dave Harper, N.J. Geological Survey
Carl Koteff, U.S. Geological Survey
Jack Ridge, Syracuse University
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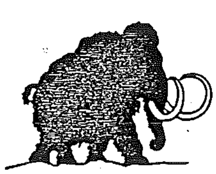
Ron Witte, Lehigh Univ. & N.J. Geological Survey

DATES: May 3-5, 1985

HOSTS: Lehigh University
New Jersey Geological Survey

HOTEL: Americana Great Gorge Resort
McAfee, New Jersey



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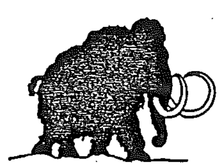

INTRODUCTION

"We are still confused -- but on a higher level"

The 1985 Friends of the Pleistocene trip marks the 48th reunion of the "Friends". This year almost ninety (90) friends will spend part of three days together in gravel pits, in cramped buses, and in the luxury of the former Great Gorge Playboy Club. The last time the friends met in northwestern New Jersey was in 1939. The group included such notables as: Dick Flint, George White, Don Chapman, Dr. and Mrs. Charlie Denny, Henry Herpers, Dr. and Mrs. Link Washburn, Meredith Johnson, Kirk Bryan, Lewis Peltier, E. T. Appfel, Peter Wolfe, Chauncey Holmes and was lead by Paul MacClintock. At that time hotels were $1/night not including the "reduced rates" MacClintock hoped to get for the group (see meeting announcement). They visited the same pits (North Church, Ogdensburg), asked the same questions and were, in the end, confused.

Forty-six years later we return with new techniques, new ideas, new maps, new friends, and we are still confused -- but on a higher level (we hope!). In reality, the basic observations of the masters like Salisbury, Lewis, Chamberlain and Leverett, were uncannily accurate and have weathered well. In many cases the revisions made by workers who followed these pioneers have not stood the test of time nearly so well and much of our recent work has simply closed the circle of revision where it started. Knowing full well the cycle is about to begin again, we welcome all of you to what we hope will be a scientifically and personally rewarding experience -- now let's see the evidence (ca. 1985).

ANNUAL INVASION
FRIENDS OF PLEISTOCENE GEOLOGY

May 20 and 21, 1939
New Jersey

The plan for this spring is to spend half a day on the Wisconsin drift phenomena in the northwestern part of the State, made classic by the early work of Salisbury. The remainder of the time will be devoted to the older drift sheets. The main topic for demonstration and discussion is the evidence for separating the Jerseyan into two or more drift sheets.

The party will assemble at the Sussex Inn, Sussex, New Jersey, (Room $2.00) to start at 8:00 o'clock Saturday, May 20. That night will be spent at Hackettstown Hotel, Hackettstown (Room $1.00-$2.00) or nearby tourist camps.

Ref:

Salisbury, "Glacial Geology of New Jersey." Geol. Surv. N. J. Vol. V, 1902.

Leverett, "Glacial Deposits Outside the Wisconsin Terminal Moraine in Pennsylvania." Penna. Topogr. and Geol. Surv. Bull. G7, 1934.

Topographic Maps. Sheets 24 and 25. New Jersey Geol. Survey (Price 50¢ each, dog gone it!)

Signed, local Committee

March 27, 1939

Henry Herpers
Meredith Johnson
Paul MacClintock, Secretary
Princeton University
Princeton, New Jersey

P. S. If the group will let me know how many are coming I can probably get reduced rates at the hotels. Also if anyone wants a route map to join us along the way I will mark one and send it to him.

P. MacClintock

ACKNOWLEDGMENTS

The Friends of the Pleistocene and the leaders of the field trip wish to thank all those who played a role in making this trip possible (and hopefully a success). Special thanks goes to those land and pit owners who allowed our group access to their property. They are: Robert C. Hummer, Foul Rift Sand and Gravel, Inc., Belvidere, N.J.; Ted Sturdevant, Grinnell Development Co., Inc., Sparta, N.J.; Tim Williams, Trans-Mix Sand and Gravel, Brainards, N.J.; "Mickey" at the truck terminal and "A.J." at A&M Auto, Ogdensburg, N.J. and several we overlooked!

Many geologists gave of their time to visit and discuss the deposits and deglaciation of the area with the trip leaders. To each and every one of them we extend our sincere gratitude. They are: Byron Stone, Jack Epstein and Bill Sevon.

Special thanks are due Lynn Murray and Garth Howe of the Americana. Without their patience, flexibility and good sense of humor the logistics for this conference would have been impossible.

Gunnar Schlieder and Susanne Pearce gave of their time to look after lunches and "refreshments" and to them we owe our comfort and "warm glow".

Lastly, and certainly not least, we thank Laura Cambiotti, who acted as meeting secretary, typist, and hostess. Without her efforts all aspects of this meeting would have fallen into chaos.

DEGLACIATION HISTORY

"... and the record low for this date is 147° below zero, which occurred 28,000 years ago during the Great Ice Age."

The Wisconsinan History of the Great Valley, Pennsylvania and New Jersey, and the Age of the "Terminal Moraine"

by

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ABSTRACT

New radiocarbon dates, detailed morphosequence mapping, and palynologic and pedogenic investigations in the Great Valley of Pennsylvania and New Jersey all support the hypothesis that the "Terminal Moraine" is Woodfordian in age and that retreat from this moraine began prior to 18,500 yr B.P.

Four major valley-constrained sublobes of the Ontario and Hudson-Champlain Lobes in the Delaware, Beaver Brook, Jacoby-Paulins Kill and Pequest Valleys deposited correlatable heads of outwash and end moraines at numerous recessional positions as the rapidly thinning, active ice margin retreated northward. In addition, these sublobes interacted with ice north of Kittatinny Mountain indicating contemporaneous occupation of the Great and Minisink Valleys.

Basal radiocarbon dates of 18,390 ± 200 (SI-4,921), and 18,570 ± 250 yr B.P. (SI-5,273) from Francis Lake, New Jersey place a minimum age on the initiation of deglaciation and the establishment of tundra vegetation in the area. Additional dates of 16,480 ± 430 (SI-5,274); 13,510 ± 135 (SI-5,300) and 11,220 ± 110 B.P. (SI-5,301) higher in the profile, and pollen stratigraphy comparable to the regional record of northeastern North America, indicate that these basal dates are correct.

INTRODUCTION

The Great Valley in Pennsylvania and New Jersey was occupied by both the Hudson-Champlain and Ontario Lobes of the Laurentide ice sheet during the maximum extent of the Woodfordian glaciation and the earliest phases of deglaciation. Documentation of the timing and geometry of the retreat of these lobes from the Great Valley is of critical importance to the development of the deglaciation history and glacial stratigraphy of New York and surrounding areas.

Surficial mapping of the Great Valley began with the tracing of the "Terminal Moraine" by Lewis (1884) in Pennsylvania and Salisbury (1902) in New Jersey. The "Terminal Moraine" was defined as a morphostratigraphic landform marking the limit of the latest phase of glaciation now known to be Wisconsinan. Subsequently, the Great Valley's surficial geology has been mapped by numerous workers (Ward, 1938; Herpers, 1961; Connally and Sirkin, 1970, 1973; Connally, 1973, 1979). Although these studies have advanced our understanding of the glacial history of the area, two principal problems have emerged. They are: 1) the delineation of the geometry and extent of the Woodfordian glaciation (Ward, 1938; Connally and Sirkin, 1973; and Crowl and Sevon, 1980), and 2) the determination of the absolute age of inception of the Woodfordian deglaciation (Crowl, 1980).

Recent mapping of the glacial deposits using the morphosequence concept (Koteff and Pessl, 1981) has documented systematic ice retreat in the Great Valley (Ridge, 1983). Basal sediment radiocarbon dates, in association with pollen stratigraphy, indicate that the Woodfordian deglaciation began in the "Terminal Moraine" area prior to 18,500 yr B.P. Together these dates form the basis for the reconstruction of the patterns and chronology of the deglaciation of the Great Valley that are presented in this paper.

LOCATION AND SETTING

The study area includes 400 km^2 of the Great Valley section of the Valley and Ridge physiographic province in Pennsylvania and New Jersey (Figure 1). This area is bordered by Kittatinny Mountain on the northwest and the hills of the Reading Prong and New Jersey Highlands on the southeast and covers parts of Warren and Sussex Counties in New Jersey and part of Northampton County in Pennsylvania.

Three northeast-southwest trending lowlands occur within the study area (Figure 2). The northernmost of these, bordered on the north by Kittatinny Mountain and on the south by slate uplands, contains Jacoby Creek, Paulins Kill, and the headwaters of Martins Creek. The Delaware River traverses this lowland southward from the Delaware Water Gap and forms the southwestern portion of the central lowland, with Beaver Brook to the northeast. The central lowland is separated from the lowland to the south by Jenny Jump Mountain. This southern lowland is drained by the Pequest River. During deglaciation these three lowlands and intervening uplands controlled the flow of ice in the Great Valley.

The Great Valley is characterized by moderate to high relief developed on resistant, clastic-rock uplands and carbonate lowlands. Within the valleys smaller scale, low relief features include carbonate-karst and shale-knoll topographies. The highest elevation in the study area is the summit of Kittatinny Mountain (Mt. Tammany: 475 m) and the lowest point is on the Delaware River (52 m).

The bedrock geology of the area includes a diverse suite of rocks which decrease in age northwestward from the Precambrian metamorphic rocks of the New Jersey highland to the Silurian sandstones which underlie Kittatinny Mountain (Figure 3).

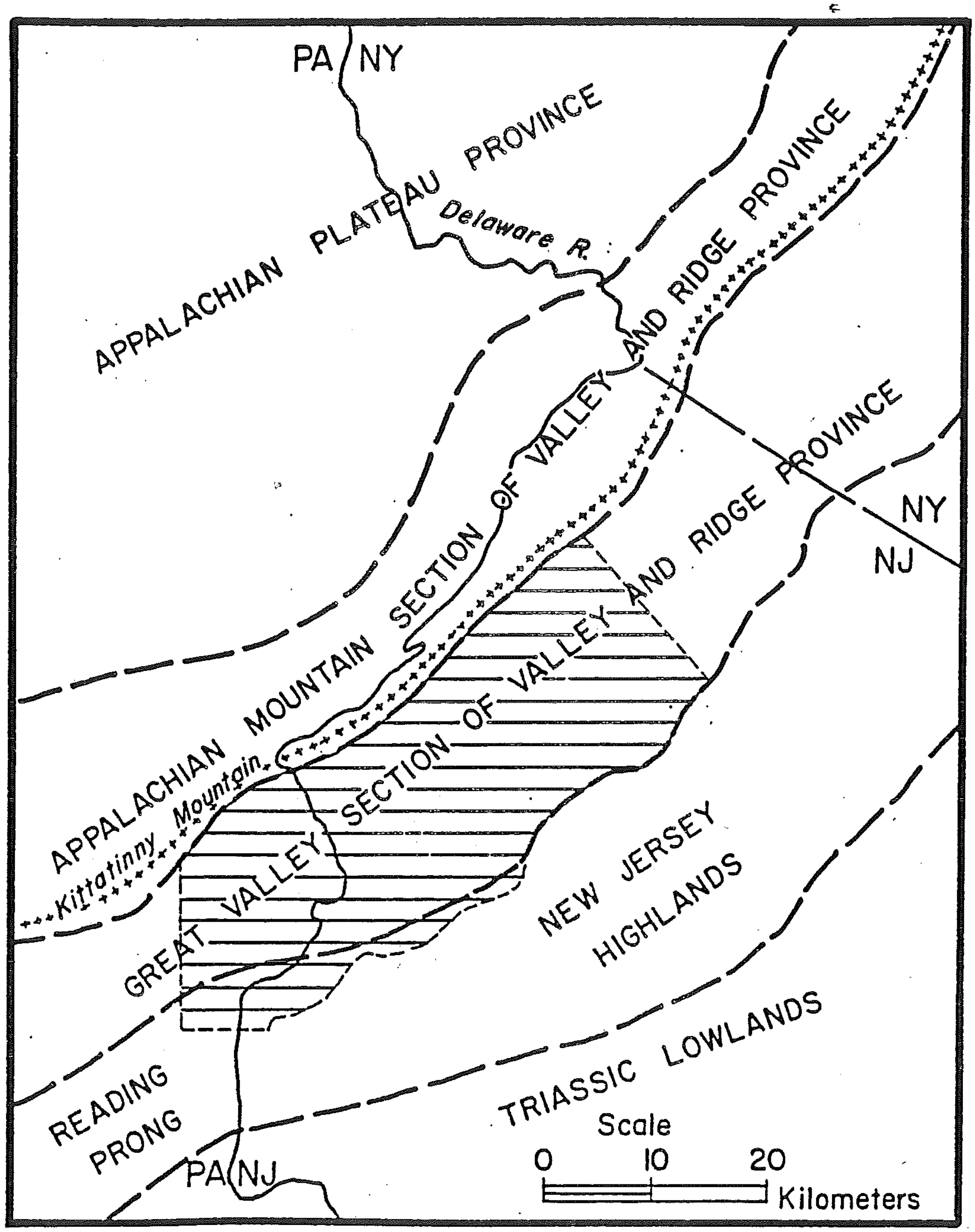


Figure 1. Location map and physiographic provinces of the study area. The study area is shown by the ruled pattern.

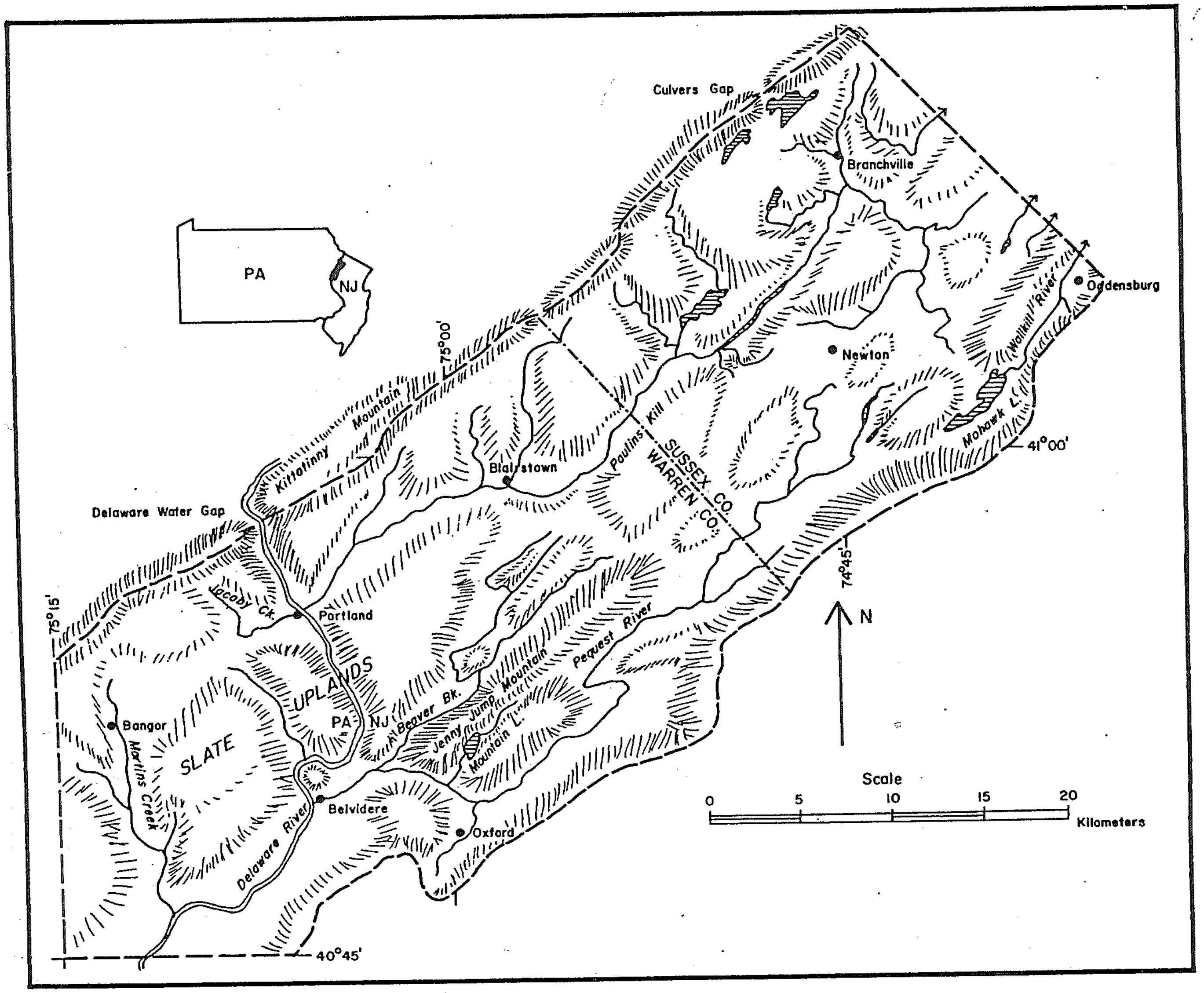


Figure 2. Physiography of the study area in the Great Valley section of the Valley and Ridge Province.

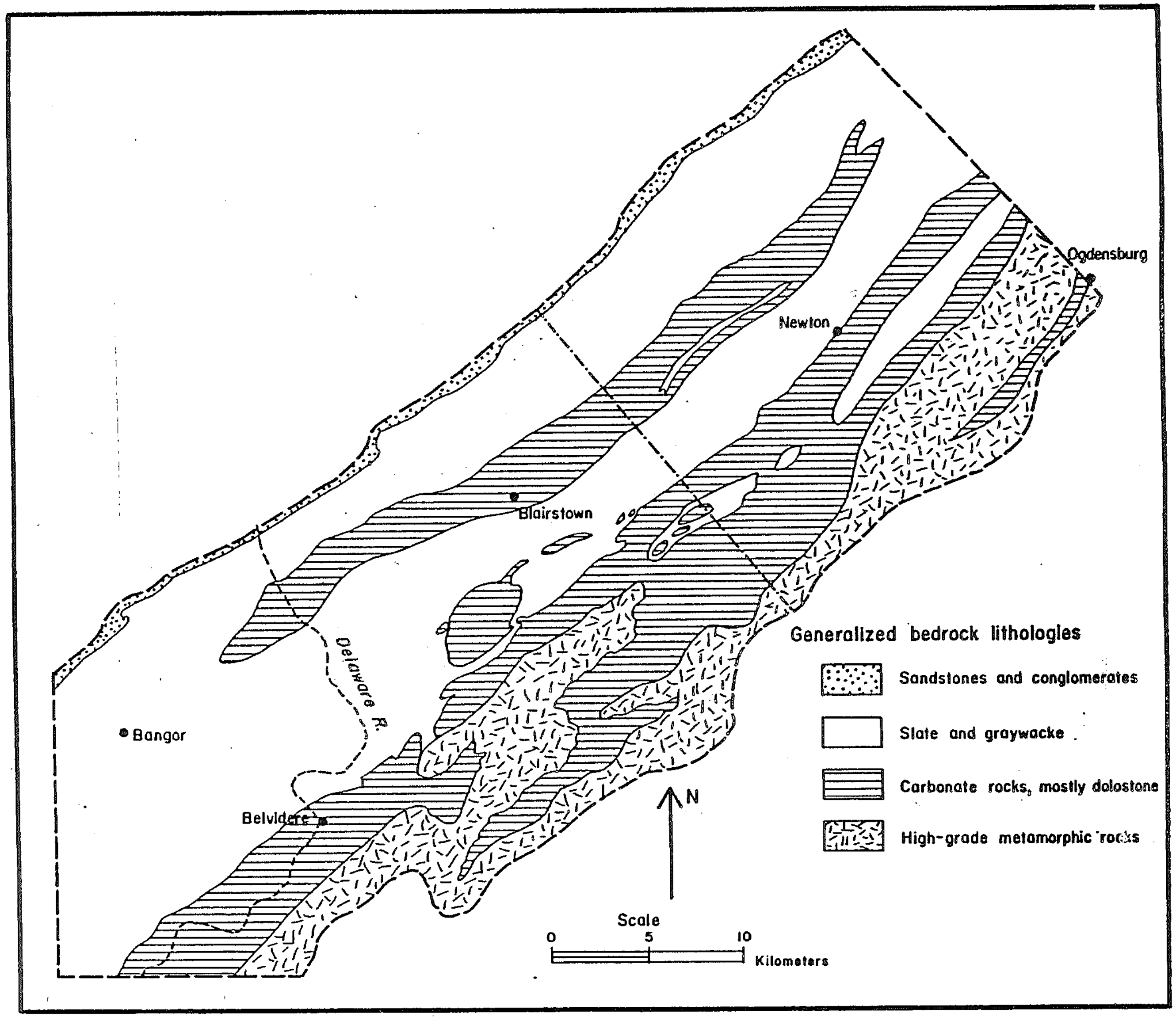


Figure 3. Generalized bedrock lithologies of the study area after Johnson (1950).

PREVIOUS WORK

The principal reference for Quaternary studies of the Great Valley is that of Salisbury (1902), a comprehensive description of all deposits in New Jersey associated with glaciation. The first published works on the glacial deposits of the Great Valley in New Jersey were those of the New Jersey Geological Survey as part of their annual report series (Cook, 1878, 1879a, 1879b, 1881). In Pennsylvania, Hall (1976), Prime (1879), and Lewis (1883a) made notes on the glaciation of Kittatinny Mountain and glacial deposits in Northampton County. Early tracings of the "Terminal Moraine" in New Jersey and Pennsylvania by Wright (1882) and Lewis (1883a, 1883b, 1884) ultimately resulted in the correlation of this morphostratigraphic unit from Massachusetts to Indiana (Chamberlain, 1883). Mapping of "extra morainic" drift by Lewis (1884), Salisbury (1892a, 1892b), Ward (1938), Wright (1893), Williams (1893, 1894a, 1894b, 1895, 1902, 1917, and 1920) and Leverett (1928, 1934) further defined the Late Wisconsinan limit of glaciation and contributed to the understanding of prior glacial events.

More recently, researchers have taken one of three approaches to the mapping of Quaternary deposits in the Great Valley region. These include; 1) moraine tracing, the mapping of moraines and kame deposits to define Woodfordian ice-marginal positions (Minard, 1961; Herpers, 1961; Crowl and Sevon, 1980); 2) surficial mapping, the mapping of all surficial deposits at a scale of 1:24,000 (Epstein, 1969; Crowl, 1972; Berg and others, 1977); and 3) morphosequence mapping; the mapping of successive ice-marginal positions in individual valleys (Connally and Epstein, 1973; Ridge, 1983). Although previous work has contributed greatly to the understanding of the glacial history of Pennsylvania and New Jersey, relatively few attempts have been made to develop a stratigraphic model for the region. Notable exceptions are, Sirkin and Minard (1972), Crowl and others (1975), Sevon and others (1975), Crowl and Sevon (1980) and Connally and Sirkin (1973, in review). The models of these workers, although disagreeing in part, conform remarkably well with the model presented by Salisbury in 1902: a morphostratigraphic model based principally on the position and characteristics of the "Terminal Moraine" and the relation of other glacial deposits to it.

The most conspicuous glacial landform in Pennsylvania and New Jersey is the "Terminal Moraine", which varies in morphology and composition throughout its length (Crowl and Sevon, 1980). These changes are due, in part, to the varying sources of glacial debris. From Kittatinny Mountain northwestward to the Salamanca Reentrant in New York state the "Terminal Moraine" and all other Woodfordian materials were deposited by the Ontario Lobe of the Laurentide Ice Sheet (Crowl and Sevon, 1980). The Hudson-Champlain Lobe deposited most of the glacial materials east of the Great Valley (Connally and Sirkin, 1973; Ridge, 1983). Interlobate deposits or deposits with till pebble lithologies characteristic of both lobes have not been recognized in the Great Valley.

In Pennsylvania, Crowl (1974; 1980), Crowl and others (1975), and Crowl and Sevon (1980) have concluded that the "Terminal Moraine" is Woodfordian in age. These workers have also suggested that the same deposit, mapped by Salisbury (1902) in New Jersey, is also of Woodfordian age. However, Connally (1979), citing his unpublished work on the Ogdensburg-Culvers Gap Moraine, has suggested that the "Terminal Moraine" in Warren County, New Jersey may be older than Woodfordian.

The Ogdensburg-Culvers Gap Moraine (Figure 4a) was originally mapped by Salisbury (1902) as a recessional moraine deposited during the

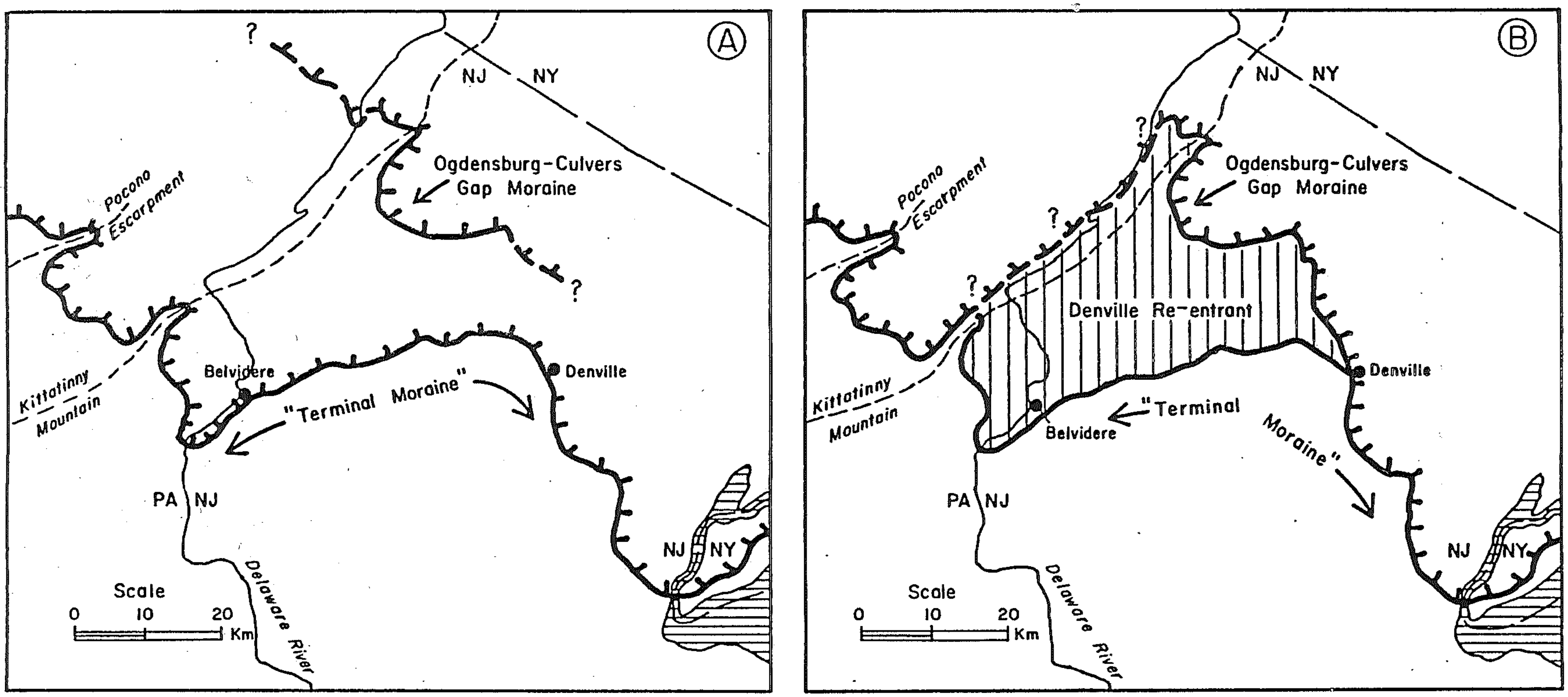


Figure 4. a) The Woodfordian terminal position and recessional moraine positions compiled from Salisbury (1902) and Sevon and others (1975).

b) The Woodfordian terminal position of Connally and Sirkin (1973) and Connally (1979).

northeastern retreat of the Hudson-Champlain Lobe. Additional remnants of moraines were mapped by Herpers (1961) who referred to the deposit as the "Ogdensburg-Culvers Gap Moraine". This name has been used by Minard (1961) and Minard and Rhodehamel (1969) in mapping additional deposits in the area, and by Sevon (1970) and Sevon and others (1975) when they mapped an assumed correlative deposit in Pennsylvania. All of these workers considered this deposit to be a Woodfordian recessional moraine (Figure 4a).

On the basis of provenance and flow direction indicators, Connally and Sirkin (1970 and 1973) suggested that the southernmost extent of the Hudson-Champlain Lobe in western New Jersey was marked by the Ogdensburg-Culvers Gap Moraine. In addition, Connally (1979) concluded "that the Culvers Gap moraine is not a recessional moraine but rather the terminal Woodfordian position" of the Laurentide ice sheet, thus suggesting that the glaciated region southwest to the "Terminal Moraine" was somewhat older (Figure 4b). The assignment of a maximum Woodfordian age to the Ogdensburg-Culvers Gap Moraine was based principally on provenance and pedologic data (Connally and Sirkin, 1973; Connally, 1979). Connally (1979) did not assign an age to the "Terminal Moraine".

Attempts to determine the absolute age of the "Terminal Moraine" in Pennsylvania and New Jersey have been inconclusive due to the paucity of organic matter in glacial deposits in the Great Valley area. Estimates of the age of the "Terminal Moraine" have been based for the most part on radiocarbon dates obtained from the base of organic-rich sediments of lakes and bogs associated with the moraine.

There are numerous ambiguities associated with "base of organic sediment" dates (which are often misnamed "bog-bottom" dates) and true "bog bottom" dates when they are used as minimum ages for deglaciation. The principal problem is that the length of time between deglaciation and the initiation of deposition of datable organic material is indeterminable unless stratigraphic control is available. Organic deficiency in basal lacustrine and bog sediments results in extreme lag time problems. Other ambiguities associated with minimum dates result from the unknown length of time between deglaciation and lake formation, and contamination by younger organic material (Cotter and others, 1983).

The difficulties inherent in the interpretation of base of organic sediment dates have resulted in the development of two different theories about the timing of regional deglaciation. Crowl (1975, 1978, 1980) and Crowl and Sevon (1980) suggest that deglaciation in northeastern Pennsylvania began at approximately 15,000 yr B.P. Crowl (1980), citing 12 "bog-bottom" radiocarbon dates ranging in age from 12,520 to 14,170 yr B.P. from sites near the glacial border, believes that the mutual consistency of these dates and the absence of older dates indicates that the influence of the Woodfordian ice sheet in the region ceased about 15,000 yr B.P. Crowl (1980) also suggests there is no evidence for a significant time lag between deglaciation and initial organic-rich sediment deposition.

In contrast, Sirkin and Minard (1972), Connally and Sirkin (1973; in review), Sirkin (1977), and Connally (this volume) suggest that northeastern Pennsylvania and northwestern New Jersey were deglaciated sometime prior to 18,000 yr B.P. These authors have attempted to solve the "lag time" problem by using sedimentation rates and thickness of sediments below the oldest obtainable radiocarbon dates to estimate the onset of deglaciation.

The resolution of this problem is essential to understanding the Late Wisconsinan stratigraphy in Pennsylvania, New Jersey, and New York. A 15,000 yr B.P. deglaciation of the Ontario Lobe from northeastern Pennsylvania is not consistent with the 14,000 "+" yr B.P. date assigned to the Valley Heads Moraine in western New York (Calkin and Miller, 1977; Muller, 1977). When compared to the well documented 18,000 yr B.P. age of deglaciation determined for the Erie Lobe (Dreimanis and Goldthwait, 1973), the problem becomes even more complex. Much of this paper is addressed to the resolution of this problem.

METHODS

FIELD MAPPING AND PEDOLOGIC ANALYSIS

Detailed field mapping was done according to the morphosequence concept of Jahns (1941) as explained by Koteff (1974). A sequence is a group of deposits associated with an interval of deposition and having a common base level. Koteff (1974) defines eight types of sequences, seven of which were observed in the study area. Because of the limited number of end moraines in the area, the morphosequence concept is extremely useful for delineating ice-recessional positions and the correlation of ice recessional positions from valley to valley. Ice flow direction indicators and provenance of glacial materials were analyzed to determine lobal (Ontario vs. Hudson-Champlain) affinities of all deposits. All mapping was done on 1:24,000-scale topographic maps with a 20-foot contour interval. An altimeter was used for determining more exact elevations than could be interpreted from topographic maps.

Soils were analyzed to determine the relative age of deposits in the study area. Soil profile sites were selected in areas that were well drained and on flat surfaces. Soil classification utilized the terminology of the Soil Survey Staff (1960). Soil horizons were differentiated on the basis of pedologic development, color, thickness, and structure. Quantification of relative age parameters was not attempted because soil parent material lithologies varied greatly. Instead, a qualitative age differentiation of glacial deposits was done using the system of Sevon (1974).

Two ages of glacial deposits were differentiated in the Great Valley on the basis of morphologic and pedologic characteristics; pre-Wisconsinan (Illinoian) and Late Wisconsinan (Woodfordian). Surficial deposits of pre-Wisconsinan age were mapped south of the Woodfordian border. These deposits possess little morphologic expression, and are intensely colluviated. Soil profile thickness greater than 2.4 m suggest a pre-Wisconsinan age for these deposits (Marchand, 1978, Crowl and Sevon, 1980). No deposits of Altonian age (early-middle Wisconsinan) were identified in the Great Valley.

Late Wisconsinan (Woodfordian) deposits were mapped as the Great Valley drift (new term, Ridge, 1983). These deposits probably correlate with the Olean Drift further north in Pennsylvania (Crowl and Sevon, 1980) and represent the last glacial episode in the Great Valley. The "Terminal Moraine", as mapped by Lewis (1884), Salisbury (1902), and Crowl and Sevon (1980) is a morphostratigraphic unit of Woodfordian age throughout the Great Valley (discussed below).

ABSOLUTE AGE DETERMINATION

During the winters of 1979, 1980, and 1981, 12 sites were cored using a Davis-type peat corer. The following factors were predetermined for each site: underlying material, closure, lake size (for feasibility or coring), relative age (morphosequence affiliation), origin of lake basin, and postglacial modification. Field description of core samples included sediment type and color, organic content, degree of compaction, macrofossil content, and sedimentary features. Core segments for pollen studies were extruded into 30 cm tubes in the field, sealed with stoppers, and stored for laboratory analysis.

Pollen analysis was used for reconstructing postglacial vegetation and stratigraphic control. The presence of an herb pollen zone similar to that described by Deevey (1949), Davis (1965), Sirkin and others (1970), Connally and Sirkin (1973), Sirkin (1977), and Watts (1979), was documented in the basal sediments of two sites. Sediment samples from these lower horizons were submitted for radiocarbon dating and the sites were re-cored to obtain material for complete pollen and additional radiocarbon analysis.

Because both the organic content of basal lake sediments and the size of the barrel of the corer (25 cm long, 2.5 cm diameter) was small, multiple core sections from the same depth were pooled for each radiocarbon sample, after the method of Sirkin and others (1970). When sampling for radiocarbon analysis, intervening sediments were left between dated levels to eliminate contamination. Samples for radiocarbon dating were packed in heavy-weight aluminum foil and refrigerated.

GLACIAL HISTORY OF THE GREAT VALLEY

WOODFORDIAN LIMIT OF GLACIATION

The "Terminal Moraine", as mapped by Lewis (1884), Salisbury (1902), and Crowl and Sevon (1980), has been assumed to represent the maximum extent of the Late Wisconsinan (Woodfordian) ice sheet. This interpretation has been questioned by both Ward (1938) and Ridge (1983).

Ward (1938), mapping in the Great Valley south of the "Terminal Moraine", noted that the deposits had an "intermediate color" due to the incorporation of "dark Illinoian till in fresh, light Wisconsinan till". Ridge (1983) recognized the presence of a narrow fringe of till and erratic boulders extending up to 5 km (Figure 5) in front of the "Terminal Moraine" in the Delaware Valley. This deposit closely mimics the pattern of the "Terminal Moraine" and consists of both freshly derived and previously weathered materials, giving the till an indeterminate age appearance. Ridge (1983) interprets these fringe deposits to be Woodfordian in age and to be the result of deposition during the maximum advance of Woodfordian ice, which subsequently retreated and deposited the "Terminal Moraine". The older appearance of Woodfordian drift in front of the "Terminal Moraine" is assumed to be due solely to the incorporation of older (weathered) material. Deposits of a similar age and origin in the Bangor area may have been incorrectly identified as Altonian in age by Sevon and others (1975).

Provenance and striation data document ice flow from the north-northeast out of the Ontario Lobe during the deposition of most of the "Terminal Moraine". The "Terminal Moraine" represents the first equilibrium position established by Woodfordian ice during its earliest retreatal phase. The position of the moraine is marked by both till deposits and heads of outwash in the Great Valley.

Following the deposition of the "Terminal Moraine", the influence of underlying topography resulted in the formation of 5 sublobes in the Great Valley: the Bangor, Belvidere, Beaver Brook, Mountain Lake, and Pequest Sublobes (Figure 5). As ice margin retreat and ice-sheet thinning progressed, tributary flow from the Ontario Lobe into the Great Valley

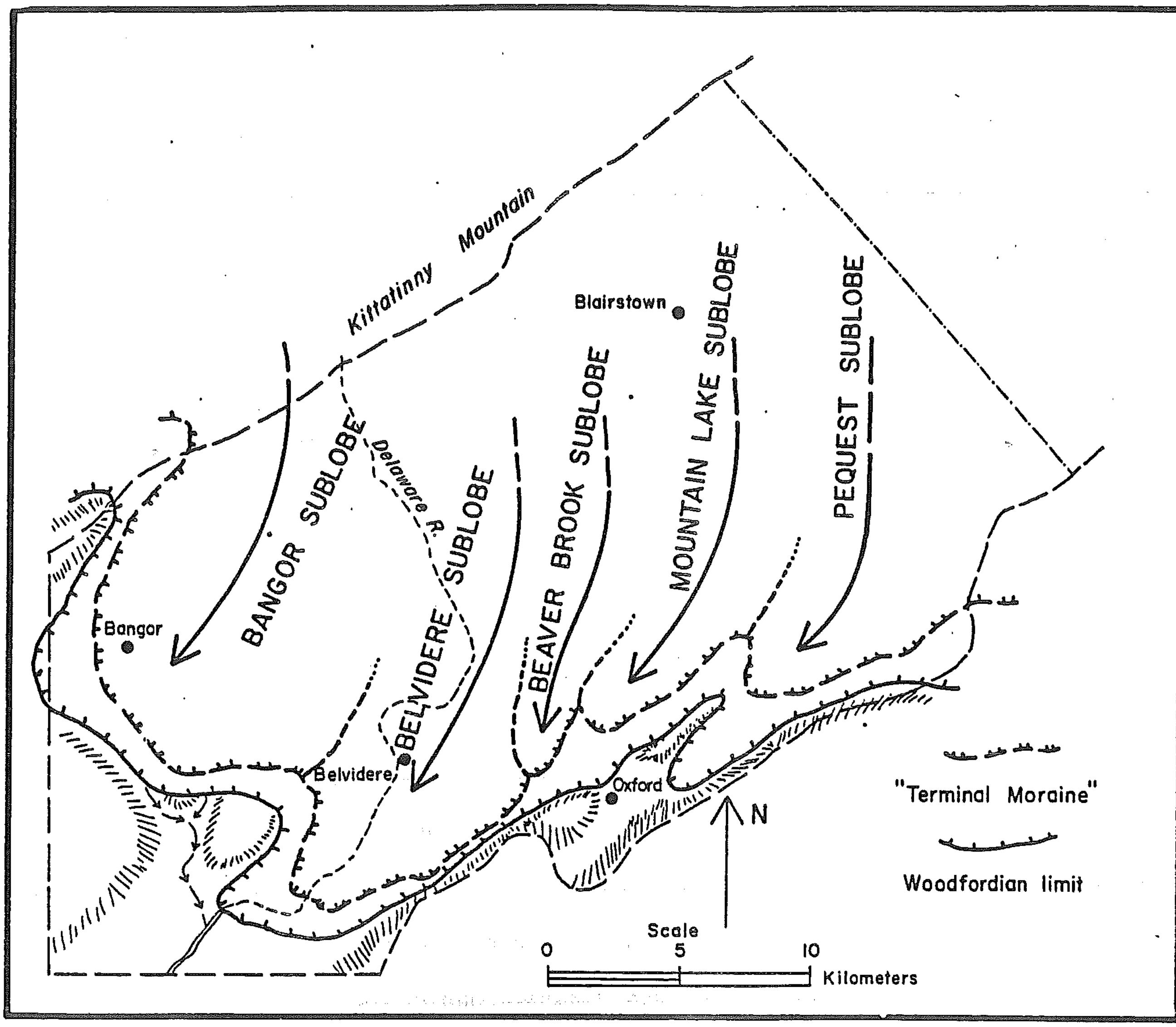


Figure 5. Sublobe nomenclature, flow patterns and geometry during the maximum extent of Woodfordian glaciation and deposition of the "Terminal Moraine" (Ridge, 1983).

diminished, and by the time of deposition of the Franklin Grove Moraine (discussed below), ice in the Great Valley was fed by the Hudson-Champlain Lobe. This resulted in major changes in sublobe deployment. The most significant changes were the development of the Paulins Kill sublobe, and the modification of the Beaver Brook sublobe (Figure 6). The Pequest sublobe underwent little change in local ice deployment and geometry even though the source area was changed (Figure 6).

DEGLACIATION OF THE BANGOR AND PAULINS KILL SUBLOBES

The Bangor sublobe occupied Jacoby Creek and the headwaters of Martins Creek (Figures 2 and 5). The Bangor Moraine, composed of bouldery red diamictons, and the Gruvertown Moraine, composed of drab, slatey diamictons, make up the portion of "Terminal Moraine" deposited by the Bangor sublobe (Figure 7). Initial deglaciation from these moraines is represented by a series of meltwater channels and morainic deposits. The Minis Lake Moraine (Figure 7), an early recessional deposit composed of both diamictons and stratified drift, is situated immediately north of an extremely bouldery till sheet. The ice-contact character, subdued topography and high boulder concentration all suggest abundant meltwater activity during this interval of ice margin retreat.

Due to ice thinning and recession, the Bangor sublobe could no longer be nourished by ice flowing through the Delaware Water Gap or over Kittatinny Mountain (Figure 8). Ice in the Jacoby Creek Valley at this time had its source to the northeast and in part from the Hudson-Champlain Lobe. This ice is referred to as the Paulins Kill sublobe on the basis of provenance (Ridge, 1983). Ice-contact outwash deposits from the earliest phase of the Paulins Kill sublobe retreat were originally mapped by Lewis

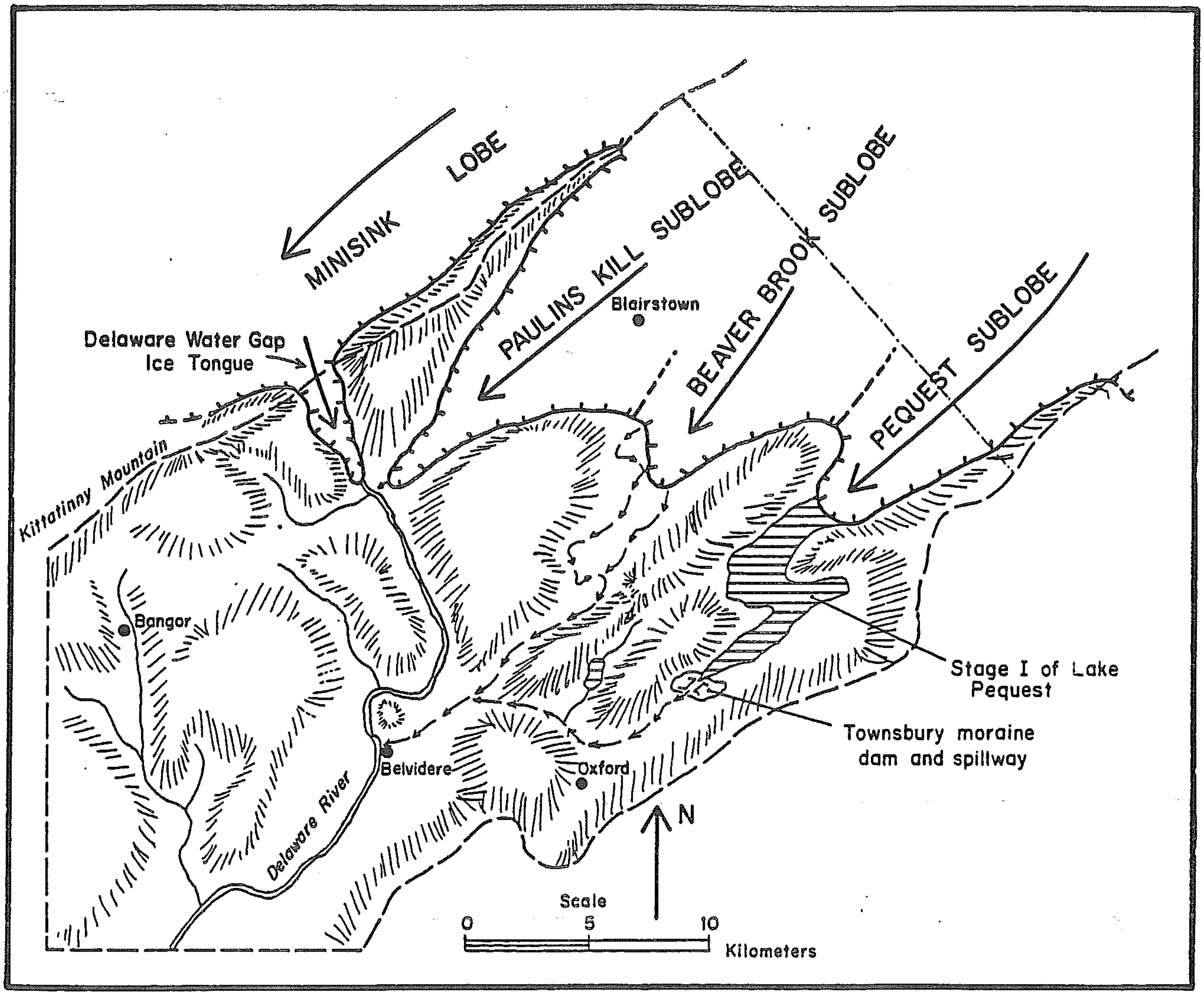


Figure 6. Sublobe nomenclature, flow patterns and geometry during early stages of the Woodfordian deglaciation (Ridge, 1983).

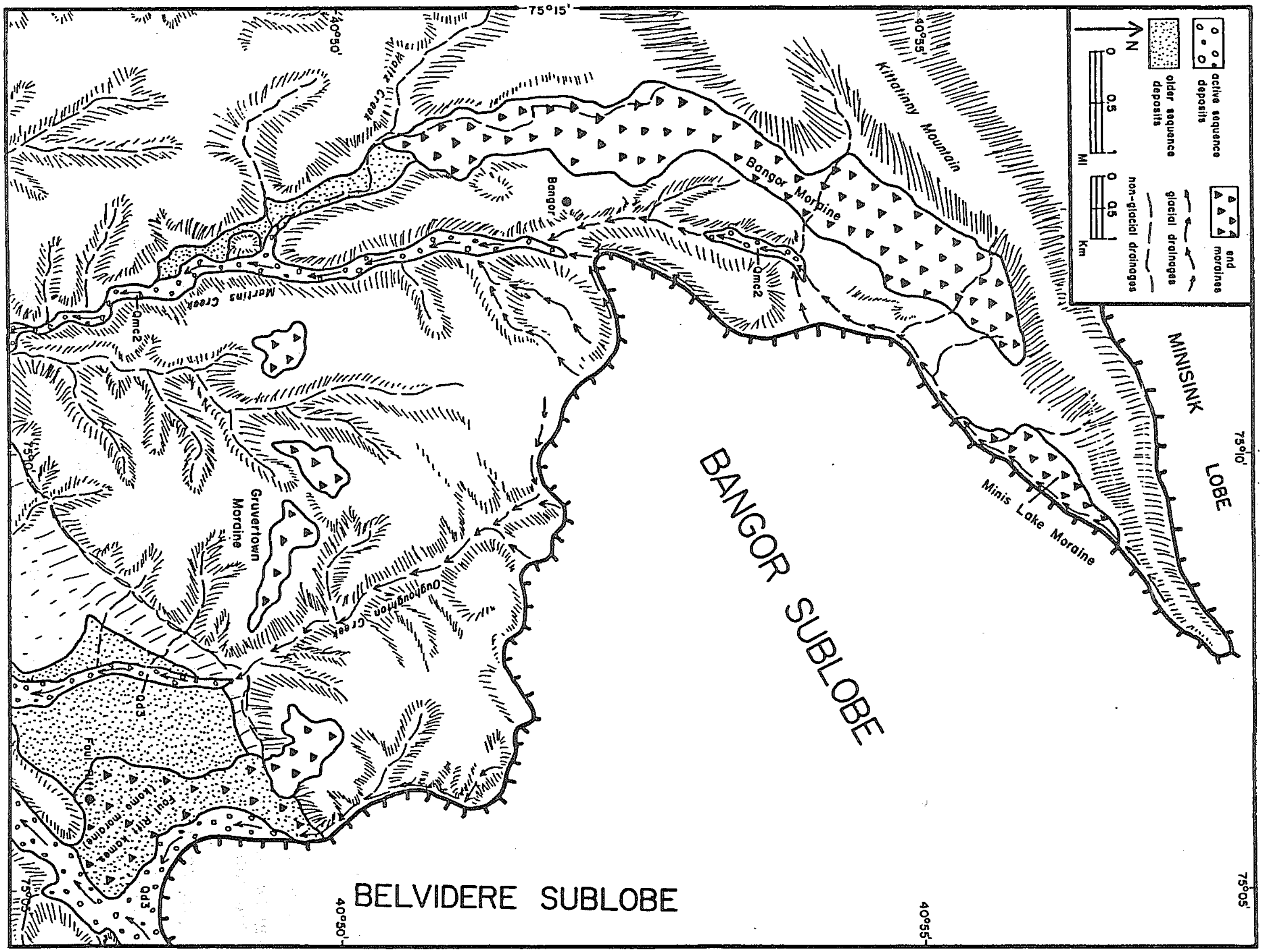


Figure 7. Location of the Bangor Moraine and reconstruction of the ice margin position during the deposition of the Minis Lake Moraine.

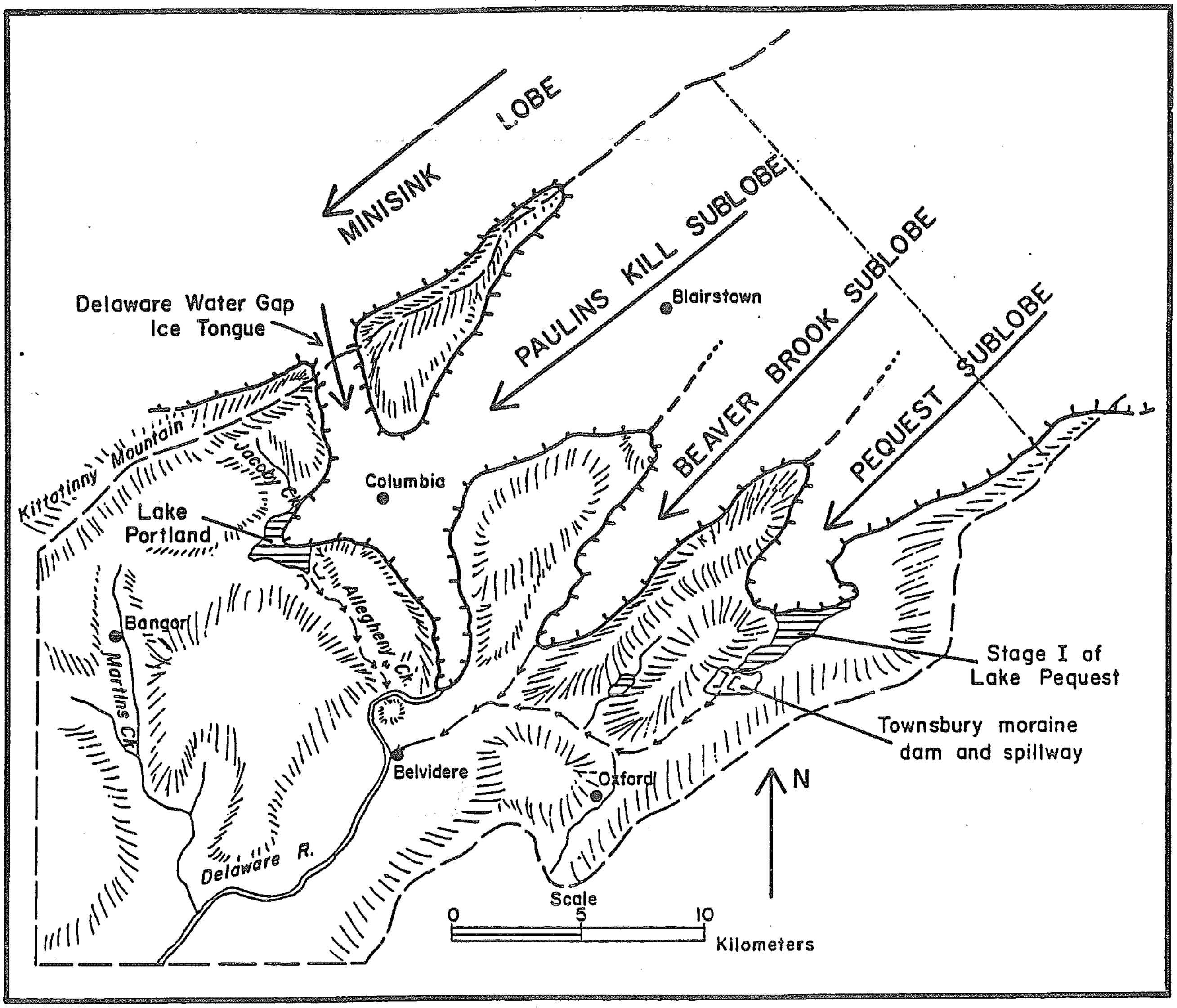


Figure 8. Reconstruction of the sublobe geometry during the formation of Lake Portland and Stage I of Lake Pequest.

(1884) who named them the Portland Kames. Ridge (1983) mapped a kame delta segment of the Portland Kames which indicates the presence of standing water early in the deglacial history of this valley, as Jacoby Creek sequence 1. This lake, named glacial Lake Portland (Ridge, 1983), formed as eastward drainage was dammed by ice in the Jacoby Creek valley (Figure 8). As retreat of Paulins Kill sublobe continued and the deposition of Jacoby sequences 2 through 5 occurred, Lake Portland drained through progressively lower spillways into Martins and Allegheny Creeks. The deposition of the Jacoby Creek sequence 5 represents a stillstand of some duration and the last evidence of the damming of Lake Portland. As the margin of the Paulins Kill sublobe retreated northeast to the town of Portland, meltwater was free to drain south through the Delaware Valley.

Provenance data and striations indicate that ice-marginal flow of the Paulins Kill sublobe in the Portland area was to the southwest (along the valley trend). Striations at Columbia, however, indicate southeast flowing ice. The source of these striations was a small lobe of ice, here called the "Delaware Water Gap ice tongue", which projected southeastward through the Delaware Water Gap (Figure 8). As the Paulins Kill sublobe retreated northeastward, ice in the Great Valley was no longer coalescent with the Minisink Lobe north of Kittatinny Mountain. Deposits of the Delaware Water Gap ice tongue, identified from provenance date (Figure 9), document the position of this extension of the Minisink Lobe and its relationship to the Paulins Kill sublobe. The Delaware Water Gap ice tongue, which was for the most part stagnant, dammed southwestward flowing meltwater of the retreating Paulins Kill sublobe (Figure 10), forming glacial Lake Paulins Kill (Ridge, 1983). Paulins Kill sequences 1 and 2 consist of ice-contact lacustrine sediments deposited in glacial Lake Paulins Kill while deltaic

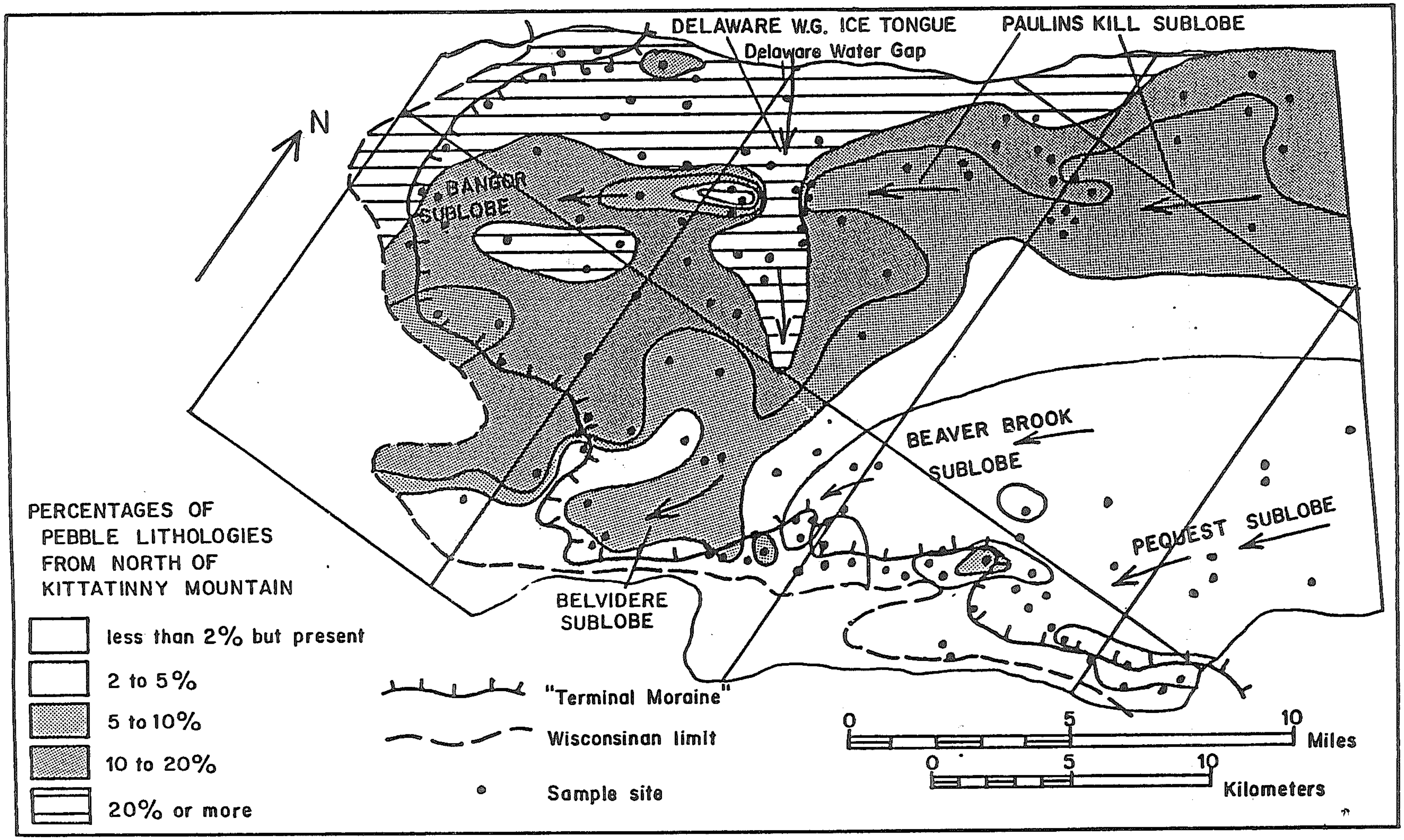


Figure 9. Quantitative distribution of exotic pebble lithologies (derived from north of Kittatinny Mountain) in Great Valley deposits.

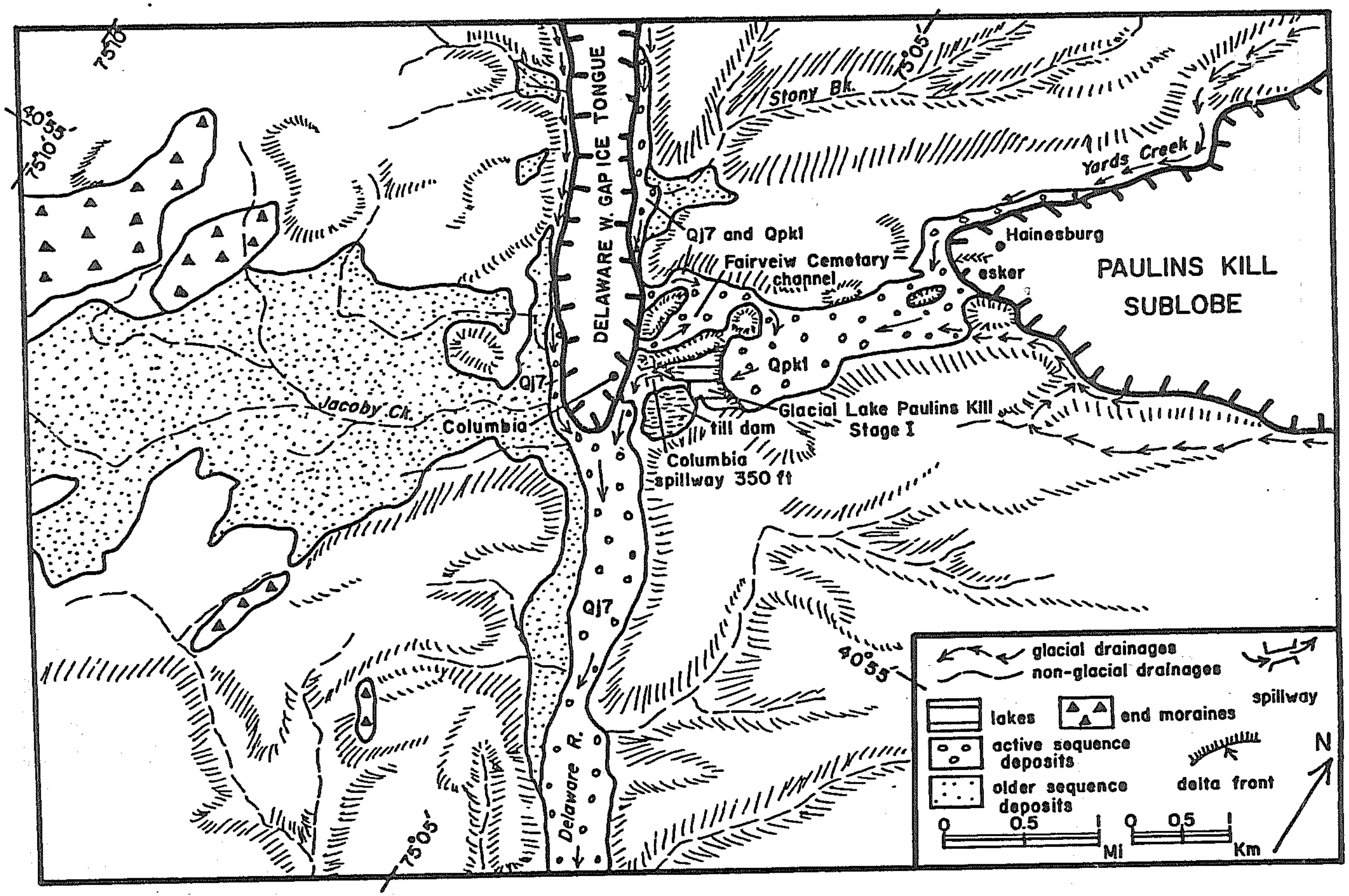


Figure 10. Reconstruction of the Delaware Water Gap ice tongue and the Paulins Kill Sublobe during Stage I of Lake Paulins Kill.

materials of sequences 3 and 4 document the lowering of the lake. During this period lake levels dropped from an initial elevation of 107 m to an elevation of 104 m before draining as the stagnant ice and drift dam at the mouth of Paulins Kill melted and was incised (Figure 10).

The deposition of the Franklin Grove Moraine is correlated with Paulins Kill sequence 5. This moraine, traceable from Spring Valley to Kittatinny Mountain, probably represents a protracted stillstand of the retreating Paulins Kill sublobe. Ice in the Great Valley consisted solely of sublobes of the Hudson-Champlain Lobe during the deposition of the Franklin Grove Moraine because Ontario Lobe ice had thinned to such an extent that it could no longer override Kittatinny Ridge.

With continued recession, the Paulins Kill sublobe retreated over the drainage divide and into the Wallkill Valley where it deposited the Ogdensburg-Culvers Gap Moraine (Figure 4a). Reconnaisance mapping indicates the presence of at least one site between the Franklin Grove and the Ogdensburg-Culvers Gap moraines at Newton, New Jersey where till was deposited. Although detailed morphosequence mapping has not been completed between the Franklin Grove and Ogdensburg-Culvers Gap Moraines, evidence of drainage from the Wallkill Valley into the Paulins Kill valley has been recognized.

DEGLACIATION OF THE BELVIDERE AND BEAVER BROOK SUBLOBES

Retreat of the Belvidere sublobe from the maximum Woodfordian position is marked by the deposition of outwash mapped as Delaware Valley sequence 1 (Qdl, Figure 11). This outwash deposit appears to be correlative with the Bangor and Gruvertown moraine segments of the "Terminal Moraine" to the west, and is the highest level of outwash in the Martins Creek valley. Further retreat was accompanied by deposition of a large kame complex and

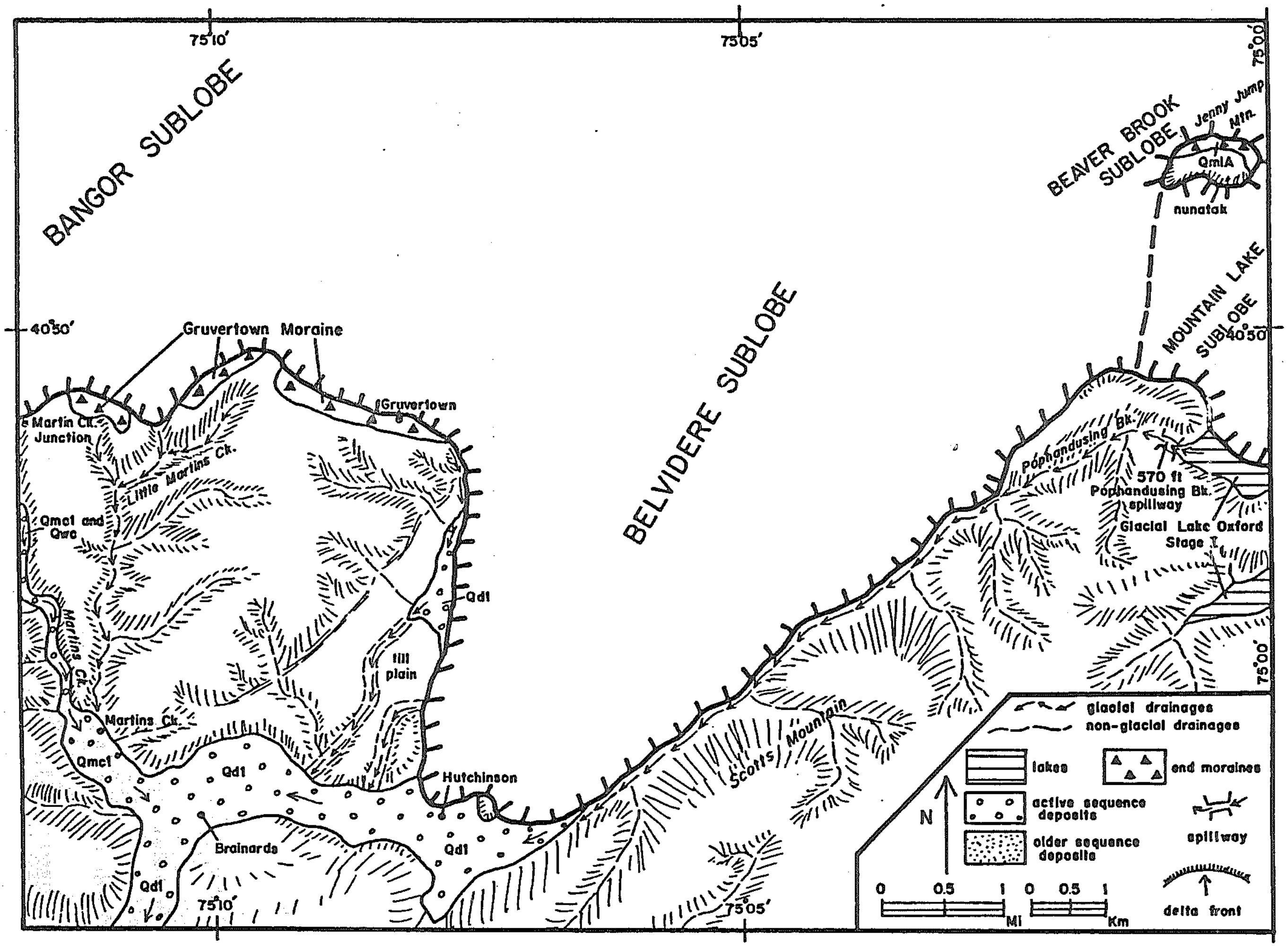


Figure 11. Reconstruction of the Belvidere sublobe during the deposition of the Gruvertown Moraine, portion of the "Terminal Moraine".

outwash plain (Qd2, Figure 12) at Foul Rift and deltaic deposits (Qd2a, Figure 12) in ponded water between the parting Belvidere, Beaver Brook, and Mountain Lake sublobes. The deposition of a series of kame deltas (Qd3, Figure 13; Qd4, Figure 14) delineate the ice margins of the Beaver Brook and Belvidere sublobes as they continued to recede and split. Subsequently, outwash was deposited in the Delaware Valley and it has been tentatively correlated with deposits of the Paulins Kill sublobe. As the Belvidere sublobe receded further, it became indistinguishable from the Paulins Kill sublobe. Ice remaining in the Delaware Valley was principally an extension of the Paulins Kill sublobe, which became completely separated from the Beaver Brook sublobe.

The region north of Belvidere consists of slate uplands and has one principal strike valley, the Beaver Brook valley (Figure 2). The Beaver Brook sublobe (Figure 14) probably retreated rapidly due to relatively thin ice cover over the slate uplands. Large isolated blocks of ice stagnated in the Beaver Brook valley during this period. No ice-marginal deposits younger than Delaware valley sequence 4 (Qd4, Figure 14) can be designated for the Beaver Brook valley.

DEGLACIATION OF THE MOUNTAIN LAKE AND PEQUEST SUBLOBES

The Buttzville Moraines (Figure 12) are interlobate segments of the "Terminal Moraine" deposited between the Belvidere sublobe and the Mountain Lake sublobe. The Mountain Lake sublobe was a short-lived appendage of the Pequest sublobe (Figure 15). As southeastward flowing ice overrode Jenny Jump Mountain and Mount Mohepinoke, a portion of the Mountain Lake sublobe became isolated in the lee of these highlands. Seven recessional positions were recognized in the Mountain Lake region, but after the deposition of

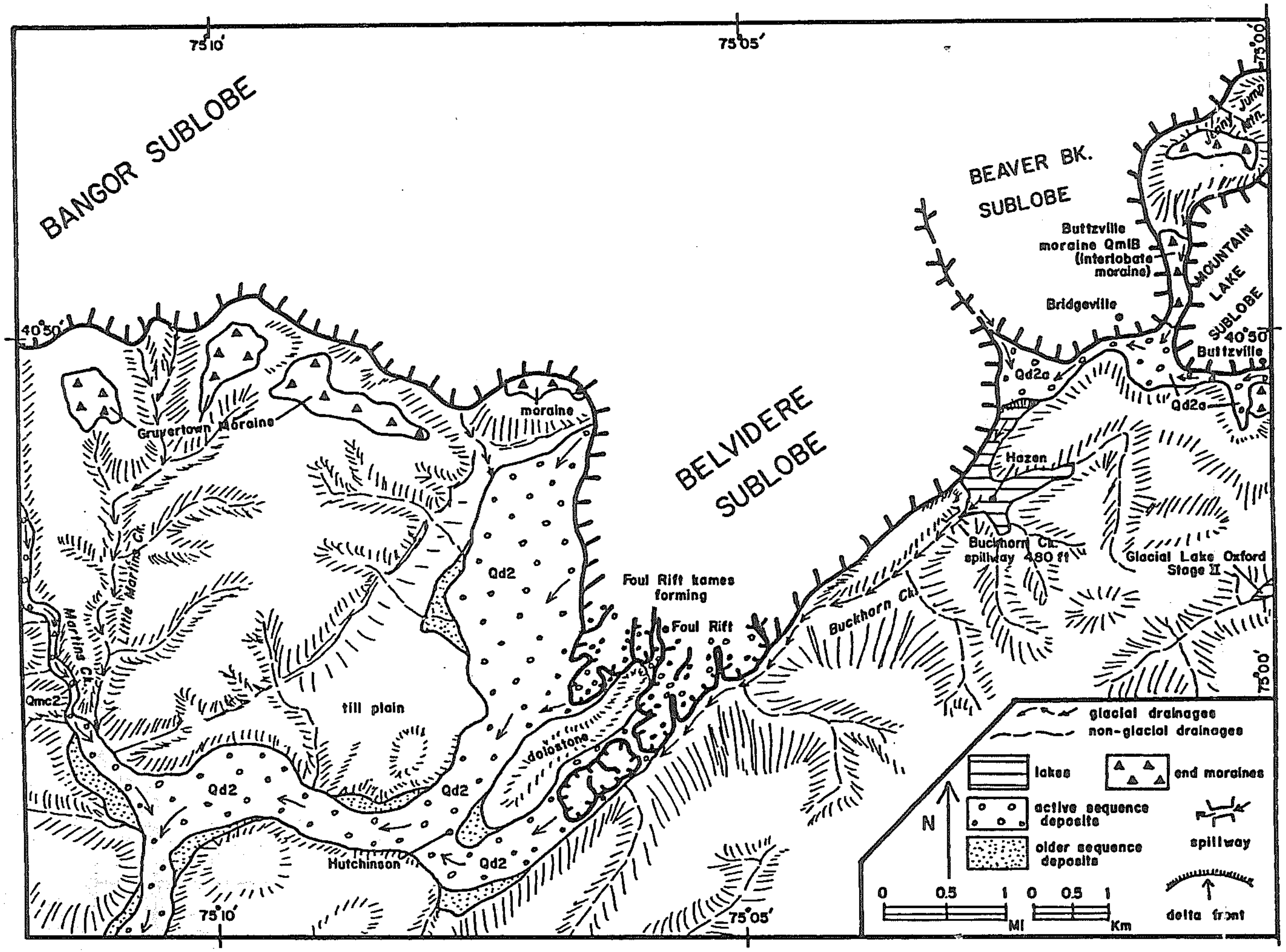


Figure 12. Reconstruction of the Belvidere and Beaver Brook sublobes during the deposition of the Foul Rift kames (early stage, Qd2).

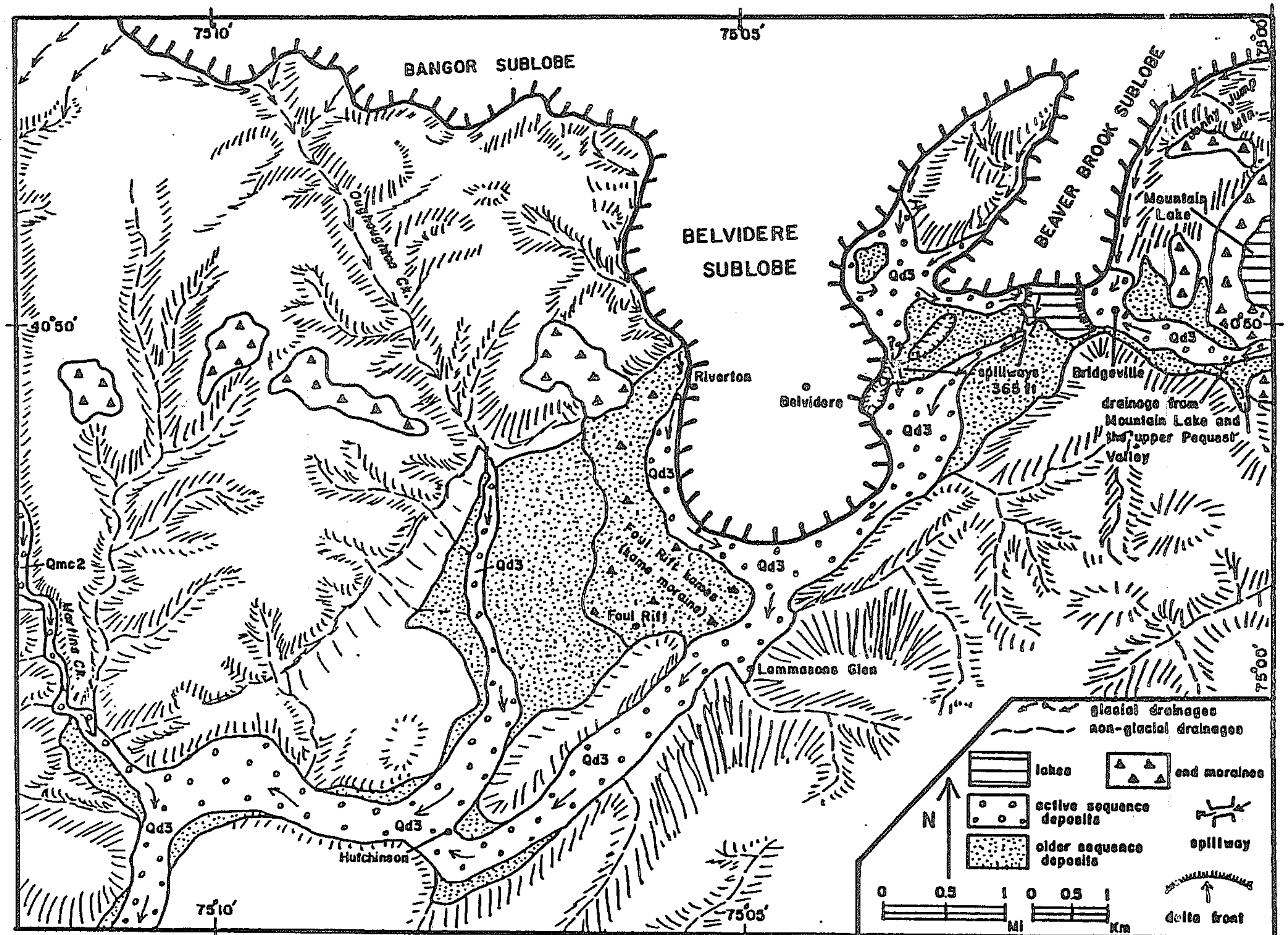


Figure 13. Reconstruction of the Belvidere and Beaver Brook sublobes during the deposition of the Foul Rift kames (late stage, Qd3).

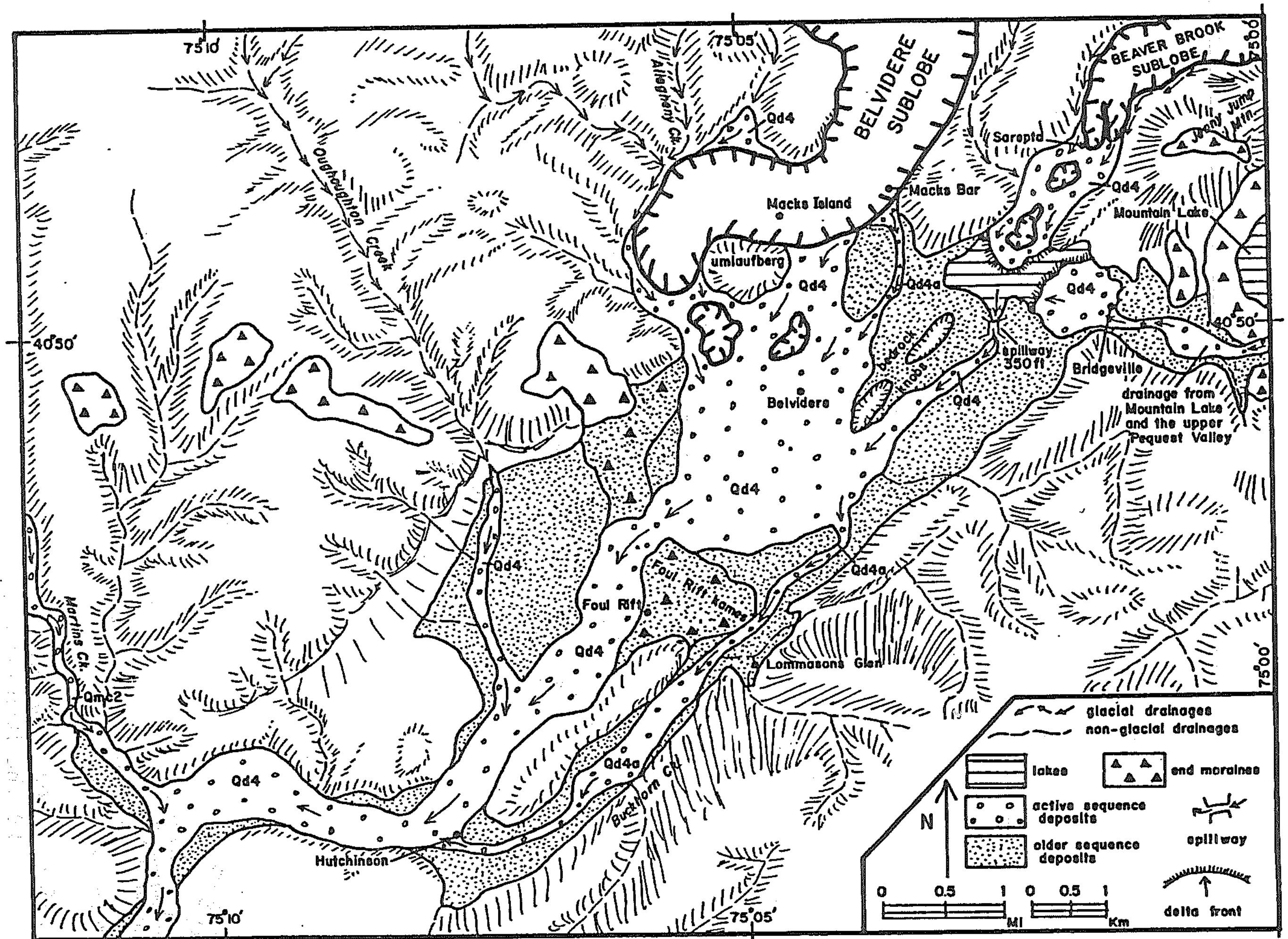


Figure 14. Reconstruction of the final stages of the Belvidere and Beaver Brook sublobes (latest stage, Qd4).

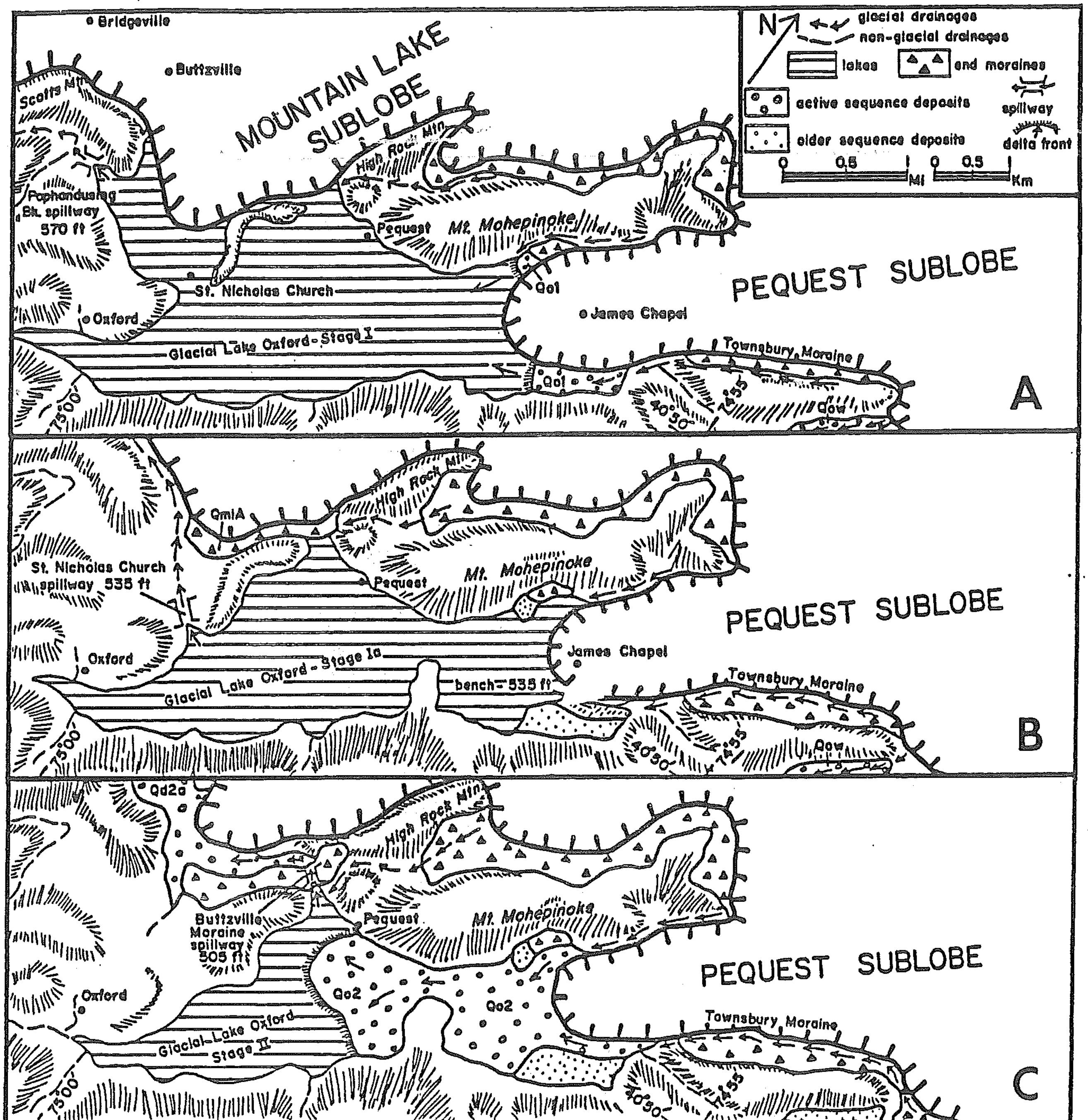


Figure 15. Three stages of development of Lake Oxford which was initially dammed by the Mountain Lake Sublobe and later dammed by end moraines in the Pequest Valley.

the last of these, the thinning ice sheet could no longer over-top the highland barriers.

The "Terminal Moraine" ice margin can be traced eastward from the Buttzville Moraines (Figure 15b) to the Townsbury Moraine, a deposit of the Pequest sublobe (Figure 15). The Townsbury Moraine complex is a well-developed moraine and, like the Bangor Moraine, deposits are preserved from a number of ice-marginal positions. Emplacement of the moraine complex began during the initial formation of glacial Lake Oxford and continued through three lake stages (Figure 15 A,B,C). The Townsbury moraine dammed southwestward flowing meltwater in the Pequest Valley, forming glacial Lake Pequest (originally recognized and named by Salisbury, 1902). Ridge (1983) has documented three stages of lake development as evidenced by several Pequest Valley sequences.

Stage I of glacial Lake Pequest had an elevation of approximately 181 m. Four successive Pequest Valley sequences consisting of ice-contact, lacustrine deposits were deposited during Stage I. The margin of the Pequest sublobe retreated approximately 6 km during this interval, and the Mountain Lake sublobe retreated through four ice marginal positions. A scarcity of ice-marginal deposits in the broad, lacustrine flats of the Pequest Valley may reflect rapid ice retreat. Striations on Jenny Jump Mountain and other upland surfaces indicate that ice flow was initially from the Ontario Lobe to the north. As the ice sheet thinned, the Mountain Lake sublobe and the southern portion of the Pequest sublobe were isolated from tributary flow of the Ontario Lobe and southwest flow developed in the Pequest Valley.

Stage II of Lake Pequest (Figure 16) drained over a spillway at an elevation of 178 m into Glovers Pond. Pequest Valley sequence 5, deposited

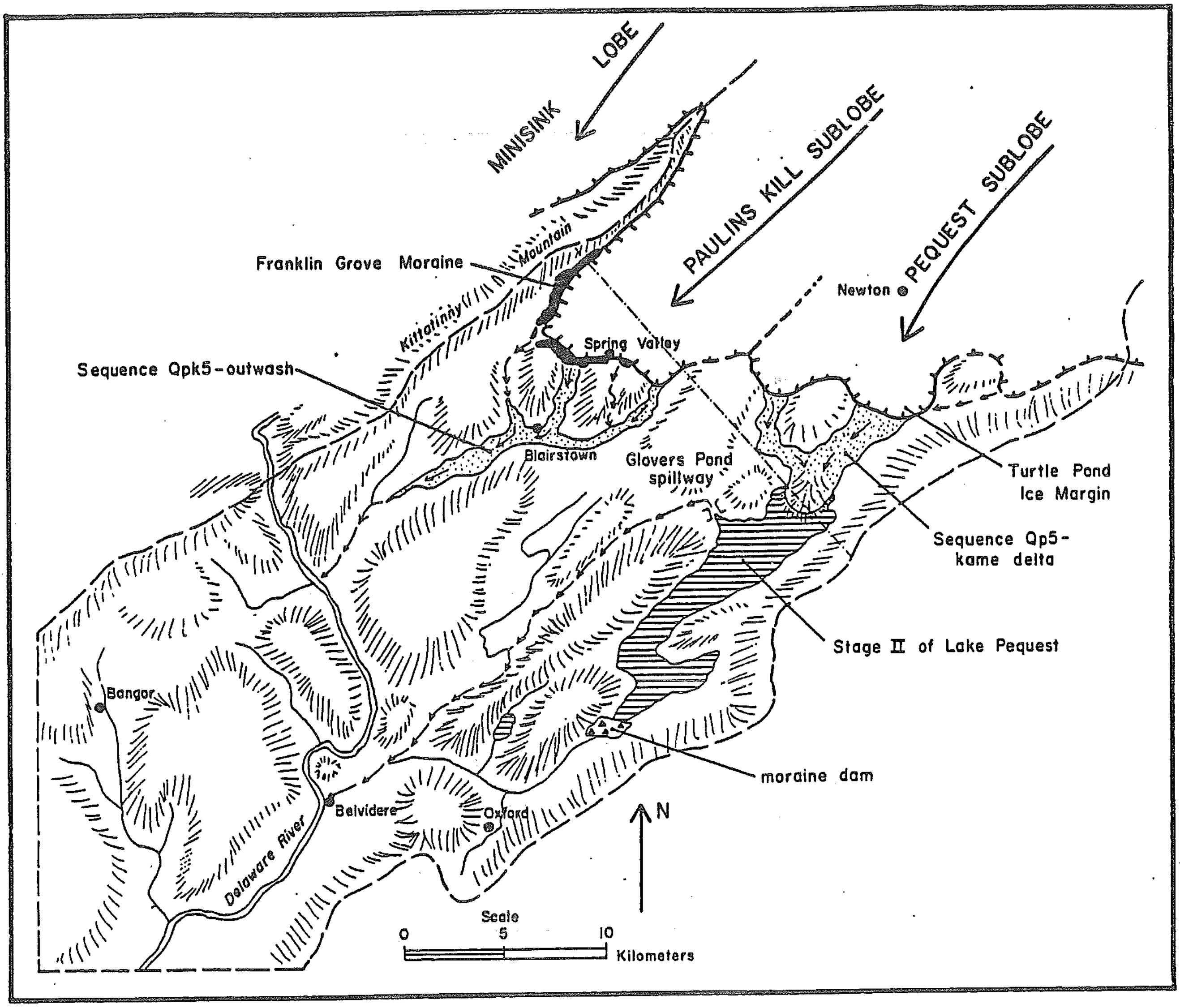


Figure 16. Reconstruction of sublobe geometry during the deposition of the Franklin Grove Moraine and the Turtle Pond Delta (stages Qpk5 and Qp5 and Lake Pequest II).

during this stage, is represented by a kame delta at Turtle Pond in Sussex County. The sediment volume of this sequence indicates that the Turtle Pond ice margin represented a stillstand of considerably longer duration than any of the previous sequences. Deposits of this ice margin position are correlated with the Franklin Grove Moraine of the Paulins Kill sublobe (Figure 16).

Stage III of Lake Pequest formed as the Pequest sublobe continued to retreat in Sussex County. During this final stage of deglaciation a new spillway at elevation 163 m was cut through older deposits at Great Meadows (Pequest Valley sequences 1 and 2). Erosional features indicative of high flow regimes and possible catastrophic drainage south of the Townsbury Moraine are evidence of the lowering of Stage II to Stage III of Lake Pequest. Deltaic deposits mapped as Pequest Valley sequence 6 are the only deposits found in association with stage III of Lake Pequest.

FRANKLIN GROVE MORAINE AND LATER EVENTS

Salisbury (1902) noted the occurrence of morainic deposits near Blairstown, but the full extent of these deposits was mapped by Ridge (1983), who named them the Franklin Grove Moraine. This moraine has a well-developed, arcuate shape and is composed of till (Figure 16). It is located 29 km northeast of the Bangor Moraine in the Paulins Kill valley. Although no ice-marginal deposits have been found on the slate uplands between the Pequest and Paulins Kill Valleys, the Turtle Pond ice margin in the Pequest Valley may be a continuation of the Franklin Grove Moraine (Figure 16). These deposits represent a significant stillstand of the retreating Woodfordian ice sheet. Ice flow indicators (striations and provenance) document southwest flow from the Hudson-Champlain Lobe (by the time the Franklin Grove Moraine was deposited). Therefore, the Franklin

Grove Moraine, and its correlative (the Turtle Pond delta) in the Pequest Valley, represent a Hudson-Champlain Lobe ice margin which lies south of the Ogdensburg-Culvers Gap Moraine which Connally and Sirkin (1973) considered the maximum position of the Hudson-Champlain Lobe in the Great Valley. As ice retreated north from the Franklin Grove Moraine and the Turtle Pond ice margin, drainage of glacial meltwater was impounded in the upper reaches of the Paulins Kill and later the Wallkill Valley (Pequest Lobe).

IMPLICATIONS FOR REGIONAL ICE FLOW

Analysis of striations by Lewis (1884), Salisbury (1902), Miller and others (1939), Ward (1939), Epstein (1969), and Ridge (1983) document southward ice flow during the maximum extent of the Woodfordian Glaciation across Kittatinny Mountain into the Great Valley (Figure 17a). As the ice sheet thinned ice flow near its terminus in New Jersey became increasingly aligned with the southwest-trending regional topography. This shift to a southwestward flow pattern is the result of a gradual decrease in the contribution of ice from the Ontario Lobe. Therefore, the Great Valley area represents a suture position during the Woodfordian maximum where the Ontario and Hudson-Champlain Lobes were in contact (Figure 17a).

Provenance investigations by Ridge (1983) document the predominance of Hudson-Champlain lobe ice in the Great Valley by the time of the deposition of the Franklin Grove Moraine (Figure 17b). The southwest flow of ice, controlled by, and parallel to, the trend of the Great Valley accounts for the distribution of Beemerville Complex erratics in the Great Valley (Ridge, 1983), and westward migration of the suture to a position in eastern Pennsylvania (Figure 17b).

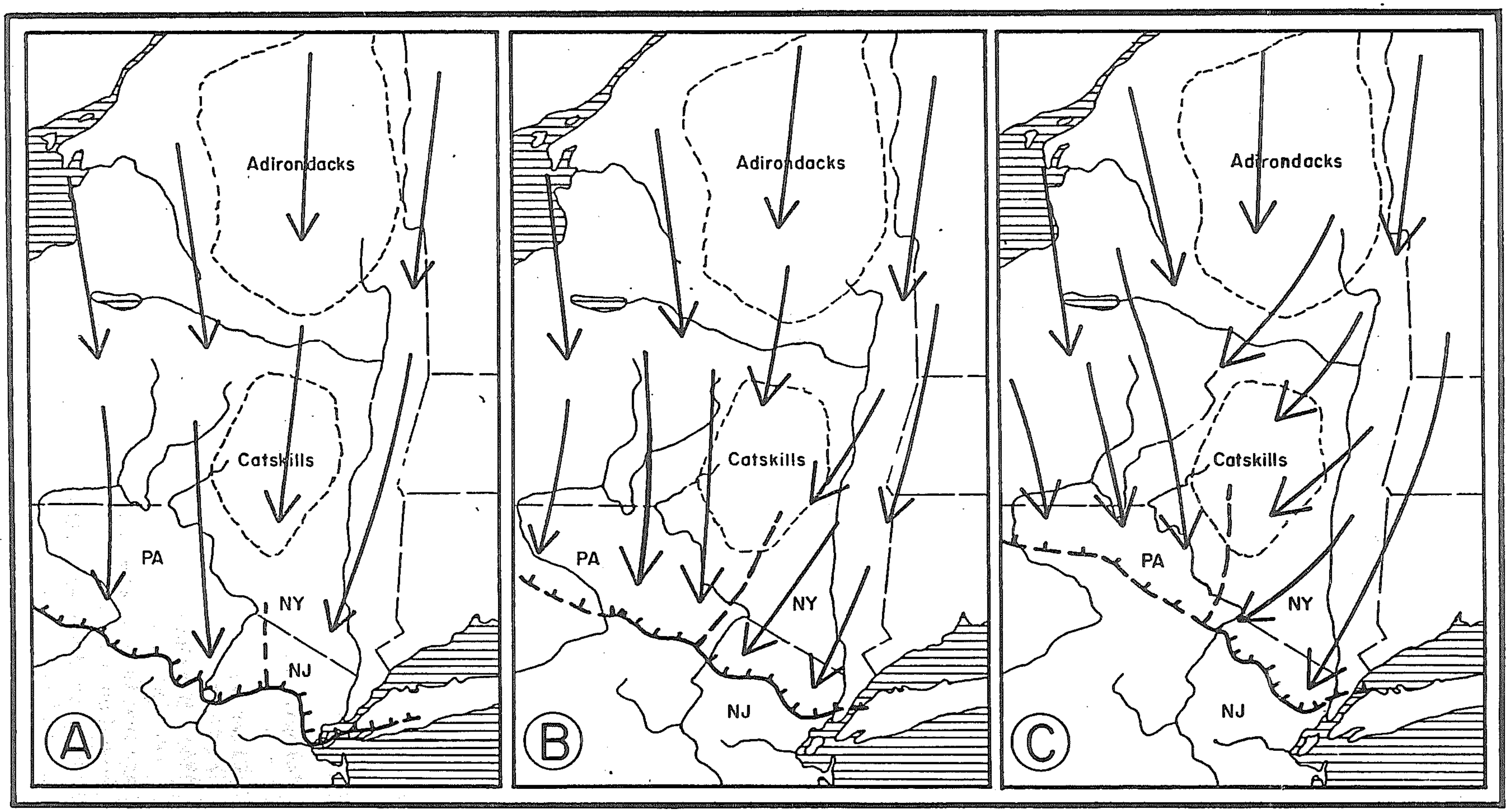


Figure 17. Inferred deployment of the Ontario and Hudson-Champlain Lobes during Woodfordian deglaciation: (A) the Woodfordian terminal position, (B) the Franklin Grove Moraine, and (C) the Ogdensburg-Culvers Gap Moraine.

We suggest here, that by the time of the deposition of the Ogdensburg-Culvers Gap Moraine, the suture had migrated northwestward, into the southern Catskills (Figure 17c). Thinning ice over the Catskills together with the migration of the suture eventually resulted in the first physical separation of the Hudson-Champlain and Ontario Lobes.

Lobate flow in the Hudson-Champlain Lobe during the deposition of the Ogdensburg-Culvers Gap Moraine explains the provenance change observed by Salisbury (1902) and Connally and Sirkin (1973) at the Ogdensburg-Culvers Gap Moraine. At this time the Hudson-Champlain Lobe would have spread across the Great Valley with a slightly more westerly trend than the local topography. This model disagrees with those presented by other workers (Coates and Kirkland, 1974).

AGE OF THE GREAT VALLEY DRIFT AND THE "TERMINAL MORAINE"

We consider the Great Valley Drift and the "Terminal Moraine" in the study area to be of Woodfordian age. This age assignment is based on: 1) pedologic development similar to that of other Woodfordian deposits in the region, 2) correlation with other deposits of Woodfordian age, 3) radiocarbon dates, and 4) palynologic studies.

Nine soil profiles in both stratified and non-stratified glacial deposits of the Great Valley drift (Ridge, 1983) were studied in detail. Although variations due to differences in parent material occur, all pedologic data indicate that the soils are of Woodfordian age. All soil profiles show weathering to depths of 125 cm or less. According to Sevon (1974), Marchand (1978), and Levine and Ciolkosz (1983), soils with these characteristics are of Woodfordian age. In the valley immediately north of Kittatinny Mountain (the Minisink Lobe of Connally, 1973) the "Terminal

Moraine" has been assigned a Woodfordian age (Epstein and Epstein, 1969; Crowl, 1971, 1972, 1975, 1980; Connally and Epstein, 1973; Berg, 1975; Crowl and others, 1975; Crowl and Sevon, 1980) as have deposits north of the "Terminal Moraine" (Connally and Epstein, 1973; Sevon, 1974; Sevon and others, 1975; Berg, 1975; Crowl and others, 1975 and Sirkin, 1977). In the Great Valley the Ogdensburg-Culvers Gap Moraine and deposits north of this moraine have also been mapped as Woodfordian by numerous workers (Connally and Sirkin, 1967, 1970, 1973; Minard and Rhodehamel, 1969; Sirkin and Minard, 1972). A comparison of the Great Valley drift to these deposits suggests that the "Terminal Moraine" in the Great Valley is also of Woodfordian age.

The reconstruction and documentation of the Delaware Water Gap ice tongue of the Minisink Lobe provides strong evidence for a Woodfordian age of the Great Valley drift. Epstein (1969) and Connally and Epstein (1973) mapped seven sequences of Woodfordian age related to the Minisink Lobe in the valley north of Kittatinny Mountain. The interaction of the Minisink Lobe (through its extension - the Delaware Water Gap ice tongue) with the Paulins Kill sublobe (Figures 8 and 10) has been clearly demonstrated in the previous discussion. Thus, if the Woodfordian age assigned to deposits of the Minisink Lobe (Connally and Epstein, 1973) is correct, then deposits of the Bangor sublobe are also of Woodfordian age.

Documentation of the detailed deglaciation history of the Great Valley, between the Franklin Grove Moraine and the Ogdensburg-Culvers Gap Moraine, is still incomplete (Ridge, 1983; Witte, in progress). However, the drift of this area was deposited by sublobes of the Hudson-Champlain Lobe and has Woodfordian weathering characteristics. Preliminary data indicate systematic retreat from the Pequest to the Wallkill Valley during which a number of outwash sequences were deposited (Witte, in progress).

ABSOLUTE AGE

As previously stated, basal sediments of lakes and peat bogs in the study area were analyzed for pollen content and radiocarbon dated in an attempt to determine an accurate minimum age of deglacation. Replicate basal dates from Francis Lake, located 3 km east of the village of Johnsonburg, Warren County, New Jersey (Figure 18) provide a minimum age of approximately 18,500 yr B.P.

Francis Lake was formed in a depression between two dolostone bedrock knobs and a till plain. The glacial deposits which surround and underly Francis Lake were deposited just prior to Pequest Valley sequence 5, the sequence which has been correlated with the Franklin Grove Moraine (Ridge, 1983). The palynology of Francis Lake and the vegetation history of the region is fully treated by Cotter (1983) and Cotter and others (1982 and in review). A brief discussion is presented here.

The pollen record of Francis Lake is similar to those of other sites in the region (Sirkin and Minard, 1972; Sirkin, 1977; Watts, 1979). Basal sediments (9.3 to 7.6 m) contain herb pollen zone spectra characterized by maximum percentages of nonarboreal pollen (NAP), and high percentages of jack pine, pine and spruce (Figure 19). During the deposition of the herb zone the climate was colder than the present. The relatively high percentages of wetland species, such as sedge and Thalictrum, indicate poor drainage typical of recently deglaciated regions (Watts, 1979). These wetland species also show that lake sediments were derived from the local area rather than subglacially.

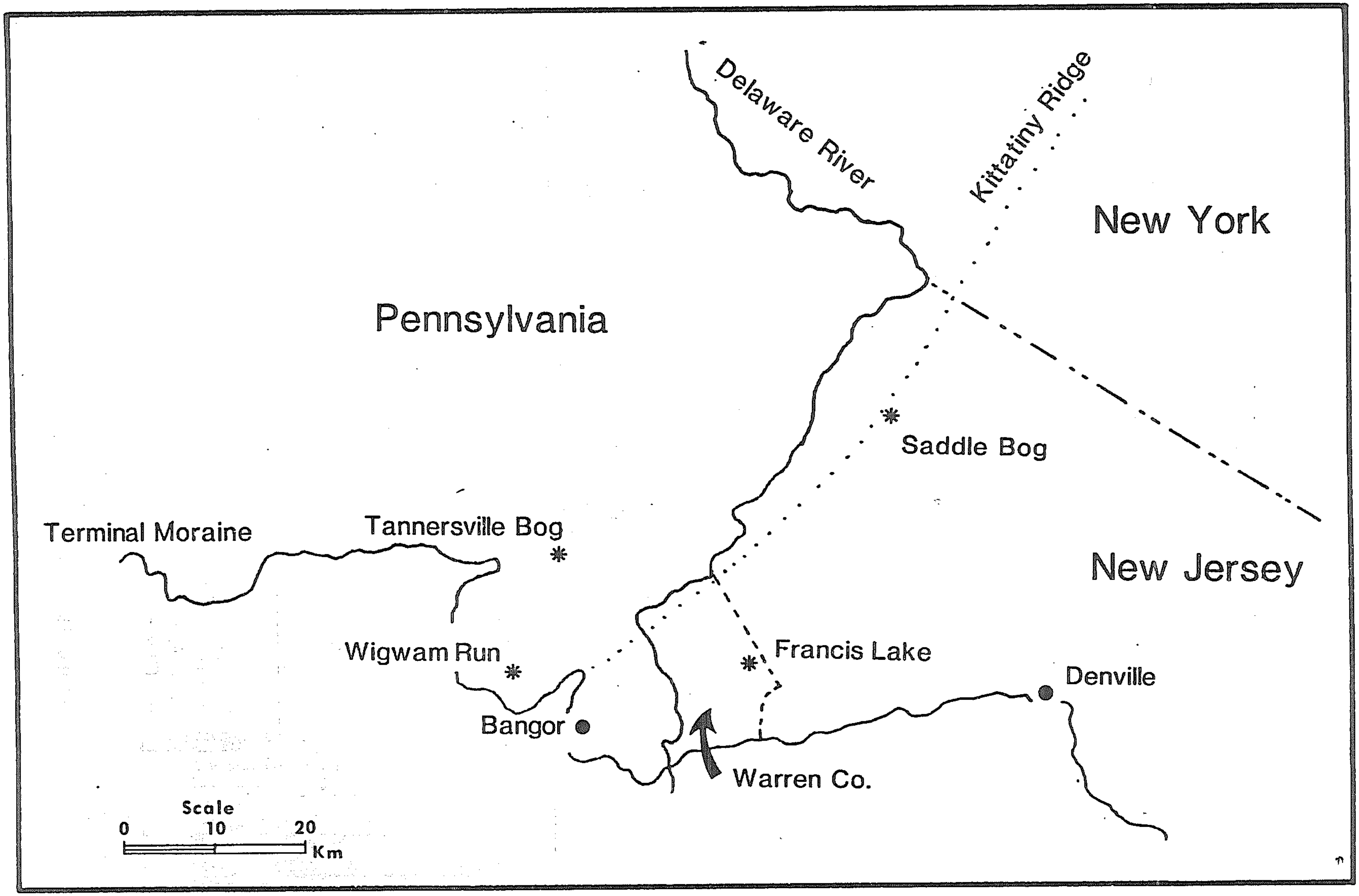


Figure 18. Pollen study sites discussed in text.

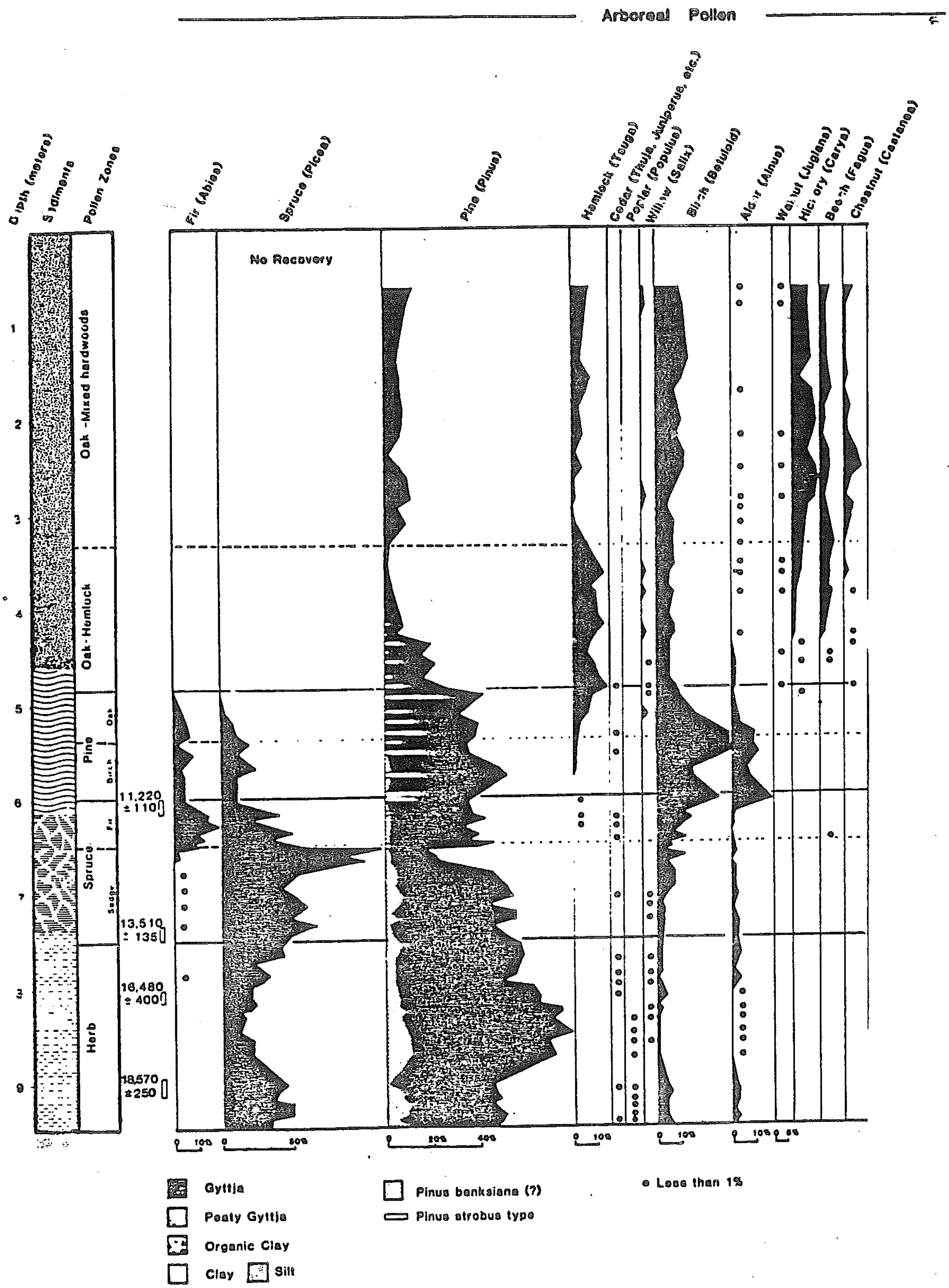
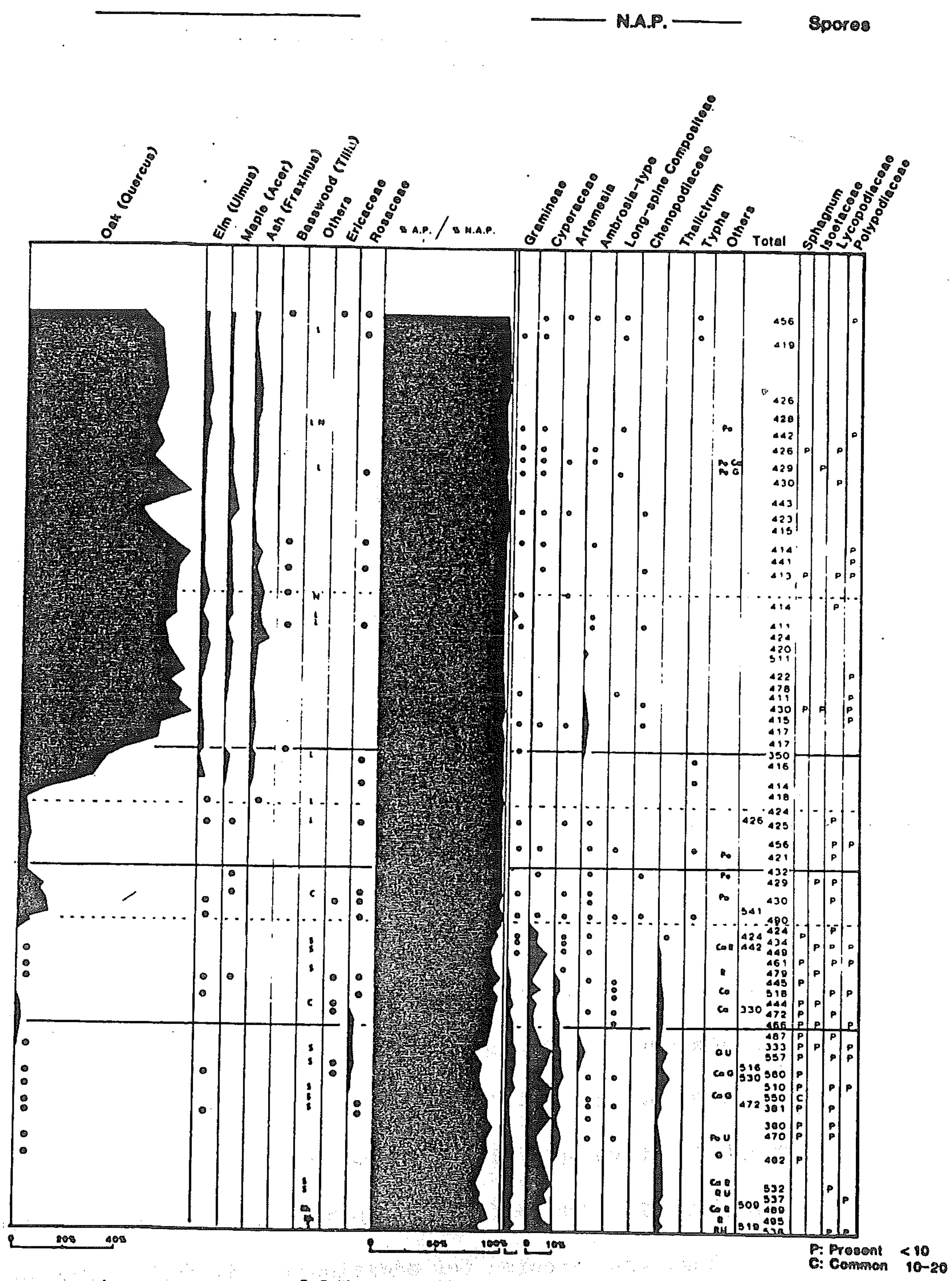


Figure 19. Francis Lake Pollen Diagram.



P: Present <10
C: Common 10-20

C: Celtis
N: Nyssa
L: Liquidambar
Rh: Rhamnaceae
S: Sanguisorba

R: Rubiaceae
G: Galium
U: Umbelliferae
Ca: Caryophyllaceae
Po: Polygonaceae

The spruce zone (7.6 to 6.0 m) is characterized by maximum percentages of spruce and fir pollen and decreasing percentages of NAP. The pollen spectra of this zone represent the establishment of forest vegetation, the result of climatic amelioration. During deposition of the spruce zone open boreal forest, indicative of cold climatic conditions, was replaced first by closed spruce-fir forest and eventually, by a pine forest marking the establishment of a slightly cooler climate than that of today. These changes occurred within about 3,000 years and reflect the waning influence of the Laurentide ice sheet.

The pine pollen zone (6.0 m - 4.8 m) is similar to pine pollen zone recognized elsewhere. Increased percentages of white pine pollen follow the rapid decline of spruce percentages. The region, during this period, was occupied by mixed pine and birch or oak forest. Climatic conditions were similar to that of today, although variations in effective moisture (drier conditions) probably did occur.

Through the oak zone, percentages of oak pollen remain high. The earliest portion of this zone, the oak-hemlock subzone, is characterized by decreasing percentages of pine pollen and relatively high percentages of hemlock pollen. The base of the oak-mixed hardwood subzone is marked by a decrease in hemlock pollen and an increase in pollen percentages of most hardwoods. A climate similar to that of today probably existed throughout the period.

Five radiocarbon dates were obtained from the sequence of samples from Francis Lake. After obtaining a basal date of 18,390 ± 200 (SI-4,921) the lake was re-cored for additional radiocarbon samples. The dates obtained are: 18,570 ± 250 (SI-5,273), 16,480 ± 430 (SI-5,274), 13,510 ± 135 (SI-5,300, and 11,220 ± 110 yr B.P. (SI-5,301) (Figure 19). The pollen and radiocarbon data demonstrate that the region was deglaciated by 18,500 yr B.P. and that tundra vegetation, as documented by the herb pollen zone, existed from 18,500 to about 14,275 yr B.P. The basal Francis Lake dates are the oldest minimum dates from the area obtained in conjunction with pollen stratigraphy. The dates document a minimum deglaciation date of 18,500 yr B.P. and provide a record of the timing and duration of inorganic and low-organic sedimentation following deglaciation.

To verify the validity of the Francis Lake dates, we have correlated pollen records of sites with similar settings located within a small area (Figure 18). The classic pollen zones described by Deevey (1949), Davis (1965), and Sirkin (1977), are used as stratigraphic horizons and as indicators of prior vegetation.

The sediment, pollen and radiocarbon data of the four sites used are summarized in Figure 20 and a composite of the local pollen stratigraphy is presented in Figure 21. Saddle Bog (Sirkin and Minard, 1972), Wigwam Run (Sirkin, 1977), and Francis Lake are in the Valley and Ridge province; Tannersville Bog (Watts, 1979) is located at the eastern edge of the glaciated low plateau section of the Appalachian Plateau. All four sites are situated between elevations of 140 and 280 m and Francis Lake is within 38 km of the other three sites. Because of the proximity of these sites, vegetational changes interpreted in the pollen records of all four locations are believed to be synchronous. The stratigraphy includes the four principal pollen zones; herb, spruce, pine, and oak, and a total of 12 radiocarbon dates (Figure 21).

Radiocarbon dates from Francis Lake of 18,570, 18,390, and 16,480 yr B.P. provide an absolute chronology for sedimentation from deglaciation to the establishment of a boreal forest (spruce pollen zone). This interval

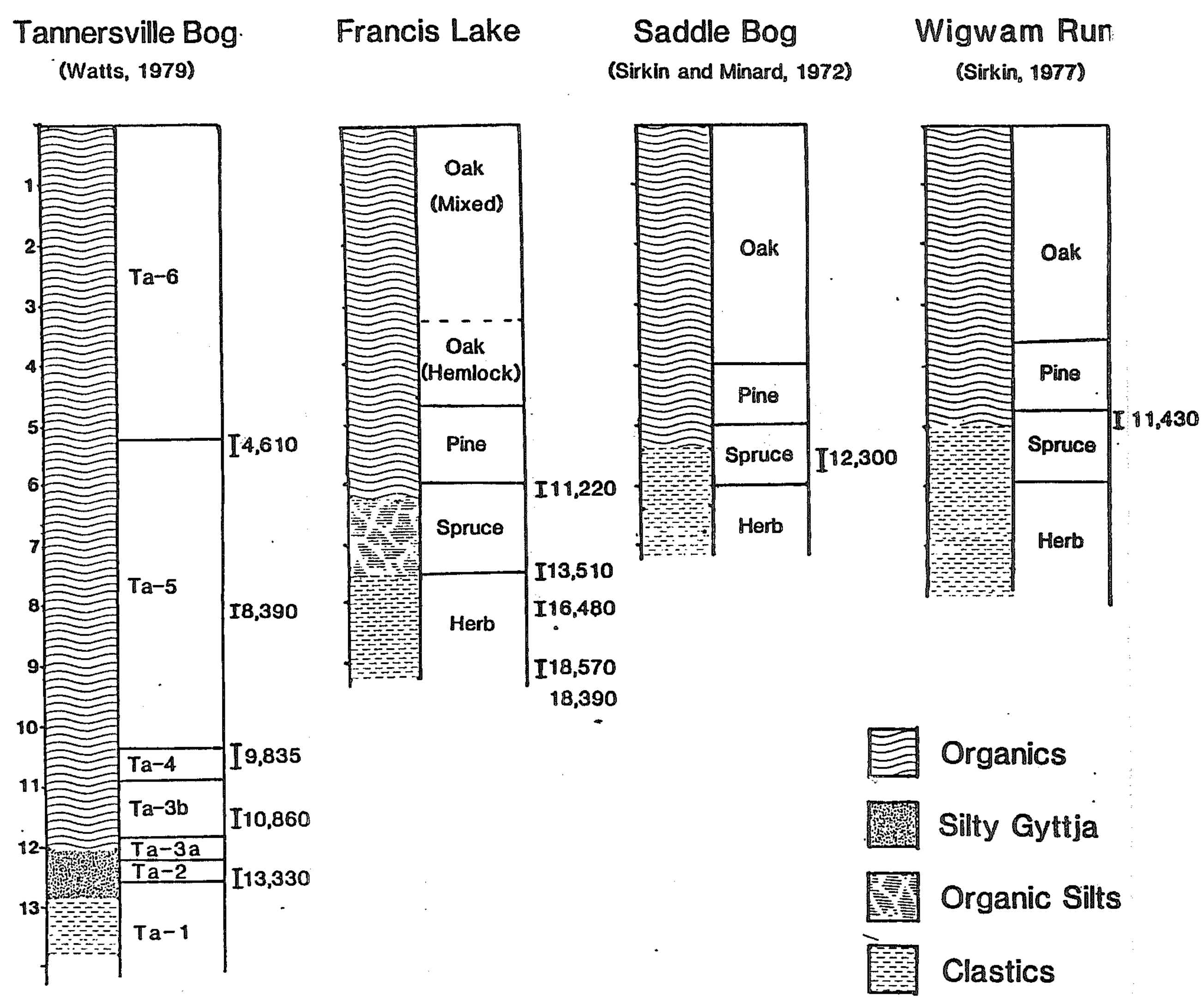


Figure 20. Generalized pollen and sediment stratigraphy and radiocarbon dates of four sites discussed in text (depth is in meters).

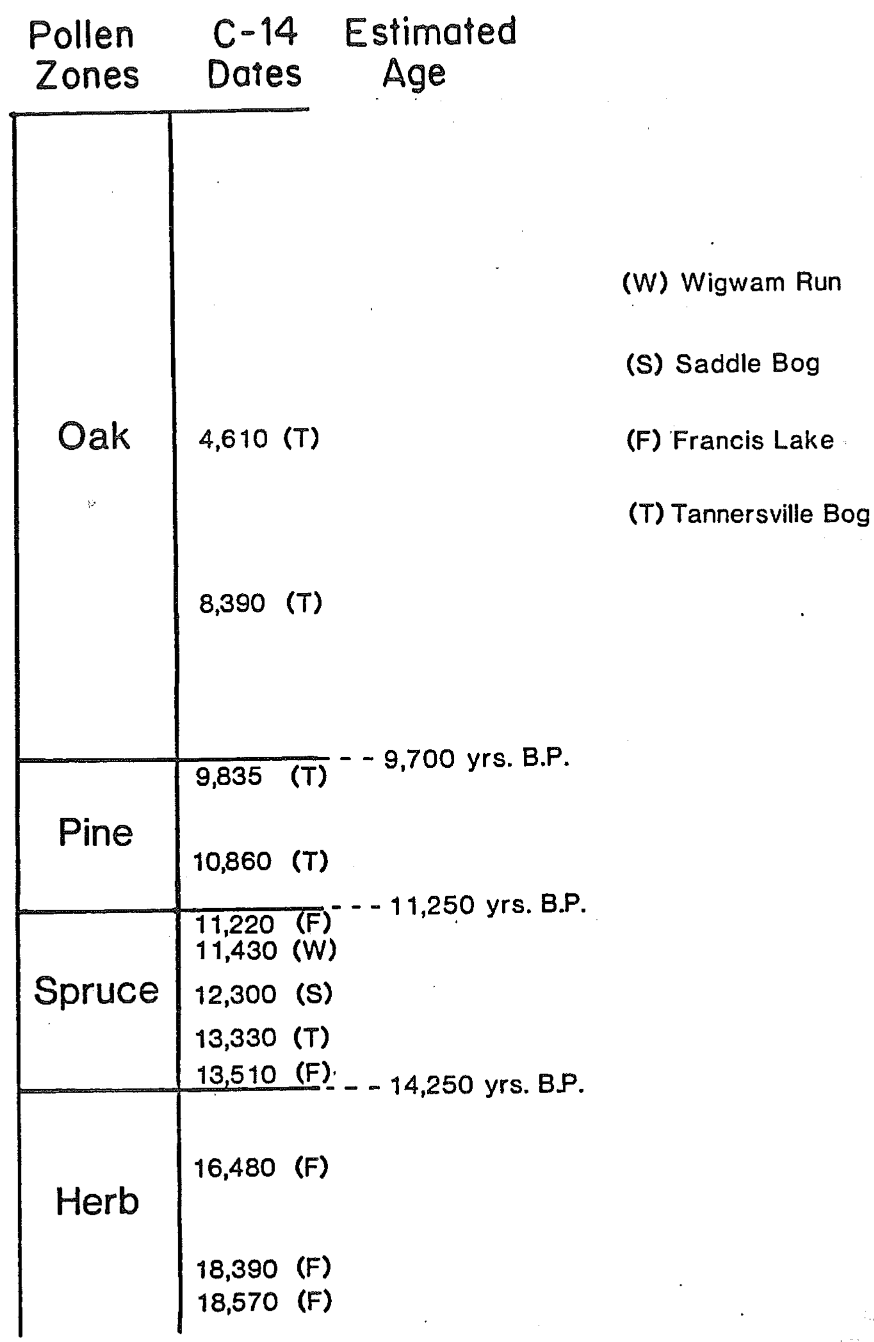


Figure 21. Diagram of composite pollen and radiocarbon stratigraphy of the study area and estimated ages of pollen zone boundary (depth is not to scale).

of tundra vegetation (herb pollen zone) lasted from prior to 18,500 yr B.P. to approximately 14,250 yr B.P. (Figures 19, 20 and 21). No radiocarbon dates were obtained for the herb pollen zone at the other four sites. At Crider's Pond in south-central Pennsylvania a date of 15,210 yr B.P. was obtained at the herb pollen zone/spruce pollen zone boundary (Cr1/Cr2 boundary), (Watts, 1979).

The vegetation during the deposition of the herb zone is very similar at all four sites. High percentages of NAP and spruce and pine indicate cold climatic conditions. The high ratio of sedge to grass pollen within the NAP percentages indicate that most of this portion of the Great Valley was poorly drained and recently deglaciated.

The spruce pollen zone at Francis Lake began approximately 14,250 yr B.P. Major floristic events during this period include a spruce peak followed by the expansion of fir into the region. A date of 13,510 yr B.P. precedes the spruce peak at Francis Lake and a date of 13,330 yr B.P. predates the fir expansion at Tannersville bog. A date of 12,300 yr B.P. was obtained from the Saddle Bog immediately above the spruce peak, and dates of 11,430 and 11,220 yr B.P. were obtained from the spruce zone/pine zone boundaries at Wigwam Run and Francis Lake, respectively. Dates of 10,860 and 9,835 yr B.P. at Tannersville provide age control of the establishment of white pine, and the peak of white pine percentages (Figure 20).

The local pollen stratigraphy described here indicates that dates of 13,510 and 11,220 yr B.P. from Francis Lake are accurate (Figure 19). Continuous sedimentation, a readily correlated pollen stratigraphy, and sequential horizon dating all indicate that the older dates (16,480, 18,390 and 18,570 yr B.P.) are both valid and accurate.

CONCLUSIONS

At the maximum extent of the Woodfordian glaciation the Ontario and Hudson-Champlain Lobes of the Laurentide ice sheet were tributary sources of ice in the Great Valley. Five sublobes: the Bangor, Belvidere, Beaver Brook, Mountain Lake, and the Pequest sublobes deposited segments of the "Terminal Moraine" just north of the Woodfordian limit. As Woodfordian deglaciation began, thinning of the ice sheet resulted in the depletion of tributary flow from the Ontario Lobe. With continued ice-margin retreat, increased influence of the Hudson-Champlain Lobe resulted in different sublobe deployment and geometry. By the time of the deposition of the Franklin Grove Moraine, tributary flow to the sublobes in the area: the Paulins Kill and the Pequest sublobes, was solely from the Hudson-Champlain Lobe.

Ice retreat in the Great Valley was characterized by both systematic ice retreat and rapid stagnation by divide cut-off. Ice retreat in the Delaware, Jacoby and Paulins Kill valleys, and portions of the Pequest Valley was by systematic ice retreat. Ice-marginal stagnation zones near both heads of outwash and deltas were present in these areas. The morphosequence concept was utilized to define retreatal positions in these areas. Lacustrine ice-contact sequences occur in the Jacoby Creek Valley (glacial Lake Portland), the Paulins Kill valley (glacial Lake Paulins Kill), the Beaver Brook and lower Pequest Valleys (glacial Lake Oxford) and the upper Pequest Valley (glacial Lake Pequest). Fluvial ice-contact sequences are present in the Delaware and upper Paulins Kill valleys. One major recessional position is marked by the Franklin Grove Moraine and Turtle Pond ice margin.

Evidence for rapid stagnation by divide cut-off exists in the Beaver Brook valley. This valley has no discernable retreatal positions or ice-contact stratified deposits. Rapid stagnation in this valley occurred when the continental ice sheet thinned and could no longer feed ice over a slate upland to the Beaver Brook sublobe.

A Woodfordian age is assigned to deposits of the Great Valley drift on the basis of pedologic analysis and comparison to Woodfordian deposits in other areas. The most significant correlation is based on the interaction of the Bangor sublobe and Minisink Lobe. At the Woodfordian maximum and during earliest phases of deglaciation southward flow of Minisink ice through the Delaware Water Gap was coalescent with the Paulins Kill sublobe. Following the retreat of the Paulins Kill sublobe into the Paulins Kill valley, a Delaware Gap "ice tongue" extended southward, damming meltwater from the Paulins Kill sublobe.

The Woodfordian deglaciation began sometime prior to 18,500 yr B.P. in the Great Valley. Radiocarbon dates of 18,570, 18,390, 16,480, 13,510, and 11,220 yr B.P., in conjunction with pollen stratigraphy, document the presence of tundra vegetation in the Great Valley until 14,250 yr B.P. The validity of the dates from Francis Lake are verified through correlation with other radiocarbon-dated pollen stratigraphies of the region.

The data presented here resolve two critical problems involving the glacial stratigraphy of northwestern New Jersey and northeastern Pennsylvania. First, the "Terminal Moraine" of the Great Valley is of Woodfordian age and the Ogdensburg-Culvers Gap Moraine is a recessional moraine of the Hudson-Champlain Lobe of the Laurentide Ice Sheet. Second, the deglaciation of both the Ontario and Hudson-Champlain Lobes in the Great Valley began prior to 18,500 yr B.P.

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LATE WISCONSINAN DEGLACIATION

FROM THE

FRANKLIN GROVE-TURTLE POND MORAINE

TO THE OGDENSBURG-CULVERS GAP MORAINE

by

Ron Witte, Edward Evenson and Carl Koteff

INTRODUCTION

Detailed mapping of morphosequences from the Franklin Grove-Turtle Pond Ice Margin (Ridge, 1983) northward to the Ogdensburg-Culvers Gap Moraine indicate deglaciation by systematic ice retreat (Koteff and Pessl, 1981). Five successively younger ice-marginal positions including the Ogdensburg-Culvers Gap Moraine have been identified north of the Franklin Grove-Turtle Pond ice margin. Lacustrine ice-contact morphosequences, described by Koteff and Pessl (1981), are the dominant depositional landforms present in the study area. These morphosequences represent deltas and sub-lacustrine fans built into proglacial lakes which formed as the ice-margin retreated northward across drainage divides.

Three separate drainage systems (Wallkill, Paulinskill and Pequest) have their headwaters in the study area (Fig. 1). Initially, the Wallkill and the Paulinskill drain northward, however, the Paulinskill turns abruptly at Augusta, New Jersey and flows southwest toward the Delaware River, whereas the Wallkill continues to flow northward toward the Hudson River in New York. The Pequest has its

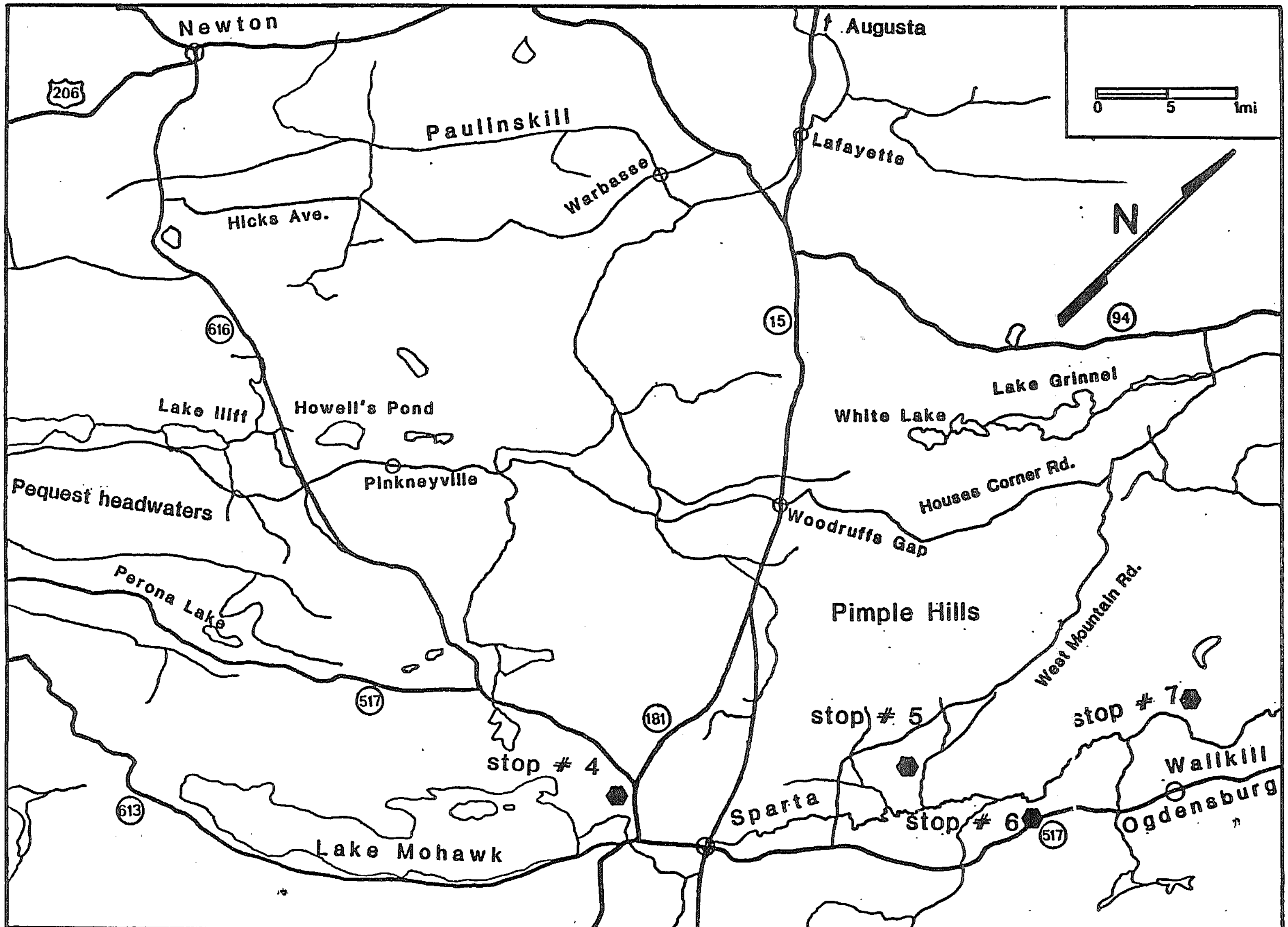

Figure 1. Field trip area with stop locations and drainage systems.

headwaters near Howell's Pond and Perona Lake and drains southwest toward the Delaware River.

Field Trip Stops 4-7 (locations shown in Fig. 1) are in the Wallkill Valley, with Stop 4 at Sparta, and Stops 5-7 northeast toward Ogdensburg. Depositional landforms in this valley represent ice-contact lacustrine morphosequences built into several stages of a proglacial lake, informally named "glacial Lake Sparta". Figure 2A shows longitudinal profiles of the four morphosequences built into various stages of glacial Lake Sparta and reflects the lowering of lake level as lower outlets were uncovered during the systematic retreat of the Wallkill Valley ice lobe.

DEGLACIATION OF THE FIELD TRIP AREA

Figure 2B illustrates the chronology of events during deglaciation. Figures 3-6 represent diagramatic reconstructions of four ice-marginal positions. These are based on the detailed mapping of morphosequences in the Wallkill Valley, Germany Flats and Newton Meadows. The unnamed ice margin which lies north of the Franklin Grove-Turtle Pond ice margin is not included in the deglaciation diagrams presented in this paper.

The numbering and identification of morphosequences in the Germany Flats and Newton Meadows is continued from Ridge (1983) with morphosequences in the Wallkill Valley starting with Qwkl.

Sparta-Lake Iliff Ice Margin (Fig. 3)

Two ice-contact lacustrine morphosequences have been identified at Sparta and Sussex Mills. The morphosequence at Sparta (Qwkl) is

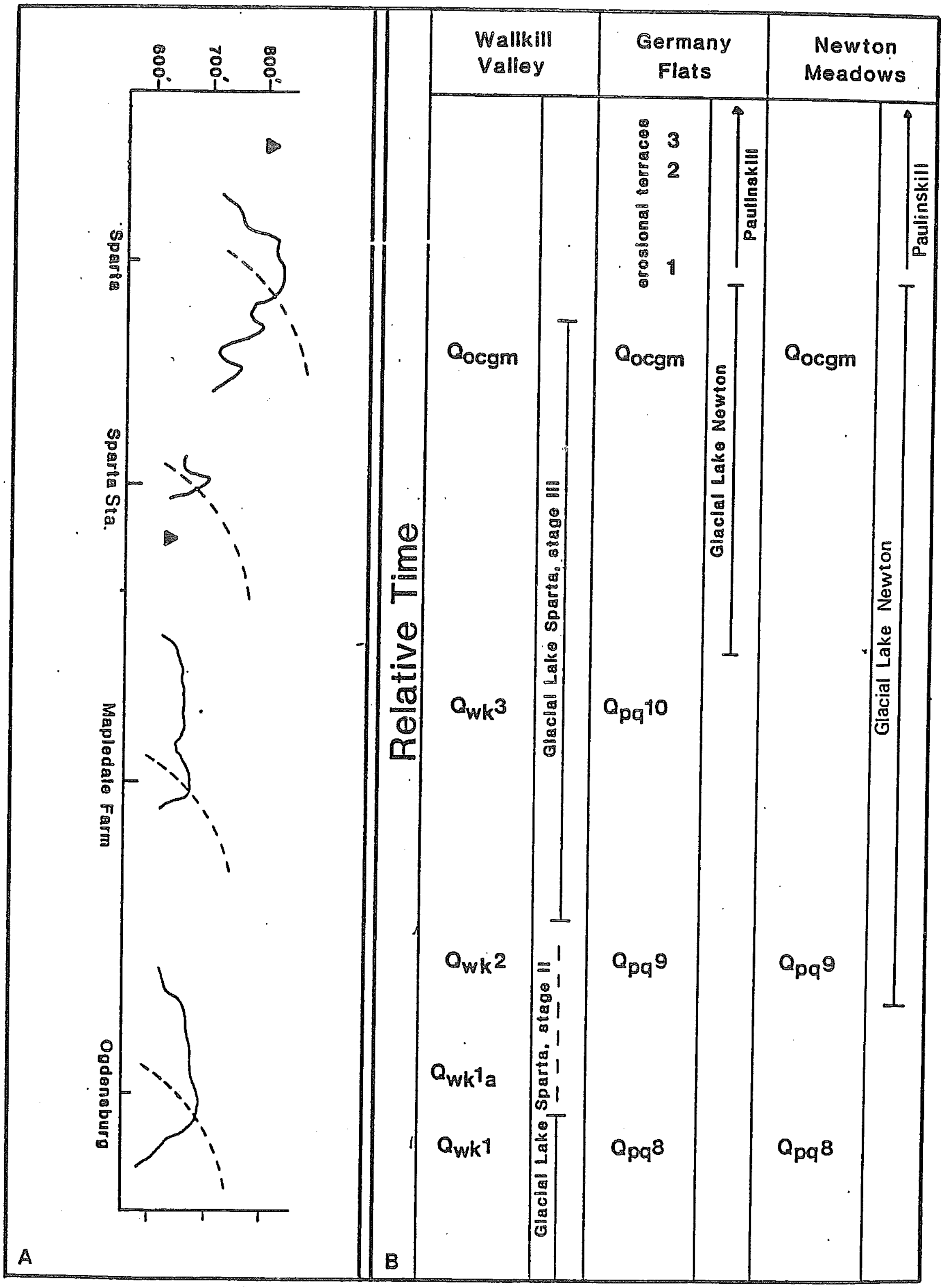

Figure 2. (A) Longitudinal profiles of four ice-contact morphosequences in the in the Wallkill Valley. (B) Deglaciation chronology of the field trip area relating the deposition of morphosequences and the formation of proglacial lakes.

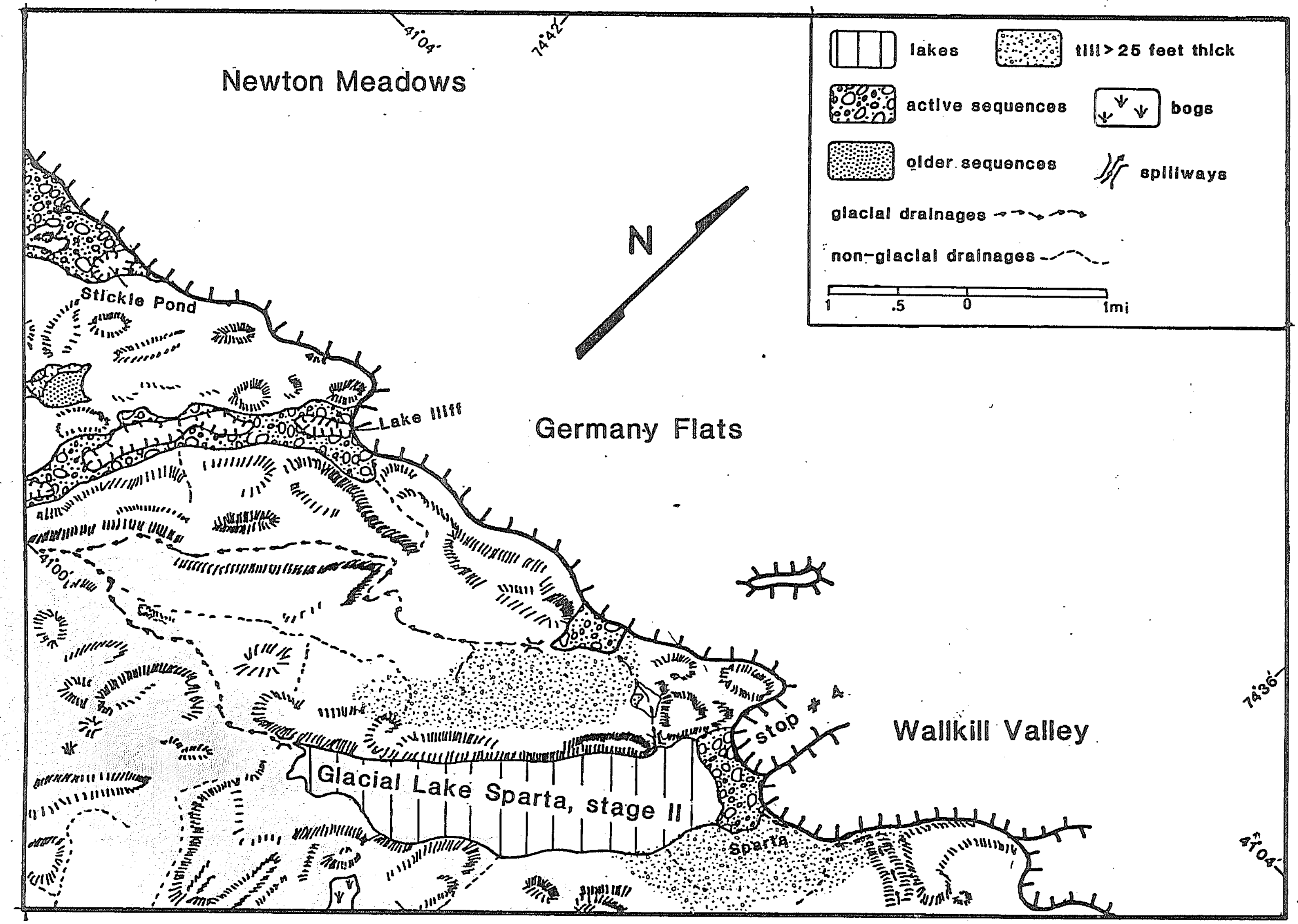

Figure 3. Sparta-Lake Iliff ice margin.

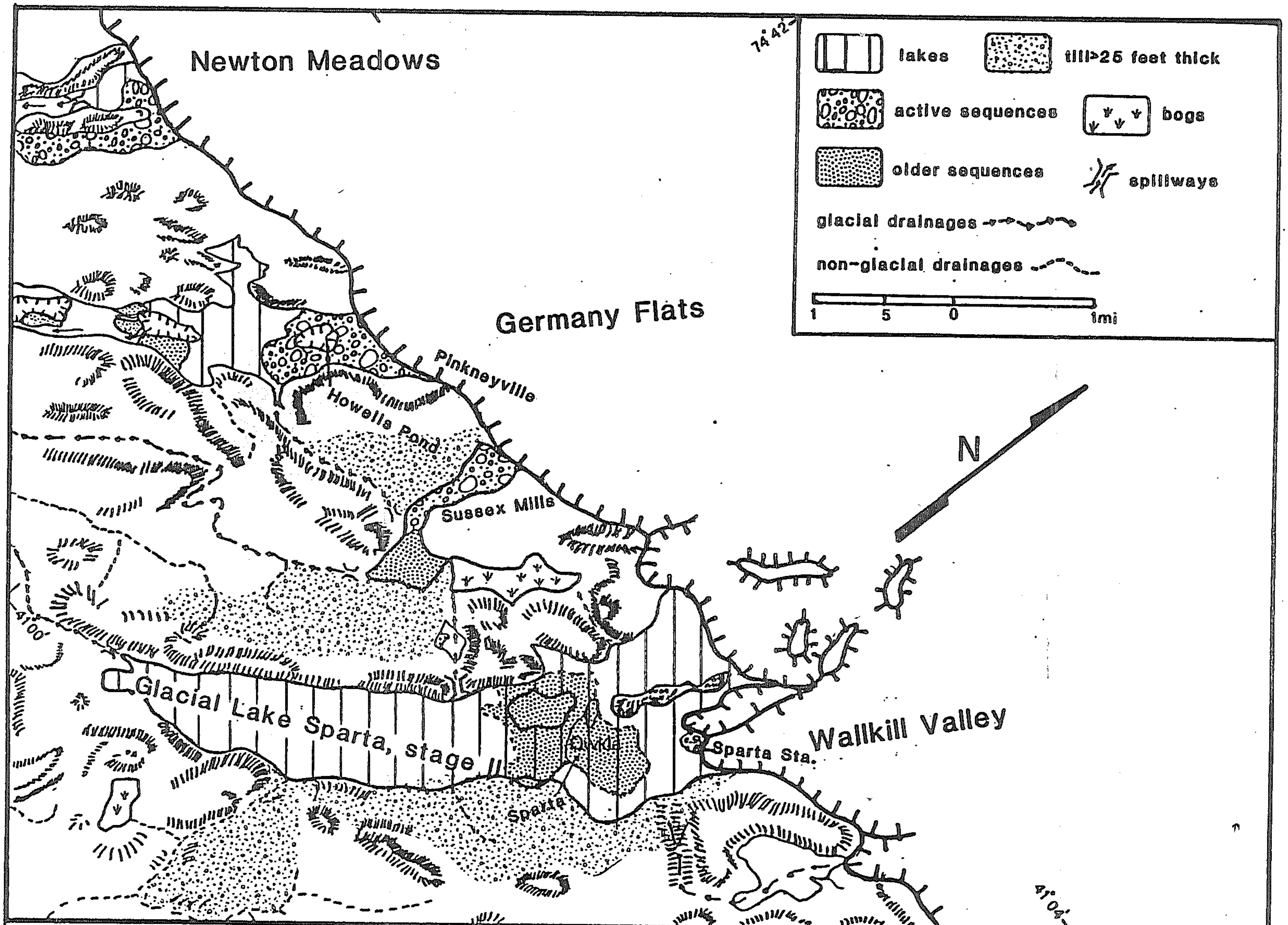

Figure 4. Sparta Station-Pinkneyville ice margin.

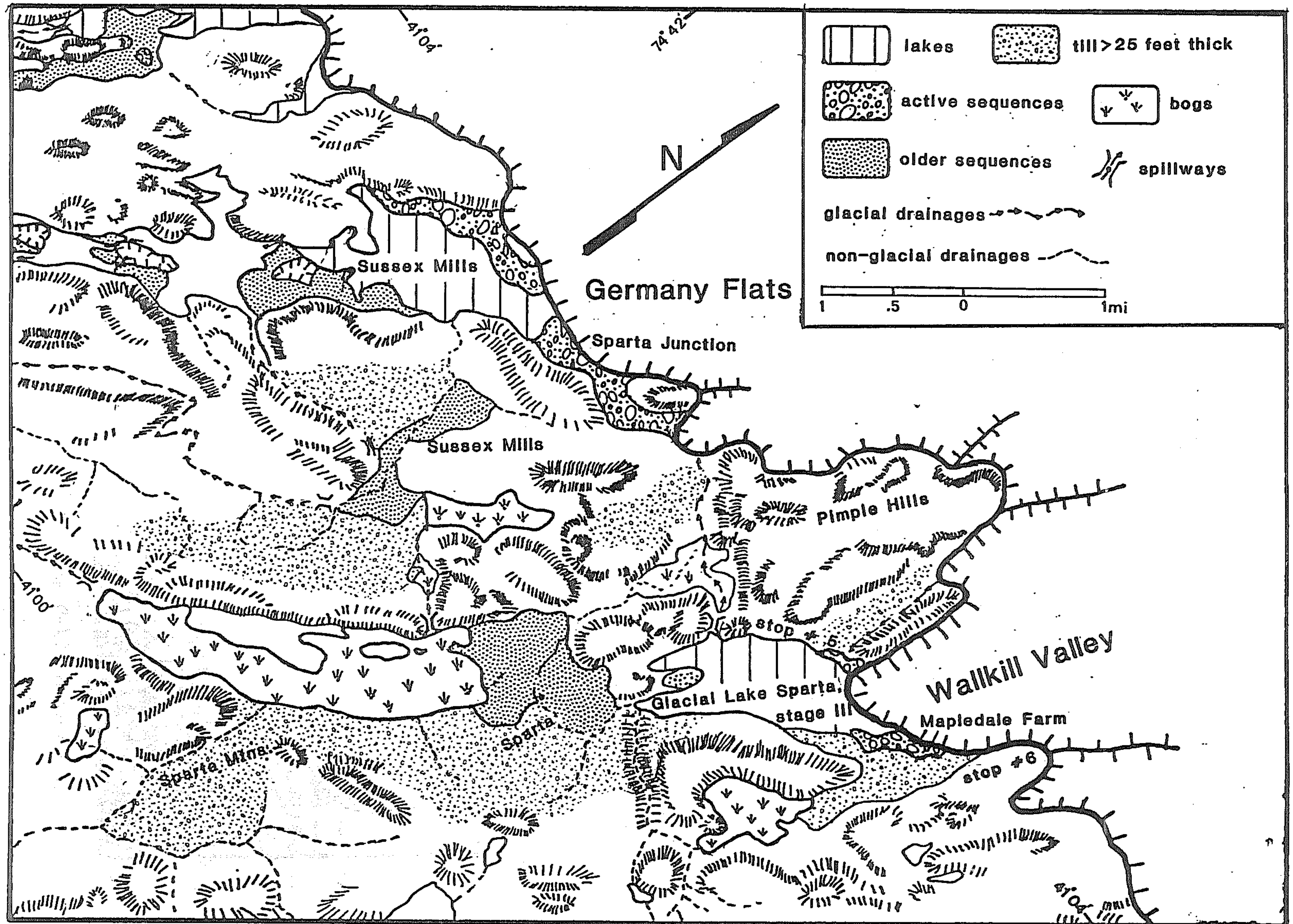

Figure 5. Mapledale Farm-Sparta Junction ice margin.

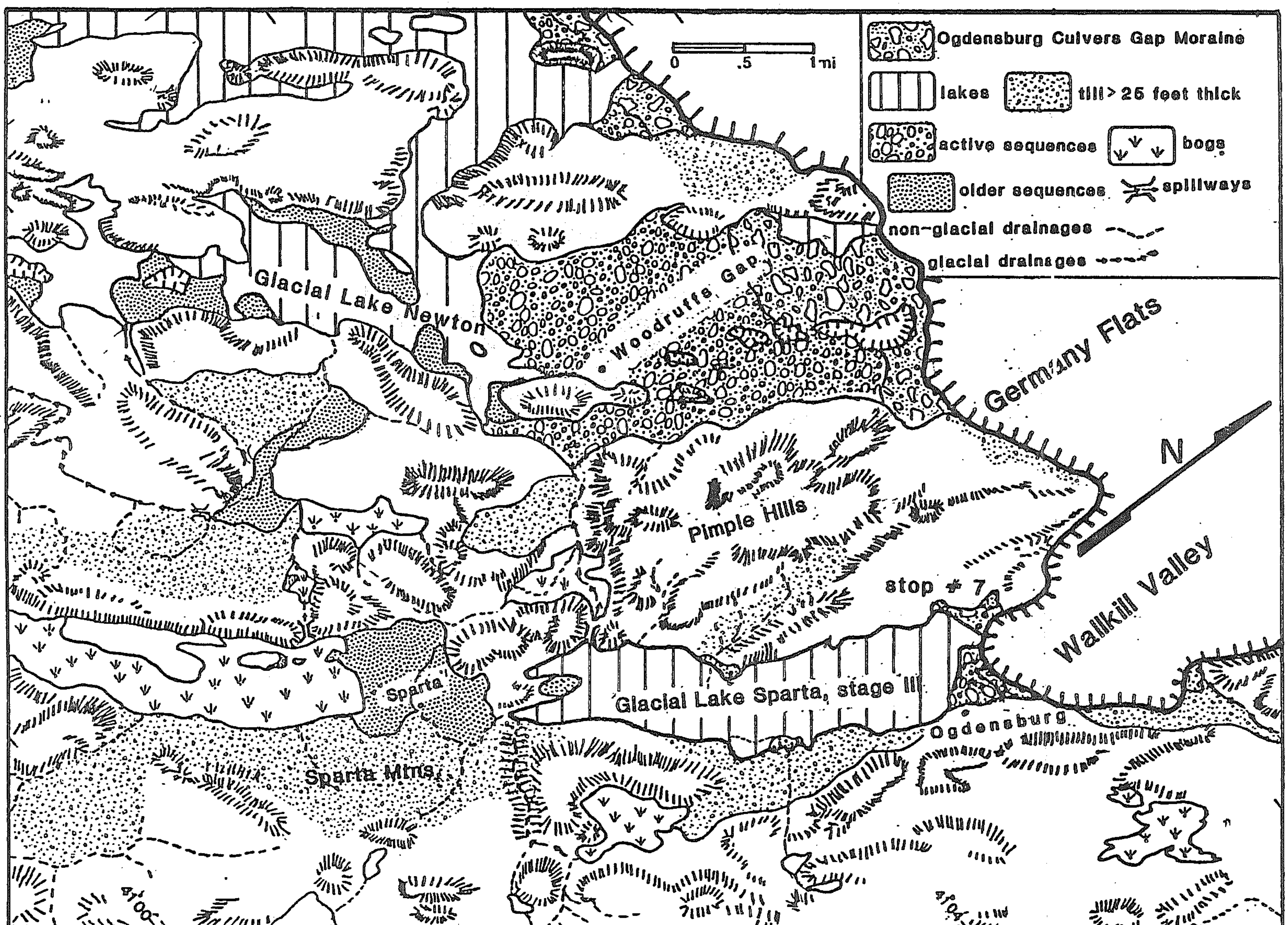

Figure 6. Ogdensburg-Culvers Gap Moraine.

the oldest deposit in the Wallkill Valley that represents a major ice margin standstill. Qwkl represents an ice-contact delta built into glacial Lake Sparta, Stage II. The spillway for Stage II is 0.25 miles south of Qwkl in a col located on the west side of the lake (elev. 825 feet). Glacial Lake Sparta, Stage I, is inferred to have existed, based on ice geometry, south of the 825 foot outlet. The outlet for Stage I would have existed at an elevation of 840 feet at the south end of the lake. However, there are no morphosequences built into this stage that represent an ice margin standstill.

An ice-contact deltaic deposit has been identified at Sussex Mills 1.5 miles west of Qwkl. This feature represents the infilling of a very small proglacial lake which had its spillway 0.25 miles south of the deposit (elev. 800 feet). This outlet also served as a drainage path for waters from glacial Lake Sparta, Stage II.

In the lower Germany Flats, near Lake Iliff, a series of kame terraces (Qpq8) are interpreted as an ice-contact fluvial morphosequence. A similar situation existed at Stickle Pond, located in the Newton Meadows.

Sparta Station-Pinkneyville Ice Margin (Fig. 4)

As the ice margin retreated northward, two ice-contact lacustrine morphosequences were deposited into the receeding waters of glacial Lake Sparta, Stage II. Qwkla represents the uncovering of a 800 foot outlet one mile northwest of Sparta. Qwk2, located at Sparta Station is graded to an outlet 1.5 miles west of Sparta Station, at an elevation of 720 feet. Ice margin retreat from Qwk2 uncovered a third outlet (elev. 650 feet), 0.5 miles north of Sparta Station; thus forming glacial Lake Sparta, Stage III.

Associated with the retreat of the ice margin to Sparta Station is the deposition of an ice-contact lacustrine morphosequence at Sussex Mills. Elevations of deposits vary from 760 feet at the south end of the complex to 720 feet at the north end and indicate an uncovering of lower outlets during ice margin retreat. In the Germany Flats, Qpq9, an ice-contact lacustrine morphosequence was built into a small proglacial lake. The spillway was regulated by ice blocks and drift of Qpq8. In the Newton Meadows, the deposition of Qpq9 coincided with the uncovering of a spillway at 599 feet and the subsequent formation of glacial Lake Newton.

Mapledale Farm-Sparta Junction Ice Margin (Fig. 5)

In the Wallkill Valley, 1.5 miles south of Ogdensburg, two ice-contact lacustrine morphosequences (Qwk3), represent deposition into glacial Lake Sparta, Stage III (Fig. 2A). The outlet for Stage III is located 3000 feet north of Sparta Station, along the Susquehanna and Western Railroad. A spillway elevation of 650 feet was determined by using an altimeter. In the Germany Flats an ice-contact lacustrine morphosequence (Qpq10), formed into a small proglacial lake, dammed by Qpq9 at Howell's Pond. No morphosequences were recognized in the Newton Meadows corresponding to Qpq10 and Qwk3.

Ogdensburg-Culvers Gap Moraine (Fig. 6)

In the Wallkill Valley, the position of the Ogdensburg-Culvers Gap Moraine is delineated by the Ogdensburg Embankment and deposits of stratified drift on the west side of the Valley. Both deposits represent ice-contact lacustrine morphosequences built into glacial Lake Sparta, Stage III. The profile of the Ogdensburg Embankment (Qocgm), graded to the spillway of glacial Lake Sparta, Stage III, is illustrated in Fig. 2A.

In the Germany Flats, a massive kame complex and outwash plain mark the position of the Ogdensburg-Culvers Gap Moraine. The outwash plain is a braided stream deposit graded to glacial Lake Newton and overlies deltaic sediments which completely filled in the Germany Flats from Lake Grinnel, south to Woodruffs Gap. Several ice blocks existed within the outwash. The depressions left behind are occupied by White Lake, Lake Grinnel and several smaller lakes. In the Newton Meadows, near Lafayette, the Ogdensburg-Culvers Gap Moraine is represented by a kame complex, bordered to the south by deltas built into glacial Lake Newton.

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LATE WISCONSINAN DEGLACIATION
FROM THE
OGDENSBURG-CULVERS GAP MORAINE TO THE SUSSEX MORAINE

by

Scott Stanford and Dave Harper

New Jersey Geological Survey

North of the Ogdensburg-Culvers Gap Moraine ice retreat was across the Paulins Kill-Wallkill divide and into the basin of the Wallkill River (refer to figure 1). The Wallkill flows north and empties into the Hudson River at Kingston, New York, approximately 60 miles from the divide. The north-draining valleys of Papakating Creek, Beaver Run, and the Wallkill River were dammed by the retreating ice to form a series of proglacial lakes. Deltas and sublacustrine fans deposited in these lakes mark ice-margin positions and lake elevations. As ice retreated, water from lakes in the Beaver Run and upper Wallkill valleys drained southward, then westward, across bedrock interfluves and drift dams to lower levels. Eventually these lakes lowered to a single large lake draining into the Paulins Kill basin across the low point on the Paulins Kill-Wallkill divide, near Augusta, at an elevation of 500 feet (figure 1). The 500-foot lake persisted until an outlet at an elevation of 400 feet near Goshen, New York (approximately 35 miles northeast of Augusta) was uncovered (Adams, 1934; Connally and Sirkin, 1967). At this stage drainage shifted eastward into the Hudson basin.

Figure 1: Study Area

The following discussion will trace the deglaciation in the Wallkill and Papakating valleys (refer to figure 1) from the Ogdensburg-Culvers Gap ice margin to the the Sussex ice margin of Connally and Sirkin (1973), ten miles to the north. This discussion is based on reconnaissance mapping in the Hamburg quadrangle (Stanford and Harper, in preparation) and on mapping by Kummel (unpublished) and Spencer and others (1908) in the Branchville quadrangle.

The first ice margin north of the Ogdensburg-Culvers Gap Moraine, termed the Augusta Moraine by Connally and Sirkin (1973), is indicated by deltas deposited in lakes in the upper Wallkill and Beaver Run valleys, and by constructional till topography and an outwash surface on the Paulins Kill-Wallkill divide at Augusta (figure 2). In the upper Wallkill valley glacial Lake North Church (Salisbury, 1902) drained southward into the Paulins Kill basin across the Germany Flats outwash surface of the Ogdensburg-Culvers Gap Moraine. An ice block at the outlet probably made for an unstable spillway. It appears that as the ice block melted the spillway was lowered. Three successively lower erosional channels cut into the Germany Flats surface (Witte, personal communication) indicate that spillway elevations fell from approximately 620 feet to 600 feet. A large delta at North Church marks the Augusta margin in this lake.

To the west of glacial Lake North Church a smaller lake in the Beaver Run valley was dammed on the south by the Ogdensburg-Culvers Gap Moraine. This lake drained westward to the Paulins Kill over a bedrock spillway at an elevation of approximately 560 feet. South of Harmonyvale a broad delta marks the Augusta margin within this lake.

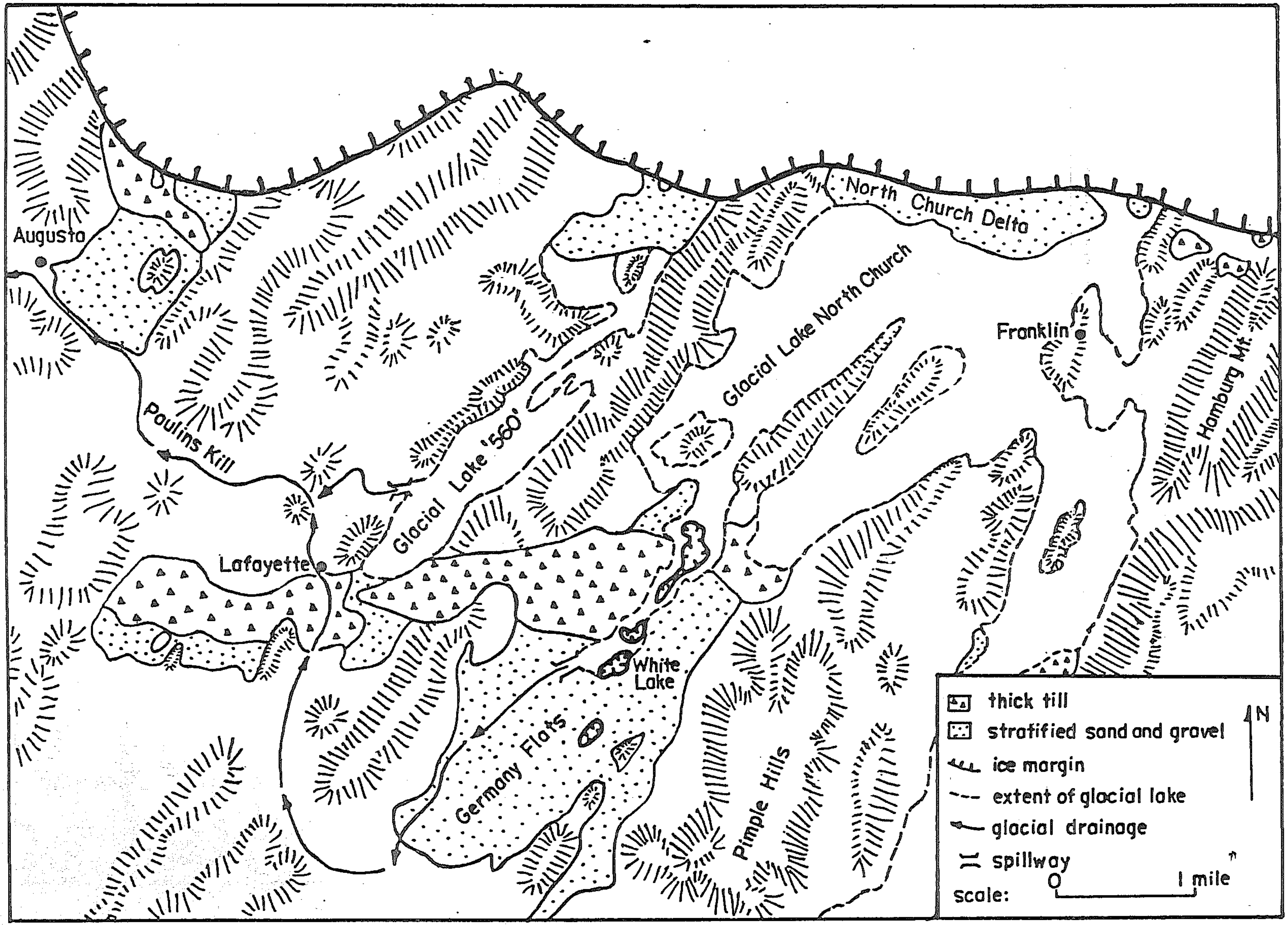

Figure 2: Ice at Augusta Margin

As the ice withdrew from the Augusta position three new lakes formed (figure 3). In the Papakating valley ice retreated north from the Paulins Kill-Wallkill divide. At the divide, the previously-mentioned 500-foot spillway near Augusta became active. As at the outlet for glacial Lake North Church, flow was across ice-margin sediment and ten to twenty feet of erosion may have occurred before the outlet stabilized at 500 feet.

At the same time in the Beaver Run valley a lake at an elevation of 575 feet spilled southward across the previously-deposited delta of the Augusta margin. This water drained to the Paulins Kill through the "560" lake, now not in contact with ice. The elevation of the topset-foreset contact in a small delta at Beaver Run confirms this outlet elevation.

In the upper Wallkill valley glacial Lake North Church appears to have lowered to 590 feet when a low divide over bedrock was uncovered just north of the North Church delta. Drainage was thus diverted from the Germany Flats outlet into the "575" lake in the Beaver Run valley. Although there are no topset-foreset exposures marking the 590 level, several sublacustrine fans within the postulated area of this lake attain but do not exceed elevations of 580 feet.

A stable ice margin position in the "575" and "590" lakes is suggested by the locations of the Beaver Run delta and by sizable sublacustrine fans in the "500" lake and in the "590" lake, as shown on figure 3.

Continued retreat of ice uncovered still lower divides across interfluves. When the ice margin was at the position shown in figure 4,

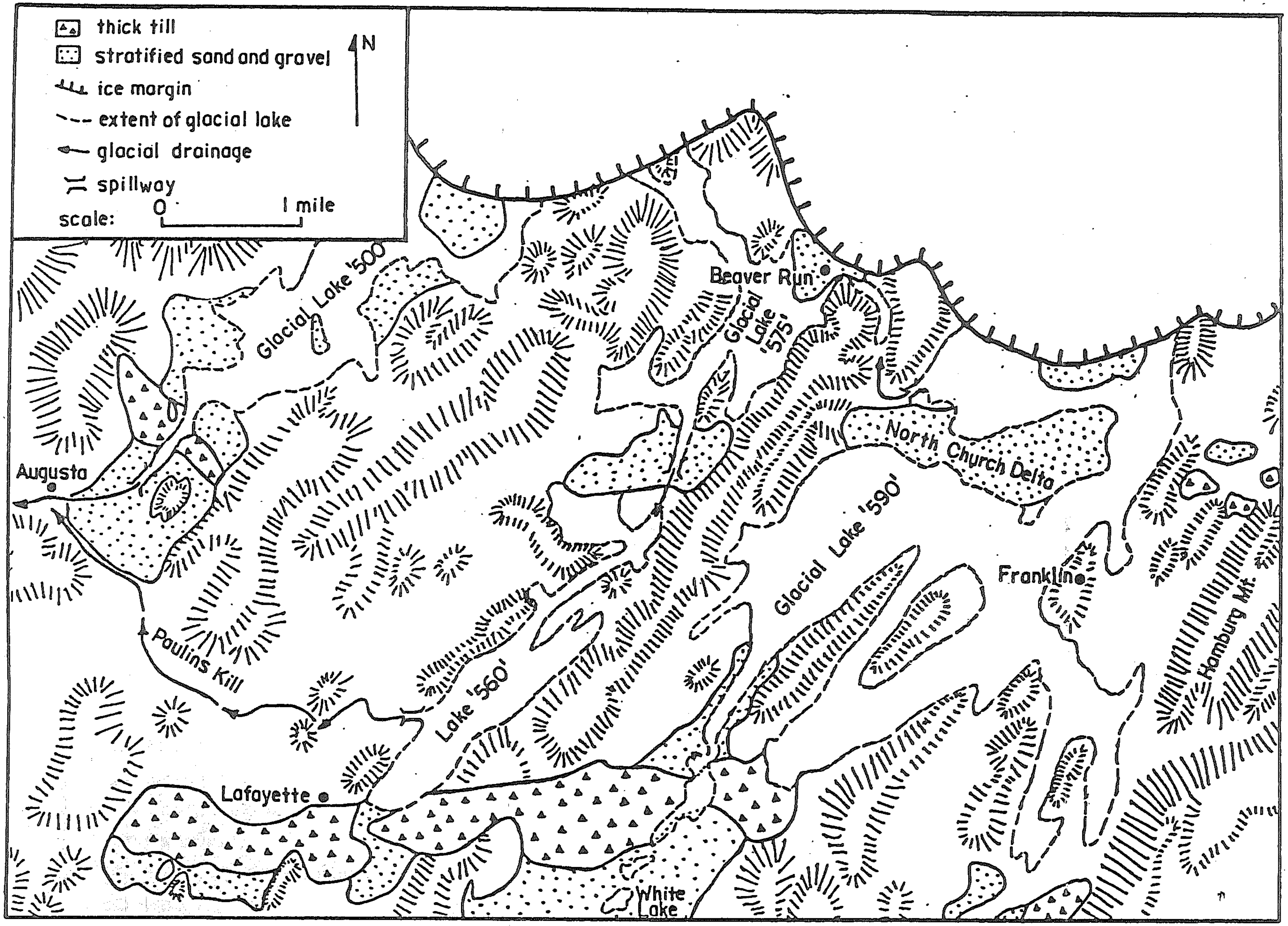

Figure 3: Glacial Lakes "590", "575", and "500"

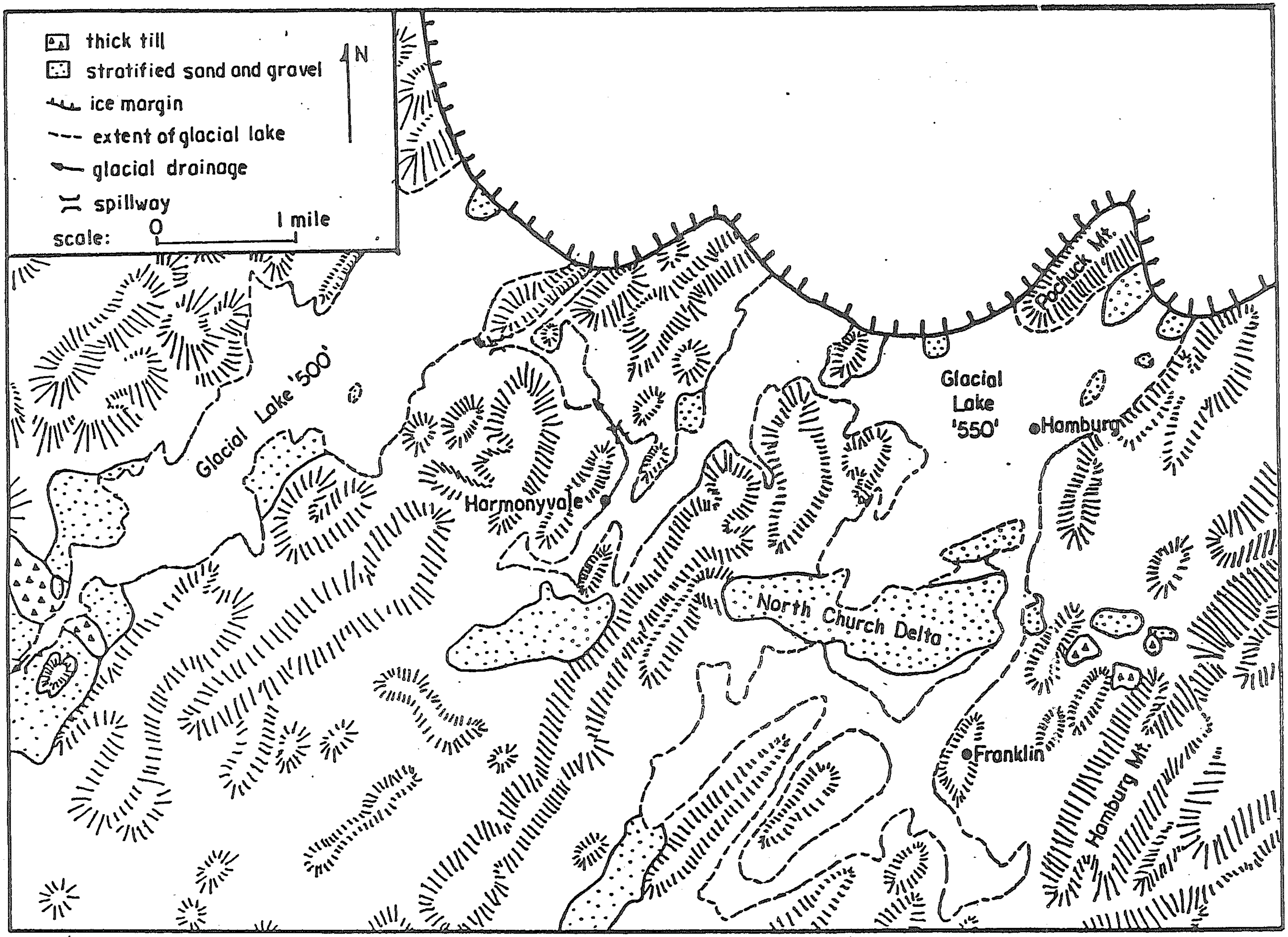

Figure 4: Glacial Lakes "550" and "500"

a divide on bedrock at an elevation of 550 feet north of Harmonyvale opened, lowering the "575" lake to 550 feet. At this time, all drainage from the Wallkill basin was channeled through the outlet at Augusta and into the Paulins Kill.

In the upper Wallkill valley the "590" lake dropped to become conterminous with the "550" lake when a gap in the Beaver Run-upper Wallkill divide approximately 1.5 miles west of Hamburg was uncovered. At this time Pochuck Mountain began to cause separation of the ice into lobes occupying Vernon Valley on the east and the Wallkill valley on the west. The relative positions of these lobes at the time of lowering of the "590" lake are indicated by the elevations of lacustrine gravels deposited by the Vernon Valley lobe in the valley northeast of Hamburg. Here, sublacustrine fans showing only foreset bedding rise to an elevation of 580 feet, indicating deposition in the "590" lake. A small delta having a topset-foreset contact at 550 feet lies a quarter mile northeast of the fans, showing uncovering of the aforementioned gap by the Wallkill lobe and lowering of the "590" lake to 550 feet during retreat of the Vernon Valley lobe over that distance.

Deposits in the "550" lake include the small delta northeast of Hamburg just mentioned and a linear string of sublacustrine fans, generally less than 520 feet in altitude, trending northward along the west side of the Wallkill valley from Hamburg to Sussex. These latter deposits may mark successive positions of the mouth of an englacial or subglacial channel.

Further melting back of the ice past the end of the Beaver Run-

Papakating Creek divide southeast of Sussex (refer to figure 1) permitted the "550" lake to lower to the "500" level.

After lowering of the "550" lake, an ice block or, possibly, the "550" delta northeast of Hamburg, maintained a short-lived lake at 540 feet in the southern end of Vernon Valley. The existence of this lake is indicated by a well-developed outlet channel leading from a low col in a bedrock ridge, cutting through the 550 delta, and terminating at the 500-foot lake level (shown on figure 5). Gravel terraces banked on the sides of Vernon Valley north from the outlet to the vicinity of Sand Hills may have been deposited in this "540" lake. By the time the ice margin was at Sand Hills, however, the dam had melted or eroded and the lake had dropped to 500 feet. At this time the only operating spillway in the Wallkill basin was at Augusta.

The final ice margin in the study area, termed the Sussex Moraine by Connally and Sirkin (1973), is shown on figure 5. This margin is marked by a large delta northeast of Sussex in the Wallkill valley and by a delta at Sand Hills in Vernon Valley, both deposited in the "500" lake. Lobation around Pochuck Mountain linking the two deltas is indicated by well-developed ice-lateral meltwater channels on both sides of the mountain and by a bench of thick till that is especially prominent on the east side.

A topset-foreset contact was tentatively identified in the Sussex delta at approximately 520 feet, and in the Sand Hills delta at approximately 510 feet, yielding an average isostatic uplift of approximately 3 feet/mile to the north from the 500-foot outlet at Augusta.

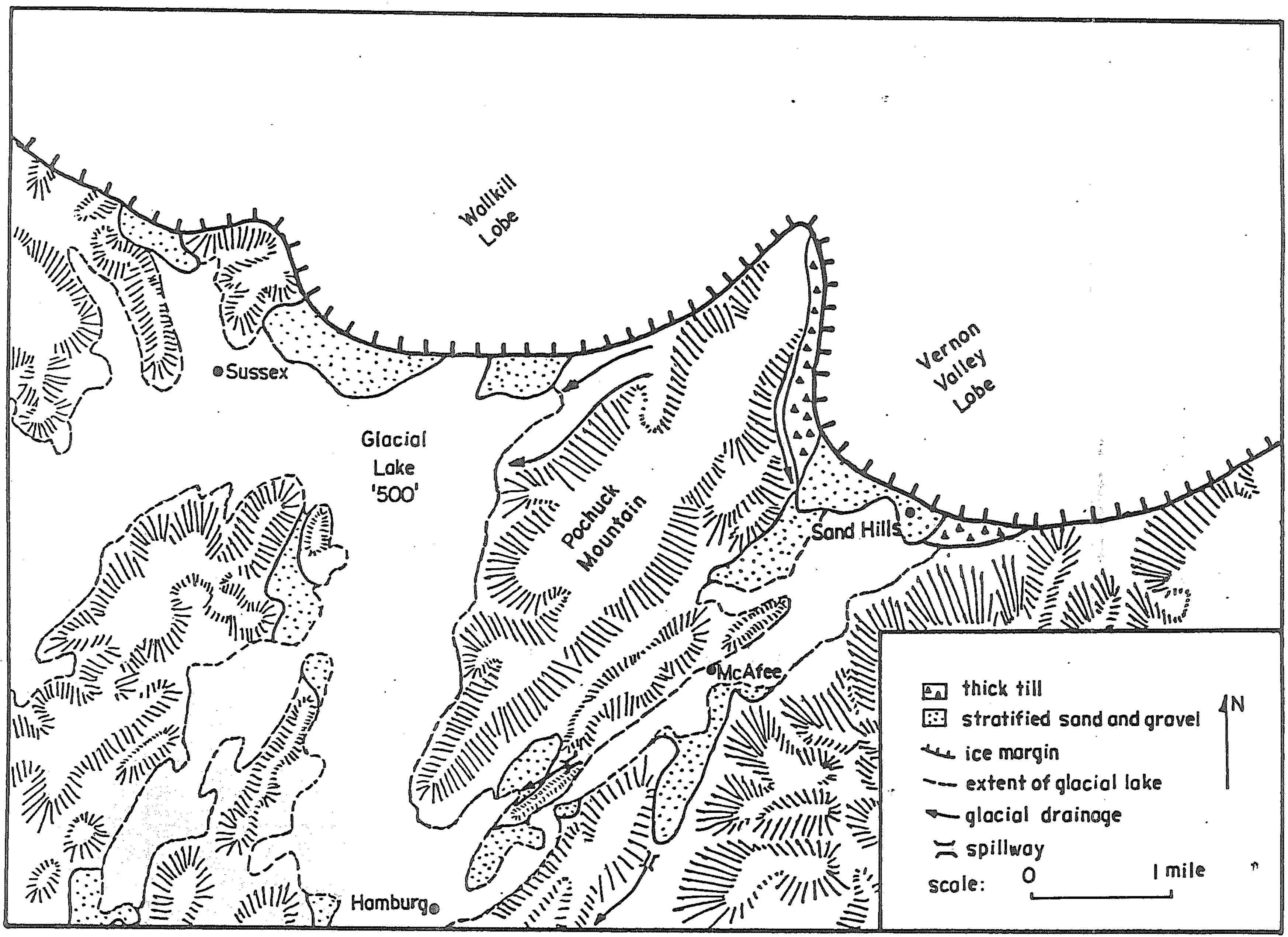

Figure 5: Ice at Sussex Margin

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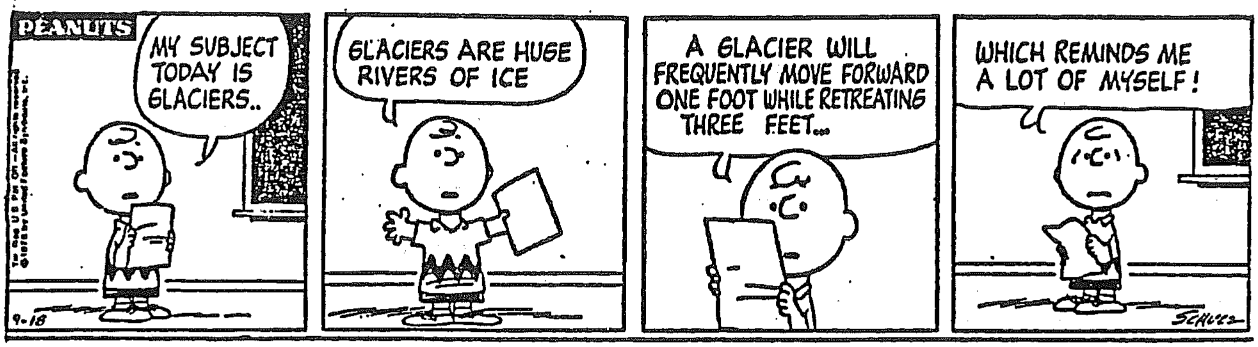
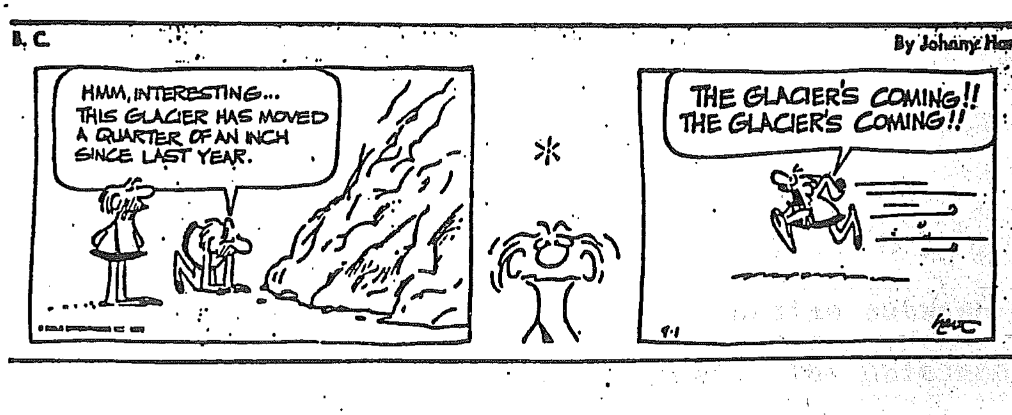
FIELD TRIP STOPS





STOP 1: OUTWASH AND COLLUVIUM AT BRAINARDS, N.J.

The gravel pit at Brainards, NJ exposes a unique cross-valley section through the guts of a Late Wisconsinan (Woodfordian) valley train terrace which houses several older stratigraphic units (Figure 1). The terrace surface reaches an elevation of 84 m (275 ft) and it is the oldest and highest Woodfordian outwash unit along the Delaware River which has an elevation of 56 m (185 ft) at Brainards. Brainards lies about 3 km south of the maximum extent of Woodfordian ice and lies 6 km south of the Woodfordian "Terminal Moraine". The uppermost Woodfordian outwash gravels at Brainards are graded to a morphosequence in the Delaware Valley which was deposited during the early deposition of the "Terminal Moraine" (see STOP 2). Brainards, NJ lies within the limits of Pre-Wisconsinan (Illinoian?) glaciation by at least 15 km. Valley train terraces associated with Pre-Wisconsinan ice recession may be seen above the Brainards gravel pit to the southeast at an elevation of 100 m (330 ft).

Four stratigraphic units may be seen in the Brainards exposure which are described and discussed below starting at the top of the sequence (Figure 2). An inferred geomorphic and depositional history of the Brainards pit is given in Table 1.

1) upper Wisconsinan gravel (Woodfordian)

The upper Wisconsinan gravel is a Woodfordian outwash deposit equivalent in age to the "Terminal Moraine". The unit consists of 0 to 10 m of cobbly to sandy gravel with fluvial current structures and a Woodfordian soil on its surface. The gravel

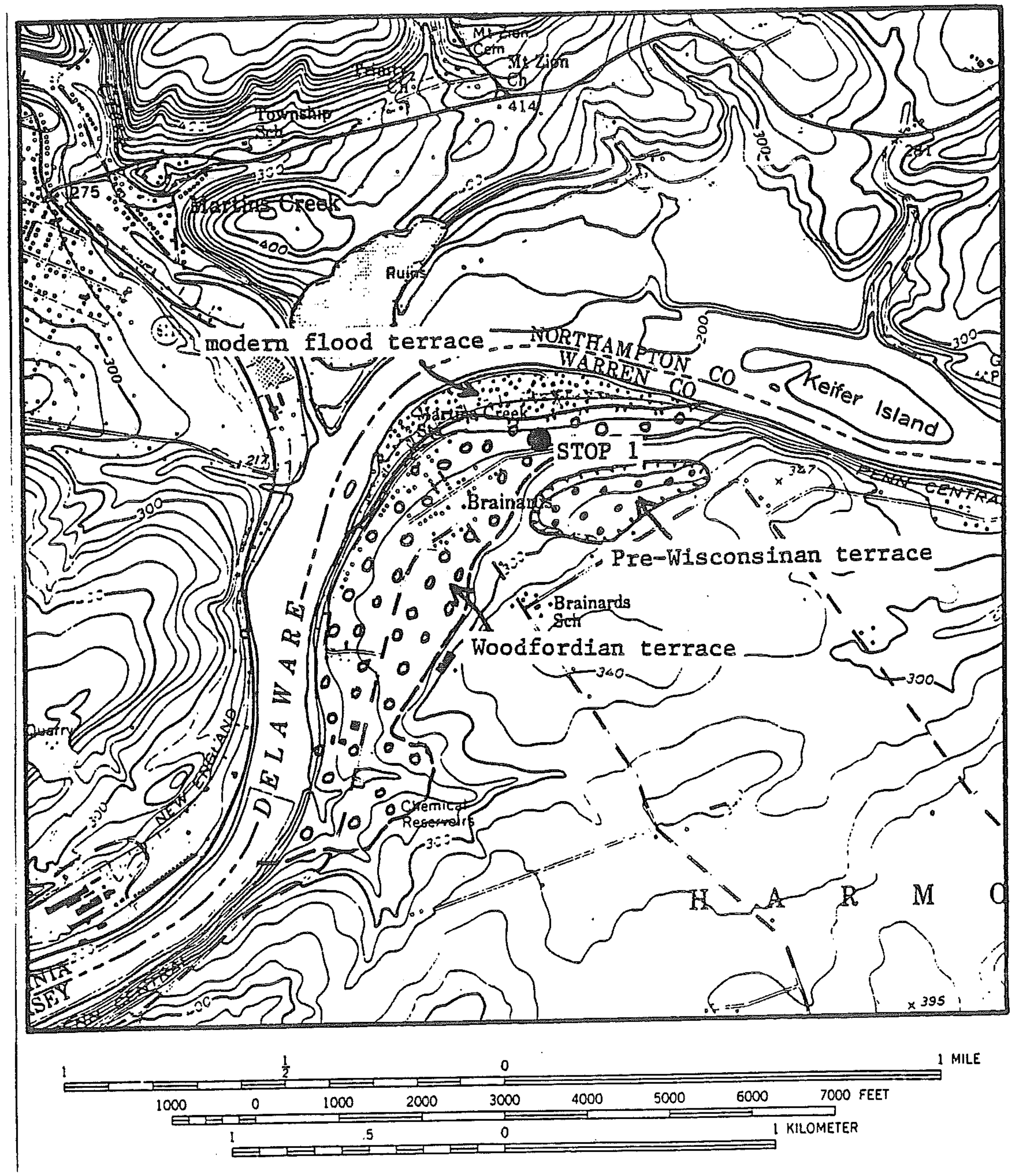


Figure 1. The location of STOP 1 at Brainards, NJ in the Bangor, PA-NJ 7½-minute quadrangle. The map shows terraces at Brainards, NJ.

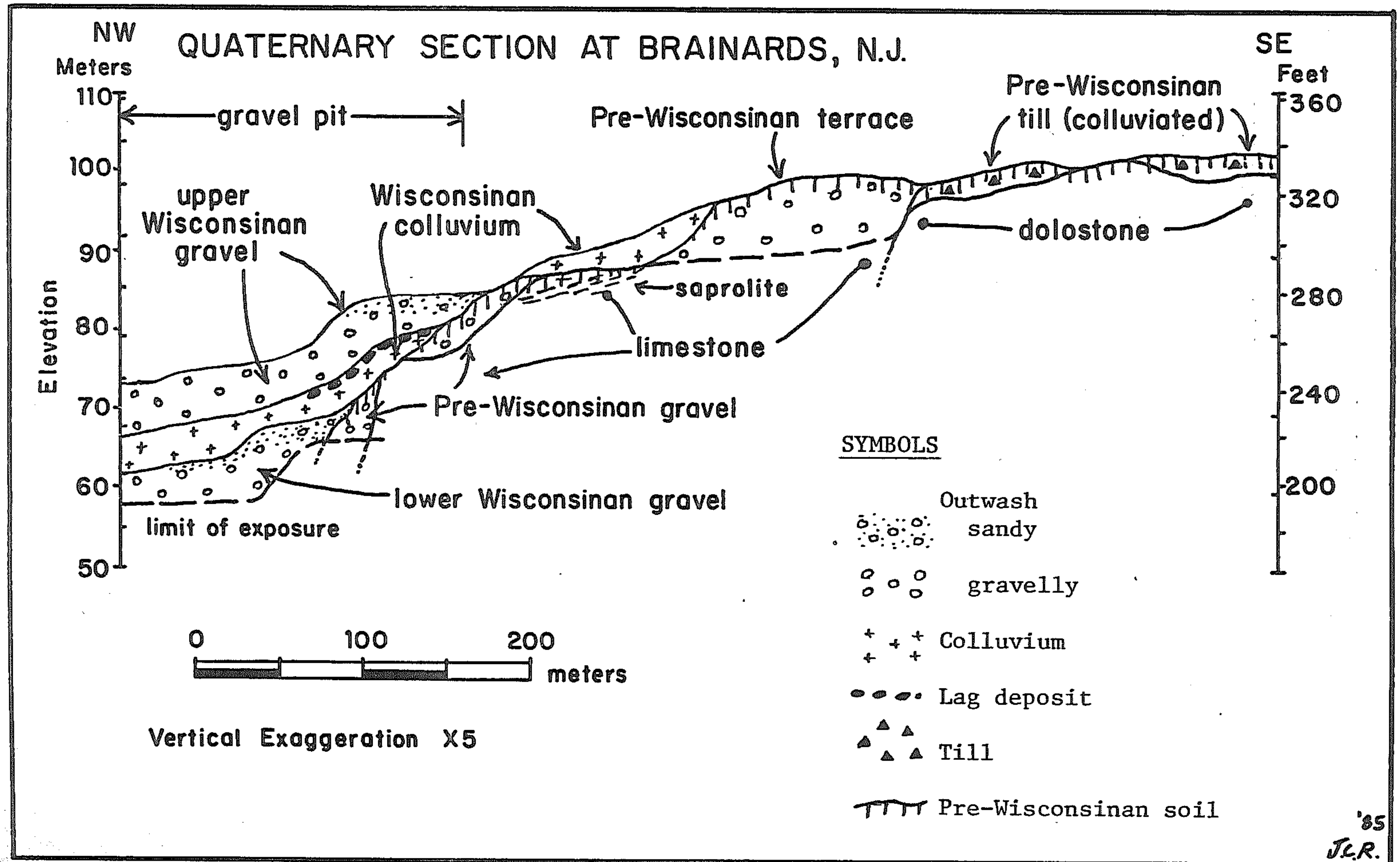


Figure 2. A cross section of the Quaternary stratigraphy exposed at the Brainards, NJ gravel pit (STOP 1) and the adjacent Pre-Wisconsinan terrace and carbonate plateau to the southeast.

Table 1. The geomorphic and depositional history of the Delaware Valley as inferred from deposits at Brainards, NJ. Elevations are recorded at the Brainards pit (Figure 1).

HOLOCENE		-- Modern flood terraces, alluvium and channel of the Delaware River associated with postglacial down-cutting and flooding in the Delaware Valley.
WISCONSINAN	Late Wisconsinan	-- Formation of multiple glaciofluvial valley train terraces during Woodfordian deglaciation. -- Aggradation by the glacial Delaware River to an elevation of 275 ft and deposition of the "upper Wisconsinan gravel" unit during deposition of the "Terminal Moraine" in Woodfordian time. -- Fluvial erosion of pre-Woodfordian deposits and lag deposit formation associated with initial increased fluvial activity during Woodfordian ice advance.
		? ? ? ?
	Early and Middle Wisconsinan	-- Colluviation of hillslopes and deposition of the Wisconsinan colluvium unit possibly triggered by periglacial activity during a time of non-aggradation by the Delaware River and low terrace levels (less than 230 ft) in the Delaware Valley. -- Formation of multiple fluvial terraces associated with waning fluvial activity. -- Aggradation by the pre-Woodfordian Delaware River to an elevation of 230 ft and deposition of the "lower Wisconsinan gravel" unit. Triggered by increased glaciofluvial activity or increased sediment discharge from inwash sources in the Delaware River catchment basin. -- Slight fluvial erosion of pre-existing deposits and soils by initial increased fluvial activity. -- Possible colluviation triggered by climatic cooling. (Not represented by deposits at Brainards)
SANGAMON		-- Development of interglacial (Sangamon) soil.
ILLINOIAN		-- Formation of multiple glaciofluvial terraces associated with Illinoian deglaciation. -- The last Illinoian glaciation in the area. Ice-cover followed by ice recession resulting in the highest elevation (330 ft) of glaciofluvial aggradation and deposition of Illinoian outwash deposits.

The upper Wisconsinan gravel truncates the colluvium with a very sharp contact along which is a dolostone lag deposit derived from fluvial erosion of the colluvium. The colluvium is generally non-layered but may exhibit indistinct stratification that appears to dip to the northwest away from the local valley wall. Layering in the colluvium is the result of sorting during downslope mass-movement and not the result of fluvial activity. The colluvium is distinguished from the overlying and underlying gravel units by the many non-fluvial clasts found in it and its deficiency of matrix material.

The exact age of the colluvium is uncertain but it is Wisconsinan because it truncates a soil developed on Pre-Wisconsinan (Illinoian?) glacial deposits and therefore probably post-dates the Sangamon Interglacial. The colluvium may reflect periglacial activity on the hillslope to the southeast and it may have been deposited during the earliest part of the Woodfordian or the Middle or Early Wisconsinan under cool climatic conditions.

3) lower Wisconsinan gravel

The lower Wisconsinan gravel unit is a sandy, pebbly gravel with fluvial current structures. It is 0 to more than 12 m thick (limit of exposure) and it pinches out against the southeast valley wall. The approximate elevation of the upper contact of this unit is 69 m (230 ft) which is below the elevation of the lowest Woodfordian valley train terrace in this part of the

Delaware Valley, but above the highest flood terrace of the modern Delaware River. The lower Wisconsinan gravel is finer-grained than the upper gravel unit suggesting a more distal sediment source. The lower Wisconsinan gravel has an abrupt contact above the Wisconsinan colluvium which is the only unit that has been found overlying the lower gravel. It is not known whether this contact is a conformable lithologic discontinuity. The lower gravel very sharply truncates Pre-Wisconsinan soils and deposits below along an erosional unconformity. Redeposited, pre-weathered Pre-Wisconsinan soil material is found within the lower Wisconsinan gravel unit and it therefore post-dates a time of Pre-Wisconsinan (Sangamon?) interglacial soil development. No soil development has been found on the upper surface of the lower gravel unit.

The lower Wisconsinan gravel unit is interpreted as a Middle or Early Wisconsinan unit as indicated by:

A) its position below Woodfordian (L. Wisconsinan) outwash and colluvium that pre-dates Woodfordian outwash, and

B) truncation of a Pre-Wisconsinan (Sangamon?) soil below the unit.

There is a temptation to correlate this unit with Altonian glaciation in Pennsylvania. Alternatively, the lower Wisconsinan gravel unit may represent fluvial infilling of the valley from inwash sources under changing climatic conditions. It does not necessarily represent increased glacial discharge and sediment load in the Delaware Valley. All that can be said at this point is that the lower Wisconsinan gravel unit represents fluvial aggradation of the pre-Woodfordian Delaware River proceeding interglacial (Sangamon?) soil development.

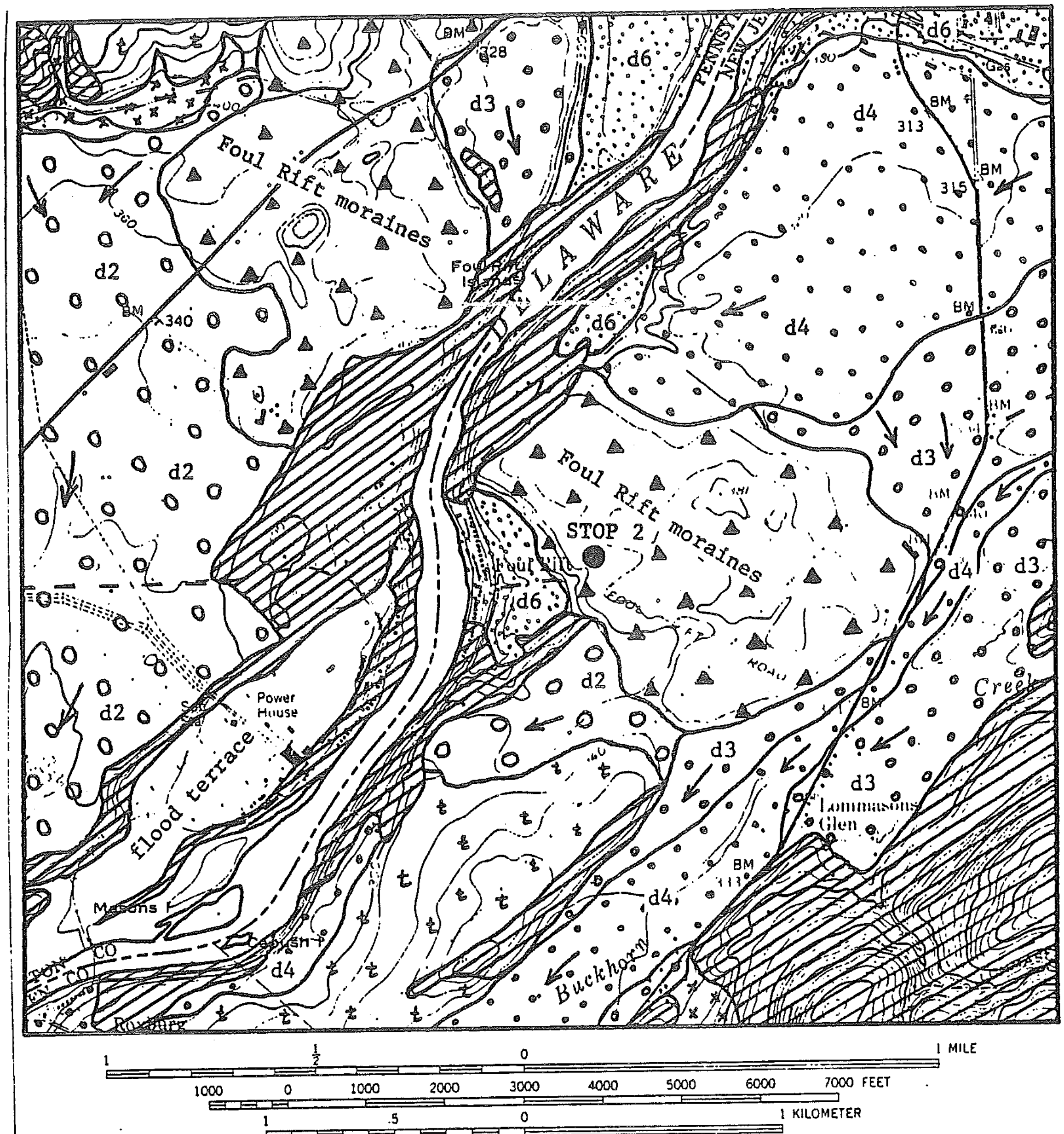
4) <u>Pre-Wisconsinan deposits and soils</u>

Adjacent to the southeast valley wall, the Brainards section exposes up to 4 m of deposits with soil development of probable Sangamon age truncated by any of the units described above. Pre-Wisconsinan weathering may be recognized by: 1) very clayey, oxidized soil material, 2) red (2.5 to 5YR) coloration, 3) moderately strong argillic B horizon development in soil profiles, 4) B horizon development to depths of 3 m or more in non-eroded or non-truncated soil profiles, 5) pebble rubification, and 6) many pebble ghosts. At Brainards, red, clayey pedogenic material also shows soil pedons or a horizontal fissile structure. Soils of this type are developed on Illinoian glacial deposits (Sevon, 1974; Sevon and others, 1975; Marchand, 1978) and are probably the result of weathering during the Sangamon Interglacial. Pre-Wisconsinan deposits (Illinoian?) are found at Brainards with a soil profile that has been eroded by Wisconsinan colluvial and fluvial processes. In addition, highly weathered Pre-Wisconsinan deposits in the Brainards Section may have undergone mixing by colluviation and cryoturbation.

STOP 2: FOUL RIFT MORAINES

The gravel pit east of Foul Rift, NJ (Figure 3) exposes the interior of what is here called the "Foul Rift moraines". The morphologic feature seen at this locality has posed problems to several workers in the Delaware Valley. There is an apparent dichotomy in morphologic and sedimentologic data. On the surface, the sediments appear to be from stratified deposits or kames while their morphologic expression as linear arcuate ridges that cross the valley floor suggest deposition as moraines at a glacier terminus.

The initial tracing of the "great Terminal Moraine" by H. Carvill Lewis included the Foul Rift deposits as part of the moraine and he proposed that these features marked "the southern limit of the great continental glacier" (Lewis, 1884, p. 55). While morphology suggested end moraine ridges, the deposits in the Delaware Valley were not "composed of true till" (Lewis, 1884, p. 55). Lewis mapped the limit of Wisconsinan glaciation in the Delaware Valley as a re-entrant that had its leading edge at Foul Rift. Subsequent workers (Ward, 1938; Ridge, 1983) place the Wisconsinan (Woodfordian) limit further down valley because hills to the south are covered with Wisconsinan till. Ward (1938) and Ridge (1983) favored a kame origin for the deposits because stratification and abundant fluvial clasts were observed in gravel pit exposures. Ward (1938) considered the deposits at Foul Rift as recessional moraines or kames behind the "Terminal Moraine".



SYMBOLS

Delaware Valley morphosequences

- outwash d6
- outwash d4
- outwash d3
- outwash d2
- colluvium
- end moraine
- till
- bedrock exposure

Figure 3. The location of STOP 2 near Foul Rift, NJ in the Belvidere, NJ-PA 7½-minute quadrangle and a surficial map of the surrounding area. The map shows the extension of the Foul Rift moraines into Pennsylvania and some of the morphosequences deposited during and after the deposition of the moraines.

Ridge (1983) considered the deposits recessional kames ("Foul Rift kames" on Figures 7, 12, 13 and 14 of text) that are equivalent in age to parts of the "Terminal Moraine" to the east in New Jersey. Foul Rift deposits are younger than the "Terminal Moraine" segment lying on hills underlain by slate to the west ("Gruvertown moraine" on Figures 7, 11 and 12 of text).

The gravel pit exposure east of Foul Rift shows two horizontal diamicton units, with thicknesses of 4 m (upper) and 2 m (lower), separated by 8 m of outwash gravel. The lower diamicton is overlain by 40 cm of laminated, sandy silts and clays which were deposited in ponded water. Outwash gravel with a thickness of more than 10 m (limit of exposure) lies below the lower diamicton. The diamicton units are interpreted as lodgement tills. They are consistent in thickness across the exposed section, are gravelly in their bases, contain many fluvial clasts which were striated subsequent to rounding, and have a sandy and silty, compact, fissile matrix. Gravels below the till units show deformation but this is not a conspicuous characteristic of the gravel units. The Foul Rift moraines are interpreted as push moraines composed of bulldozed outwash gravel. The till units seen in the gravel pit represent till sheets put down during minor oscillations of the receding Woodfordian ice-front (Figure 4). The Foul Rift moraines are a part of the "Terminal Moraine" that were deposited during the final phases of moraine deposition by at least 5 minor readvances. They are equivalent in age to a large ice-contact glaciofluvial morphosequence in the Delaware Valley (Figure 12 of text). This is in keeping with the general character of the

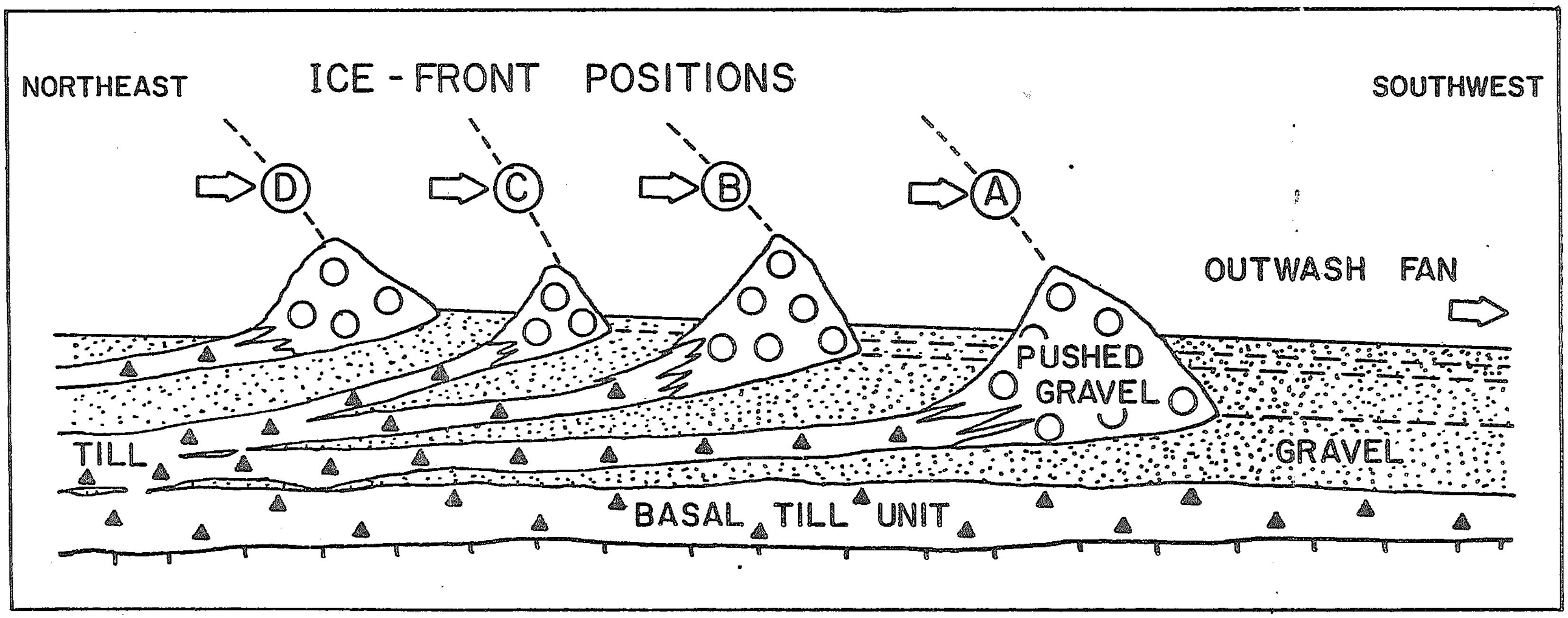


Figure 4. An idealized cross section of the Foul Rift moraines (STOP 2) showing stacked till sheets separated by outwash gravels and moraine hummocks composed of pushed outwash gravel. Moraines were deposited at small readvance positions marked as A (oldest) thru D (youngest). Dashed lines in the outwash fan represent the progressive aggradation of outwash sediment in the Delaware Valley (morphosequence d2) during the construction of the Foul Rift moraines at positions A thru D. The basal till unit extends south of the Foul Rift moraines as shown on Figure 3.

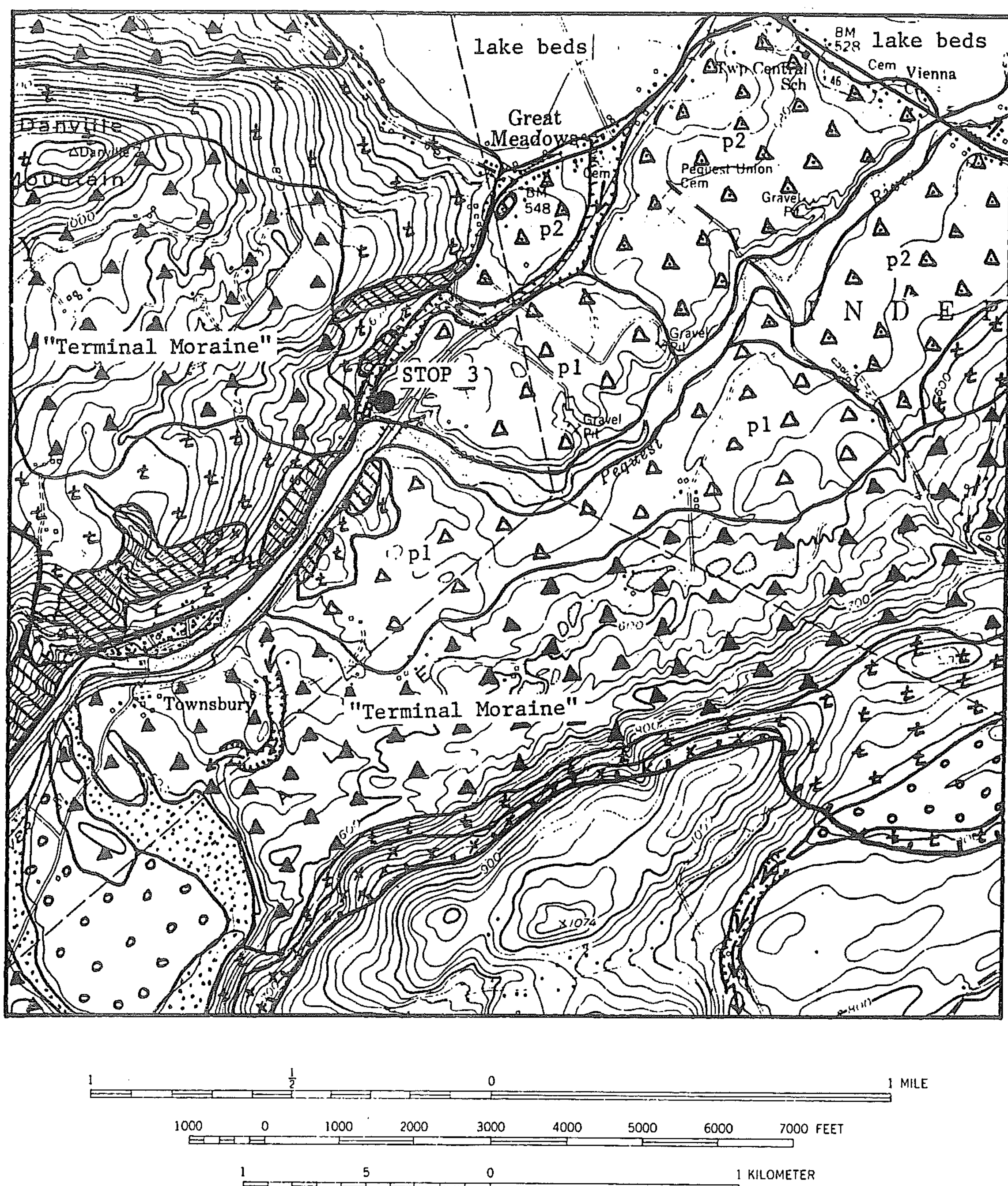
"Terminal Moraine" which is not a single ice-marginal position but rather a composite of several end moraines constructed in a zone that can be as much as 4 km wide. The composite character of the "Terminal Moraine" may be seen on the flank of Kittatinny Mountain near Bangor, Pa ("Bangor moraine" on Figure 7 of text) and also between Buttzville and Hackettstown, NJ (moraines of the Mountain Lake and Pequest Sublobes on Figure 15 of text). In New Jersey, ice-front positions within the "Terminal Moraine" may be tied to as many as four separate morphosequences.

Deposits of the Foul Rift moraines have implications for the development of ice-marginal deposits and morphosequences in the Delaware Valley. They testify to the existence of an active ice margin that had very little stagnant ice. The Foul Rift moraines served as a barrier to later meltwater drainage in the Delaware Valley. The moraine ridges do not continue east to the valley wall because they were dissected by later meltwater drainage (Figure 13 of text). Drainage was forced to the east into the Buckhorn Creek Valley and then to Hutchinson, NJ where it joined the present Delaware River channel. Later meltwater breeched the Foul Rift moraines at Foul Rift and allowed flow along the present Delaware River channel (Figure 14 of text).

STOP 3 (LUNCH STOP): GREAT MEADOWS DELTA

This sand pit exposes beds of the older of two ice-marginal deltaic morphosequences (p1 and p2) deposited into Stage I of Glacial Lake Pequest in the Pequest River Valley (Figure 5). Lake Pequest was dammed by and had its spillway across the "Terminal Moraine" at Townsbury, NJ. In 1979, an exposure of a topset-foreset contact above the present sand pit exposure indicated an elevation of approximately 171 m (560 ft) for Stage I of Lake Pequest and its nearby spillway. The spillway probably occupied an area now cut by the present Pequest River or was across a small divide in the "Terminal Moraine" as shown on Figure 5. The course of Rt. 46 adjacent to the sand pit heads north through a spillway channel occupied by drainage from Stage III of Lake Pequest when ice in the Pequest Valley had receded into Sussex County, NJ.

Several deltaic sediment facies may be seen in the sand pit which include ice-proximal gravels, bottomset beds, and toeset or lower foreset beds. Ice-proximal gravels occur in the base of the section and were deposited prior to the progradation of finer-grained deltaic facies seen in the overlying units. A coarsening-upward deltaic assemblage overlies the gravels beginning with bottomset beds. Laminated, silty fine sands with fewer ripples than units above mark bottomset sediment which was mostly a suspended load. Many silt drapes may be seen in the bottomset beds. The overlying toeset or lower foreset strata



SYMBOLS

meltwater channels

outwash

Lake Pequest morphosequences
delta p1

delta p2

limit of Woodfordian glaciation

colluvium

end moraine

till

bedrock exposure

Figure 5. The location of STOP 3 on Rt. 46 between Great Meadows and Townsbury, NJ along the Pequest River in the Washington, NJ $7\frac{1}{2}$-minute quadrangle. The first deltaic morphosequences (p1 and p2) deposited in glacial Lake Pequest which was dammed by the "Terminal Moraine" at Townsbury are shown.

consist of fine to coarse sand dominated by climbing ripples. Other features seen in the exposure are ball and pillow structures and stagnant ice hollows and collapse features. The stagnant ice hollows contain very angular clasts dropped from melting ice blocks.

STOP 4: SPARTA ICE MARGIN, SPARTA, N.J.

This stop is located within the Nicole Heights housing development, Sparta (Fig. 1). The section exposed represents the proximal facies of an ice-contact delta built into glacial Lake Sparta, Stage II. The units exposed in the northeast section of the housing development, shown in Fig. 2, represent an ice-contact depositional environment. Units exposed are deltaic foresets, some of which are collapsed and faulted and debris flows which are generated from the ice-contact slope. The many large striated boulders piled next to the exposure also indicate close proximity to the ice margin.

Descriptions of Units Exposed in Section A-A' in Figure 2

Deltaic Foresets: alternating beds of fine to coarse sand and gravel, interbedded with coarse clast supported gravels. Individual beds range from 0.25 inches to 36 inches thick and commonly display grading and inverse grading. Silty clay beds less than 0.25 inches thick are occasionally found draped over an individual foreset bed. Foresets, where not collapsed, dip toward the south.

Debris Flow: matrix supported diamicton containing lens and stringers of sand and gravel. The matrix material is composed of fine to medium, poorly sorted sand and is loosely consolidated. The boundary between the debris flow and underlying sediment is erosional and represents a zone of shear produced by the debris flow as it moves down the foreset slope. This erosional contact is very evident in the section where the debris flow overlies a fine sand unit (colored black in the cross-section).

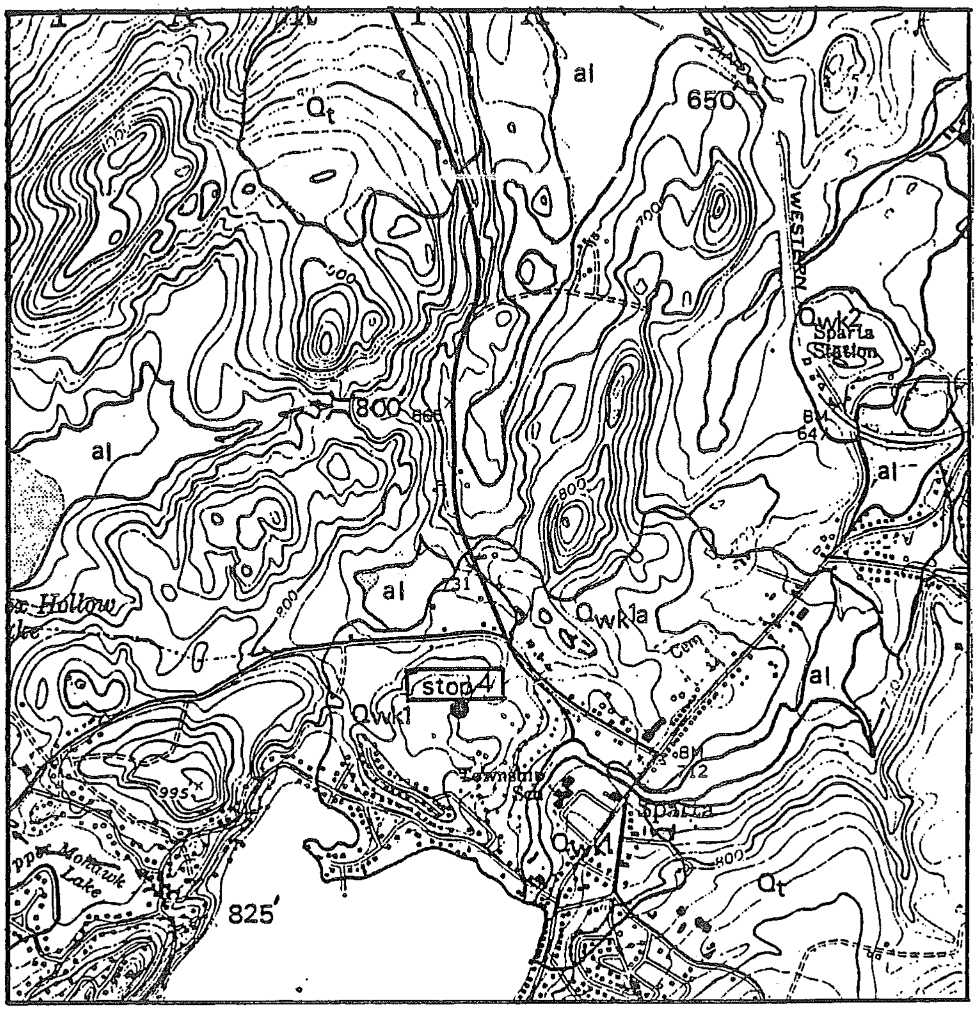


Figure 1. Surficial map of morphosequences near Sparta showing location of stop 4. Qwk1, Qwk1a and Qwk2 represent ice-contact lacustrine morphosequences. Qt represents till deposits in which bedrock outcrop is less than 5% of the surface area. Spillways with elevations are also shown (Newton east 7½ minute topographic Quadrangle).

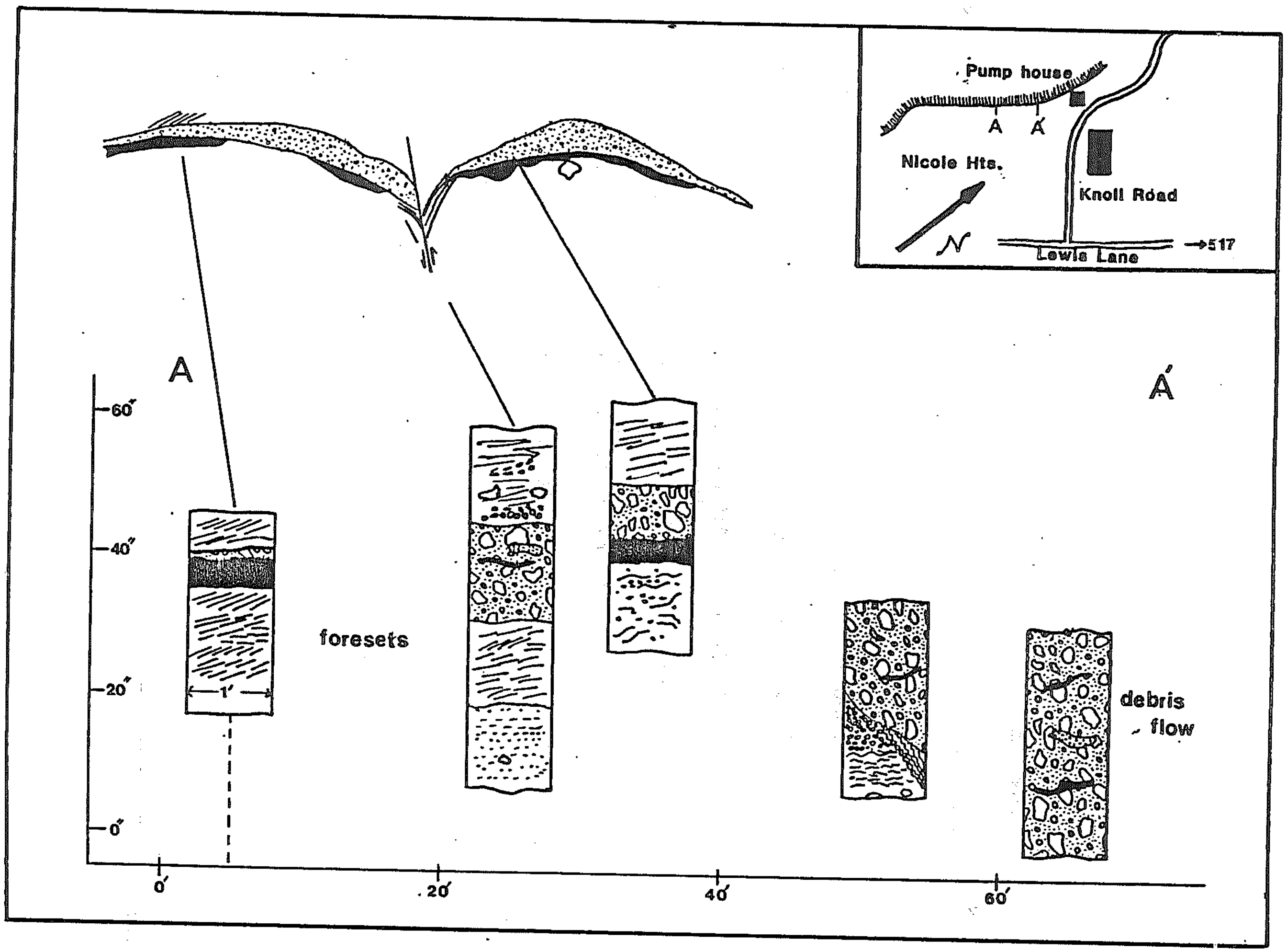


Figure 2. Stop 4, location of exposure at Nicole Heights, Sparta. Stratigraphic sections in A-A' represent collapsed deltaic foresets interbedded with a debris flow. The cross section shows a normal fault cross-cutting a debris flow. The underlying fine sand unit has been deformed and eroded by the overlying debris flow.

STOP 5: OVERVIEW OF MAPLEDALE FARM-SPARTA JUNCTION ICE MARGIN

Stop 5 is located on the western slope of the Wallkill Valley at the base of the Pimple Hills (Fig. 3). The Wallkill Valley is underlain by Cambro-Ordovician carbonates of the Leithsville and Allentown formations. The Sparta Mountains which form the ridge and hills to the east and the Pimple Hills to the west are underlain by pre-Cambrian gneiss. Interbedded within the gneiss, in the Pimple Hills, is the Franklin Marble, also pre-Cambrian in age.

Qwk3 represents two ice-contact deltas built into glacial Lake Sparta, Stage III. The housing development near the overview stop is built partially on Qwk3, as is the farm across the valley. Exposures throughout the deposit reveal deltaic sedimentation with an ice-contact facies similar to that seen at Sparta, Stop 4, exposed in a gravel pit at the Mapledale Farm.

Pebble lithologies (Fig. 4) reveal that each delta is composed of sediment derived from different source areas. Samples 149 and 150 reflect the lithology of the till found in the Pimple Hills (sample 22). Sample 136, found at the Mapledale Farm, reflects the lithology of the till embankment (samples 2 and 147), bordering Qwk3 on the east side of the Wallkill Valley. Deltaic foresets measured also indicate that sediment has been derived from lateral sources.

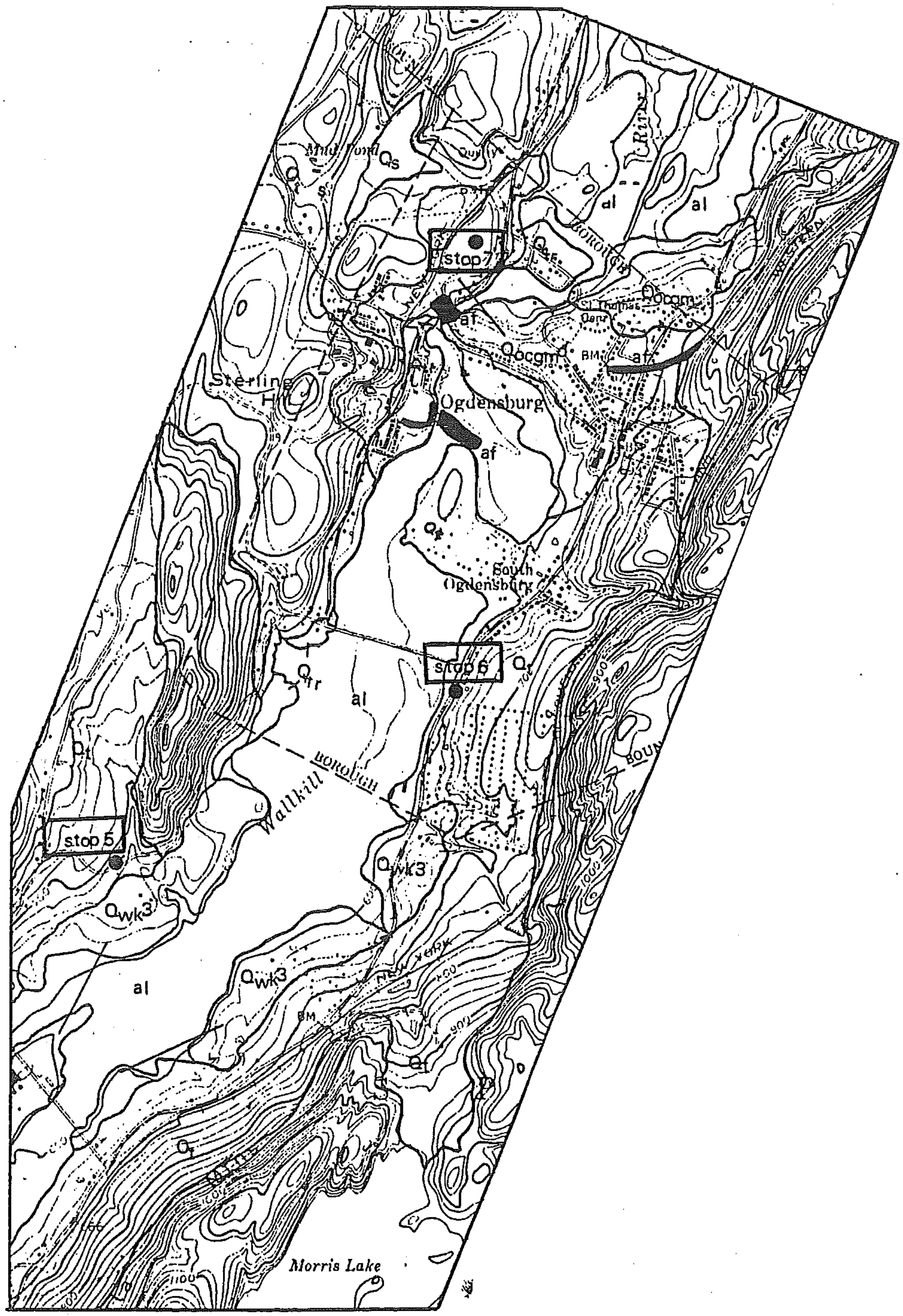


Figure 3. Surficial map of morphosequences in the Wallkill Valley showing location of stops 5, 6, and 7. Qwk3 and Qocgmd represent ice-contact lacustrine morphosequences. Qocgm represents the Ogdensburg moraine (Franklin 7½ minute topographic Quadrangle, reduced by 25%).

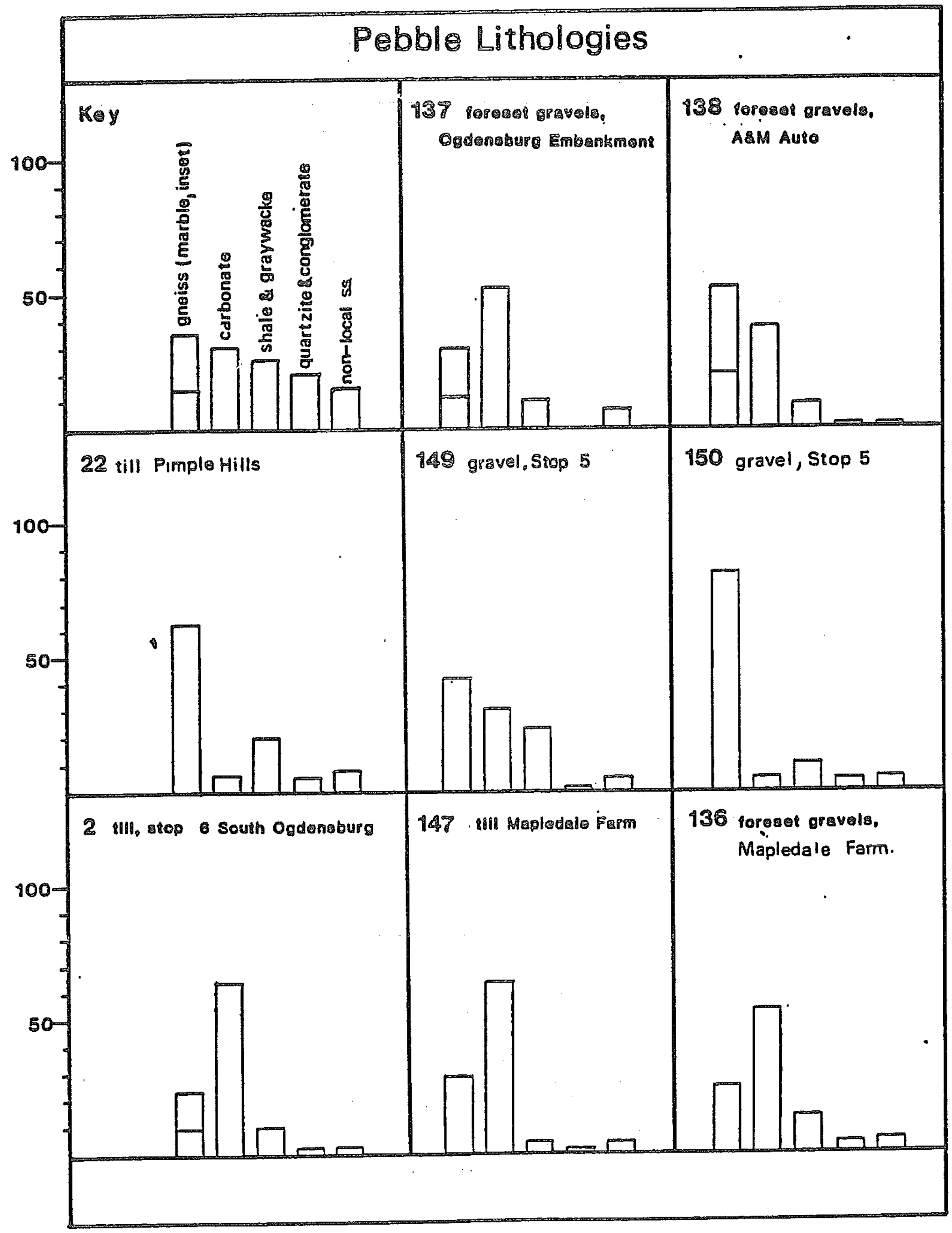


Figure 4. Provenance data based on the lithologies of 125 pebbles at each sample site. Vertical scale is in %.

STOP 6: WOODFORDIAN LODGEMENT TILL: 1 mile south of Ogdensburg, Route 517

The section of till exposed is a lodgement till and is part of a very thick till ramp deposited on the eastern slope of the Wallkill Valley (Fig. 3). Maximum thickness of the till determined from well records is 100 feet. There are some stratified sands and gravels at the southern end of the exposure. However, information by owner has indicated that only till was removed from the pit.

Till description - The color of the till ranges from 2.5-5 Y 5/3. The matrix is composed of very fine sand with some silt. Rock fragments are common within the matrix. Calcium carbonate content ranges from moderate to extreme, which results in cementation of the till. Consolidation where not affected by cementation is moderate. Striated dolostones and sandstones are found throughout the exposure.

Provenance of the till, samples 2 and 147 in figure 4, strongly reflects the carbonate lithology of the Wallkill Valley as well as the Franklin Marble of the Pimple Hills. The above provenance data suggests that ice flow was oblique to the strike of the Wallkill Valley (S35°W) in a more southernly direction. The occurrence of a magnetite boulder found on the Mapledale Farm, has been traced from a mine near Franklin and indicates an ice flow direction of S10°W. The distribution of till on the east slope of the Wallkill valley (Qt in figure 3) also seems to indicate ice flow across the Wallkill and not along its strike. Till fabric (figure 5) and striations (S28°W) found one mile northeast of Ogdensburg indicate that ice flow was parallel to the strike of the Wallkill Valley. The above data and observations

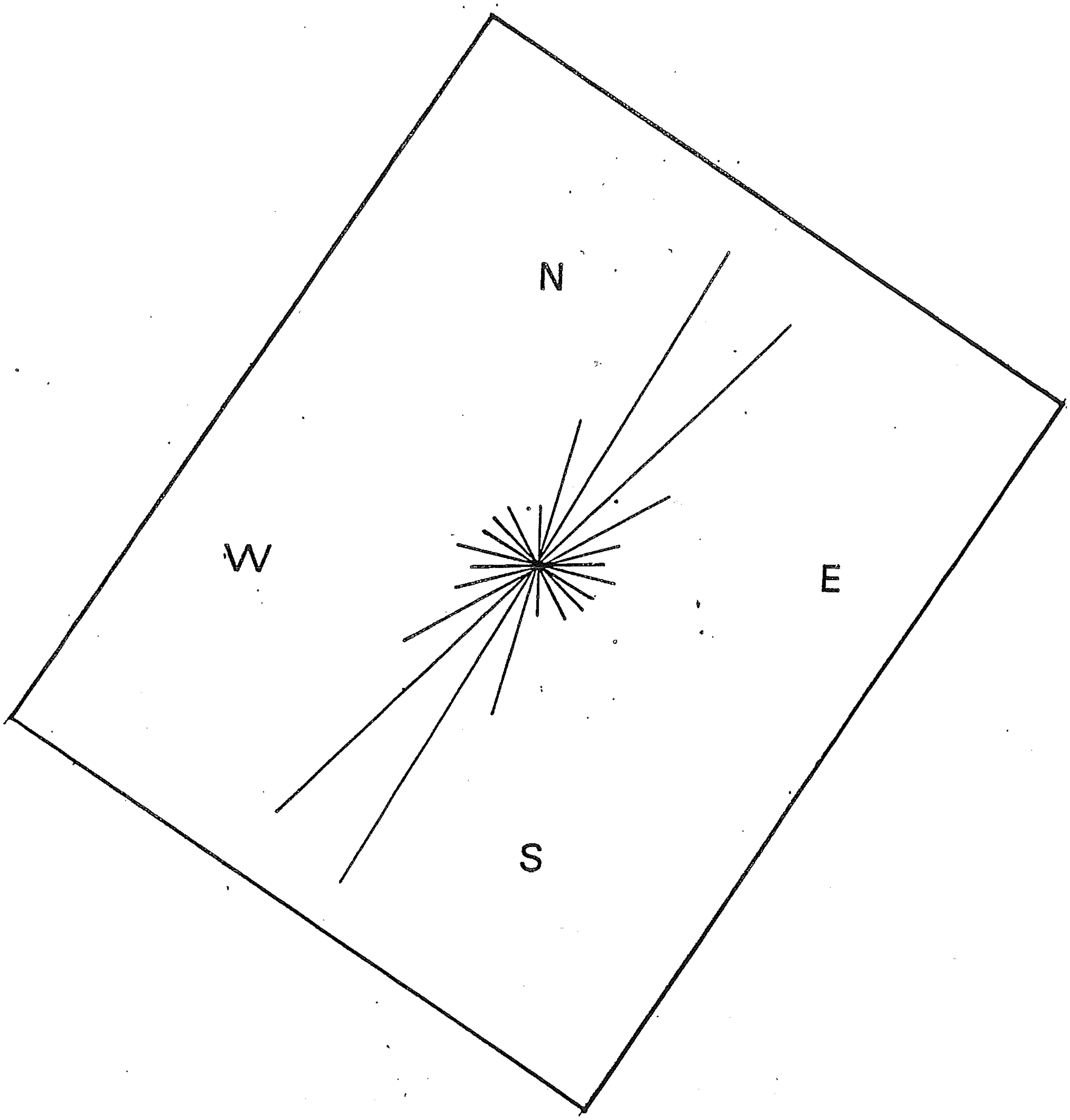


Figure 5. Fabric of lodgement till, stop 6.

indicate that (1) ice flowing into the Wallkill Valley was moving in a direction almost due south, and (2) ice in the Wallkill Valley was constrained by the small valley width and flowed parallel to the strike of the valley. The above model accounts for the provenance of the till and its distribution as ice flow was across bedrock formations northwest of the Wallkill and accounts for ice flow within the valley as ice became channeled within the narrow confines of the Wallkill Valley.

STOP 7: OGDENSBURG-CULVERS GAP MORAINE, OGDENSBURG

Stop 7 is located in a gravel pit at A&M Auto, Ogdensburg, Figure 3. The exposure is a three-dimensional view of an ice-contact delta. Figure 6 represents the floor plan of the pit as well as several measured cross-sections and stratigraphic sections. The general stratigraphy of the pit is that of lower deltaic foresets overlain by middle and upper deltaic foresets. Topset gravels are also exposed in several sections. Interbedded within the foreset units are debris flows.

The following is a description of units found in the exposure (Fig. 6).

Lower Delta Foresets: rhythmically bedded, planar to ripple laminated, very fine sand, interbedded with clayey silt laminae. Convolutions, faulting and injection features are found throughout these units which indicate overloading by the coarser gravel foresets above. Drop stones are commonly found throughout the unit.

Middle and Upper Delta Foresets: Alternating beds of fine to coarse sand and gravel interbedded with coarse open-framework gravels. Individual beds range from 0.5 to 36 inches thick. Upper deltaic foresets are composed of sand to gravel-sized material and have an average dip of 25°. Middle deltaic foresets are composed of fine sand and dip less than 10°. Commonly, the middle deltaic foresets exhibit climbing ripple sequences.

Topset Gravels: poorly sorted cobble gravel, bedding is indistinct.

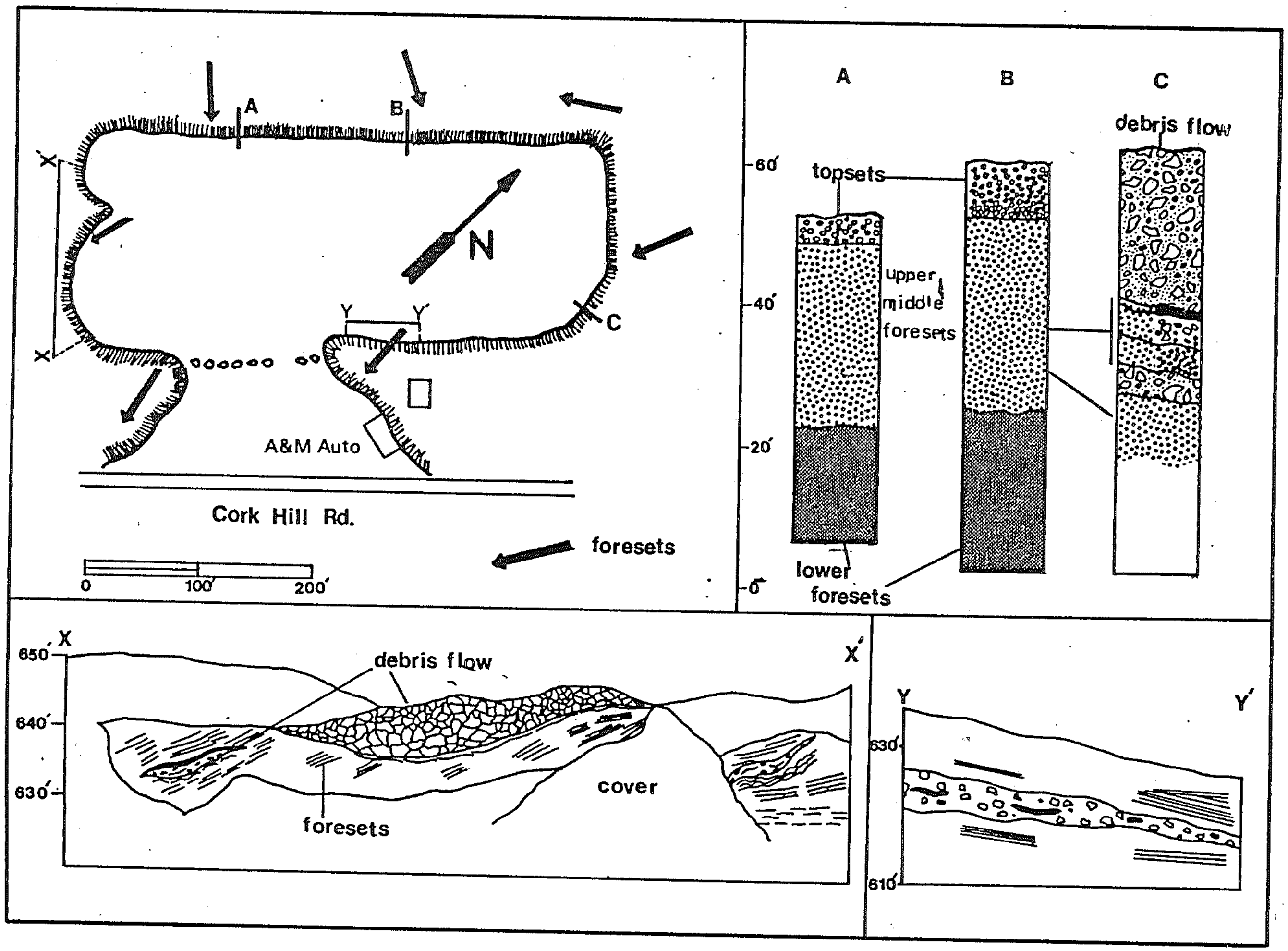


Figure 6. Floor plan of the gravel pit at A&M Auto, Ogdensburg. The units in stratigraphic sections A, B and C are discussed in the text. Cross sections X-X' and Y-Y' represent debris flows interbedded within deltaic foresets.

Debris Flows: matrix supported diamicton containing lens and stringers of sand and gravel. Matrix material is composed of fine to medium, poorly sorted sand and is loosely consolidated. Foreset units underlying the debris flow exhibit faulting and folding produced by shearing of the over-riding debris flow. Rip-up clasts are commonly found within the debris flows. An exception to the above description is the debris flow in cross-section X-X'. It is clast supported and contains very little matrix. The matrix seems to have been washed out of the deposit during its emplacement.

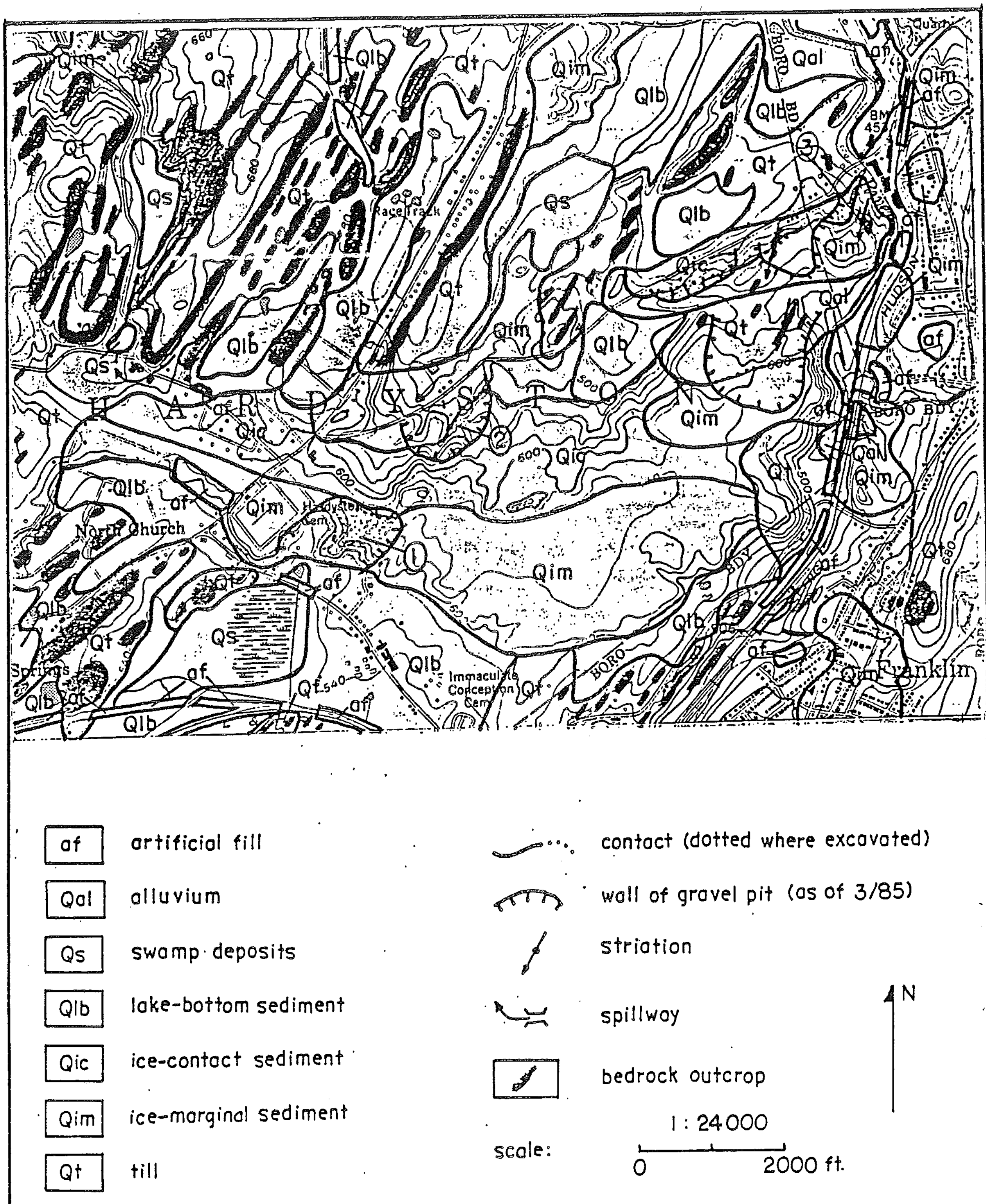
Deltaic foresets measured throughout the exposure (Fig. 6) indicate two fans building into the lake and a sediment source from the Pimple Hills. Foreset directions in the Ogdensburg Embankment dip south to southwest and represent a sediment source from the east side of the valley. Pebble lithologies (Fig. 4) also reflect different sediment source areas. Samples 138 and 139 represent a Pimple Hills source; sample 137 represents sediment derived from the erosion of till on the east valley slope.

STOP 8: DISTAL SIDE - NORTH CHURCH DELTA/AUGUSTA ICE MARGIN

The North Church delta is a segment of the Augusta ice margin. It was deposited in a proglacial lake that drained over the Germany Flats outwash surface near White Lake and into the Paulins Kill basin (refer to figure 2 in Stanford and Harper, this volume). This was an unstable outlet around an ice block, and three erosional channels at successively lower elevations from approximately 620 feet to 600 feet have been identified here by Witte (personal communication).

The North Church delta (refer to figure 1) may be a composite delta reflecting the lowering spillway elevation. The broad, southernmost portion of the delta (marked by the 640 contour on the map) is a separated from a lower, more northerly surface on the flanks of the delta by a distinct fifteen- to twenty-foot scarp on both the east and west sides. This difference in elevation may indicate an earlier, higher segment of the delta deposited at a more southerly ice front and graded to an earlier, higher outlet; and a later, lower segment of the delta deposited at a more northerly ice front and graded to a later, lower outlet.

Pebble counts in the delta (figure 2) show an interesting pattern. Pebbles in the topset and foreset beds (termed "ice-marginal sediment" in figures 1 and 2) show much more carbonate than those in the poorly-sorted gravel and sediment-flow deposits on the



SURFICIAL GEOLOGY OF THE NORTH CHURCH DELTA
HAMBURG QUADRANGLE
NEW JERSEY

Figure 1: North Church Delta

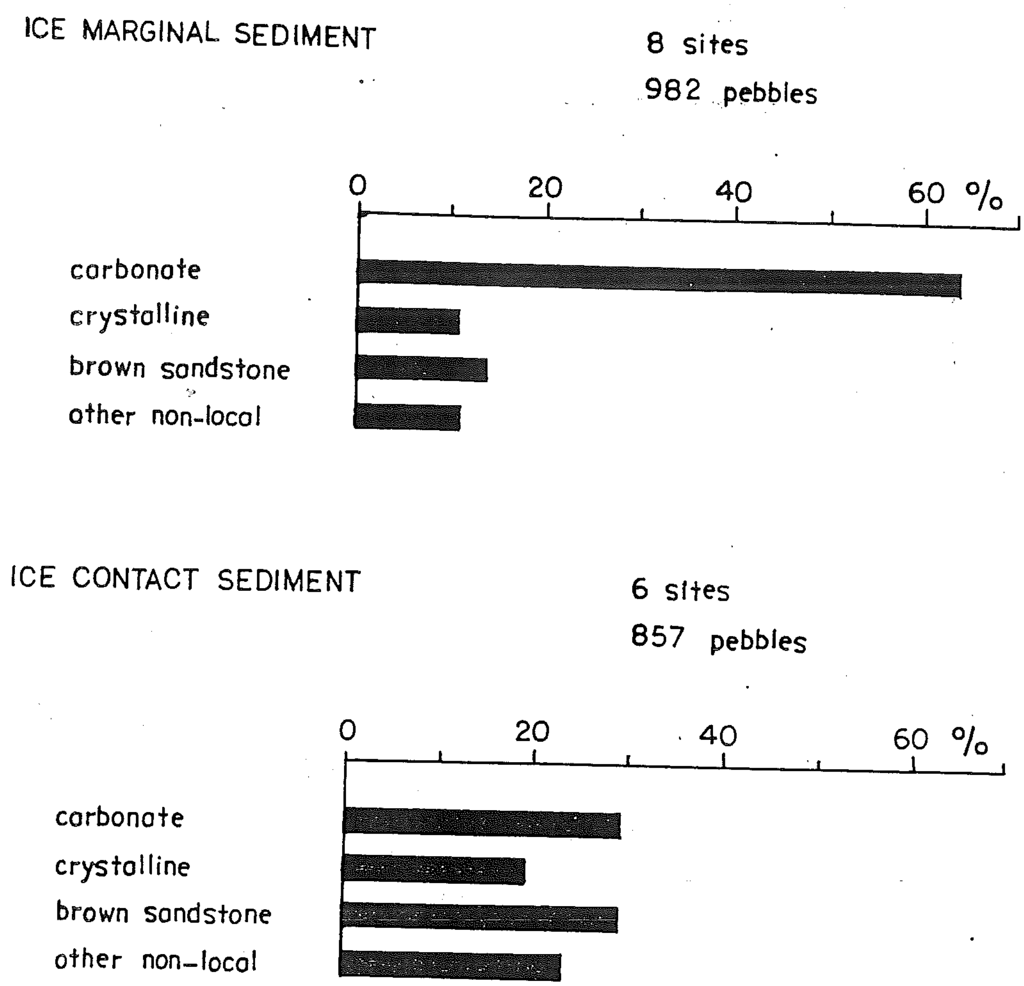


Figure 2: Pebble Lithologies in the North Church Delta

ice-contact side of the delta (termed "ice-contact sediment" in figures 1 and 2). One interpretation of this pattern is that the topset and foreset beds were fed by subglacial channels that eroded the local carbonate bedrock and carbonate-rich local till. The ice-contact sediment, on the other hand, contains a much greater percentage of material sliding from the surface and upper portion of the ice and thus reflects a more regional pebble lithology.

The North Church sand and gravel pit (indicated by "1" on figure 1) is dug through almost the entire length of the earlier, higher portion of the delta. From the south (distal) end of the pit fine sand bottomsets lead upward and northward into sand and pebble gravel foresets with minor beds of cobble gravel. The foresets may be overlain by sand and pebble gravel topsets, but topsets are not clearly exposed. At the north (proximal) end of the pit a fifteen-foot thick clast-supported pebble and cobble diamicton overlies sand foresets and seems to be the surface unit on the ice-contact slope to the north. In the pit, this unit rises and pinches out within a short distance to the south. It may be collapsed or colluviated topset and foreset material, sediment-flow deposits from the ice surface, or a coarse, massive, closed-work gravel deposited by meltwater feeding the delta.

STOP 9: PROXIMAL (ice-contact) SIDE - NORTH CHURCH DELTA/AUGUSTA ICE MARGIN

This pit ("2" on figure 1) is excavated in the ice-contact slope of the delta. Dipping packages of interlayered sand, pebble gravel, and matrix-supported fine sand diamicton containing pebbles and cobbles are exposed. Dips and strikes of these packages are variable, but some small-scale reverse faulting is visible, suggesting that collapse has disrupted the original structure. The sand units show recumbent folds and load structures in places, perhaps due to dragging and loading from sediment flows represented by the diamicton units, or from dragging and loading by ice. Note also the decreased percentage of carbonate lithologies as compared to the lithologies in the foresets at the last stop.

The deposits exposed may represent interlayering of sediment flows from the ice front and sand and gravel delivered by subglacial or englacial channels as the ice retreated northward from the delta. Alternatively, the sediments may represent subglacial stacking of till and meltwater-deposited sand and gravel while the delta was forming. The layering and structure of the units, however, seems to rule out colluviation or collapse of the ice-contact face as possible origins for these sediments.

STOP 10: STRIATED, POLISHED DOLOMITE

Dolomite bedrock of the Allentown Formation of Cambrian age is exposed in a former gravel pit ("3" on figure 1). Before excavation the bedrock was overlain by cobble and pebble gravel foresets of a large sublacustrine fan deposited in the "590" lake (refer to figure 3 in Stanford and Harper, this volume). Till atop the bedrock is absent here, but is present just to the east and south where the bedrock surface drops off 40 to 80 feet.

The bedrock surface is polished and striated. Roches moutonees, crag-and-tail features, and crescentic marks are all well-displayed. Ice flow directions range from S20W to S30W and average approximately S25W. The attitude of the bedrock here is N45E;47NW, indicating that ice flow at this location is not strictly controlled by bedrock structure. The ice flow direction here and at numerous other locations in the valley bottoms indicates channeling of ice within valleys during formation of striations.

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