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The Wisconsinan Stage of the First Geological District, Eastern New York

Donald H. Cadwell, Editor

Bulletin Number 455

New York State Museum

The University of the State of New York
THE STATE EDUCATION DEPARTMENT

Albany, N.Y. 12230
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INTRODUCTION

Donald H. Cadwell
State Geological Survey
New York State Museum
The State Education Department

This volume contains papers presented in a symposium held during the annual meeting of the Northeastern Section, Geological Society of America, at the Concord Hotel, March 24, 1983. These papers describe the glacial geology of certain areas within the First Geological District of 1837. The volume is to appear during the Sesquicentennial Celebration of the founding of the New York State Geological Survey; hence the title.

The Geological Survey of New York was established in 1836 by Governor William L. Marcy. The State was partitioned into four districts, subsequently redefined in 1837, each with a State Geologist (Fig. 1). The mapping of these four districts led to the publication of the first geologic map of New York in 1844.

The current surficial geologic mapping program is one of the major projects of the New York State Geological Survey and will result in the publication of surficial data at a scale of 1:250,000. This concentrated mapping effort has brought to light at least three major problems associated with the interpretation and publication of surficial data. First, there is a need for standardization of terms and mapping units. Second, there is a need for the definition of ice margin positions and the associated environments of deposition. Third, correlation of ice margins cannot be done adequately without at least reconnaissance surficial mapping. The need to solve these problems and summarize the status of research led to the organization of a symposium entitled "The Pleistocene Time-, Rock-, and Morpho- Stratigraphy of Eastern New York" at the 1983 Northeastern Section Meeting of the Geological Society of America held at Kiamesha Lake, New York. We are pleased that ten of the eleven papers given at the symposium appear in this volume.

The papers included in this volume discuss the style of Wisconsinan deglaciation between the Terminal Moraine in New York and New Jersey and the Mohawk Valley in New York. The general study area of each paper is indicated in Figure 2. The east-west concentration of research across the Hudson Valley and the Catskill Mountains has enabled tentative correlations of ice margins. Table 1 summarizes the ice margin positions presented by the authors in this volume; locations of the major retreatal positions are indicated in Figure 3.

The first paper in this volume, by L. Sirkin, is a revision of the surficial geology of Long Island and suggests these deposits are the result of two Wisconsinan glaciations with lobate fronts. The two drifts are distinct and separated by fluvial, marine and lacustrine sediments deposited during the Portwashingtonian (mid-Wisconsinan) warm interval. In the second paper, a discussion of the Woodfordian Terminal Moraine in the Great Valley of New Jersey, J. Cotter and others suggest valley constrained sublobes of the Ontario and Hudson-Champlain glacial lobes. These glacial lobes deposited correlatable heads of outwash and end moraines during retreat of the active ice margin. All of the remaining papers discuss events associated with retreat of the Woodfordian Ice Sheet. Recessional events in the Mid-Hudson Valley, discussed by G. G. Connally and L. Sirkin, define ice margins in association with a gradually expanding Glacial Lake Albany, north of the Hudson Highlands. D. H. Cadwell describes the deglaciation history of the Catskill Mountain region and delineates ice margins in the Schoharie Valley. Identification of ice margins in the Hudson Valley between Hyde Park and Albany is made by R. J. Dineen

D. Ozsvath and D. R. Coates describe the glacial history of the western Catskill Mountains, with ice margins along the West Branch of the Delaware River. Topographic control of retreatal positions in the upper Susquehanna River drainage basin is the topic of a paper by P. J. Fleisher. E. H. Muller and others, in their discussion of the western Mohawk Valley, suggest that the Mohawk and Ontario glacial lobes attained their maximum extent at different times. G. C. Kelley and W. B. Thompson present the deglaciation of the eastern

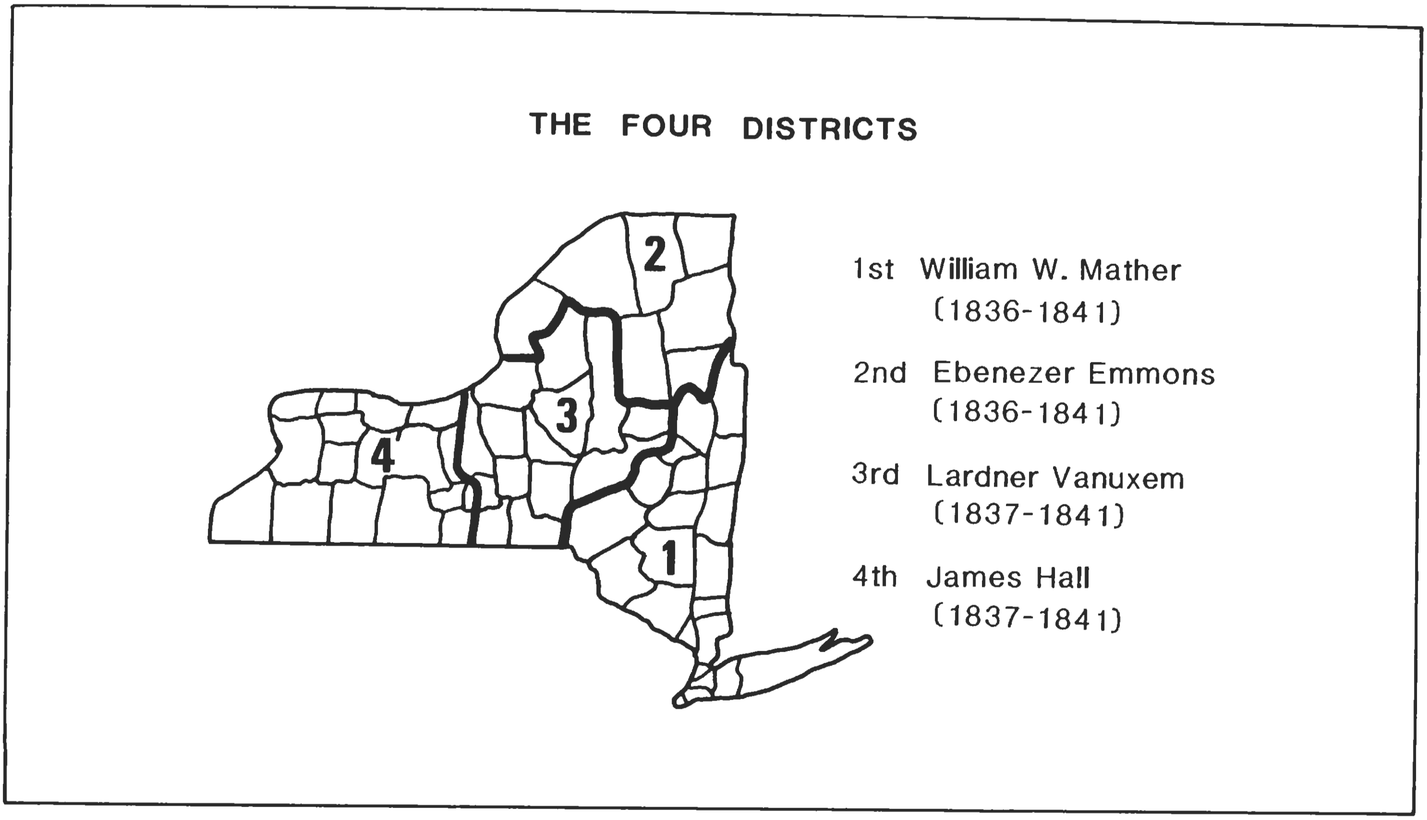

Figure 1. The Geological Districts of New York State, as defined in 1837, each with a State Geologist.

edge of the Hudson-Champlain glacial lobe in western Connecticut. The final paper, by C. R. Warren and B. D. Stone, suggests a chronology for western Massachusetts, with the development of proglacial lakes in association with stagnant zone retreat of the glacier margin.

I gratefully acknowledge the thorough reviews of this entire volume by R. P. Goldthwait and D. E. Lawson. In addition, each manuscript was reviewed by the authors of the other papers. The editor expresses his appreciation for the cooperation of all contributing authors.

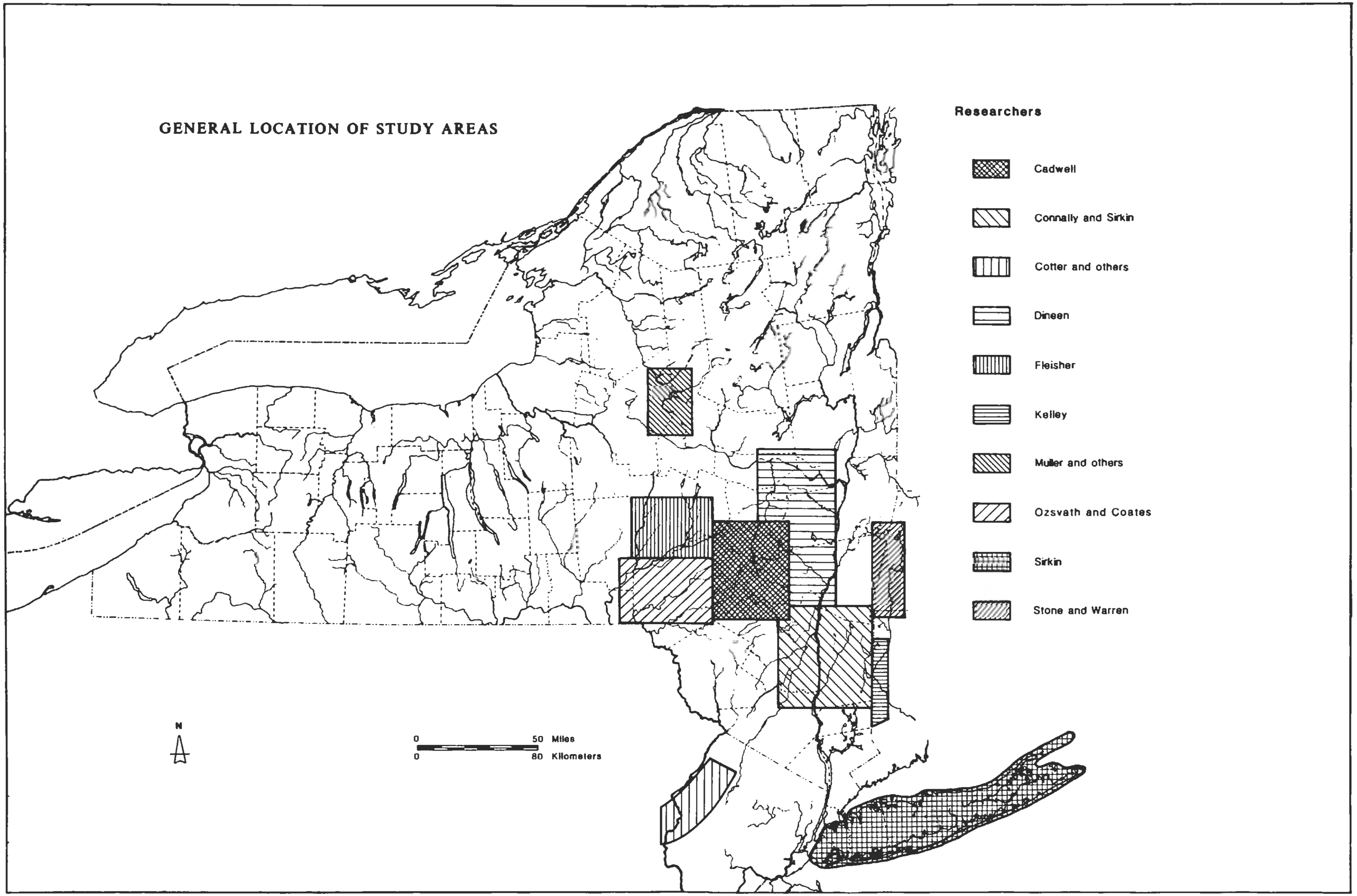

Figure 2. General location of study areas of contributing authors.

SUMMARY CORRELATION OF ICE MARGIN POSITIONS

West | East

yBP*	MULLER	OZSVATH	FLEISHER	CADWELL	DINEEN	CONNALLY	COTTER	SIRKIN	KELLEY	STONE	yBP*
13,000	Rome										13,000
	Holland Patent										
	Norway										
	Hawthorne				Delmar						
16,000											16,000
	West Canada		Middleburg	Middleburg	Middleburg					Western Massachusetts	
			Cassville-Cooperstown								
			New Berlin	North Blenheim Haines Falls	Rosendale 2						
				Grand Gorge	Rosendale 1	Rosendale					
						Pine Plains					
						Hyde Park			Sharon		
17,000		Delhi	Oneonta	Wagon Wheel Gap	Wagon Wheel Gap	Wallkill			Salisbury		17,000
		Sidney	Wells Bridge			Pellets Island-Shenandoah			Gaylordsville		
						Sussex					
						Augusta					
						Culvers Gap	Culvers Gap				
21,000							Terminal Moraine	Harbor Hill Ronkonkoma			21,000

*years Before Present

Table 1. Summary correlation of ice margin positions of contributing authors.

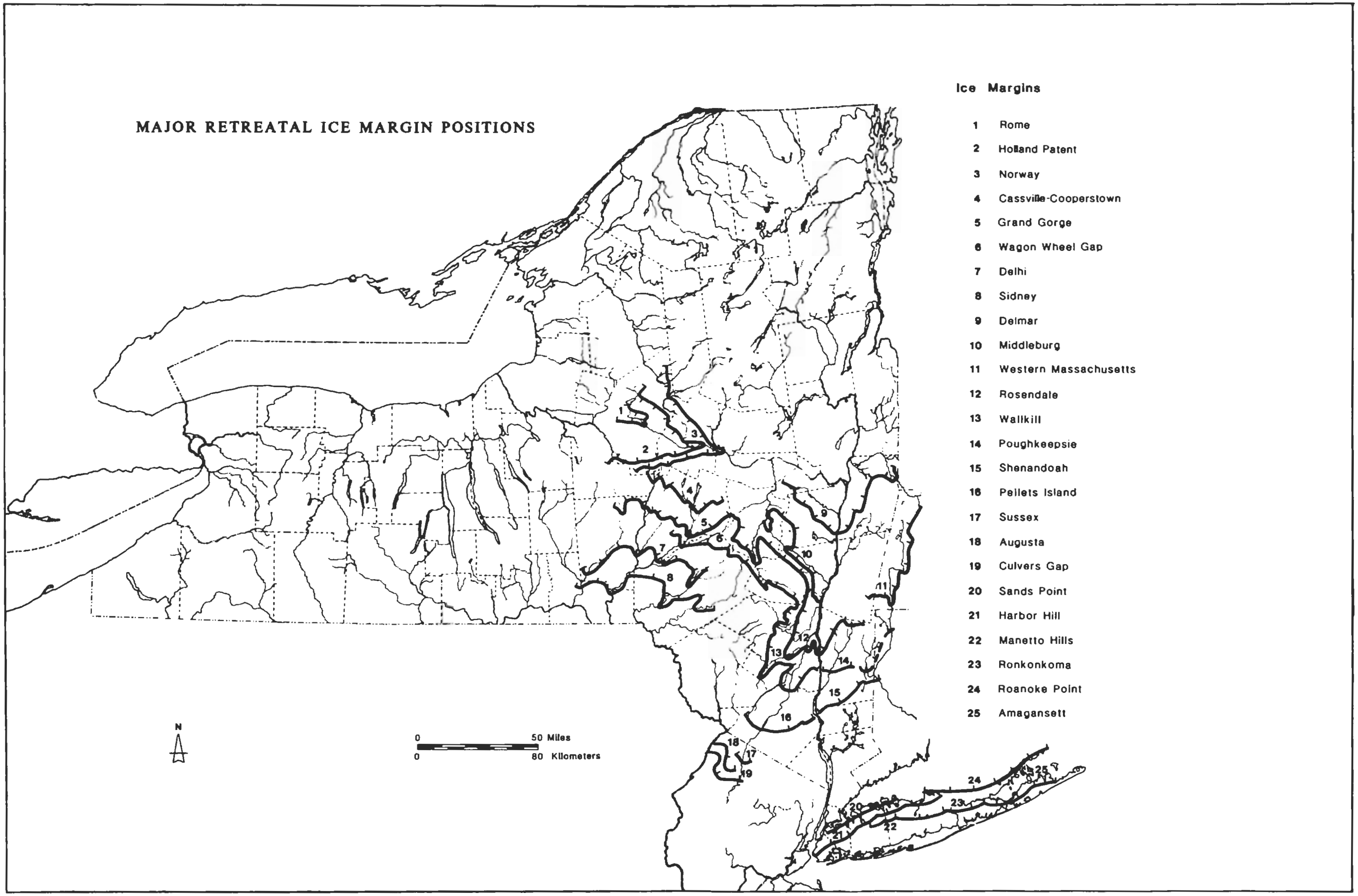

Figure 3. Location of major retreatal ice margin positions of contributing authors.

PLEISTOCENE STRATIGRAPHY OF LONG ISLAND, NEW YORK

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ABSTRACT

The Pleistocene stratigraphy of Long Island has evolved slowly from Fuller's comprehensive work and his adherence to the classical, mid-North American plan of multiple glaciations. Recent studies now assign the Pleistocene deposits to several glacial lobes of two Wisconsinan glaciations, as well as to one interstadial, deglaciation during the late Woodfordian, and possibly one interglacial. As a result, several of Fuller's stratigraphic units are believed to be invalid.

Differentiation of two drift sheets was facilitated by the presence of marine deposits dated in the 42,000 to 21,750 yrs BP range. The older glacial deposits have been placed in the Altonian Substage of the Wisconsinan Stage; the younger drift in the Woodfordian. Altonian moraines are recognized by increasing relief on the Altonian surface, even though the Altonian has been deformed by the last ice advance. Prominent Altonian highs are found in Port Washington, Port Jefferson, and the South Fork of Long Island. The deposits generally consist of sandy till over outwash. During the mid-Wisconsinan, Port Washingtonian warm interval, climate and sea level were similar to the present.

The Woodfordian glacier reached Long Island around 21,750 yrs BP and recession from the end moraines took place shortly thereafter. Three glacial lobes controlled ice margins on Long Island – the Hudson Lobe in the west, the Connecticut Lobe in central Long Island, and the eastern Connecticut-western Rhode Island Lobe in the east. Several recessional positions and morainal envelopes are documented for each of the glacial lobes and appropriate terminology established for stratigraphic units and ice margins.

INTRODUCTION

The Pleistocene stratigraphy of Long Island has been the subject of varying interpretations for most of this century. While the early workers in this field, up to and including Fuller (1914), sought to accommodate the stratigraphic framework of the mid-continental Pleistocene to the unconsolidated deposits, or drift, of Long Island, the main issues today deal with the development of the lobate ice fronts during two glaciations, the timing of the emplacement of the moraines and the recession of the last glacier, the age relationship between the two drift sheets and an intervening interstadial unit, and the validity of a pre-Wisconsinan interglacial unit in the Long Island Pleistocene sequence.

The early history of geologic study on Long Island is outlined in Fuller (1914) and depicts the gradual emergence of the glacial theory as an explanation of the origin of the deposits that overlie the Cretaceous strata. Fuller's interpretations were derived from the classical works in New England, for example, Woodworth (1901), in which up to five glacial stages were recognized between Long Island, Block Island, Martha's Vineyard, Cape Cod, and Nantucket. Early workers also recognized the inner and outer moraines of Long Island and the continuity of the moraines across the southern New England islands.

It is interesting to note that an alternative interpretation appeared at about the same time. In the United States Geological Survey's New York City Folio Number 3, Merrill (1902) concluded that only two glacial advances had occurred in this region. However, in his field studies on Long Island in the early 1900's, Fuller was able to locate deposits, both existing and conjectured, in the surface exposures and in subsurface well data that he could fit into each of the postulated glaciations. He also found erosion surfaces and marine or other clays that represented the interglacial events (Figure 1).

The result was a profound layer cake that could also accommodate the Cretaceous, Tertiary, and the Pleistocene of the coastal plain south of Long Island. Each change in lithology or even in grain size was taken to represent a stratigraphic unit or a time unit that had been described elsewhere, and a correlation with a similar horizon found in eastern or central North America. In the absence of both a biostratigraphic approach and absolute dating methods, the results of Fuller's efforts on Long Island were quite thorough for the time.

Central to the stratigraphy and to the validity of his rock units is the presence in the subsurface and, according to Fuller (1914) in many exposures, of such units as Cretaceous clays and sands, at least two weathered gravel deposits of Pleistocene age, a fossiliferous marine clay and sand sequence, and a gravel-till-gravel formation of glacial origin (Figure 1). In the perspective of more recent studies that have had the benefit of biostratigraphy and radiometric dating, it is essential to test the validity of Fuller's (1914) stratigraphy. It is also important to consider an updated stratigraphy based on the two glaciation-several glacial lobe model and the more recently described rock units.

Wisconsin	(G)	Harbor Hill Moraine Ronkonkoma Moraine
Sangamon	(I)	Vineyard erosion surface
Illinoian	(G)	Manhasset Formation
Yarmouth	(I)	Jacob Sand Gardiners Clay
Kansan	(G)	Jameco Gravel
Aftonian	(I)	erosion surface
Pre-Kansan	(G)	Manetto Gravel

G-Glacial I-Interglacial

Figure 1. Pleistocene stratigraphy of Long Island, after Fuller, 1914.

COMMENTS ON FULLER'S STRATIGRAPHY

In Fuller's time, the moraines that form the upland in the northern half of Long Island were seen as the terminal positions of two glaciations. Coarse-grained deposits that mantle the north-south trending hills in west-central Long Island presumably represented early Pleistocene glaciations. His work became the definitive source of a morpho-stratigraphy for southern New England, as in Woodworth and Wigglesworth (1934) and Kaye (1964), and it has been incorporated with little question into many subsequent studies. For example, the glacial map of eastern North America, compiled by Flint and others (1959), depicts a simplistic two moraine system between Long Island and Nantucket and along the southeastern New England coast. As such these moraines represented only the younger surficial deposits of the "Wisconsin" age.

Manetto Gravel

According to Fuller (1914), the Manetto Gravel represents outwash deposited in front of an early Pleistocene, pre-Kansan glacier. This weathered and cross-bedded, granite and quartz pebble gravel was found in the crest and distal slope of the Manetto Hills in central Long Island. The gravel unit was more recently exposed in the West Hills-Manetto Hills region where over the years a variety of coarse to fine glaciofluvial and glaciolacustrine facies have emerged, rather than a single gravel unit. Furthermore, the stratified drift generally is covered by or intercalated with varying thicknesses of sandy melt-out till. While of glacial affinity, the gravel is more likely a coarse facies of the outwash and deltaic sequences that formed in a zone of meltwater discharge during the last glaciation. It is not a Tertiary gravel, as suggested by Lubke (1964).

In light of the two-glaciation hypothesis, two alternative interpretations of the age of the Manetto Gravel may be postulated. First, field studies show that a yellow-stained, quartz and rock pebble gravel does mantle the margins of the interlobate zones. In Manetto Hills the gravel surrounds collapsed deposits of glaciofluvial and glaciolacustrine origin. A second gravel, a polymict gravel comprised of cobbles of glacial origin lies in the outwash plains adjacent to the hills of the interlobate deposits and the yellow, pebble gravel. Between the High Hill Interlobate Moraine in the West Hills (geographic) area and the interlobate zone to the south, one kame-like hill has a deep (2.0 $\pm$ m) red (2.5 YR 5/8) soil developed on it. In the Half Hollow Hills, the southern portion of the Dix Hills interlobate zone, yellow pebble gravels surround brown, sandy till and local clay lenses.

The yellow-stained pebble gravels, the till and the red soil-capped outwash might be interpreted as remnants of the early Wisconsinan (Altonian) glaciation. In this model the Woodfordian interlobate zones were formed over and around these older glacial deposits. In this case, the Manetto Gravel could be an outwash of Altonian age. However, the south-trending Manetto Hills and Dix Hills are now mapped as part of an interlobate zone of Woodfordian age (Sirkin, 1982). Apparently, meltwater was dammed by lobes of glacial ice to form proglacial lakes into which deltas were deposited. The deltas were engulfed by later outwash deposits, while meltwater deposited outwash adjacent to the interlobate deposits and in the valleys.

Pollen analysis of bog sections on the surface of the Harbor Hill, Manetto Hills, and Ronkonkoma moraines reveals only postglacial pollen zones (Sirkin, 1967, 1971), precluding a greater age for the basal surface on which the bogs formed. Furthermore, a bog and lacustrine sequence buried beneath outwash on the southwestern base of Manetto Hills yielded radiocarbon dates and a pollen microflora corresponding to the Portwashingtonian Oak Pollen Zone (Mesticky, 1977; Sirkin and Stuckenrath, 1980; Sirkin, 1982). These deposits represent a pre-Woodfordian lake sequence buried by the outwash deposits of the Woodfordian.

These constraints limit the age of the interlobate deposits, including the yellow pebble gravels, to the Woodfordian Substage of the Wisconsinan. The weathered gravel probably was derived from weathered bedrock in the Connecticut upland to the north or perhaps in part from the Cretaceous. The deep, red soil may be a locally altered product of the brown till that caps the nearby Manetto Hills Interlobate Moraine.

Jameco Gravel

Fuller (1914) found that another weathered gravel, the Jameco Gravel, generally lay beneath the Gardiners Clay (Figure 1). He indicated that the unit was glacial or glaciofluvial and represented outwash from an ice front located somewhere in Long Island Sound. Unlike the Manetto, the Jameco Gravel does not crop out at the surface, but usually is found in deep wells and is described as a very coarse, dark-colored gravel at depths between 55 and 284 feet below the surface (after Veach, as cited in Fuller, 1914).

The unit was reassessed by Soren (1978) who supported Fuller's (1914) description and suggested that the Jameco could be Kansan or Illinoian in age. Soren (1978) concluded that the Jameco filled a buried valley cut into the underlying Cretaceous and that it extended southward toward Jamaica Bay at depths from 80 to 200 feet below sea level. Perlmutter and Geraghty (1963) determined that the Jameco was as much as 180 feet thick with between 50 and 75 feet of relief. They believed that it represented "pre-Wisconsin" outwash.

A major difficulty with the interpretation of this unit is that its age relies heavily on the age of the Gardiners Clay, which has not been dated. Depending on the outcome of radiometric analysis of the various marine sediments referred to as Gardiners Clay, the gravels cannot be restricted in age, other than older than the overlying clay. This leaves several options including ages of pre-Wisconsinan, mid-Wisconsinan, or even Woodfordian. The deposit could be late Wisconsinan if the channels in which it is found were cut as the ice advanced, and the deposits were graded from the ice front to the low Woodfordian sea level. The overlying marine clay could then be the transgressive marine sequence of the late Wisconsinan, similar to the marine sediment deep in the Hudson River channel that was analyzed by Weiss (1971). It also is possible that the unit is mid-Wisconsinan in age and can be correlated with the well-dated Portwashingtonian beds.

Gardiners Clay, Jacob Sand

The Gardiners Clay in the cliff sections on Gardiners Island and Long Island is a marine clay interbedded with sand and fine silty sand that grades upward into fine sand, the Jacob Sand, as described by Fuller (1914). Both units, however, are ice-shoved. Fuller placed the Gardiners Clay in the "Yarmouth" Interglacial and the Jacob Sand in the Illinoian, even though they are conformable units. He also indicated that the Gardiners could vary in color from blue to red or gray and that it could be found as a sand unit bounded by clays, as long as it was in gradational contact with the Jameco Gravel, below. The type section also contained a few marine fossils.

The Gardiners Clay was correlated by Fuller (1914) with the dark clays of Block Island, Nantucket, and other areas of New England, some of which are lacustrine. Thus, the Gardiners was originally defined from a surface exposure and thought to be stratigraphically comformable with the under- and overlying rock units. Subsequently, MacClintock and Richards (1936) placed the Gardiners, in part, in the Sangamon Interglacial and correlated it with the Cape May Formation of New Jersey.

The Jacob Sand coarsens upward into sands and gravels of the overlying "Manhasset Formation." It contains a marine fauna of cool to cold water affinity, similar to conditions found today north of Cape Cod. The Jacob Sand also displays some deformation in proximity to the upper contact. The cold water faunas of the Gardiners Clay and the Jacob Sand from Gardiners Island were believed by Gustavson (1976) to represent a Wisconsinan interstadial rather than the Sangamonian.

Gustavson also discovered a warm-water fauna in an ice-shoved mass of sediment in the moraine at Bridgehampton, New York, and he correlated these sediments with the marine beds that generally are found in the subsurface in central Long Island. This latter unit, which also is called the Gardiners Clay, underlies the southern half of central Long Island between 2 and 200 feet below sea level.

The Gardiners Clay of south-central Long Island was sampled in test wells at Brookhaven where it was found between 77 and 92 feet below sea level. This unit contains foraminifera representative of a shallow, brackish water by or lagoonal environment (Weiss, 1954). Modern

Hudson Lobe	Interlobate Zone	Connecticut Lobe	Eastern Connecticut-Western Rhode Island Lobe, eastern Long Island
Sands Point Moraine at Lloyd Neck		Roanoke Point Moraine	South Fork end & recessional moraines
Target Rock Till		Sound Shore Formation: Wildwood Till, Centerville Sand	Hither Hills Formation: Heatherwood Till, Upland Road Sand
Harbor Hill, Northport & Sands Point Moraines	High Hill Moraine	Ronkonkoma & Stony Brook Moraine	Amagansett Moraine
East Hills Formation: Roslyn Till, Hempstead Sand	West Hills Formation: Melville Till, Plainview Sand	Farmingville Formation: Hauppauge Till, Sagtikos Sand	South Fork Formation: Devon Till, Acabonack Sand

Figure 2. Stratigraphy of Woodfordian deposits in Long Island, after Sirkin, 1982.

foraminiferal assemblages of the Great South Bay of Long Island proved to be comparable, leading Weiss to suggest an analogous setting. Elsewhere in southern Long Island, a marine clay was mapped between 50 and 120 feet below sea level and correlated with the Gardiners Clay at Brookhaven (Perlmutter and Geraghty, 1963; Perlmutter and others, 1959).

The many surface exposures of the Gardiners Clay and Jacob Sand that were mapped by Fuller (1914) in the cliffs along the north shore have been shown to be nonmarine deposits with some of the clay units resembling varves (Weiss, 1954; de Laguna, 1963; Gustavson, 1976; and Lonnie, 1977). Gustavson (1976) identified the nonfossiliferous clay as glaciolacustrine deposits in sequences that coarsened upward into outwash.

In a study of the geochemistry of twelve presumed Gardiners Clay sites, Lonnie (1977) differentiated between two clay types: Group I clays, which are illitic and high in magnesium and iron, and Group II clays, which are high in kaolin and low in magnesium and iron compounds. Group I clays are believed to be of proglacial origin, while Group II clays are marine. In general, the north shore clays were shown to be proglacial, while the type Gardiners is marine.

The Gardiners Clay in western Long Island overlies the Jameco Gravel in the large, north-south buried valley at a depth of 60 feet, according to Soren (1978). The Jameco and Gardiners units pose a number of problems: they vary considerably in color, clastic content, sequence of beds, fossils, and chemistry; their ages have not been determined; the Gardiners type section does not show the unit in place, rather as an ice-shoved clast and the cold-climate fauna of the Gardiners does not compare with the warm water faunas of the Sangamonian. These factors give the type Gardiners Clay a closer resemblance to an interstadial or proglacial deposit that was ice-shoved and emplaced in the moraine during the last glaciation. Thus, both the Jameco and the Gardiners units from the deep channel in western Long Island could be much younger.

Stage	Substage	Yrs. B.P.
Wisconsinan	Woodfordian (G)	18,000
	Farmdalian (I)	22,000
		28,000
	Portwashingtonian (I)	33,000
		>43,000
	Altonian (G)	
Sangamonian ?	marine deposits	

G-Glacial I-Interstadial

Figure 3. Wisconsinan stratigraphy of Long Island, after Sirkin, 1982.

Major differences also exist between the type Gardiners Clay and the subsurface Gardiners in southern Long Island. There are, however, no radiometric ages or intergrated pollen data from the various deposits to adequately differentiate them. Pollen assemblages in samples from four wells in central Long Island indicate temperate climatic conditions and vegetation for the subsurface Gardiners. In three samples pine and birch are dominant, while in the other, oak is most abundant with birch and other temperate forest elements (Sirkin, in press). This pollen evidence would not correlate with the cold water faunas of the type Gardiners and suggests that perhaps these units are not coeval.

Manhasset Formation

The Manhasset Formation of Fuller (1914) represented a major glacial event of Illinoian age (Figure 1). Its basal gravel, the Herod Gravel Member, heralded the approach of the glacier. The succeeding unit, the Montauk Till Member, is a basal till that attests to the actual presence of the ice on Long Island. The uppermost member, the Hempstead Gravel, is an outwash that indicated the recession of the ice. The Manhasset comprised most of the glacial section on the Island and only the younger "Wisconsin" moraines masked this unit, according to Fuller (1914). The latter moraines presumably were separated from the Manhasset by an erosion interval, the Vineyard Interglacial of Sangamon age that supposedly created much of the topography of northern Long Island.

The Montauk Till, the equivalent of Woodworth's (1901) boulder clay bed of the Columbia Formation, is mostly a compact mass of sediment that can vary considerably in grain size and clay content. In some exposures it consists of a very tight, sandy diamicton; in others it is a complex of diamicton, sand lenses, and deformed varve-like beds. In most sections, the Montauk or equivalent tills have been deformed by later ice-shove, like the Hempstead outwash. As has been demonstrated, the upper outwash that has been stratigraphically repositioned and in some instances renamed (Figure 2), incorporates ice-shoved masses of marine sediments that range in age from >43,000 to about 22,000 radiocarbon yrs BP (Figure 3).

These marine beds represent the Portwashingtonian interval and contain a cold-warm-cold climatic sequence based on the pollen spectra. This unit has been interpreted as evidence of a mid-Wisconsinan interstadial that separated an earlier, Wisconsinan glaciation from the Woodfordian (Sirkin and Stuckenrath, 1980; Sirkin, 1982). Accordingly, the Montauk and equivalent lower tills and the underlying outwash are assigned to the early Wisconsinan glaciation of the Altonian substage. The name, Manhasset Formation, is restricted here to the early Wisconsinan till and outwash in western Long Island (Figure 4).

Western & Central Long Island		Eastern Long Island, South Fork		Montauk Peninsula	
Manhasset Formation	West Shore Till	East Hampton Formation	Springs Till	Camp Hero Formation	Montauk Till
	Flower Hill Sand		Long Pond Sand		Ditch Plains Sand

Figure 4. Stratigraphy of Altonian deposits in Long Island, after Sirkin, 1982.

20-Foot Clay

An additional fine-grained, late Pleistocene, marine unit is encountered in wells in the southern margin of the outwash plain. This unit, the 20-Foot Clay, named by Perlmutter and Geraghty (1963), is found between 0 and 40 feet below sea level, generally between -20 to -35 feet. Its silt and clay lithology, gray and greenish-gray color, and fossil fauna are much like that of the Gardiners Clay of south central Long Island, and where superposed, the two are indistinguishable.

A possible equivalent of the 20-Foot Clay was cored off southern Long Island (Rampino and Sanders, 1981). They describe a tripartite sequence consisting of a lower outwash comparable to the early Wisconsinan outwash of the upland and informally named the Merrick Formation, and an upper outwash, the Bellmore Formation, corresonding to the Woodfordian outwash. The middle unit, the Wantagh Formation, is a marine silt and clay that may correlate with the 20-Foot Clay, according to the authors. Alternatively, in the absence of a chronology and biostratigraphy, both the 20-Foot Clay and the Wantagh Formation could be a facies of the Gardiners Clay, or they could be mid-Wisconsinan in age, equivalent in part to the Portwashingtonian beds, or even late Wisconsinan in age, as suggested by Sirkin (1982).

The "Wisconsin" Moraines

Fuller (1914) believed that the Harbor Hill and Ronkonkoma moraines represented separate "Wisconsin" glacial advances, the Ronkonkoma being the older of the two, and the Harbor Hill Moraine actually crossing the Ronkonkoma in western Long Island. More recent studies favor a single, late Wisconsinan glaciation and the sequential emplacement of morainal envelopes by the advance and subsequent recession of a lobate ice margin (Sirkin, 1976; Sirkin and Buscheck, 1977; Sirkin and Stuckenrath, 1980; Sirkin, 1981; and Sirkin, 1982).

Thus, the surface moraines represent Woodfordian end and recessional positions of the ice margin that conform to the lobate shape of the ice front, and rather than crossing, the moraines intersect in reentrant angles or interlobate positions (Figure 5). The moraines consist of basal or meltout till over outwash and, in some sections, Woodfordian deposits clearly mantle the older, probably Altonian age, morainal deposits and topography. The latter sediments also consist of till over outwash.

Intermorainal Deposits

Fuller (1914) suggested that the area between the Ronkonkoma and Harbor Hill moraines, west of Smithtown in the Nissequogue drainage basin in central Long Island was driftless because it lacked conspicuous "Wisconsin" deposits. Analysis of well sections showed that fine-grained, lacustrine deposits underlay fluvial sediments. Lubke (1964) called these sediments the clay unit of Smithtown and suggested that the clay might overlie or include the Gardiners Clay. He identified the upper clay as glaciolacustrine and described the deposit as representing up to 200 feet of fill in a large, buried valley.

In a subsequent study, Foord, Parrott, and Ritter (1970) identified the clay as a fresh-water deposit based on its kaolin content. They determined that the clay unit overlies outwash and in the northern portion is partly covered by coarser clastics.

It is probable that the clay unit of Smithtown is equivalent to the clay at Manorville (Weiss, 1954; de Laguna, 1963) and to the "unidentified unit" at Brookhaven (Weiss, 1954). The Manorville clay unit was described by de Laguna (1963) as being varved, in part, while the "unidentified unit" consists of 25 to 50 feet of greenish colored clay and sand that overlies the Gardiners Clay. The "unidentified unit" was not shown to be either fresh water or marine in origin. None of these clays has been dated. However, it is probable that the fine-grained, fresh-water deposits were laid down in one or more proglacial lakes that formed during the recession of the Woodfordian ice from the end moraines in the interval between approximately 19,000 and 21,750 years ago (Sirkin, 1982).

THE WISCONSINAN STAGE OF LONG ISLAND – A REVISION

In an updated analysis of the Wisconsinan glaciations of Long Island and Block Island, new formation and member names were assigned to both Altonian and Woodfordian till and outwash sequences. The new names were based on the substage, glacial lobe, lithology and provenance (Figures 2 and 4; Sirkin, 1982). This new stratigraphy is a means of emphasizing regional variations in rock units and replacing stratigraphic names that have been shown to be inappropriate. The Montauk Till of Montauk Peninsula in eastern Long Island is dissimilar to the corresponding Altonian tills in central and western Long Island. Eastern Long Island names, like Herod and Montauk, are not relevant elsewhere in Long Island, particularly in the restricted definition of the Manhasset Formation.

The new terminology also differentiates glacial deposits from geomorphic features. The till and outwash of Woodfordian age that comprise the Harbor Hill Moraine are given distinct stratigraphic names to avoid the ambiguity of similar stratigraphic and geomorphic names. Terms like Harbor Hill Till and Ronkonkoma Advance Outwash are eliminated.

In mapping the Pleistocene geology of the South Fork of Long Island, Sirkin and Buscheck (1977) found that many of Fuller's (1914) cross sections were based on stratigraphic units that did not exist, and the sections could not be duplicated. Nor was the earlier work cognizant of shallow subcrops of older glacial sediment, for example, Altonian drift that was buried by Woodfordian deposits, or proglacial sediments in the so-called "driftless" terrains. The new interpretations incorporate a substantially revised stratigraphy and recognize the glacial lobes, interlobate moraines and interlobate angles, as well as extensive glaciofluvial and glaciolacustrine sediments.

In this paper, the new model for glaciation in this region is developed, and field evidence is cited to support the designation of new moraines or stratigraphic entities. Supporting biostratigraphic and radiometric information are also included where necessary. The presence of the mid-Wisconsinan, Portwashingtonian beds and the accompanying pollen and radiocarbon data provide a means of separating the drift sheets stratigraphically. The field evidence emphasizes the presence of a complex of arcuate moraines that conform to the outline of the lobate ice front. Where mapped in detail the older drift has considerable topographic expression beneath Woodfordian deposits as shown by Sirkin and Mills (1975), Sirkin and Buscheck (1977), and Sirkin (1982).

Glacial Lobes and Moraines

Recent studies have shown that several major glacial lobes were active in the deposition of the moraines, including two or three lobes of the early Wisconsinan or Altonian glacier and at least three lobes during the Woodfordian advance and recession. The new model for the deposition of Wisconsinan moraines on Long Island, therefore, incorporates two episodes of glaciation, two drift sheets, and a lobate ice margin. There is no evidence at this time of glacial deposition older than that described here.

End and recessional moraines of the three major glacial lobes, as well as interlobate deposits and intersections comprise the dominant glacial morphology of Long Island. In western Long Island, deposition was controlled by the Hudson Lobe (Hudson-Champlain Lobe of Connally and Sirkin, 1973) of the Woodfordian glacier. The ice advanced as far south as southern Staten Island and extended east, northeastward at least to Manetto Hills. The Harbor Hill Moraine is the end moraine of this lobe (Figure 5).

In central Long Island, the Connecticut Lobe of the Woodfordian glacier deposited the Ronkonkoma Moraine, an end moraine that extends from the Dix Hills area eastward and includes two sublobes prior to its reentrant junction with the Shinnecock Moraine. The latter moraine intersects the Amagansett Moraine, the end moraine of the eastern Connecticut-western Rhode Island Lobe. To the east in the Montauk Peninsula, it is probable that the Narragansett Lobe reached this region during the Altonian glaciation but not during the Woodfordian. Recessional Woodfordian moraine of the eastern Connecticut-western Rhode Island Lobe caps the Montauk upland.

The advancing Woodfordian glacier also deformed and displaced the existing Mesozoic and Cenozoic sediments of the coastal plain. The ice picked up and reorganized these sediments and then redeposited them as till and outwash. Large masses of strata were transported, reoriented and deposited in the drift (Mills and Wells, 1974; Sirkin and Stuckenrath, 1980). Woodfordian sediments also were intercalated with the underlying drift, filling spaces opened by deformation. Folding and faulting with high angle dips are common within the thrusted blocks of strata. Extensive ice-shove deformation has complicated the stratigraphic relationships, even though the major thrusts are from north to south.

Furthermore, the till that caps the moraines generally is flat-lying and relatively undeformed. This presents an apparent unconformable boundary with the underlying outwash. While Fuller (1914) placed the outwash and the underlying Montauk Till in the Manhasset Formation, the mid-Wisconsinan sediments clearly show that the outwash and the upper till are Woodfordian in age.

Altonian Substage

The oldest glacial deposits of Wisconsinan age, perhaps the oldest glacial deposits in the end moraines, occur beneath outwash and till of Woodfordian age and in two localities below radiocarbon-dated mid-Wisconsinan sediments. The Montauk Till and its equivalents and associated outwash are believed to be of Altonian age (Sirkin, 1982). These units overlie presumed Sangamonian deposits or Cretaceous strata. The till, till stones and outwash are relatively unweathered. The pollen record in the overlying sediments of mid-Wisconsinan age shows a transition from boreal to temperate vegetation dating from >43,800 yrs BP, with the cold climatic interval lasting until about 33,000 yrs BP (Sirkin and Stuckenrath, 1980). The cold interval persists through the duration of the Cherry Tree Stadial of the Erie Lobe (Karrow and others, 1978) and the glaciation that deposited the Chaudiere Till in southeastern Quebec (McDonald and Shilts, 1971), although this ice did not reach Long Island.

The best known tracts of Altonian deposits are found in western Long Island in the Port Washington sand pits, in the South Fork of eastern Long Island, and in north-central Long Island underlying recessional morainal deposits. In western Long Island the Altonian may have been deposited by a Hudson lobe of the glacier, although till fabrics show a north to northeast direction

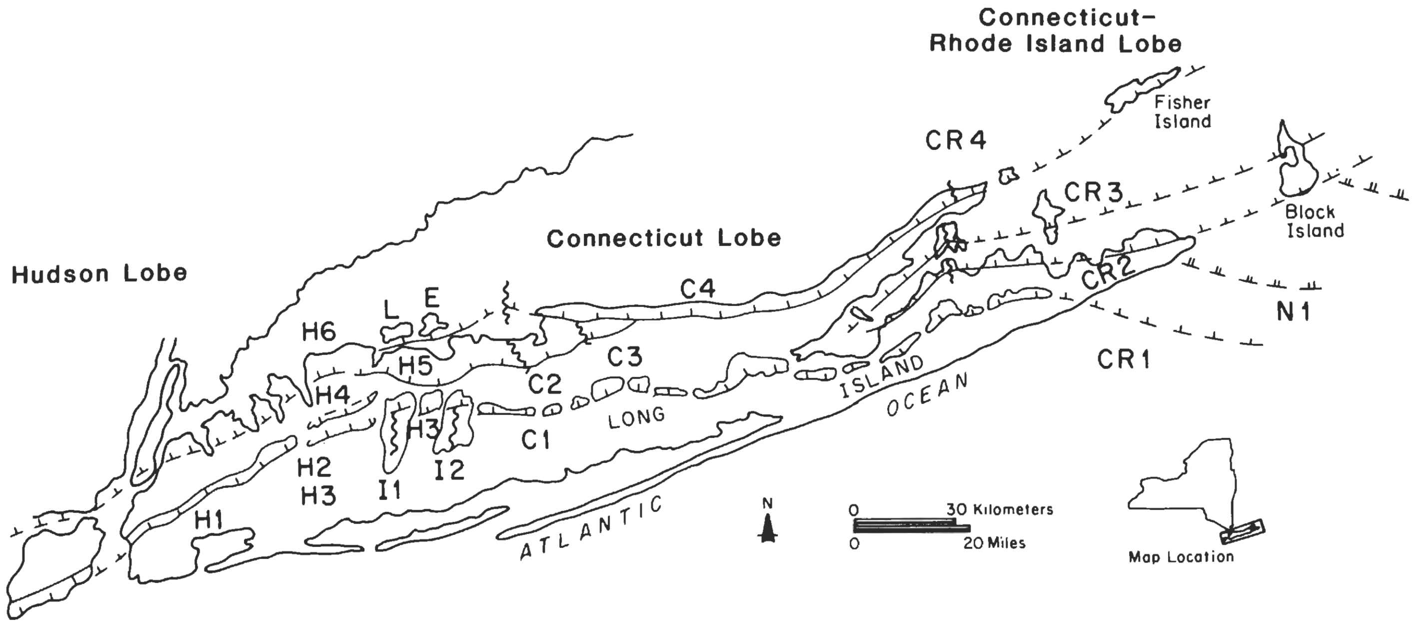

Figure 5. **Map of Long Island glacial end and recessional moraines and the relative positions of glacial lobes, keyed as follows:**

H1 **Harbor Hill Moraine**
H2 **Jericho Moraine**
H3 **Old Westbury Lobe**
H4 **Oyster Bay Moraine**
H5 **Northport Moraine**
H6 **Sands Point Moraine**

Interlobate Moraine

I1 **Manetto Hills Lobe**
I2 **Dix Hills Lobe**
I3 **South Huntington Lobe**
I4 **High Hill Interlobate Moraine**

Connecticut Lobe Ice Margins

C1 **Ronkonkoma Moraine – Shinnecock Moraine**
C2 **Stony Brook Moraine**
C3 **Mount Sinai Moraine**
C4 **Roanoke Point Moraine**

Connecticut Lobe and Eastern Connecticut – Western Rhode Island Ice Margins

CR1 **Amagansett Moraine**
CR2 **Sebonack Neck – Noyack – Prospect Hill Morainal Envelope**
CR3 **Robins Island – Shelter Island – Gardiners Island – Morainal Envelope**
CR4 **Roanoke Point – Orient Point – Fishers Island Moraines**

Narragansett Lobe

N1 **Montauk Point (Altonian)**

of ice flow (Sirkin and Mills, 1975). In central Long Island, northerly fabrics indicate that deposition of the Altonian may have been controlled by the Connecticut lobe.

The Altonian tills of Long Island are mainly lodgement but with minor melt-out components. They range in color from 5 YR 4/4 to 4/1 for the reddish brown and brown-gray tills to 10 R 4/2 for the red clay till at Montauk; 5Y 4/1 and N4 for the dark gray tills and 10 YR 5/4 for yellow-brown tills (Sirkin and Mills, 1975; Sirkin and Buscheck, 1977). The Altonian tills, with the exception of the clay till, have a sandy matrix with the mean size in the fine sand range. Individual sections of brown and brown-gray till may be 3 to 5 m thick. The upper contact between the Altonian till and the overlying Woodfordian drift is complicated by Woodfordian glacial tectonics. Sheets of till have been uplifted, thrusted and separated along joints and faults, and have been intercalcated or filled-in with yellow-brown sand and gravel from the Woodfordian ice. Some clasts of Altonian till are even incorporated in the younger outwash (Sirkin, 1982).

In the Port Washington sand pits the till is thrusted over itself increasing its thickness to nearly 10 m (Mills and Wells, 1974); Sirkin and Mills, 1975). Facies of this till, now named the West Shore Till, include a southward gradation into outwash or alternating melt out till and lacustrine sand with dropstones in thin beds (Sirkin and Mills, 1975, Figure 3F).

Mid-Wisconsinan

The moraines in western Long Island contain datable mid- Wisconsinan sediments in a sequence of marine clay, salt marsh and fresh water peats, and oyster reef beds that occur folded, thrusted and intercalated with outwash and melt out of tills of Woodfordian age (Sirkin and Stuckenrath, 1980, Figure 2). These beds range in age from 43,000 to about 21,750 yrs BP and include pollen evidence of a warm interval prior to 42,500 yrs BP, a cold interval which began more than 43,000 yrs BP, lasting until about 33,000 yrs BP, the Portwashingtonian warm interval as indicated by the Portwashingtonian Oak Pollen Zone between 33,000 and 28,000 yrs BP, and a succeeding cold episode that corresponds to the Farmdalian Interstadial and contains the Farmdalian Spruce Pollen Zone. The Farmdalian in this reconstruction lasted until about 21,750 yrs BP, the inferred beginning of the Woodfordian glaciation (Sirkin and Stuckenrath, 1980).

Woodfordian Substage

The late Wisconsinan glacier advanced to its terminal position around 21,750 yrs BP (Sirkin and Stuckenrath, 1980). The ice front was formed of at least three major lobes, the Hudson, Connecticut and eastern Connecticut-western Rhode Island lobes. The end moraines in turn are lobate in form and have sublobes and interlobate angles. The Harbor Hill Moraine is the end moraine of the Hudson lobe from Staten Island to western Long Island. The Ronkonkoma Moraine is the end moraine of the Connecticut lobe. These glacial lobes met in west central Long Island to form massive, south trending hills that make up the Manetto Hills Interlobate Zone (Figure 5). The interlobate zone is comprised of several sublobes that will be differentiated in the text. All of the surface north of the end moraines is occupied by some type of recessional deposits of glacial origin or glacio-fluvial or glacio-lacustrine sediments. Many recessional deposits form linear features that are described herein as recessional moraines and ice positions. The classical moraines of Fuller (194) are now subdivided according to this new scheme of differentiating recessional deposits and identifying them with glacial lobes.

The Manetto Hills Interlobate Zone consists of a thick sequence of outwash intercalated with melt-out till, collapsed glaciofluvial and glaciolacustrine silt and clay beds, and locally a thin upper till. In the northern part of the zone the outwash coarsens and may be deformed. Over 50 m of interlobate deposits have been exposed in a large sand pit. The yellow gravels prograde over the distal surface of the zone.

Hudson Lobe – Western Long Island

Prior to the redefinition of the moraines in terms of their lobate relationships, the Harbor Hill Moraine was believed to be the end moraine of western Long Island, to cross over the Ronkonkoma Moraine at Lake Success, and to extend along the north shore of eastern Long Island to Orient Point. The new lobate model, of course, changes this interpretation. It is also apparent that there is a kame moraine that lies south of the Harbor Hill in this area (Figure 5).

In fact, a line of kames lies at a lower elevation and to the south of the Harbor Hill Moraine in the Hicksville and Huntington Quadrangles. This moraine was believed to be the westward extension of the Ronkonkoma Moraine and a product of an earlier glaciation (Fuller, 1914). However, both this southerly kame moraine, now named the Jericho Moraine (Sirkin, 1982), and the Harbor Hill Moraine are moraines of the Hudson lobe, and they both abut the Manetto Hills Interlobate Zone, as seen in the Huntington Quadrangle. There is also a southwesterly trending morainal lobe, the Old Westbury Lobe, that lies between the Jericho and Harbor Hill moraines. This lobe may have originated from ice overriding the Harbor Hill position. The Jericho Moraine incorporates a medium gray till beneath outwash and till as observed by Sirkin (1968). Otherwise, the two moraines are separated by large outwash heads or fans and recessional moraine. Thus, the Jericho Moraine is the actual end moraine of the Hudson lobe in this area, although to the west the Harbor Hill Moraine is the terminal moraine of the glaciation.

Generally, the thin till cap on the kames of the Jericho Moraine extends continuously northward as the melt-out till covering the Old Westbury Lobe, the Harbor Hill Moraine and the recessional moraine to the north. This till was named the Roslyn Till by Sirkin (1968). The recessional moraine with its Roslyn Till cover, that may be over 5.0 m thick, is continuous northward to the northern boundary of the Oyster Bay Recessional Moraine (new name here) except where the till is covered by younger outwash, deltaic or lacustrine sediments. The Old Westbury Lobe forms a south trending morainal mass between the Harbor Hill and Jericho moraines.

In the Harbor Hill and Oyster Bay moraines, the till thickens and appears to be undeformed. The Roslyn Till of the Hudson lobe is a more clay-rich till than those of eastern Long Island. In most sections it appears to be an ablation or melt-out till, particularly in the upper portions of most exposures, with flow structures and shear planes, but with little ice-contact deformation (Sirkin, 1968). In the lower portions of some exposures, especially in the end moraines, the Roslyn Till is more compact and yellow-brown in color. It contains a variety of rock types including gneisses and schists of the New York Group and the Hartland Formation, as well as many other Hudson Valley rock units. There are also abundant erratics derived from the Palisades basalts. Till fabrics for the Roslyn Till in western Long Island indicate a northwest to southeast flow of the ice, which would be expected on the east side of the Hudson lobe. Corresponding north-northwest striations on the bedrock are cited by Woodworth (1901).

The thick outwash beneath the till was folded, faulted, and thrusted by the overriding ice. The emplacement of large masses of mid-Wisconsinan sediment into the outwash supplies basic evidence for having the Woodfordian outwash as a deposit preceding the advancing ice. Thus, the Woodfordian outwash is the matrix in which much of the deformation is contained. Much of the faulting involves thrusting with large-scale overturning as well as small-scale contortions (Sirkin and Mills, 1975). In some examples the thrusts are responsible for morainal topography, particularly where large masses of strata have been displaced. Intermorainal and supermorainal deposits include small kames, and in the Sands Point Moraine, deltaic deposits like the Port Washington and College Point deltas (of Woodworth, 1901) that are now mostly mined out. The deltas and the prominent west-southwest to east-northeast meltwater channel and associated lineations mark the distal margin of the Sands Point Moraine that formed during an ice stand on the northern margin of western Long Island (Figure 5; Sirkin, 1982).

Recession of the ice from the Harbor Hill Moraine to the Sands Point position may have occurred as early as 21,000 years ago, and the formation of the proglacial lakes dates from about that time (Sirkin, 1982). In the emplacement of the Sands Point morainal envelope, some sequential deposition is seen in the alternation of low morainal ridges with small east-west drainage lineations. The low areas occasionally contain lacustrine clays that may be rhythmically banded (Sirkin and Mills, 1975). A thin till, the Target Rock Till, also is a result of this recession. The Kings Point Bog (Sirkin, 1967) is situated on the distal margin of the Sands Point Moraine.

The Woodfordian age, Hempstead Sand encloses the mid-Wisconsinan, Portwashingtonian beds and overlies the Altonian, West Shore Till. The Hempstead outwash was deformed by the ice as it overran its deposits. Outwash extends southward from the moraine as outwash heads and pitted outwash to form an extensive outwash plain. The pitted outwash indicates the presence of ice south of the moraine and accounts for a barrier south of the moraine that could hold back local ponding and account for the fine-grained deposits in the distal part of the moraines.

Presumably, Woodfordian outwash overlies Altonian outwash, but with little differentiation. The outwash plain is incised by a dendritic or subparallel network of fluvial channels. In general, these channels are relict and in some cases intermittent, except in their southern reaches where they become tidal. Large fans of alluvium deposited by several larger streams today form peninsulas projecting into the south shore bays of Long Island. Holocene marine and bay sediments, as well as salt marshes onlap the outwash, which to the south is considerably finer. Clay deposits occur in areas of local meltwater ponding in both interfluves and channels. The channels have numerous boggy and ponded tracts, although many existing ponds are cultural.

Manetto Hills Interlobate Zone – Central Long Island

The Hudson and Connecticut lobes intersected in west central Long Island and formed a large interlobate moraine which consists of several north-south trending hilly uplands. The Connecticut Lobe, as indicated by the Ronkonkoma Moraine, extended as far west as the northeastern edge of the Dix Hills portion of the interlobate region. The Hudson Lobe, as indicated by the Harbor Hill Moraine, extended at least as far east as the northwestern edge of the Manetto Hills interlobate zone. Ice probably engulfed the interlobate zone which in the Manetto Hills filled in with glacio-lacustrine and glaciofluvial deposits and thin melt-out tills. Perhaps the interlobate zone existed originally as a series of more or less coalescing hills from Manetto Hills to Dix Hills. Valleys between the hills were widened by meltwater during deglaciation and filled in with outwash with large-sized, rock fragments derived from the moraine. This polymict gravel forms the upper reaches of the outwash plain and surrounds the interlobate zones.

Cold Spring Harbor and the valley heading to the south represent the north-to-south valley that marks the interlobate angle between the Harbor Hill Moraine to the west and the High Hill Interlobate Moraine to the east. In this northern portion, the interlobate moraine consists of over 3.0 m of a strong, brown (7.5 YR 5/6) till over outwash, as in the Hudson lobe moraines.

In the large sand pit at Melville, a melt-out till is interlayered with fine-grained stratified outwash to the north, while clays are stratified with silts and sands to the south, indicating the presence of a proglacial lake. The interlobate zone, including the yellow gravels, extends several km southward through the Huntington quadrangle into the Amityville quadrangle resulting in prominent north-south topography.

The Jericho Moraine, which consists of a line of kames lying south of the Harbor Hill Moraine, abuts the southwestern margin of the interlobate zone, and two kamic masses actually overlie the latter feature east of the boundary. The Jericho Moraine, like the Old Westbury lobe, may have formed when the Hudson lobe ice expanded south of the Harbor Hill Moraine, and perhaps both moraines formed during the same episode. East of the kames the interlobate deposits become more deltaic with only thin, sandy melt-out till and cross-stratified sands and clays.

The interlobate zone is bisected to the east by another outwash-filled valley that cuts between the lobes from Huntington southward through the Route 110 valley, isolating the South huntington Lobe of interlobate deposits to the northeast of the Manetto Hills interlobate zone. The South Huntington lobe, in turn, is separated from the eastern portion of the interlobate zone by another outwash-filled valley. The interlobate zone to the east, the Dix Hills-Half Hollow Hills lobe, extends southward to form another prominent north-south ridge. The deposits of this lobe resemble those of the Manetto Hills lobe with local, thin melt-out till over outwash, including the yellow-stained pebble gravels, and clay lenses increasing to the south. However, the Half Hollow Hills segment is comprised of till surrounded by the yellow gravel. Yet another north-south valley separates the interlobate zone from the east trending Ronkonkoma Moraine. The outwash-filled valley is now occupied by the Sagtikos Parkway.

Recessional Moraines

The interlobate zones are separated by outwash gravels derived from the moraines as the ice front began to disintegrate. Recessional moraines truncate the interlobate deposits as drainage valleys emerge, especially in the interlobate angles where much of the meltwater seems to be directed. The valleys also break up the linear trend of the moraines.

In the Hudson lobe to the west, the Oyster Bay Moraine, the recessional moraine north of the Harbor Hill Moraine, is truncated by an ice margin that extends from west of Mill Neck in the Bayville quadrangle through Northport to the Nissequogue River in the Saint James quadrangle. An east-west lineation of meltwater channels forms the southern boundary of this stillstand. The moraine, the Northport Moraine, is comprised of an equally distinctive alignment of north-south elongate dumlinoid masses of glacial sediment that lie north of the east-west channels and north of the outwash and deltaic deposits that partially engulf the northern margins of the interlobate zones.

A thick section of the Northport Moraine is exposed in a sea cliff at Long Beach in the Saint James quadrangle. Here over 25 m of cross-stratified outwash are capped by a meltout till and Holocene sand dunes. To the west, brown clays of untested origin have been observed at beach level. The Northport Moraine also appears to document the eastward spreading of the Hudson lobe during recession to form an interlobate angle at the position of the present Nissequogue River. This position also emphasizes the northeasterly migration of the interlobate positions from the previous interlobate zones.

North of the Ronkonkoma Moraine the terrain appears to be essentially driftless although cut through the meandering valley of the Nissequogue River. The Connecticut Lobe must have receded rapidly here leaving a large open valley between the Ronkonkoma Moraine and the recessional moraine to the north. Actually this basin has been shown to hold lake clays beneath about 3.0 m of sand (Krulikas and Koszalka, 1983). Apparently, the recession of the ice left a proglacial lake in its wake. After deposition of the lake sediments, a fluvial, meltwater drainage system occupied the lowland and breached the moraine to the south.

The moraine bounding the basin to the north, the Stony Brook Moraine, extends northeastward from the interlobate angle with the Northport Moraine toward Port Jefferson. Both the Stony Brook and Northport moraines have a recessional washboard type of topography with kamic highs and intervening swales filled with finer sediments, or in some instances, clays in small proglacial lakes.

There appear to be three recessional positions within the envelope of the Stony Brook Moraine as it widens to the east in the Port Jefferson quadrangle. The southernmost, the Nesconset lobe, appears to be a small remnant of recessional moraine that has been eroded by meltwater in a northeast to southwest valley that heads in the interlobate angle to the east. The remnant also appears to have been masked by outwash from the moraine to the north.

North of the valley there are two segments – the South Setauket segment and the East Setauket segment. The South Setauket segment is a kamic moraine that held in a proglacial lake were a delta formed when the ice receded to the East Setauket position. Of considerable interest is the presence of a lower gray till presumably of Altonian age beneath the recessional deposits and the thick (3 to 4 m), brown Woodfordian till. Both tills have northerly till fabrics, but differ in color and lithology.

An additional small, recessional-morainal segment lies east of the interlobate angle that trends north-northwest to south-southeast from Northport Harbor. This moraine, the Mount Sinai Moraine, was deposited as a recessional position of the Connecticut lobe prior to recession to the Roanoke Point Moraine's position to the north. This moraine is characterized by a brown sandy till over outwash.

The Roanoke Point Moraine of the Connecticut glacial lobe and the Sands Point Moraine of the Hudson lobe both represent the final ice margins on northern Long Island. Both truncate the earlier moraines and interlobate positions. The interlobate angle between these moraines is projected north of Smithtown Bay. The position of the glacial lobes at this time is also north of the major north-south drainage through central Long Island, the Nissequogue-Connetquot river drainage.

While today the rivers are independent north and south flowing drainages, respectively, during deglaciation there was a south flowing meltwater channel that breached the moraines and cut terraces into the outwash south of the Ronkonkoma Moraine where there are broad terraces at the present time. The headwaters of the two rivers are separated today by a divide of less than 3.0 m, using the channel of the Northeast Branch of the Nissequogue.

The Roanoke Point Moraine clearly truncates the Stony Brook Moraine in its westerly trend. The Sands Point Moraine, however, has a deep east-west meltwater channel along its southern border. The channel's components stand out today as part of the flooded harbor system and they conveniently demark the ice margin and the moraine. West of Bayville the lineation of the meltwater channel is seen in deep east-west valleys incised into the morainal surface. the deposits of the Sands Point Moraine are mainly kamic and till capped, similar to the deposits of the moraines to the south. However, the outwash and till often lie above recessional deposits.

In Lloyd Neck on the east side, the washboard topography is pronounced with each small moraine segment or kame separated from the next by a small swale, channel, or proglacial lake. In at least one such lake nearly 1.0 m of clay were deposited in a section with a thick till at the base (the Roslyn Till?), 1.0 m of clay, 2.0 m of laminated and rippled silts, 0.8 m of deformed gravelly outwash, and 1.0 m of meltout till at the top. At the south end of the exposure there are only 1.8 m of till over 1.2 m of outwash. Between these two exposures the laminated and rippled silt and fine sand overlie the till, and in turn are covered by a thin gravel layer. A similar gravel layer also crops out at the north end of the section, overlying both till and outwash. These sections indicate a minor sequence of recessional deposition.

At Caumsett, in the northwest of Lloyd Neck and on the west side of Eatons Neck to the east, large masses of Cretaceous sediment have been incorporated into the moraine and the outwash has been folded. In a 20.0 m section on the west side of Eatons Neck, thin till overlies cross stratified sand that contain a 1.0 m clay unit. The lower part of the section is comprised of deformed Cretaceous beds.

Connecticut Lobe – East Central Long Island

Between the Dix Hills-Half Hollow Hills Interlobate segment and Riverhead on the eastern side of the study area, the Connecticut lobe end moraine, the

Ronkonkoma Moraine, there are several meltwater channels and erosional breaks in the moraine, in addition to the Connetquot valley mentioned earlier. In a large sand pit in kamic sands on the distal slope of the Ronkonkoma Moraine, in the segment between the Sagtikos channel and the Connetquot, there is a lower gray till that may be Altonian in age.

The Ronkonkoma Moraine is a series of larger kames and thrust sheets generally capped by sandy till, the Hauppauge Till in this lobe. The moraine divides the north-flowing Nissequogue drainage from the south-flowing Connetquot. The streams may occupy an older, proglacial channel cut into the outwash as the ice advanced to the Ronkonkoma position. Lake Ronkonkoma occupies a kettle in the Ronkonkoma Moraine. In the intermorainal area north and northwest of Riverhead, there are a number of arcuate wetlands that may mark west-east drainage channels that developed during glacial recession (Bartunek, 1982). South of the Ronkonkoma Moraine a number of alluvial peninsulas project into Great South Bay and are associated with the major drainage channels of the outwash plain. For example, the Heckscher Park peninsula is flanked to the east by the Connetquot River, and the Smith Point peninsula by the Carmans River on the west and the Forge River on the east.

In the next morainal segment to the east of the Connetquot River the trend of the moraine becomes northeasterly. This segment also has the large twin kettle complex of Lake Ronkonkoma and an adjacent deep peat bog to the north. This bog yielded the postglacial history of central Long Island that showed that the moraines were essentially coeval and late Wisconsinan in age (Sirkin, 1971). A thrust located on the proximal side of this lobe of the moraine east of the lake near Selden in the Patchogue quadrangle involves a red sand and clay sequence that appears to be Cretaceous in age. The next major channel to the east, that of the Carmans River cuts through the moraine, deeply cliffing the east side of the morainal lobe, and cutting terraces in the outwash to the south. Again it appears that a meltwater channel corresponds to an interlobate angle as seen in the change in the direction of the moraine in the Bellport quadrangle; east of this angle the moraine again trends eastward.

The next breach in the moraine is at Forge River and it appears to be an indication of another interlobate angle as the moraine again trends northeasterly. In the Riverhead area the morainal band broadens and the elevation increases to nearly 88 m. Deposits of Altonian till may underlie portions of the southeastern segment of this moraine which is largely sandy outwash capped by sandy till. A major thrust also is found in the proximal slope of this segment near Manorville, incorporating a thin marine silt with highly oxidized clam shells into the outwash.

The Connecticut lobe in east central Long Island retreated from the Ronkonkoma Moraine leaving a broad outwash plain and at least one additional major proglacial lake basin north of the moraine in Manorville. The main still-stand of the ice, however, is at the Roanoke Point Moraine along the north shore of the Island. From the interlobate angle north of Smithtown Bay, through its first landfall to the east at Old Field, this moraine forms a smooth arc as it continues easterly that appears to mirror the final arcuate shape of the Connecticut lobe on Long Island. In conforming the glacial lobe origin of the arc, till fabrics range from northeast at Hallock Landing to the west, north on the center of the arc near Roanoke Landing, and northwesterly at points located to the northeast of Riverhead (Corley, 1976).

Port Jefferson Harbor and Mount Sinai Harbor are the main heads for meltwater channels that breach the moraine. There are also a number of kettle valleys that extend southward from the interlobate angles, like the lineation of depressions southeast of Mount Sinai and south of Fresh Pond Landing. In fact, pitted outwash is common in the outwash plain especially near the moraine. Furthermore, the outwash plain adjacent to the Roanoke Point Moraine is made of coalescing fans of outwash similar to alluvial fans, attesting to the immediate source of the sediments from the ice at this still-stand. Typical sections in the moraine have 1-2 m of till over outwash. Occasional melt-out till and massive silt lenses also are found in the outwash. Usually the moraine is covered by a thin layer (0.3 m) of loess. The till is generally 25% gravel, 65% sand, and 10% silt and clay with a light brown (10 YR 5/6) color.

South of Riverhead, the Ronkonkoma Moraine overlies the Altonian, Springs Till, the latter forming a buried moraine of considerable relief. The Shinnecock Hills Moraine, east of the Altonian high, is also a kame moraine consisting of a series of small kames, now largely mined, that overlie thin outwash and Springs Till. Here the surface of the till lies at about sea level. The Springs Till surface is relatively flat laying, but cut by drainage channels, and it supports the recessional Woodfordian deposits and the Holocene wetlands that developed in kettles and channels in the recessional terrain north of the Shinnecock Hills Moraine. The moraine becomes more massive to the east and incorporates some large, thrusted masses of Altonian drift and smaller blocks of marine sediment as cited by Gustavson (1976) and Sirkin (1982).

Eastern Long Island

The intersection of the Connecticut Lobe and the eastern Connecticut-western Rhode Island Lobe of the Woodfordian glacier is marked by the reentrant of the Shinnecock Hills Moraine with the Amagansett Moraine southwest of Sag Harbor. The intersection is breached by the Long Pond meltwater channel lineation. The Amagansett Moraine strikes east and southeast and is sharply truncated above Napeague Beach. This moraine represents the end position of the lobe (Figure 5).

As the lobes receded from the end moraines, the intersections appear to have migrated northward and then to the northeast as indicated by the reentrants. The Connecticut Lobe receded to the Sebonack Neck position, the eastern Connecticut-western Rhode Island Lobe to the Noyack-Prospect Hill envelope and the Beacon Hill Moraine of Block Island. The recessional moraines are comprised mainly of kames, while the deposits between the end moraines and the kame moraines are outwash. The Springs Till underlies the kames and is seen in some sections along Peconic Bay. Woodfordian tills cap the end moraines and some of the kames, but the tills do not form a continuous blanket of ground moraine (Figure 2). The recessional moraines develop a washboard-like ridge and channel morphology, as for example in the Hither Hills of the Montauk Peninsula.

Recession of the two lobes formed a reentrant or interlobate position at Shelter Island with the Robins Island-Shelter Island segment to the west and the Shelter Island-Gardiners Island segment to the east. The recessional moraine is traced west of Robins Island to Southport, just east of Riverhead. The Gardiners Island position may represent another interlobate intersection of sublobes east of Shelter Island, and the moraine is correlated to the east to the Corn Neck Moraine of Block Island (Figure 5).

A reentrant angle between moraine segments occurs near East Marion and marks the intersection of the lobes on the northeastern margin of Long Island. Here the Roanoke Point Moraine meets the Orient Point-Fishers Island Moraine of the eastern Connecticut-western Rhode Island Lobe. The moraines consist of till over outwash (Figure 2), and the outwash incorporates masses of ice-shoved alluvium. Outwash and lacustrine sediments occur between the recessional moraines.

SUMMARY

This paper presents a discussion of the existing stratigraphic models for the Pleistocene of Long Island, particularly questioning the validity of the stratigraphic units developed by Fuller (1914).

1. Several stratigraphic units established in prior studies are reinterpreted. This group includes the Manetto Gravel and the Manhasset Formation. Several others appear to be stratigraphically afloat, pending more definitive study. These include the Jameco Gravel, Gardiners Clay, and Jacob Sand. A third group includes rock units that are restricted to type areas with new unit names proposed for local equivalents. The Montauk Till, for example, is in this category.
2. Pleistocene deposits provide evidence of two glaciations, both Wisconsinan in age and assigned to the Altonian and Woodfordian substages. The drifts are physically separated by marine, fluvial, and lacustrine sediments that were deposited during the midWisconsinan, Portwashingtonian, warm interval. Marine deposits, such as those equated with the Gardiners Clay, generally are placed in the Sangamonian Interglacial. Some marine beds may be of mid-Wisconsinan age. The mid-Wisconsinan pollen record also indicates that a cold interval preceded the Portwashingtonian.
3. During both glaciations, a lobate ice front deposited end moraines on Long Island. Remnants of the Altonian moraines may be found beneath Woodfordian drift at Port Washington, East Setauket, and in the south fork of Long Island. The Woodfordian end and recessional moraines were deposited by at least three glacial lobes. An interlobate zone separates the major end moraines in west-central Long Island, and reentrant angles mark the interlobate positions during recession.

Ground moraine, outwash, and proglacial lake deposits form the intermorainal terrain. Meltwater deposited extensive outwash plains in valleys south of the moraines, with the outwash separating the interlobate zones. Postglacial deposits include large fans of alluvium that spread southward from the major fluvial systems, as well as stream deposits, loess, lake and bog sediments, and soils on the glacial strata.

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THE WISCONSINAN HISTORY OF THE GREAT VALLEY, PENNSYLVANIA AND NEW JERSEY, AND THE AGE OF THE "TERMINAL MORAINE"

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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 yrs BP.

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 yrs BP (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 yrs BP (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.

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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 yrs BP. Together these data 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 600 km² 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).

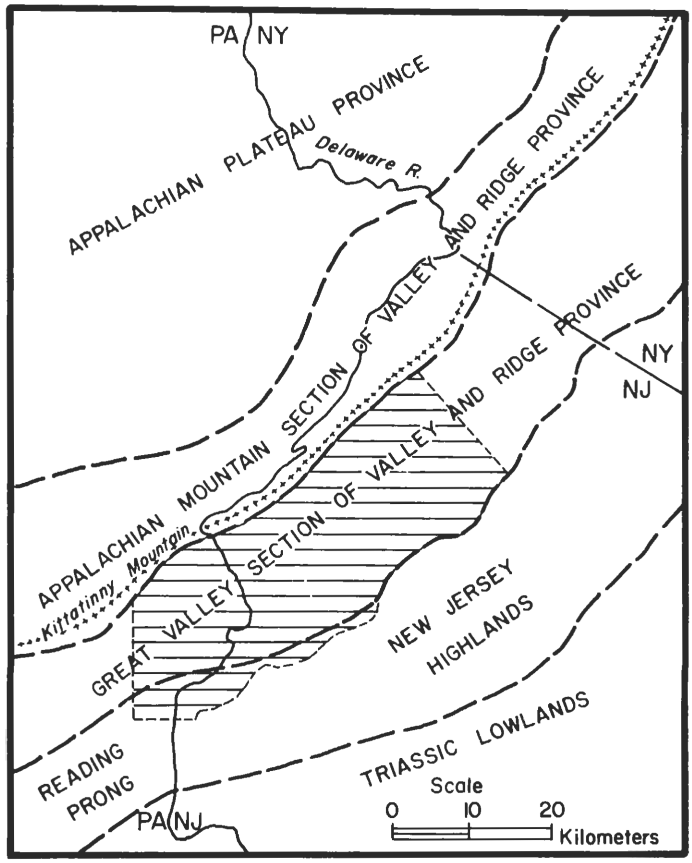


Figure 1 **Location map and physiographic provinces of the study area. The study area is shown by the ruled pattern.**

PREVIOUS WORKS

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 its annual report series (Cook, 1878, 1879a, 1879b, 1881). In Pennsylvania, Hall (1876), 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 Indi-

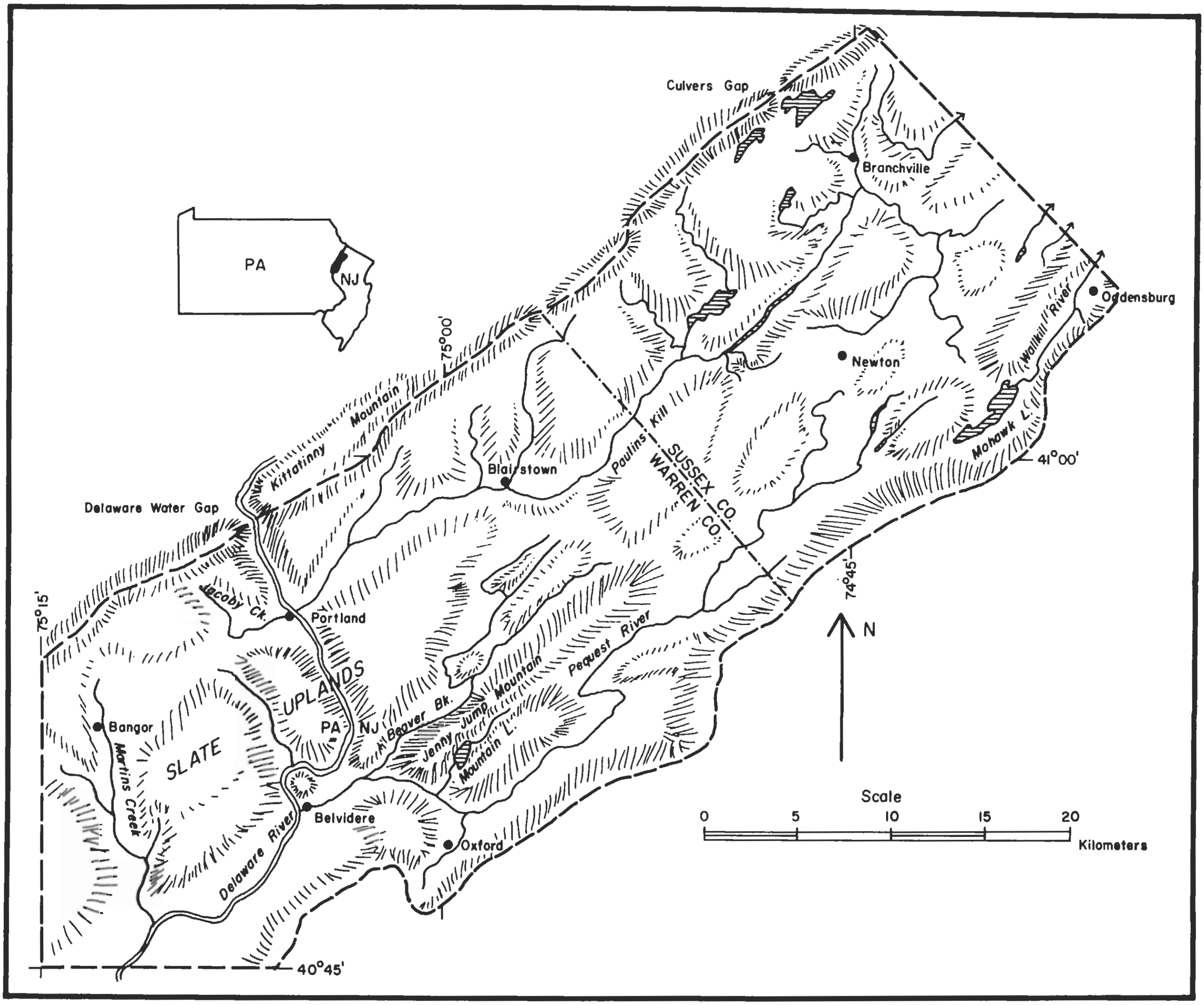


Figure 2 Physiography of the study area in the Great Valley Section of the Valley and Ridge Province.

ana (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 margin 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 margin 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 ex-

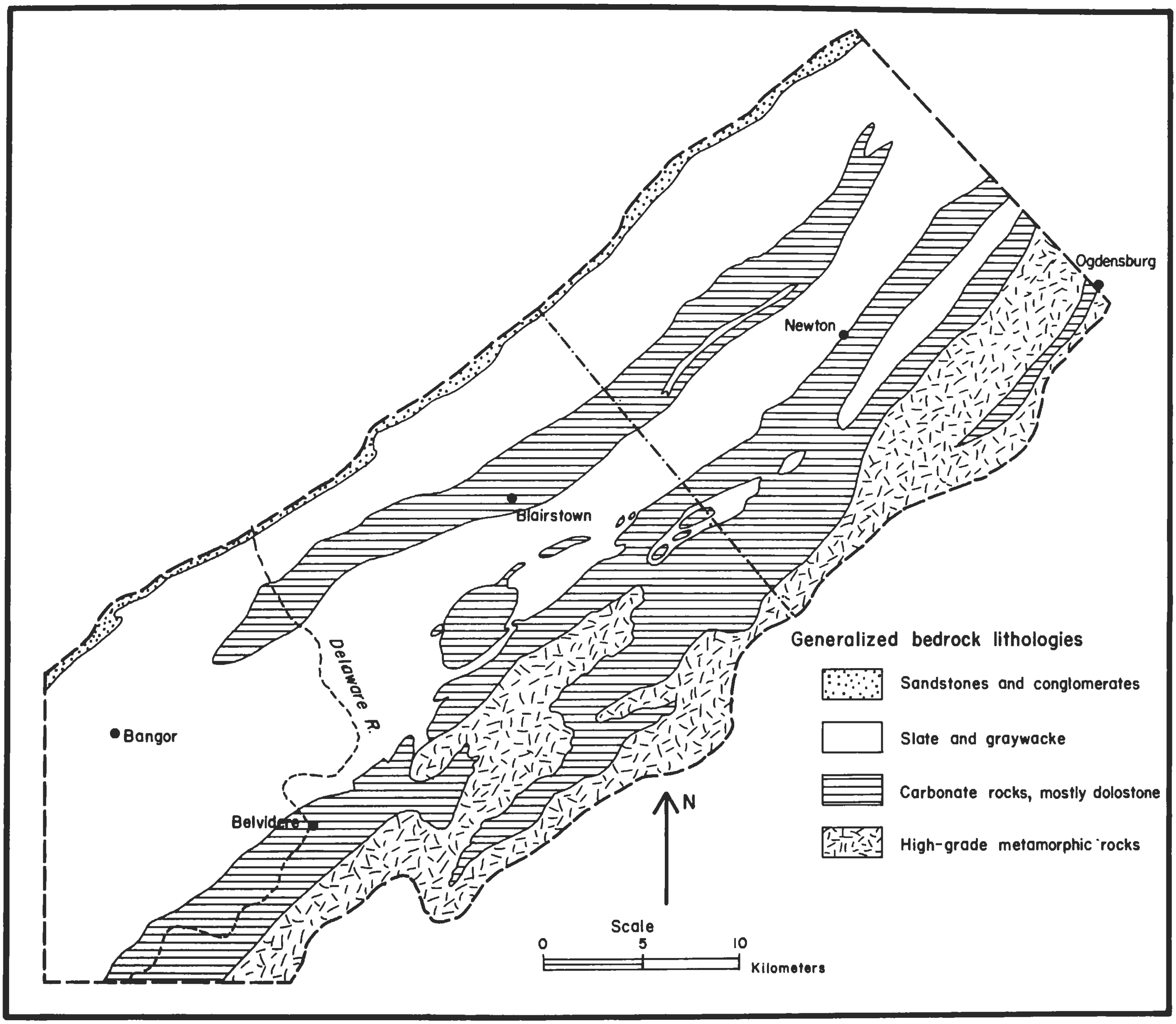


Figure 3 Generalized bedrock lithologies of the study area after Johnson (1950).

ceptions 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

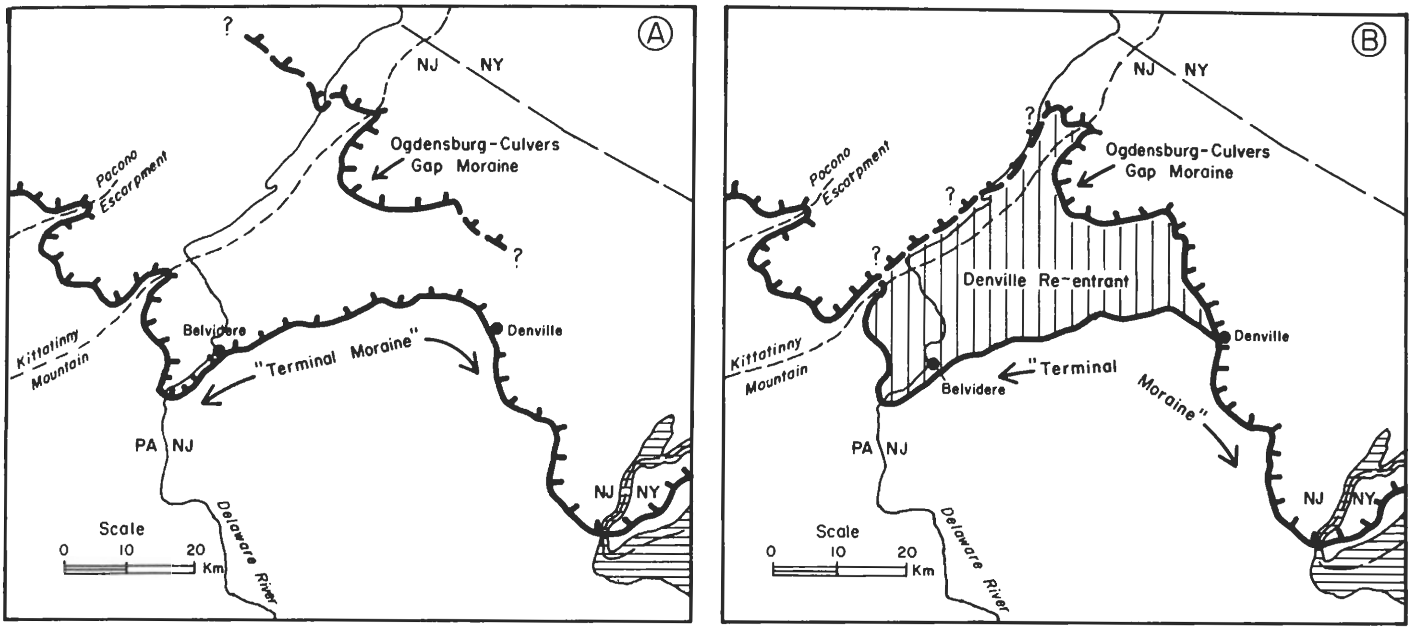


Figure 4 a) The Woodfordian terminal position and recessional moraine positions compiled from Salisbury (1902) and Sevon and others (1975).
b) The Woodfordian terminal position of Connally and Sirkin (1973) and Connally (1979).

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 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 Rhodehammel (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 because of 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 yrs BP. Crowl (1980), citing 12 "bog-bottom" radiocarbon dates ranging in age from 12,520 to 14,170 yrs BP 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 yrs BP. 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 yrs BP. 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 yrs BP deglaciation of the Ontario Lobe from northeastern Pennsylvania is not consistent with the 14,000" + " yrs BP 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 yrs BP 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 deosits (Marchand, 1978, Crowl and Sevon, 1980). No deposits of Altonian age (Early or 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 west 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 through 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 of coring), relative age (morphosequence affiliation), origin of lake basin, and postglacial modification. Field description of core samples included sediment type and color, organic

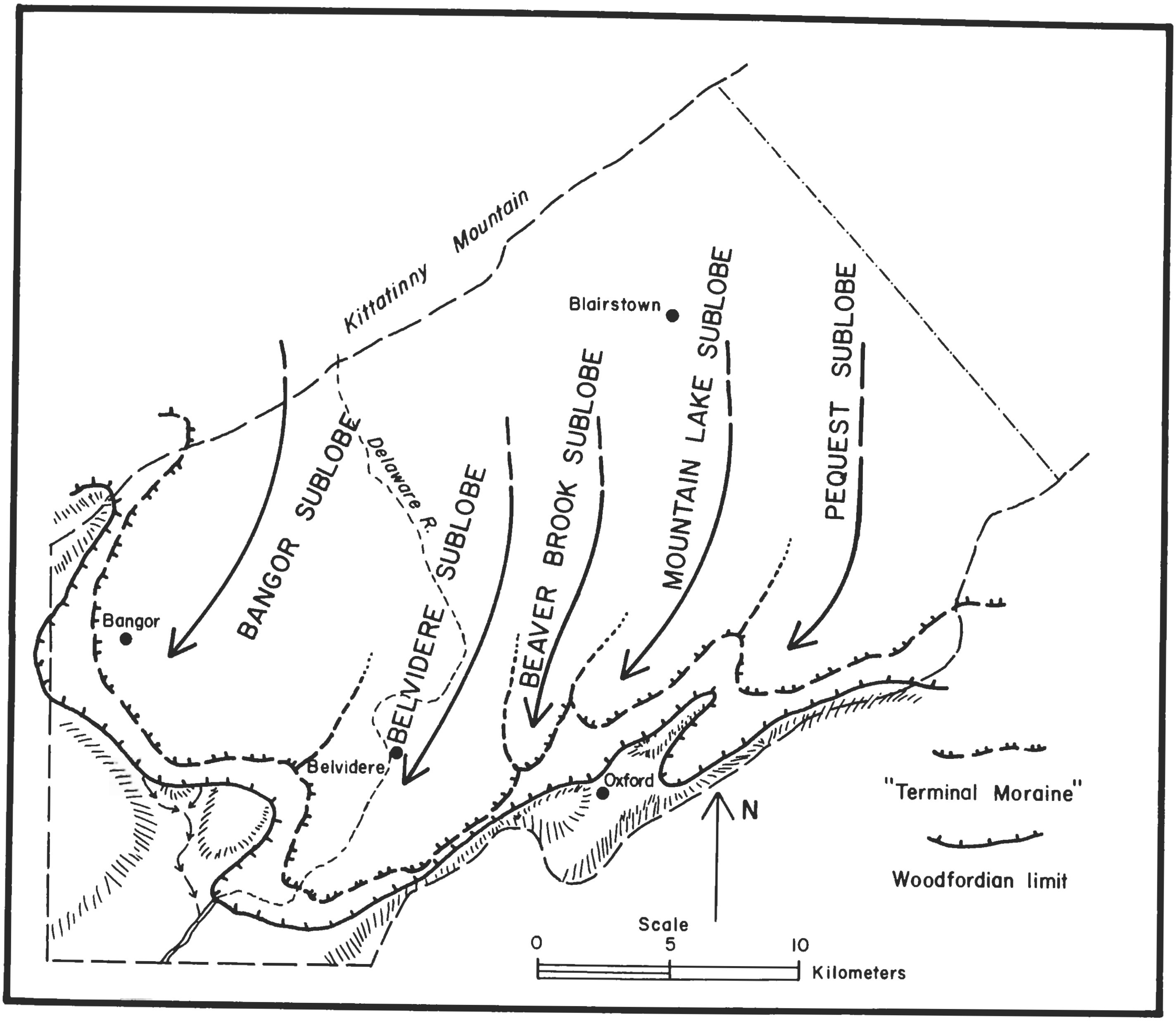


Figure 5 Sublobe nomenclature, flow patterns and geometry during the maximum extent of Woodfordian glaciation and deposition of the "Terminal Moraine" (Ridge, 1983).

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 sedi-

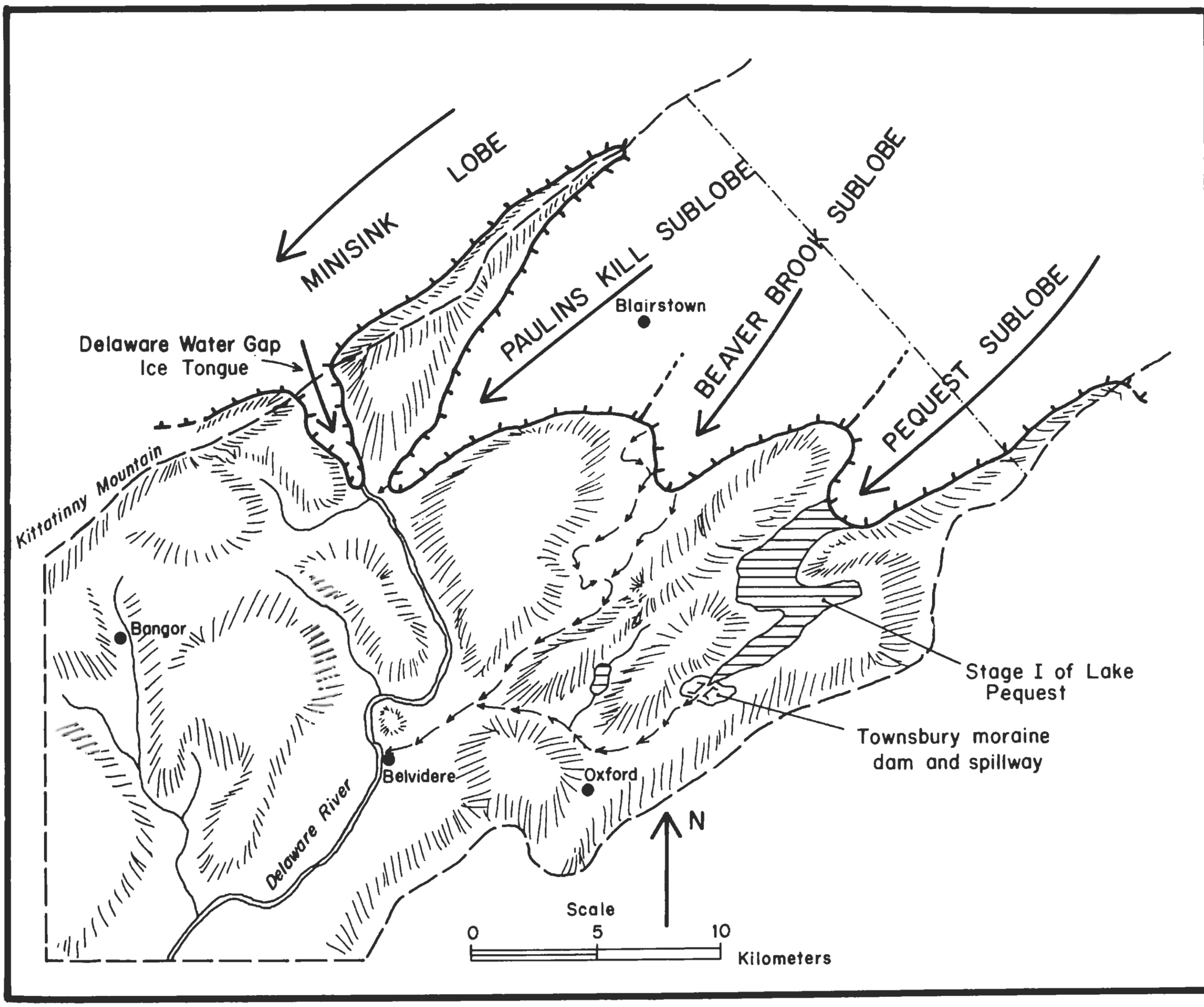


Figure 6 Sublobe nomenclature, flow patterns and geometry during early stages of the Woodfordian deglaciation (Ridge, 1983).

ments 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 previ-

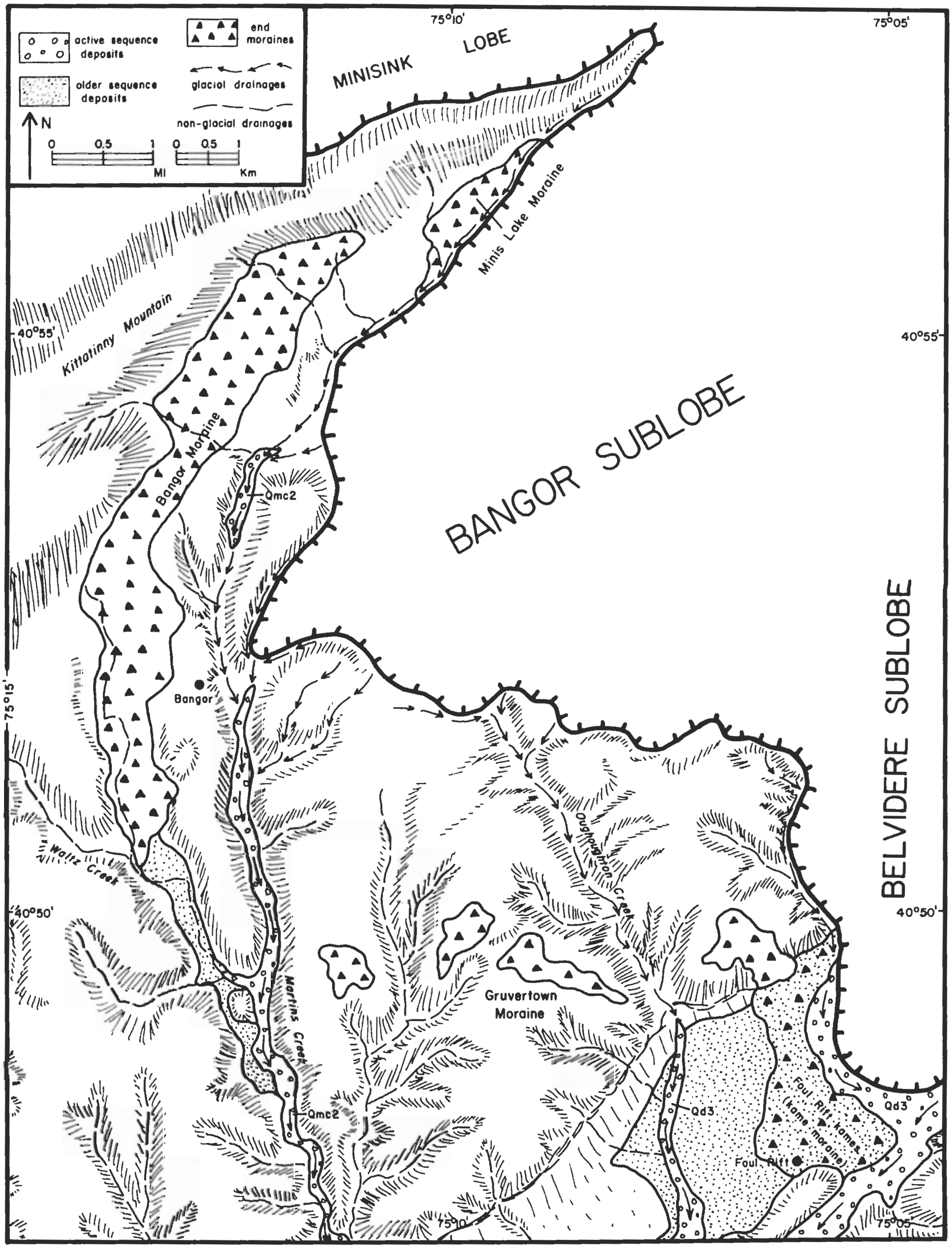


Figure 7 Location of the Bangor Moraine and reconstruction of the ice margin position during the deposition of the Minis Lake Moraine.

ously 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 duirng 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 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 called the Paulins Kill sublobe. The reorganization of sublobes was recognized on the basis of provenance (Ridge, 1983). Ice contact outwash deposits from the earliest phase of the Paulins Kill sublobe retreat originally were mapped by Lewis (1884) who named them the Portland Kames. Ridge (1983) mapped a kame delta segment of the Portland Kames that 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 recession of the Paulins Kill sublobe continued and 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 existence 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). Striae at Columbia, however, indicate southeast flowing ice. The source of these striae 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 data (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 sublobe (figure 10) forming Glacial Lake Paulins Kill (Ridge, 1983). Paulins Kill sequences 1 and 2 consist of ice-contact lacustrine sediments deposited in Lake Paulins Kill while deltaic materials of sequences 3 and 4 document the lowering of the lake. Dur-

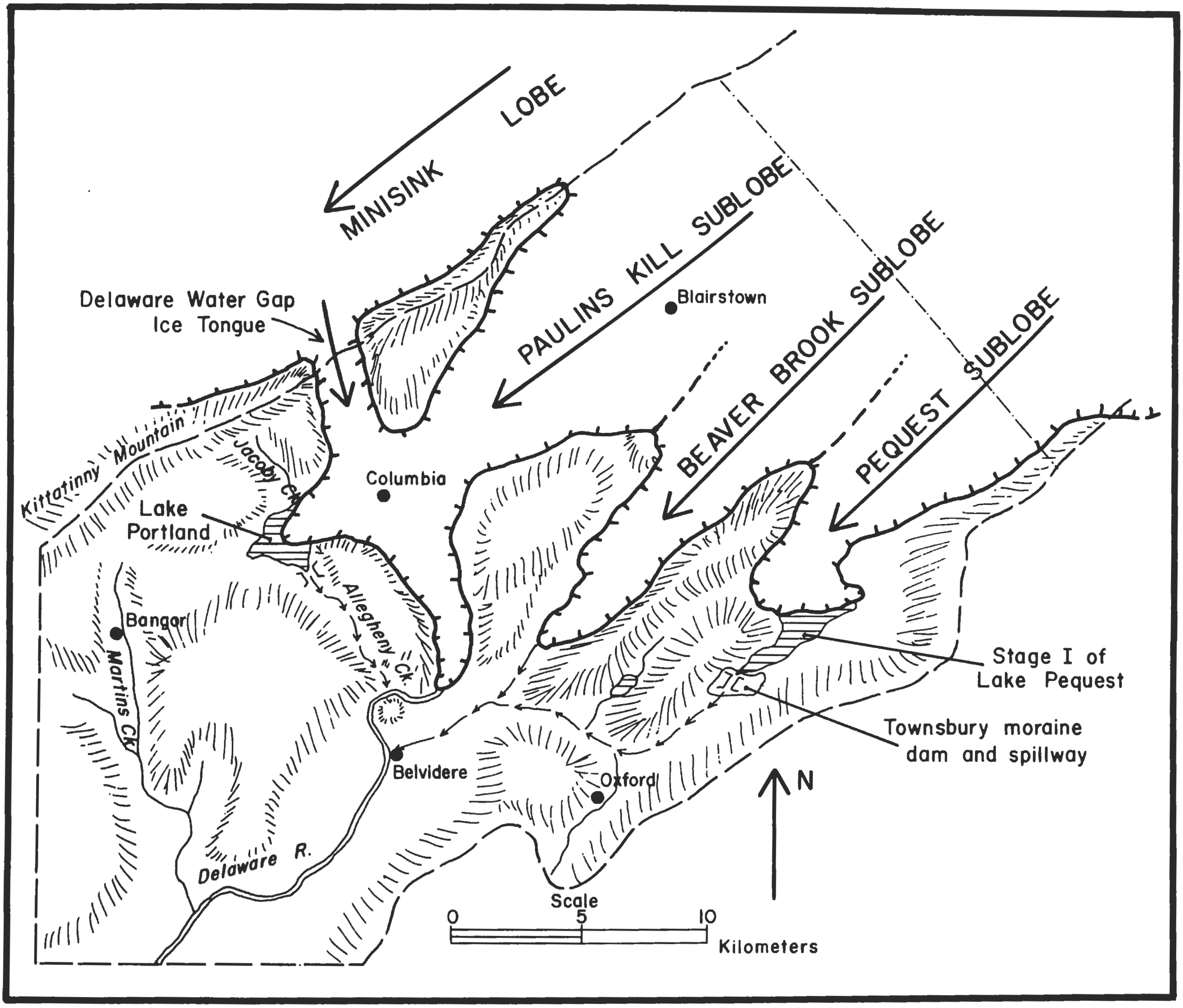


Figure 8 Reconstruction of the sublobe geometry during the formation of Lake Portland and Stage I of Lake Pequest.

ing 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). Reconnaissance 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

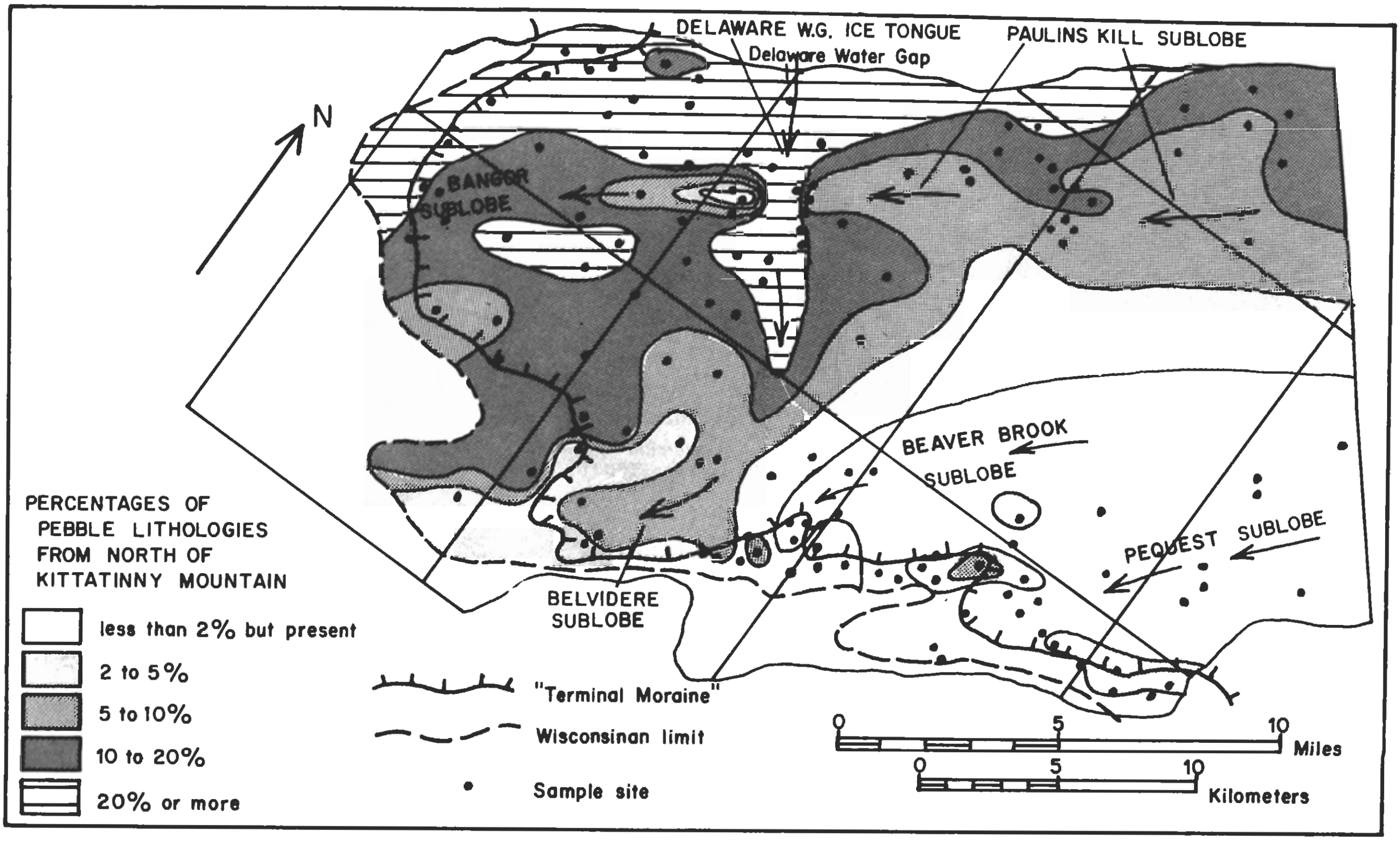


Figure 9 Quantitative distribution of exotic pebble lithologies (derived from north of Kittatinny Mountain) in Great Valley deposits.

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 (Qd1, 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 and end moraine complex and 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, 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 thinning 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.

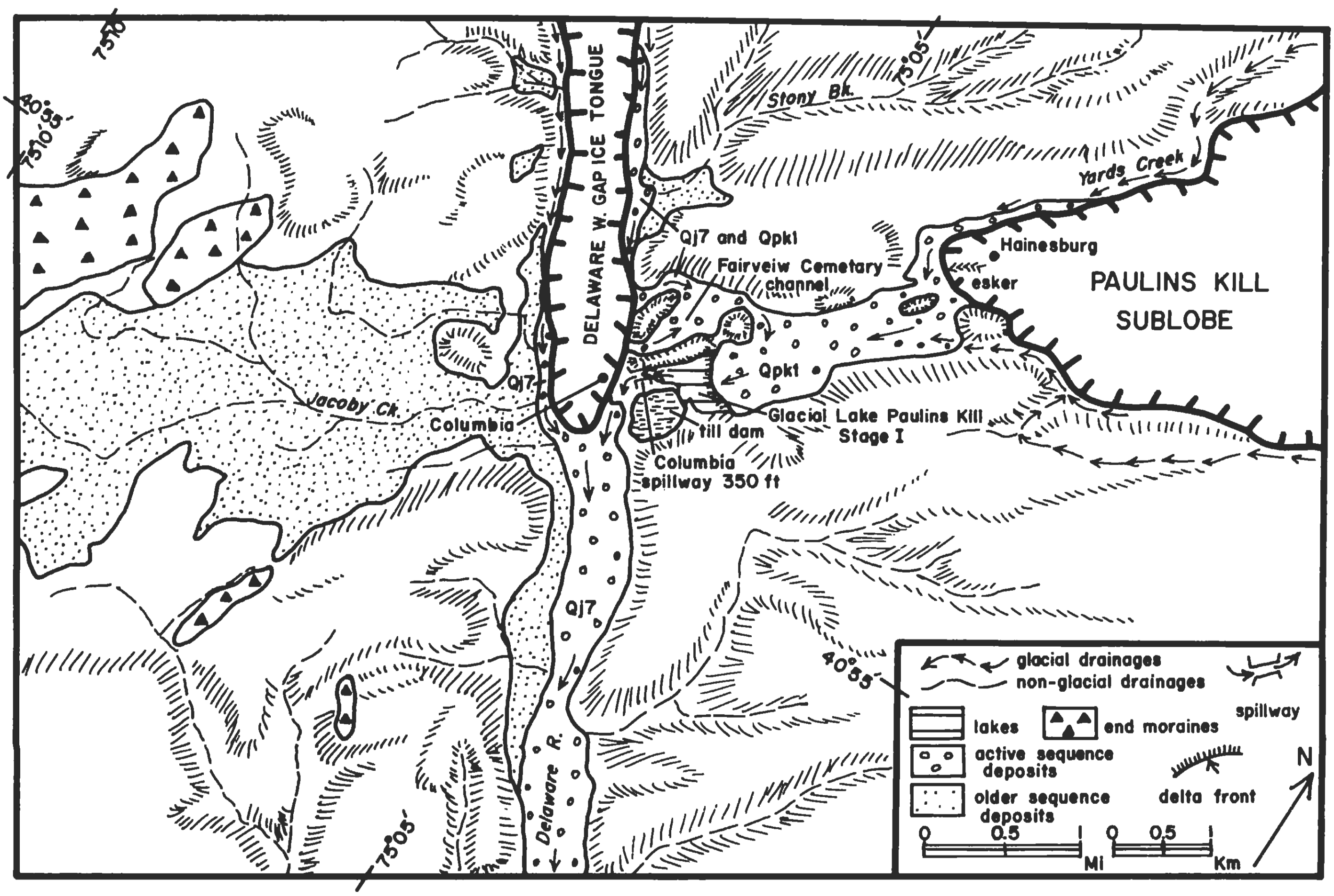


Figure 10 Reconstruction of the Delaware Water Gap ice tongue and the Paulins Kill Sublobe during Stage I of Lake Paulins Kill.

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 the last of these, the thinning ice sheet could no longer overtop 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 ice margin 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

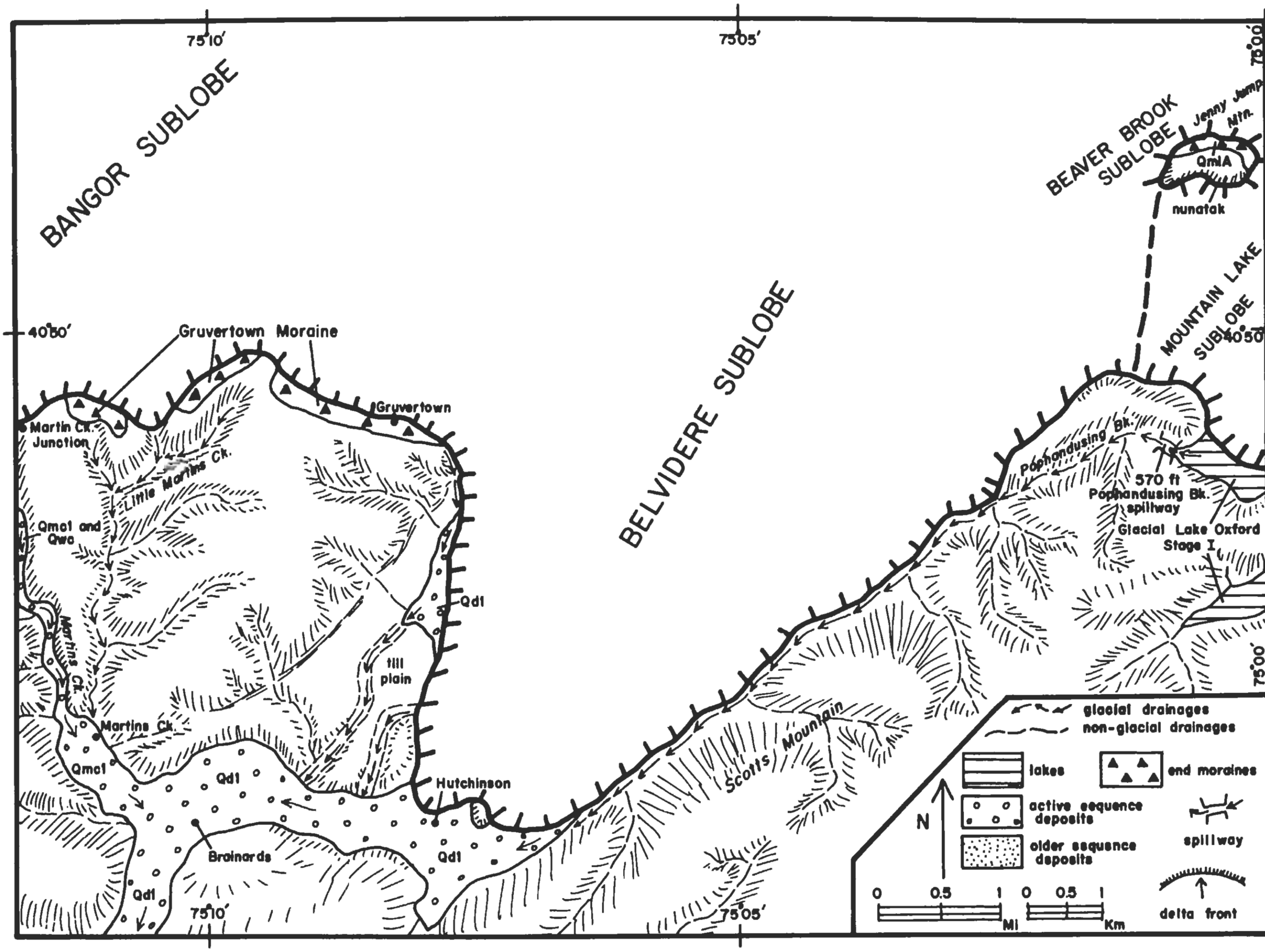


Figure 11 **Reconstruction of the Belvidere sublobe during the deposition of the Gruvertown Moraine, portion of the "Terminal Moraine."**

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 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 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.

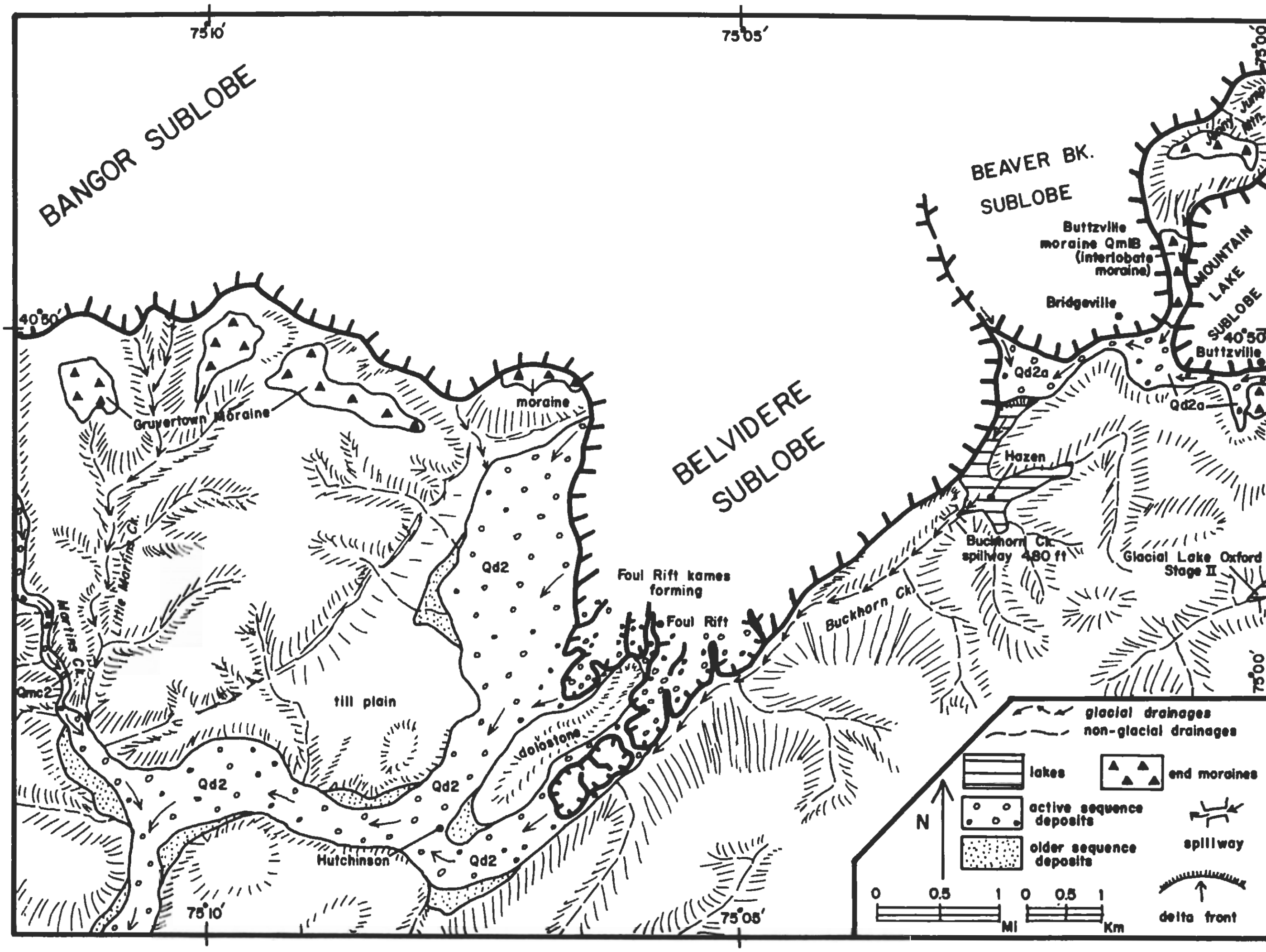


Figure 12 **Reconstruction of the Belvidere and Beaver Brook sublobes during the deposition of the Foul Rift kames (early stage, Qd2).**

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 or minor readvance 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 that lies south of the Ogdensburg-Culvers Gap Moraine. 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).

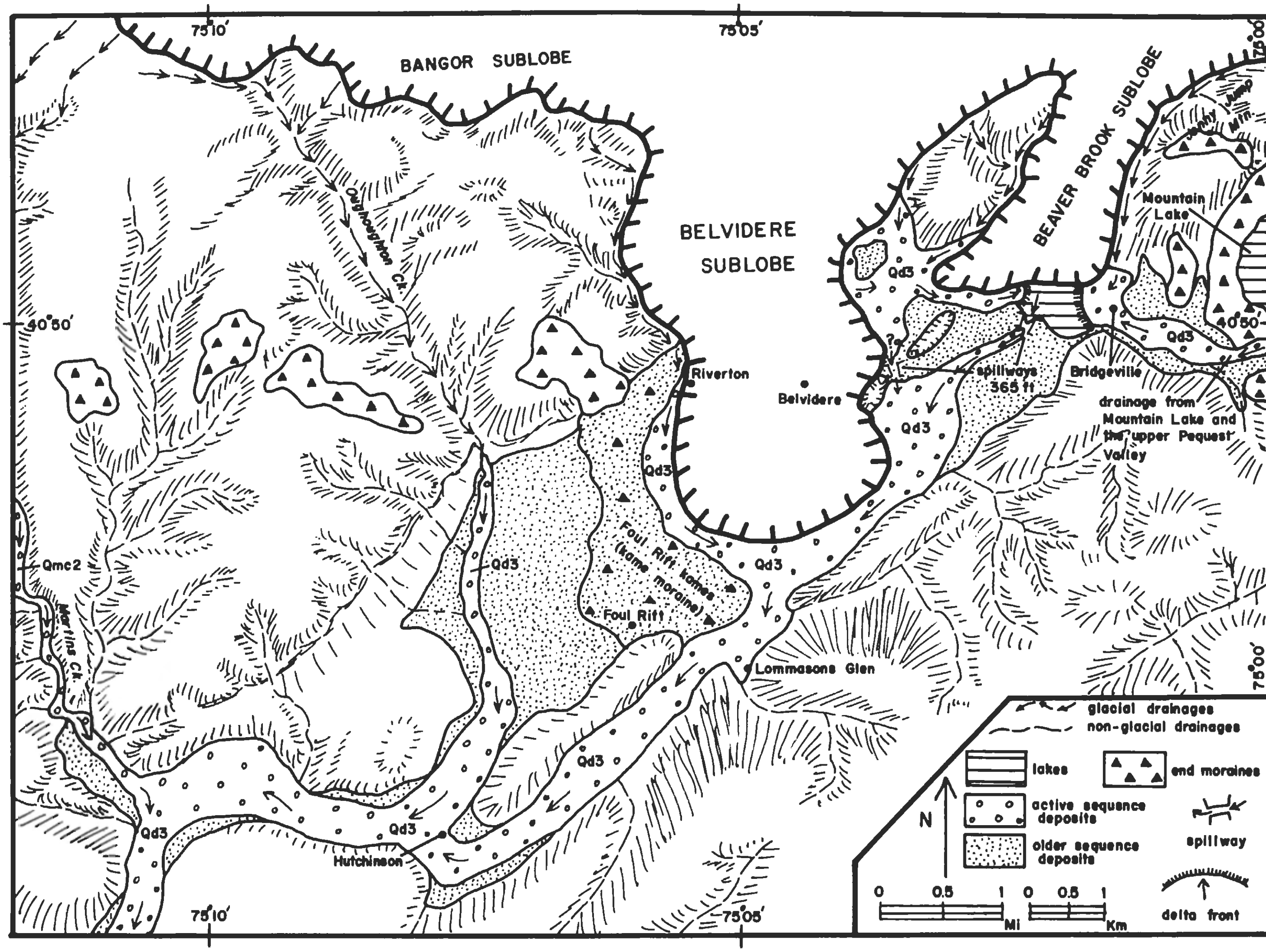


Figure 13 Reconstruction of the Belvidere and Beaver Brook sublobes during the deposition of the Foul Rift kames (late stage, Qd3).

IMPLICATIONS FOR REGIONAL ICE FLOW

Striae reported 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 the 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 Ontario-Hudson-Champlain suture to a position in eastern Pennsylvania (Figure 17b).

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.

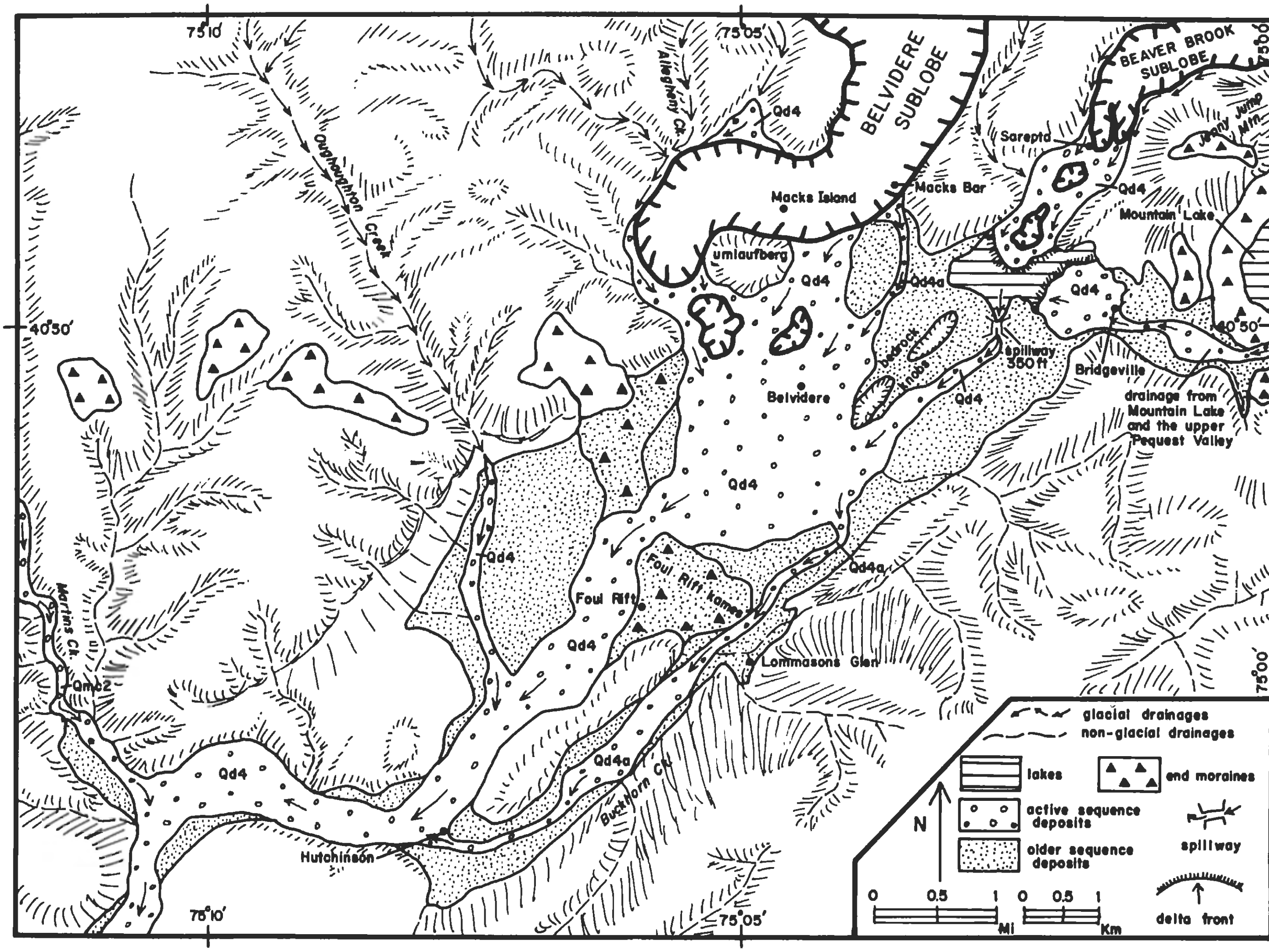


Figure 14 Reconstruction of the final stages of the Belvidere and Beaver Brook sublobes (latest stage, Qd4).

Lobate flow in the Hudson-Chamlplain 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.

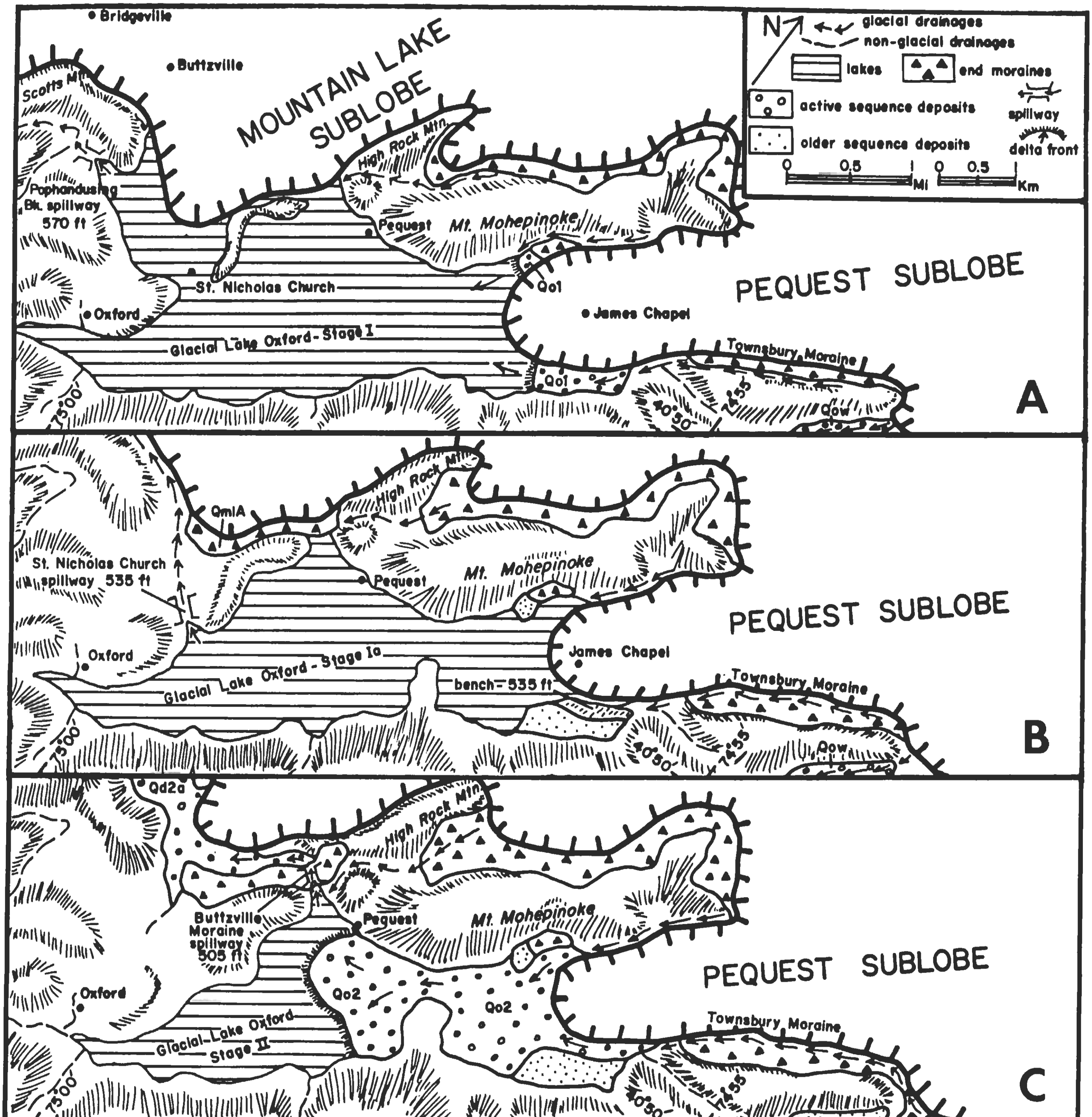


Figure 15 Three stages of development of Lake Oxford which was initially dammed by the Mountain Lake Sublobe and later dammed by end moraines in the Pequest Valley.

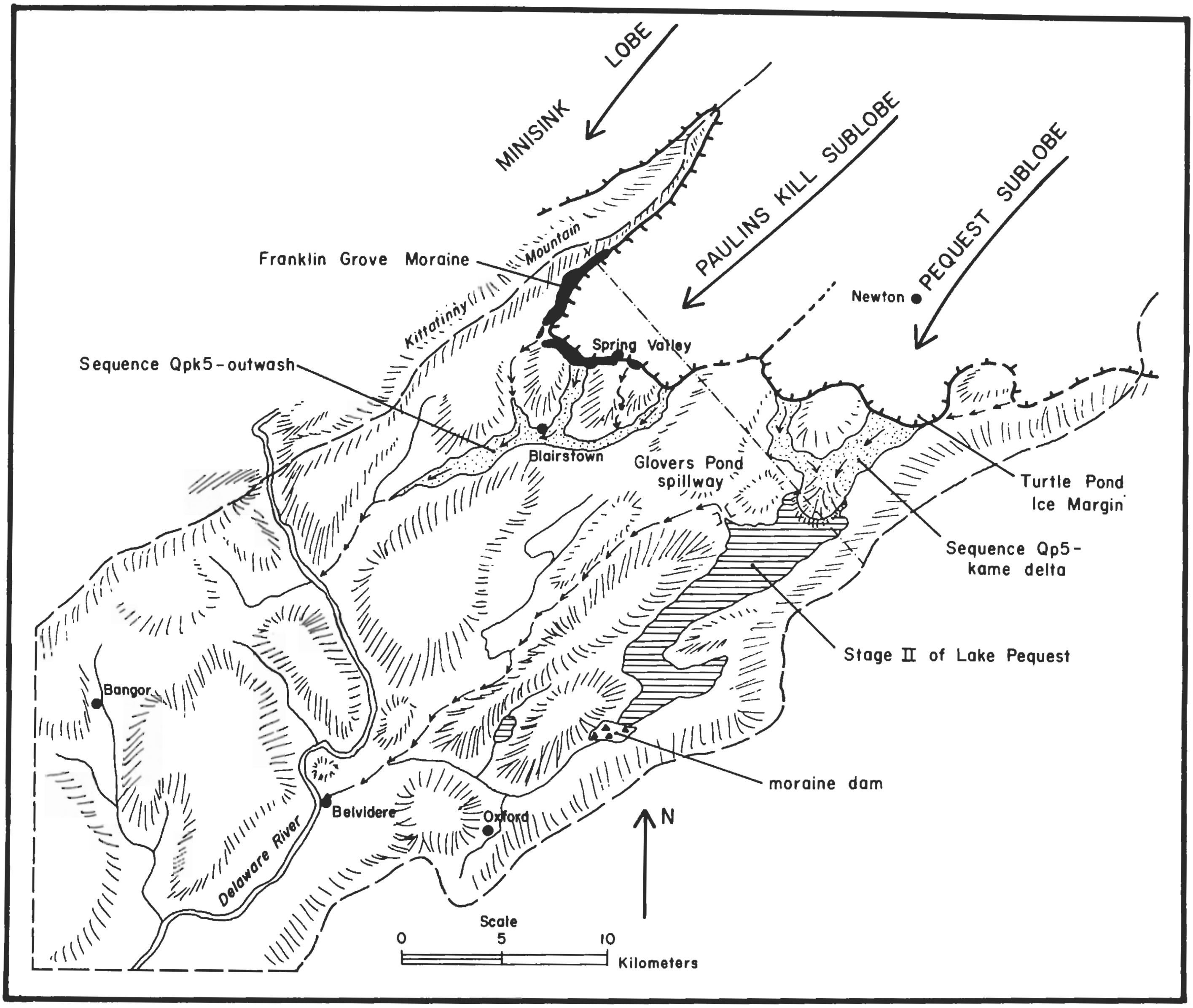


Figure 16 Reconstruction of sublobe geometry during the deposition of the Franklin Grove Moraine and the Turtle Pond Delta (stages Qpk5 and Qp5 and Lake Pequest II).

Nine soil profiles in both stratified and nonstratified 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 Rhodehammel, 1969; Sirkin and Minard, 1972). A

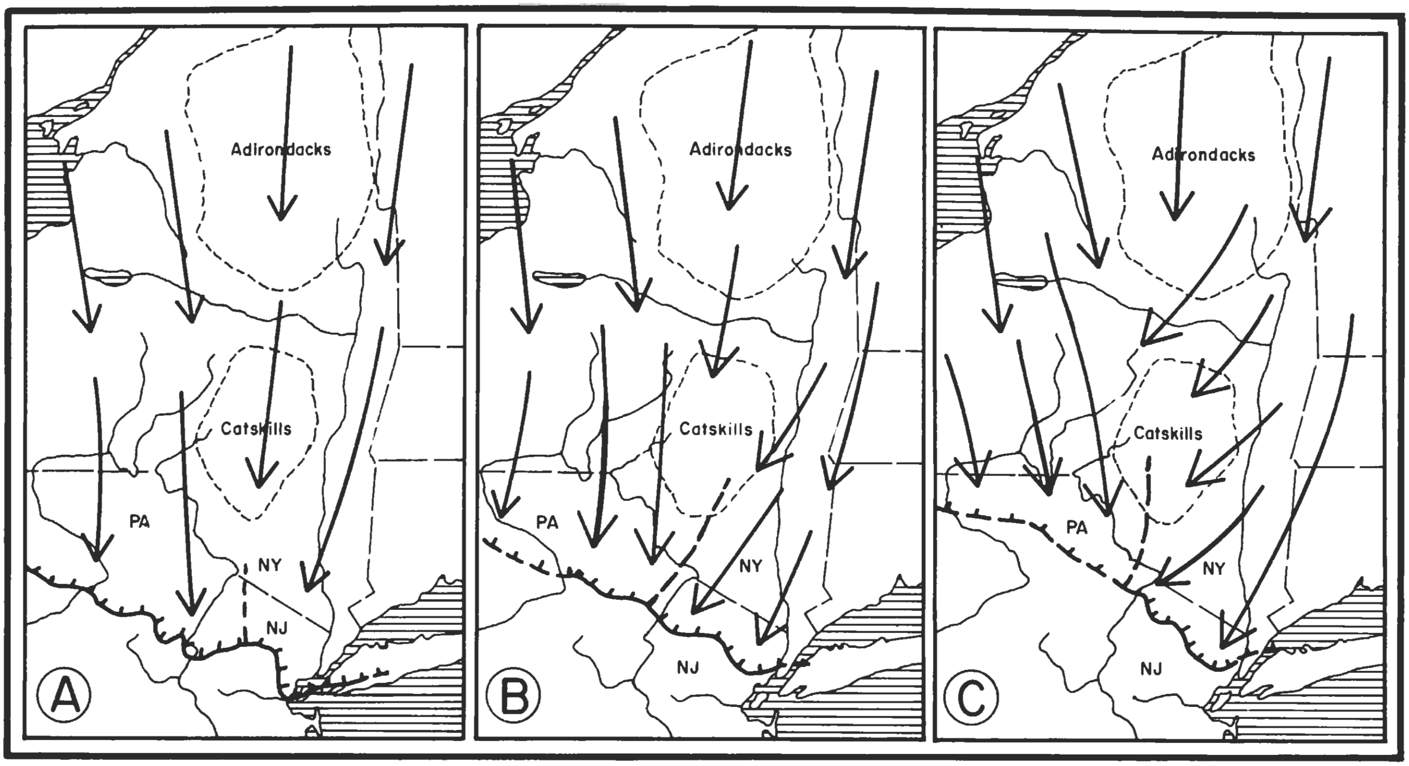


Figure 17 Inferred deployment of the Ontario and HudsonChamplain Lobes during Woodfordian deglaciation: (A) the Woodfordian terminal position, (B) the Franklin Grove Moraine, and (C) the OgdensburgCulvers Gap Moraine.

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 deglaciation. 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 yrs BP.

Francis Lake was formed in a depression between two dolostone bedrock knobs and a till plain. The glacial deposits which surround and underlie Francis Lake were deposited as part of 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, in review). A brief discussion is presented here.

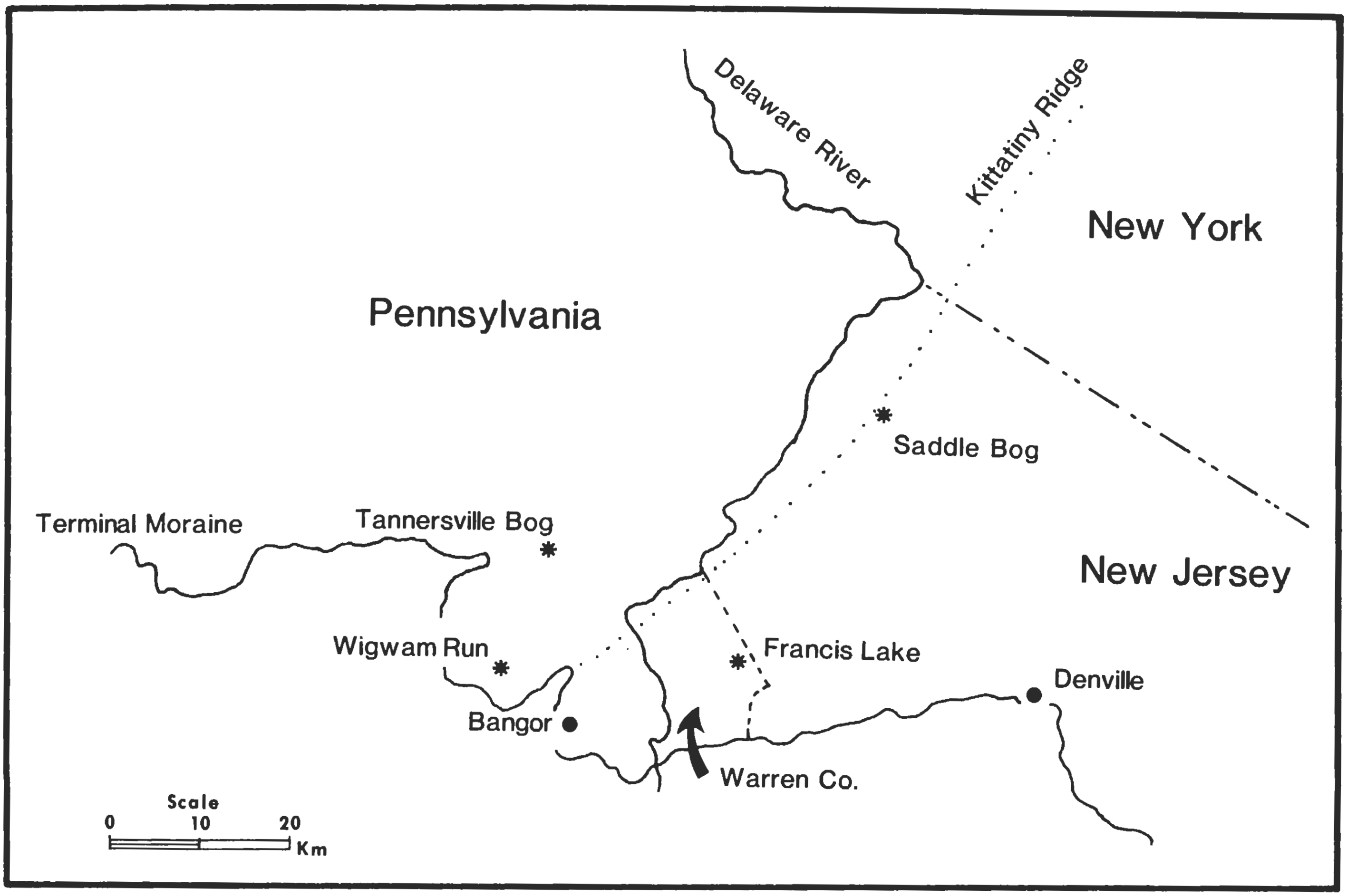


Figure 18 Pollen study sites discussed in text.

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.

The spruce zone (7.6 to 6.0 m) is characterized by 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 an open boreal forest, indicative of cold climatic conditions, was replaced first by a 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 zones 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 those of today, although variations in effective moisture resulted in drier conditions.

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-4921) 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 yrs BP (SI-5,301) (Figure 19). The pollen and radiocarbon data demonstrate that the region was deglaciated by 18,500 yrs BP and that tundra vegetation, as documented by the herb pollen zone, existed from 18,500 to about 14,275 yrs BP. 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 yrs BP 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 interpeted 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 yrs BP provide an absolute chronology for sedimentation from deglaciation to the establishment of a boreal forest (spruce pollen zone). This interval of tundra vegetation (herb pollen zone) lasted from prior to 18,500 yrs BP to approximately 14,250 yrs BP (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 yrs BP 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 yrs BP. Major floristic events during this period include a spruce peak followed by the expansion of fir into the region. A date of 13,510 yrs BP preceded the spruce peak at Francis Lake and a date of 13,330 yrs BP predates the fir expansion at Tannersville bog. A date of 12,300 yrs BP was obtained from the Saddle Bog immediately above the spruce peak, and dates of 11,430 and 11,220 yrs BP were obtained from the spruce zone/pine zone boundaries at Wigwam Run and Francis Lake, respectively. Dates of 10,860 and 9,835 yrs BP 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 11,200 yrs BP 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 yrs BP) are both valid and accurate.

CONCLUSIONS

At the maximum extent of the Woodfordian glaciation the Ontario and Hudson-Chamlplain 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.

The deglaciation of the Great Valley was characterized by both systematic ice retreat and rapid stagnation by divide cut-off. Deglaciation of the Delaware, Jacoby and Paulins Kill valleys, and portions of the Pequest Valley was by systematic ice retreat alone. 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

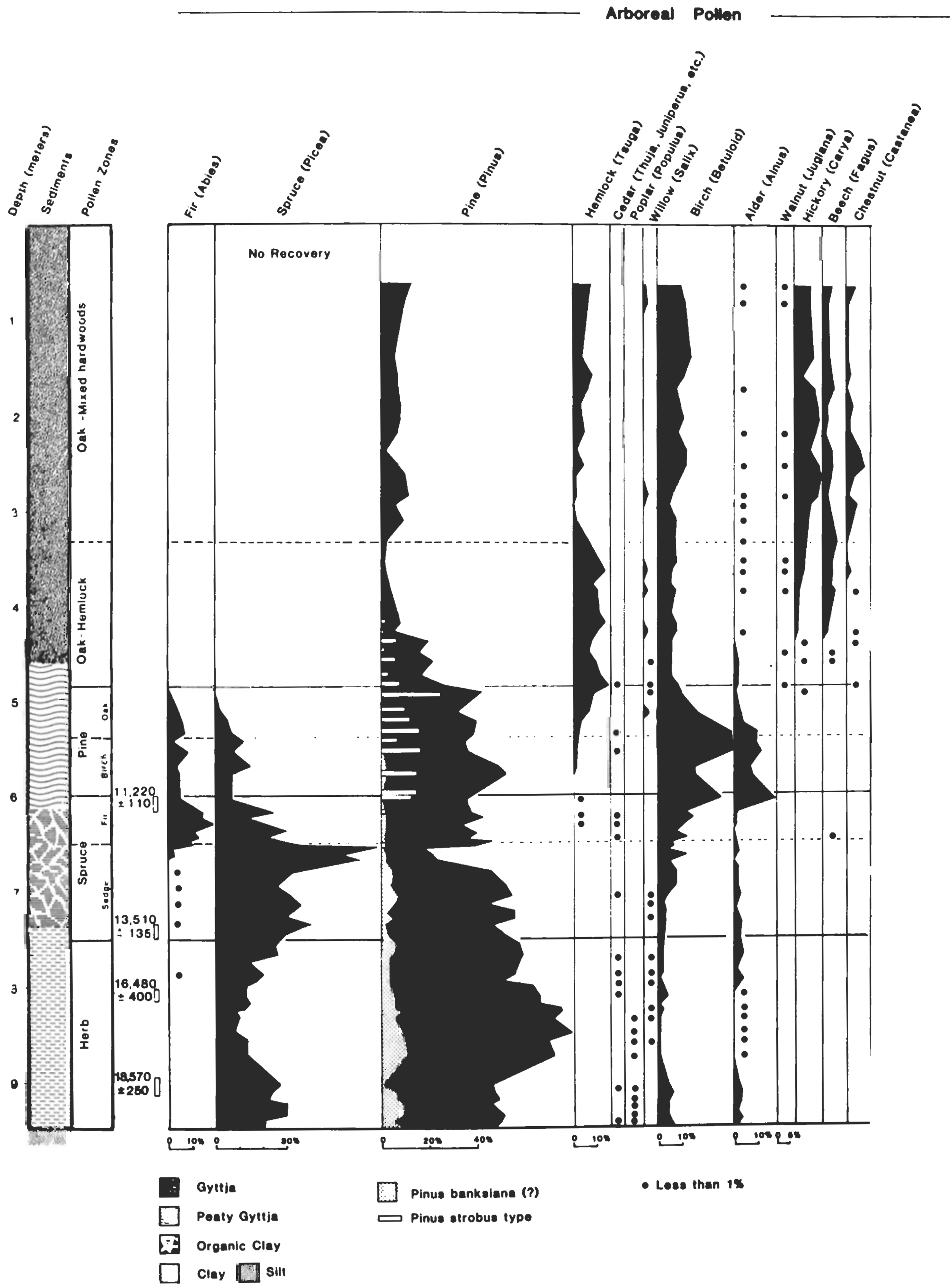
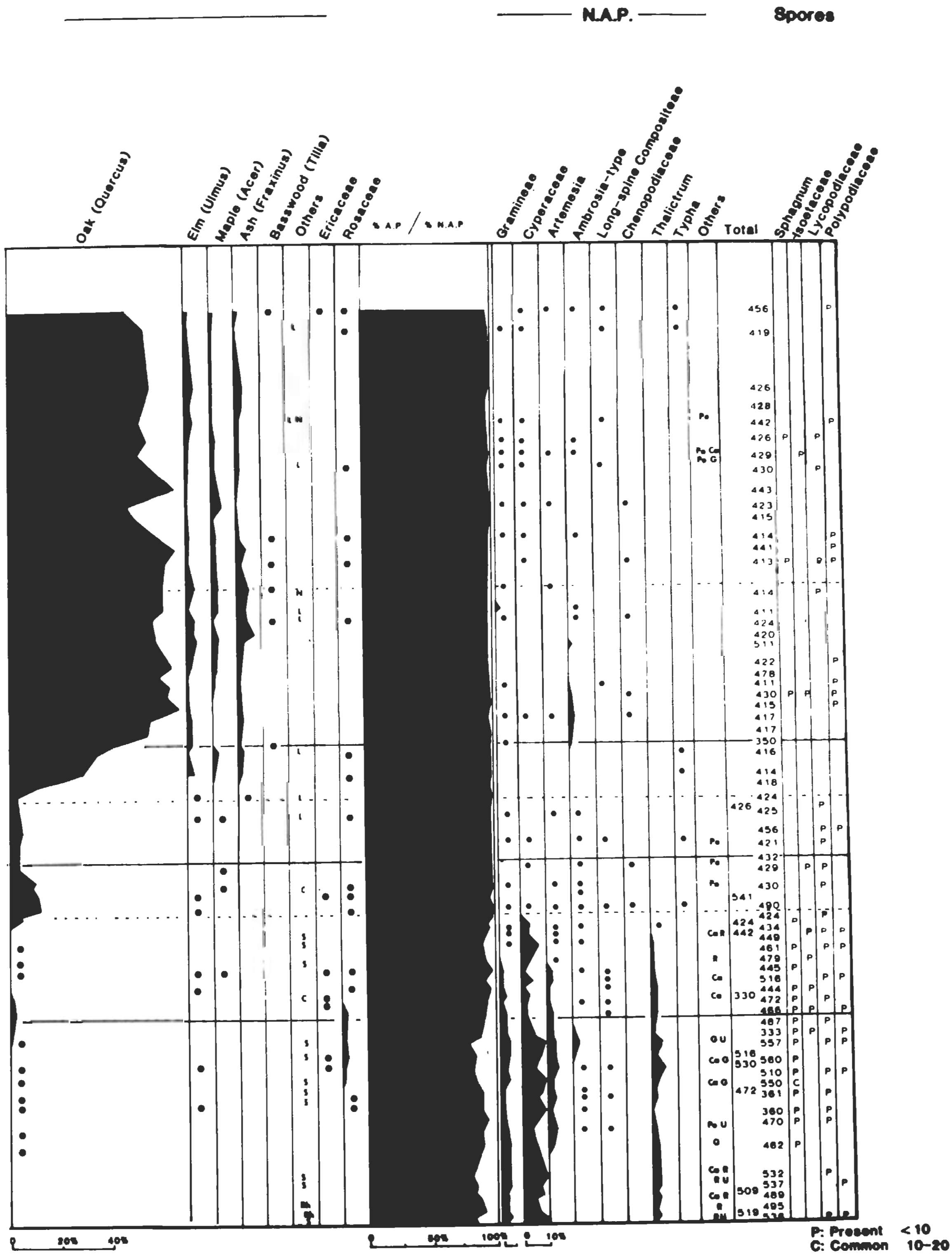


Figure 19 Francis Lake Pollen Diagram.



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

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

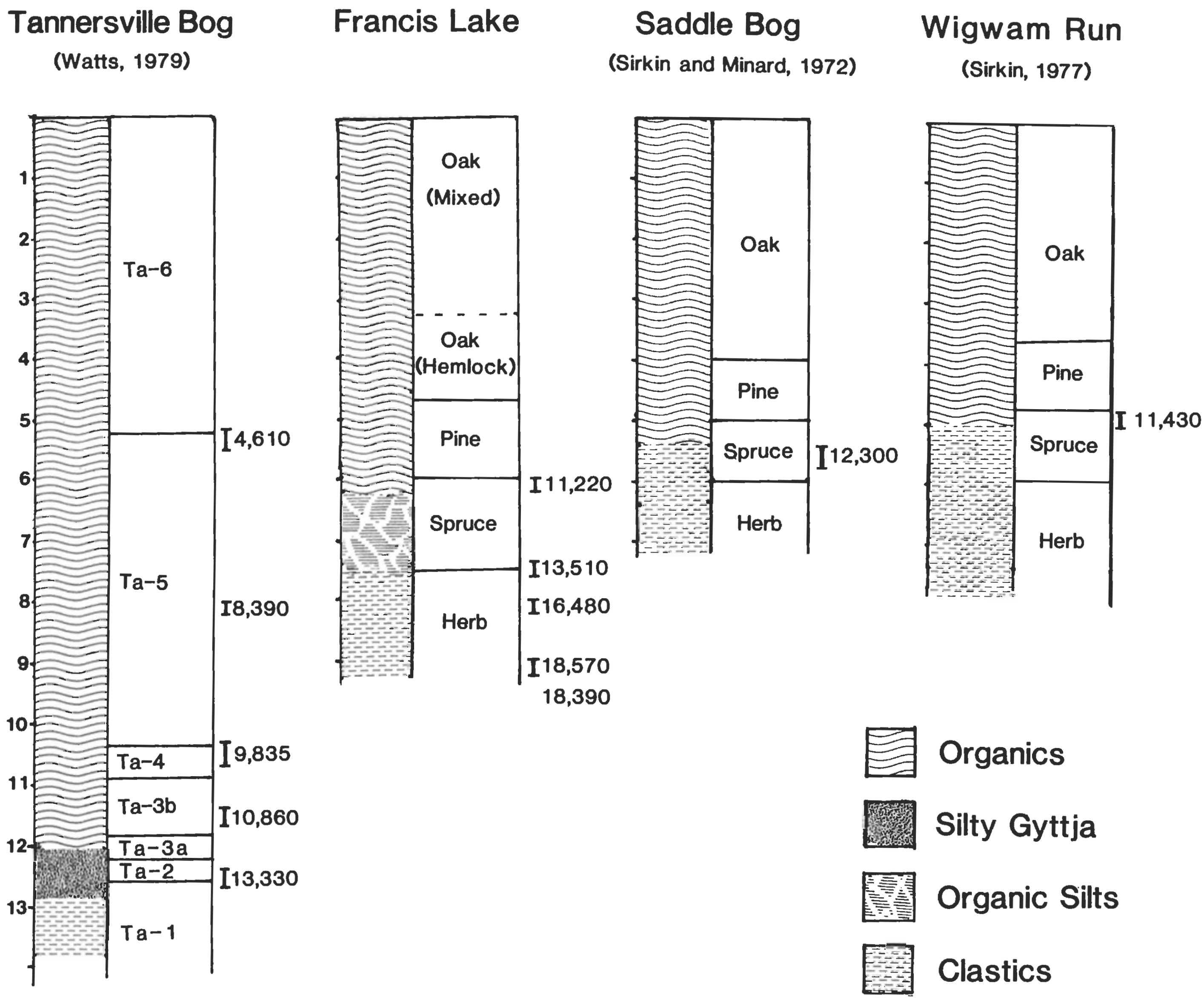


Figure 20 Generalized pollen and sediment stratigraphy and radiocarbon dates of four sites discussed in text (depth is in meters).

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 discernible 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 sub-

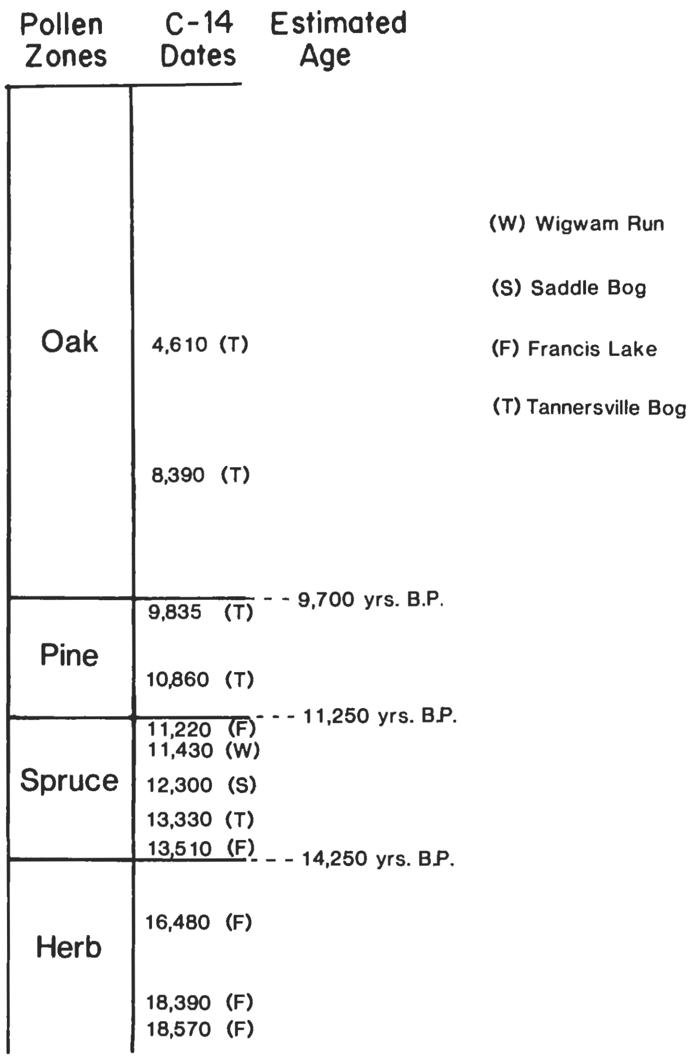


Figure 21 Diagram of composite pollen and radiocarbon stratigraphy of the study area and estimated ages of pollen zone boundary (depth is not to scale).

lobe 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 yrs BP in the Great Valley. Radiocarbon dates of 18,570, 18,390, 16,480, 13,510, and 11,220 yrs BP, in conjunction with pollen stratigraphy, document the presence of tundra vegetation in the Great Valley until 14,250 yrs BP. The dates from Francis Lake are verified through correlation with other radiocarbon-dated pollen stratigraphies of the region.

The data presented here resolve three 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. Second, the Ogdensburg-Culvers Gap Moraine is a recessional moraine of the Hudson-Champlain Lobe in the Great Valley. Third, Woodfordian deglaciation from the "Terminal Moraine" began prior to 18,500 yrs BP.

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Woodfordian Ice Margins, Recessional Events, and Pollen Stratigraphy of the Mid-Hudson Valley

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ABSTRACT

The Mid-Hudson Valley, bounded on the west by the Marlboro Mts., on the south by the Hudson Highlands, and on the east by the New England Uplands, comprises the Hudson River Lowlands, the low Taconics, high Taconics, and Harlem Valley in Dutchess County and southern Columbia County, New York. It includes western Ulster County, adjacent to the Hudson River trench and the northeast corner of Orange County, near Newburgh. Four ice margins, traced west to east across the region, are named and described, from south to north, the Shenandoah, Poughkeepsie, Hyde Park and Pine Plains Moraines.

As the Woodfordian glacier retreated from the Shenandoah Moraine to the Poughkeepsie Moraine, Lake Albany expanded northward in the Hudson River trench, as did Lake Fishkill in the Fishkill Creek, Sprout Creek, and Whortlekill valleys. At the Hyde Park Moraine the ice dammed Lake Washington in the North Branch of Wappinger Creek. At the Pine Plains Moraine the ice dammed Lake Attlebury in the headwaters of Wappinger Creek that expanded into the valley of the Roeliff Jansen Kill during recession.

The Red Hook Moraine is the fifth major ice margin but it was traced from northeast to southwest, cross cutting the trend of the Hyde Park and Pine Plains Moraines. At the outermost position the glacier dammed Lake Eighmyville in the Landsman Kill valley and Lake Jansen in the valley of the Roeliff Jansen Kill. There is no Lake Albany delta associated with the outermost position but recessional positions were traced to the Rhinebeck and Red Hook deltas. Lake Albany continued to expand northward during the disintegration of the glacier following the Rosendale readvance.

It is suggested that Glacial Lake Hudson, south of the Hudson Highlands, expanded northward into the Mid-Hudson Valley to become Glacial Lake Albany. The projected level rises northward due to regional isostatic rebound following deglaciation. However, the "water plane" defined by the ice margin deltas that were deposited sequentially during recession, raised above the stable level of Lake Albany, is inferred to be an artifact of differential local rebound of the Hudson Lowlands.

Accumulation of the sub-organic portion of Eagle Hill Camp bog began when the ice margin retreated to the innermost position of the Red Hook Moraine and deposition of the Red Hook delta began. A radiocarbon date of 13,670 $\pm$ 170 yrs BP enabled us to calculate an age of 16,070 yrs BP for the base of the pollen stratigraphic section. A truncated T zone is present at the base, followed by the A, B, and C pollen zones.

We propose that the Woodfordian glacier had receded to the Pellets Island/Shenandoah Moraine position by 17,950 yrs BP and to the Poughkeepsie/Wallkill Moraine position by 17,210 yrs BP. The Rosendale readvance took place about 16,100(?) years ago, crossing the trends of the earlier Hyde Park and Pine Plains Moraines. Recession from the Red Hook Moraine positions probably was completed by 16,070 yrs BP, following which the vegetation of the herb, spruce, pine, and oak pollen zones migrated into the region from the south.

INTRODUCTION

The Mid-Hudson Valley, or Mid-Hudson Region, is a loosely defined, quasi-political, economic entity (Fig. 1A). The southern boundary is readily defined as the mountains of the Hudson Highlands. The more diffuse northern boundary, separating the Mid-Hudson Region from the Capital District, probably is the Rip Van Winkle Bridge that crosses the Hudson River just south of the City of Hudson. It comprises most of Dutchess

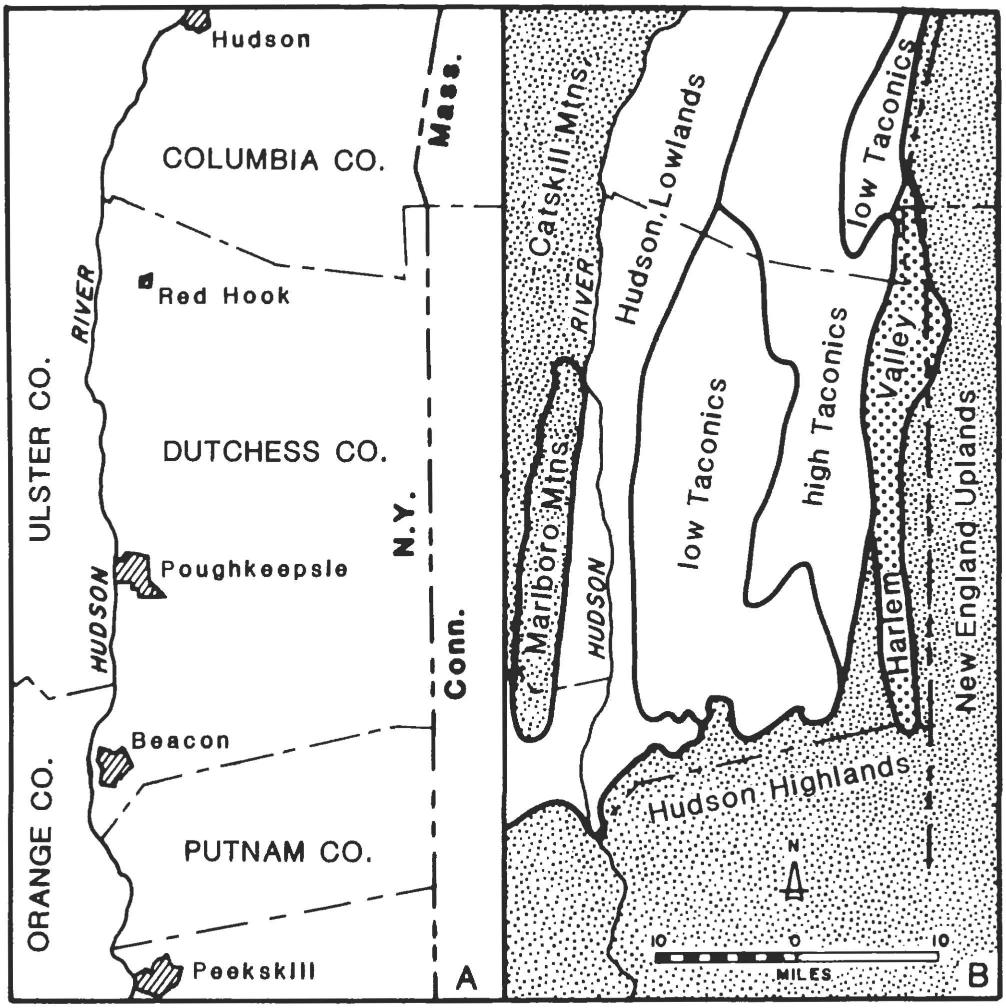

Figure 1 Location map for the Mid-Hudson Region. 1A, at left, shows the political units in, and south of the region. lB, at right, shows the physiographic subdivisions as used in this report.

County and southern Columbia County, extreme eastern Ulster County, and perhaps the northeast corner of Orange County. The region stretches westward to the Marlboro Mountains, including the west bank of the Hudson River, but generally excluding most of Orange and Ulster Counties in the Wallkill Valley and beyond. Most of the population is concentrated between the city of Poughkeepsie and the Highlands. On the west, the New York Thruway establishes a transportation link that ties Newburgh, in the south, and Kingston, in the north, more closely with the New York City and Capital District markets, respectively.

The Harlem Valley, in extreme eastern Dutchess and southeast Columbia Counties is a stepchild. Because of the lack of adequate transportation routes to the west, it is more closely allied with New York City, to the south, or New Haven, Connecticut, to the southeast. However, the Harlem Valley is included in this presentation even though its geologic ties with the Mid- Hudson Valley are at least as tenuous as the transportation links.

Physiography

For the purposes of this report, the Mid-Hudson Valley is subdivided into four physiographic units; from west to east, the Hudson Lowlands, the Taconic Hills (low Taconics), the Taconic Mountains (high Taconics), and the Harlem Valley (Fig. 1B). The Hudson Lowlands consist of the Hudson River trench and the low lying, glacially modified, berms on either side. The surface of the Hudson River is at sea level but within 1/2 km on either side, the berms rise to 30 to 90 m. Relief on the berms is generally less than 120 m with maximum hilltop elevations of less than 150 m. The west berm is only 3 to 5 km wide, but the east berm widens to 15 km in central Columbia County. The low Taconics reach rather gentle summits with crest elevations between 150 and 250 m. In eastern Dutchess County and central and eastern Columbia County, the high Taconics have rugged, rocky summits that reach elevations between 330 and 420 m.

The topographic grain is NNE to SSW, paralleling bedrock structures. Most of Dutchess County is drained by southerly flowing Casper Creek, Little Wappinger Creek, Wappinger Creek, Sprout Creek, and the Whortlekill. In the south, Fishkill Creek cuts across the topographic grain ENE to WSW along the base of the Hudson Highlands and empties into the Hudson River at the city of Beacon. Roeliff Jansen Kill, along the Dutchess/Columbia County line, also cuts across the bedrock grain from ESE to WNW, emptying into the Hudson south of the city of Hudson. The remainder of southern Columbia County exhibits deranged radial drainage that eventually flows around the high Taconics and then northwest across the broad eastern berm of the Hudson River trench.

The Harlem Valley, in the extreme east, is bordered on the east by the New England (Housatonic) Uplands and on the west by West Mountain, which gives way northward to a series of high knobs, ending at Fox Hill in southeastern Columbia County. The northern Harlem Valley is drained by Webatuck Creek which is joined by Wassaic Creek, near Amenia, and then becomes the Tenmile River until its confluence with the Housatonic River in western Connecticut.

Geology

Bedrock of the Mid-Hudson Valley comprises sedimentary and metasedimentary rocks deposited during the Ediacarian(?), Cambrian, and Ordovician Periods. Between the Marlboro Mts. and Wappinger Creek, bedrock consists of deformed pelitic rocks with weak to strong slaty cleavage. East of Wappinger Creek, the high Taconics owe their relative resistance to higher rank schists while the lowlands are underlain by Cambrian and Ordovician limestones and dolostones. The carbonates were metamorphosed to marble in the Harlem Valley. The Hudson Highlands consist of Prepaleozoic granitic gneisses. The bedrock geology is illustrated in Fisher and Warthin (1976, Fig. 1 and Fig. 2).

During Pleistocene glaciations, bedrock was scratched, polished, and eroded by repeated glacial advances. Thick deposits of glacier-derived diamict are present on the berms flanking the Hudson River trench. Drumlins commonly have 60 m of relief. Yet, the low Taconics exhibit only a thin veneer of diamict and the high Taconics generally are barren of glacial drift. Stratified drift is relatively common in stream valleys. Present-day topography dates from recession of the most recent, Woodfordian glacier.

Previous Work

Prior to the quadrangle maps produced for this study, no surficial maps existed for the Mid-Hudson Valley. The best approximations were Plates 5-7 of Woodworth (1905, p. 115-128) who depicted a few ice contact terraces and glaciolacustrine deposits along the Hudson River and the sketch map by Cook (1942, Fig. 78) for selected deposits north and south of the village of Elizaville in southern Columbia County.

During the early part of this century, the New York State Geological Survey sponsored a series of mapping projects which were reported in New York State Museum Bulletins. The maps were prepared at a scale of 1:62,500 on topographic maps of the old 15' series. Five of those maps (Fig. 2) concern the Mid-Hudson Region.

Holzwasser (1926, p. 71-76), who mapped the Newburgh 15' Quadrangle, and earlier Ries (1895), who mapped Orange County, mention the same "terraces" on the west bank of the Hudson that were reported by Woodworth (1905), but the discussions are less precise. Similarly, Berkey and Rice (1921, p. 145-147), who mapped the West Point 15' Quadrangle, noted the Peekskill delta, but little else. In contrast, Gordon (1911, p. 96-104), reporting on the Poughkeepsie 15' Quadrangle, gave excellent descriptions of a few features and documented them with photographic plates. Finally, Cook (1942, 1943, 1944), who mapped the Catskill, Coxsackie, and Katterskill 15' Quadrangles, respectively, illustrated selected features that supported Cook's interpretation of Glacial Lake Albany. In fact, the features cited earlier by Woodworth (1905) and Fairchild (1919) were those that supported their disparate views of the "water planes" in the Hudson Valley, with Woodworth's observations proving to be the more reproducible of the two.

More recently, Murray (1976) made a careful study of the Millbrook area in Dutchess County for the New York State Geological Association field trips hosted by Vassar College. The work of our study has been reported by Woodward-Clyde (1980) and Connally (1983) in preliminary form.

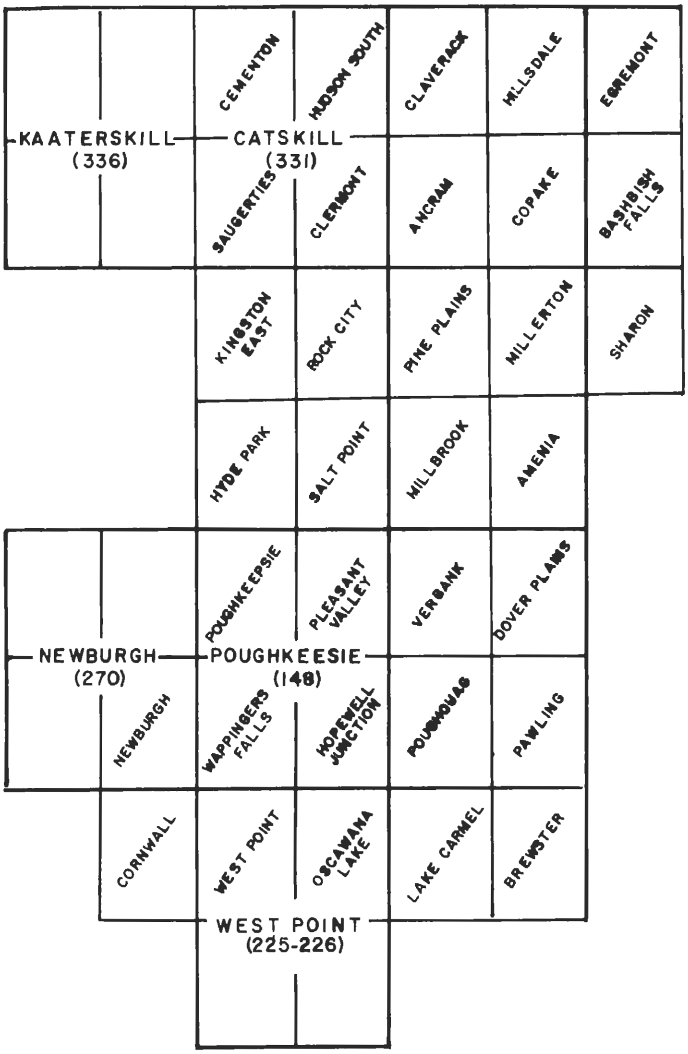

Figure 2 **An index to the quadrangles of the Mid-Hudson Region. Surficial maps for all 7 1/2 minute quadrangles except Brewster, Cornwall, Lake Carmel, and Newburgh are on open file with the New York State Geological Survey. The geology of the Kaaterskill, Catskill, Newburgh, Poughkeepsie, and West Point 15 minute quadrangles is reported in New York State Museum Bulletins 336, 331, 270, 148, and 225-226, respectively.**

The Present Study

This report is based on detailed mapping of the surficial geology of the region that has been underway since 1964. Stratified deposits were mapped first, at a scale of 1:24,000, and in great detail due to their economic potential. During 1977 and 1978, Connally mapped the near-site area for the proposed Red Hook-Clermont power site (Woodward-Clyde, 1980) in great detail. In the spring of 1979, we cored Eagle Hill Camp (EHC) bog, retrieving an 11.15 m core and Sirkin performed a detailed pollen analysis of that core. During 1980 and 1981, Connally completed reconnaissance mapping of the entire area (Fig. 2) on a quadrangle basis for the New York State Geological Survey. All 29 7-1/2' quadrangles, shown in Figure 2, are on open file with NYSGS.

The primary purposes of this report are (1) to document the field work, (2) to summarize the conclusions resulting from the field work, and (3) to serve as a reference for those who visit specific field sites. Because it is not possible to publish individual quadrangles, nor to include all geographic details in the figures of this report, reference is made to each quadrangle as an aid to those who visit the field sites. Also, all figures for vertical elevations (but not horizontal), in the sections on ICE MARGIN POSITIONS and GLACIAL LAKE ALBANY, are given in feet rather than meters, in deliberate violation of modern editorial guidelines. Because New York quadrangles give elevations in feet, it seems senseless to convert those figures from feet to meters, thus forcing the reader to reconvert from meters to feet, just for the sake of editorial consistency.

ICE MARGIN POSITIONS

Many ice margin positions are present in the Mid-Hudson Valley, dating from the retreat of the late Wisconsinan, Woodfordian glacier. They range from isolated heads of outwash to bands of discontinuous constructional topography to constructional end moraines with enough lateral extent to warrant formal names. Three ice margins have been traced from the Hudson River to West Mt., the high Taconic ridge that separates the Harlem Valley from the rest of the Mid-Hudson Region. A fourth ice margin position is continuous across the north end of West Mt. into the Harlem Valley. The fifth, and youngest, ice margin cross cuts the others along the edge of the low Taconics. These five ice margins, illustrated in Figure 3, are continuous enough to receive formal names as follows:

Red Hook Moraine
Pine Plains Moraine
Hyde Park Moraine
Poughkeepsie Moraine
Shenandoah Moraine

The use of the formal term "Moraine" to define these ice margin positions is open to criticism and the names are proposed advisedly. We have chosen only ice margin positions that consist of at least one major segment of constructional topography traceable for 8 to 25 km. To a midwesterner, who insists that moraines be continuous till (or diamict) ridges, the Mid-Hudson positions might appear merely as fragmentary deposits of stratified drift. To a New Englander, who is accustomed to disconnected heads of outwash that may be impossible to relate from one valley to another, the Mid-Hudson positions may offer gratifying evidence of lateral continuity. To someone from western New York, where most named moraines are defined by lines of stratified drift, the Mid-Hudson moraines should prove acceptable, if spotty in places.

Where topographic relief disappears, or is subdued, proglacial drainage features, drainage derangements, and soils were used to establish continuity of the positions. When all else failed, we relied on the intuition gained through 20 years of field work in the area. No gaps of more than 5 km exist in any of the named moraines. Although most of the names for these moraines have been used informally (e.g., Connally, 1983; Woodward-Clyde, 1980) they never before have been named formally. As the ice margin positions are discussed, the names of 7-1/2' quadrangles are given in parentheses, following descriptions of key features, as a further reference to the reader. Geographic features not shown on Figure 3 will be found by the interested reader on the 7-1/2' topographic maps.

Shenandoah Moraine

The oldest and southernmost ice margin position is here named the Shenandoah Moraine for a ridge of stratified drift at the base of Shenandoah Mountain (Hopewell Junction) that is 4.7 km long and has 180 feet of relief. Gordon (1911, Plate 21) referred to this ridge as the Shenandoah kames. Stratified drift was traced westward along the north slope of the Hudson Highlands to the east flank of Honess Mt. Drift appears to be absent from the west flank of Honess Mt., but a large valley stopper (MacClintock and Apfel, 1944, p. 1149) kame delta, exposed in a pit operated by Southern Dutchess Sand and Gravel Co., marks the terminal position in the valley of Clove Creek. Kame deposits at Dutchess Junction, and associated outwash at $\pm$ 140 feet at Beacon, both just south of the confluence of Fishkill Creek and the Hudson River (West Point) are correlated tentatively with the Shenandoah Moraine.

To the east, the Shenandoah Moraine was traced to thick stratified deposits at Greenhaven State Prison and to similar deposits around the village of Poughquag. Although the moraine was not traced across West Mountain, we suggest tentative correlation with the thick pitted outwash at Pawling (Pawling) in the Harlem Valley. The flat-topped outwash was deposited in a lake with its surface at $\pm$ 480 feet and an outlet somewhere to the south, in the Croton River drainage system.

A recessional position was traced along the flank of Clove Mt. (Verbank) and across into the Clove Valley adjacent to the hamlet of Clove. This position probably is related to pitted outwash at Fishkill Plains (Hopewell Junction) even though the position cannot be traced back west, across the Whortlekill and Sprout Creek lowlands.

Lake Fishkill

As the glacier began to retreat from the Shenandoah Moraine, a proglacial lake formed in the Fishkill lowland with a dam, at $\pm$ 220 feet, since removed by erosion between the city of Beacon and Fishkill village. Evidence for this lake is a sheet of outwash, up to 20 feet thick, over lacustrine clay. The outwash-choked lake is here named Glacial Lake Fishkill.

As the ice margin retreated northward, Lake Fishkill expanded up the Sprout Creek and Whortlekill lowlands. A period of quiet water sedimentation preceded deposition of an outwash valley train that must eventually have filled the lake. Outwash deposition is documented at the recessional position at the head of the Whortlekill valley where the valley train rises to 350 feet. In the Sprout Creek lowland the glacier continued to furnish outwash even when the ice margin stabilized at the Poughkeepsie Moraine. The outwash of the Sprout Creek valley train has a lower gradient than in the Whortlekill lowland, reaching a height of only 330 feet, 6 km north of the head of outwash in the Whortlekill. The fact that thick, extensive clay deposits underlie the prograded outwash suggests either a readvance to the Poughkeepsie Moraine or, more likely, that that ice margin was continuously occupied long enough for Lake Fishkill to have become filled with sediment.

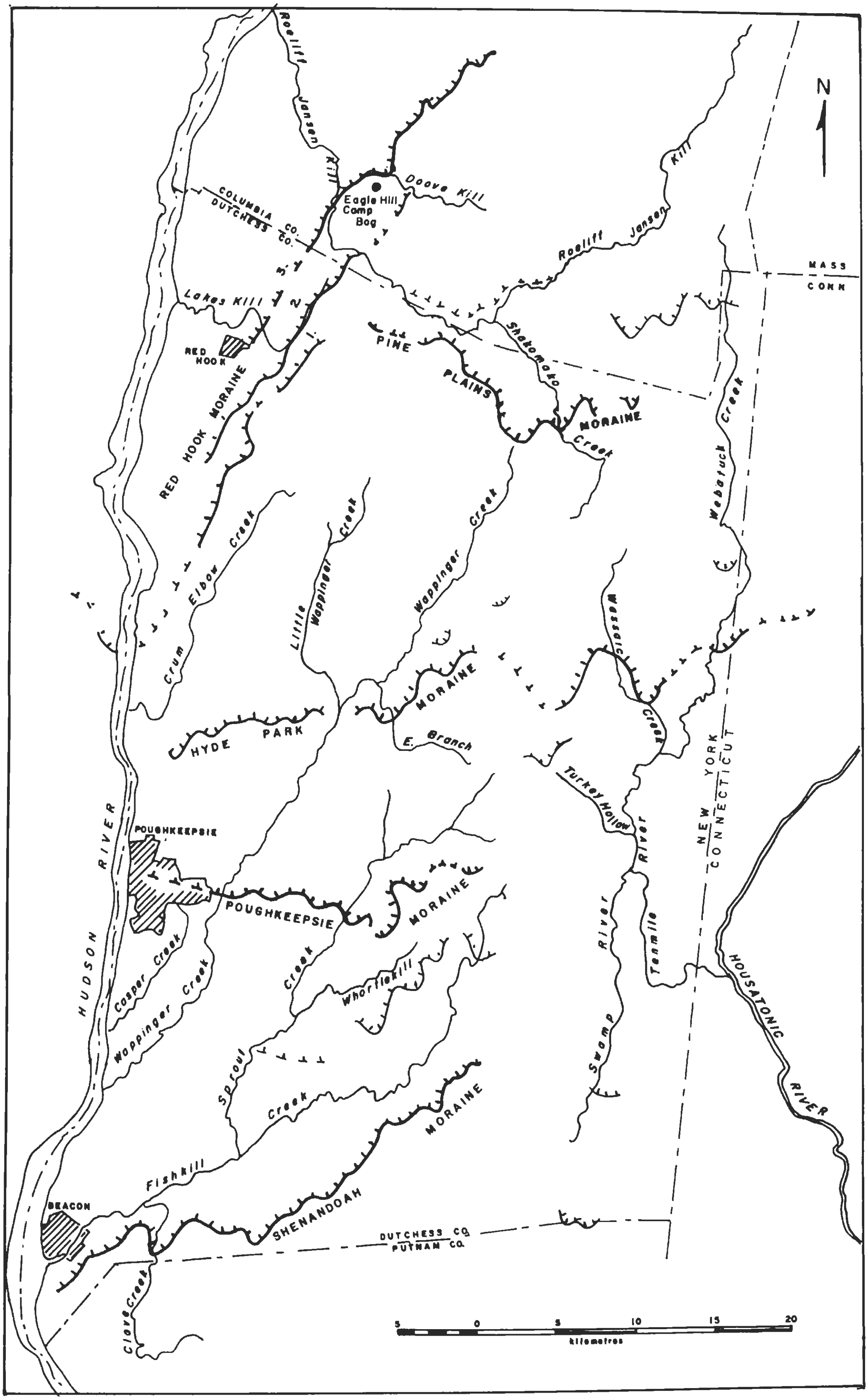

Figure 3 Moraines and major stream systems in the Mid-Hudson Valley.

Poughkeepsie Moraine

Stratified drift covering 3 km^2, with $\pm$ 90 feet of relief was traced from James Baird State Park (Pleasant Valley) to the city of Poughkeepsie and is here named the Poughkeepsie Moraine. Though constructional topography is best displayed in the Town of LaGrange, to the east, the moraine is named for three critical deltas in the Town of Poughkeepsie. The deltas approximate the edge of this ice margin position at Manchester Bridge on Wappinger Creek, the Vassar College campus at the head of Casper Creek, and along South Road south of the city of Poughkeepsie. The Manchester Bridge delta was mined out in the 1950's but Gordon (1911, Plate 20 and p. 103) documents its form. Gordon places the elevation at 160 feet but remnants occur today at 180 feet, which is the approximate elevation assigned to it in this study. The Poughkeepsie Moraine was traced from Baird Park to Manchester Bridge via a till ridge. Although it was not possible to determine the precise position of the ice margin at each of the deltas, there was no other possible source for the deltaic sediments except the retreating glacier while it constructed the Poughkeepsie Moraine and Casper Creek and Wappinger Creek were estuaries of Lake Albany (Fig. 4).

There is a barren bedrock upland separating the moraine at Baird Park from the Sprout Creek valley to the east. However, kames at the hamlet of Moores Hill (Pleasant Valley, Verbank) correlate with this ice margin position. From Moores Hill, the ice margin probably trended northeast toward Verbank and then east along the south side of Willow Brook. There is no evidence of the ice margin on Clove Mountain. Though there may be related deposits at the hamlet of North Clove, in the Clove Valley, the margin is absent from West Mountain and can not be traced into the Harlem Valley.

A recessional position exists at Rochdale (Pleasant Valley) along Wappinger Creek, about 3 km north of the Manchester Bridge delta. Outwash grades from this position, and through it from the Hyde Park Moraine, at about 200 feet. It is possible that there was a reentrant in the Poughkeepsie Moraine, documented by the Rochdale kames. However, the coarseness of the gravel remnants at Manchester Bridge argues for a source closer than 3 km away.

Ice disintegration topography between Sheldon Hill and Platt Hill (Verbank), in a tributary to the North Branch of Wappinger Creek, probably relates to the Rochdale position. About 7 km east, at Littlerest, a massive head of outwash supplied sediment to the Harlem Valley at Dover Plains, via Mutton Hollow. Though there is no evidence to suggest simultaneity of the three recessional deposits, they probably were closely related and document orderly retreat of active ice from the Poughkeepsie Moraine to the Hyde Park Moraine.

When the glacier stood at the Poughkeepsie Moraine, it shed outwash onto the Hudson River, Casper Creek, and Wappinger Creek deltas, all at $\pm$ 180 feet. The massive moraines at Baird Park and Moores Hill were sources of outwash into Lake Fishkill in the Sprout Creek lowland. As the ice margin retreated northward, it continued to supply outwash to valley trains in Wappinger Creek and Sprout Creek. When the glacier front crossed the divide separating Sprout Creek from the North Branch of Wappinger Creek (Verbank), it dammed a lake in the North Branch at $\pm$ 430 feet. Extensive lacustrine terrace remnants occur throughout the North Branch tributaries (Millbrook) in the vicinity of Washington Hollow. Murray (1976) described three successive lake levels as the ice receded from the head of outwash at Littlerest:

Lake Littlerest 900 feet
Lake Mabbitsville 700 feet
Lake Washington 440 feet

The 900 foot elevation marks the flat-topped upper surface of the head of outwash and the threshold elevation at Mutton Hollow; undeniably waters were dammed at 900 feet during retreat, but the lake probably was minor and short-lived. Erosional features at 750 feet and lacustrine clay in bottom lands document the $\pm$ 740 foot level. However, neither of the upper two levels represents much time or sedimentation. Murray's Lake Washington appears to have been a much more significant event during recession from the Poughkeepsie Moraine.

Hyde Park Moraine

A continuous east-west line of stratified drift and associated outwash was traced from 2 km east of the village of Hyde Park to Salt Point, at the confluence of Wappinger and Little Wappinger Creeks. It is here named the Hyde Park Moraine for an area of bold constructional topography south of Fall Kill (Hyde Park) with 170 feet of relief. While constructing this ice margin the Woodfordian glacier deposited the Hyde Park deltaic terrace into Lake Albany at $\pm$ 185 feet. The village of Hyde Park is built on the Hyde Park delta which was mapped originally by Woodworth (1905, Plate 6). Pitted outwash from this ice margin forms a divide between Maritje Kill and Fall Kill. The former, once a tributary of Fall Kill, was forced eastward into a bedrock gorge by deposition of outwash and the presence of the glacier at the Hyde Park ice margin (Fig. 5).

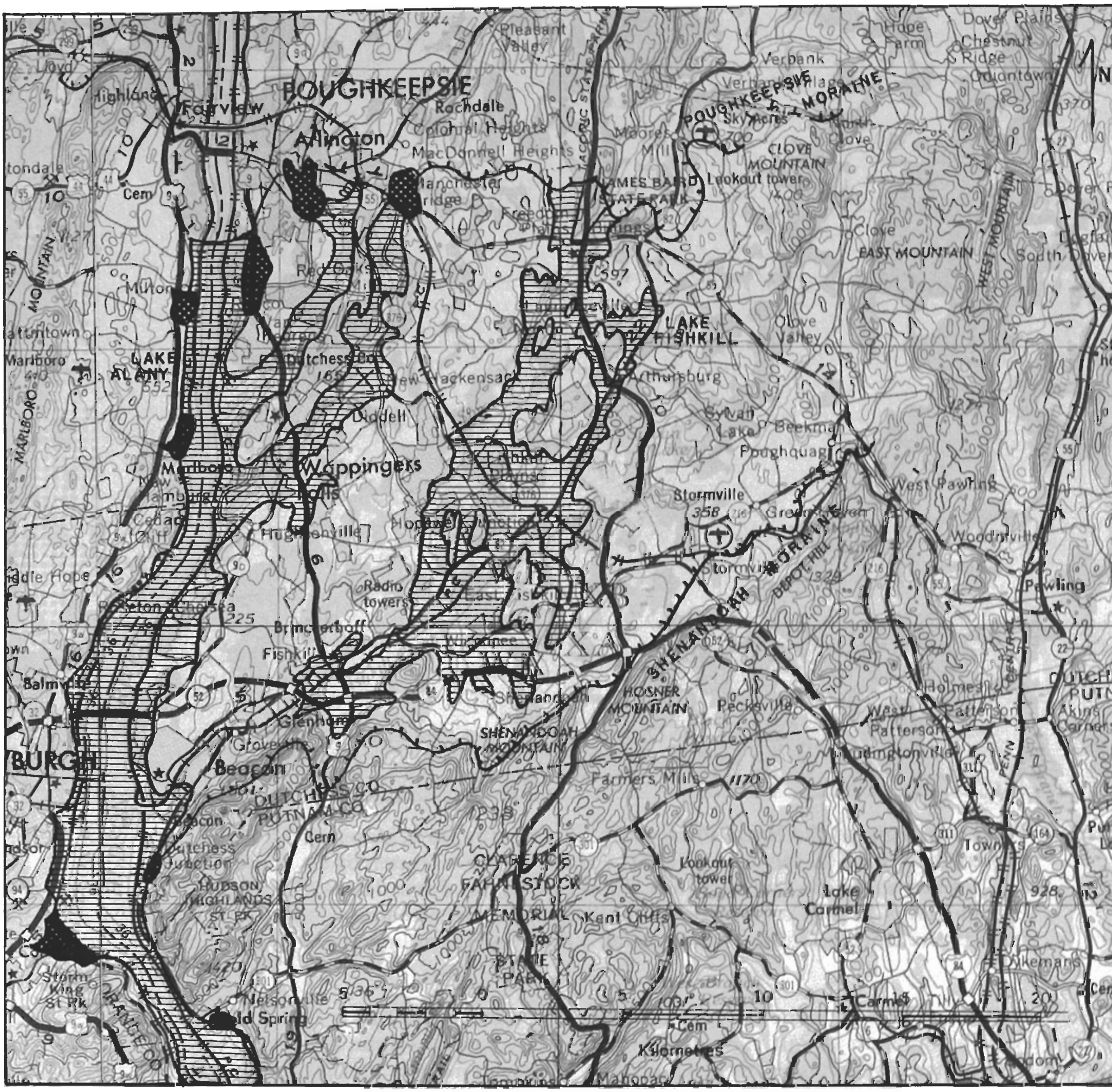

Figure 4 The ice margin at the Poughkeepsie Moraine with the Shenandoah Moraine shown to the south. Lake Albany is defined in the Hudson River trench, from lower left, by the Moodna Creek, Marlboro, Milton, and South Road deltas and in the Casper Creek and Wappinger Creek estuaries by the Vassar and Manchester Bridge deltas, respectively. The Beacon outwash and Cold Spring delta are shown within the Hudson Highlands. Lake Fishkill, to the east, occupies the valleys of Fishkill Creek, Sprout Creek, and the Whortlekill.

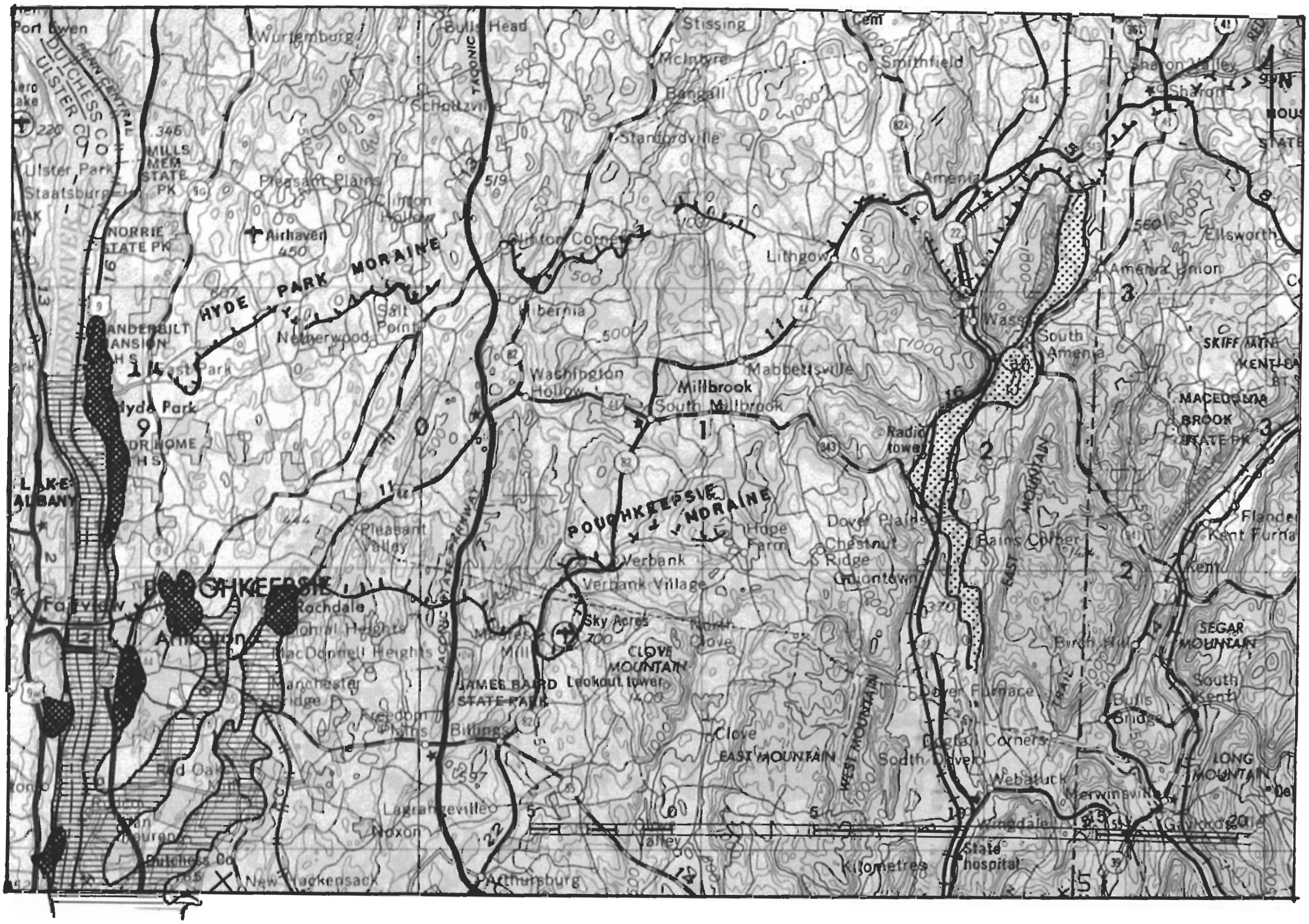

Figure 5 **The ice margin at the Hyde Park Moraine with the Poughkeepsie Moraine to the south. Lake Albany is defined on the west side of the Hudson River trench by the Marlboro and Milton deltas and on the east side by the South Road and Hyde Park deltas. The Vassar and Manchester Bridge deltas, in the Casper Creek and Wappinger Creek estuaries also are shown. The outwash train in the Harlem Valley to the east also may represent lacustrine deposition.**

A barren bedrock promontory of the low Taconics separates the Hyde Park margin at Salt Point (Salt Point) from the North Branch of Wappinger Creek, 3 km to the east. In the North Branch, there is massive ice disintegration topography north of Washington Hollow (Millbrook) which, in turn, is correlated with a head of outwash at Shunpike, 2 km farther to the northeast. Constructional topography on the uplands marks this position between Shunpike and extensive stratified deposits west of Lithgow (Amenia) on the north end of West Mountain.

Although the moraine was not traced across West Mountain to the Harlem Valley, we correlate the deposits at Lithgow with those in the Wassaic Creek valley (Amenia) north of the village of Wassaic. The superposed channel of the Turkey Creek gorge, 400 feet deep, probably owes its origin, at least in part, to drainage along this ice margin. Ice contact drift in the Webatuck Creek lowland shed outwash southward into a valley train of foreset beds with topset beds at 480 feet and certainly is equivalent to the drift at Wassaic. Thus, the Hyde Park Moraine was traced all the way from the Hyde Park delta, at the Hudson River, to the Harlem Valley and the New York/Connecticut border near Sharon, Connecticut.

Between the Hudson River and Wappinger Creek, there is little constructional topography for 20 km north of the Hyde Park Moraine. There are three morphosequences in the Wappinger Creek lowland near Stanfordville (Millbrook), within 7 km of the end moraine and they continued to nourish the Wappinger Creek valley train. A single head of outwash is present in the valley of Shekomeko Creek. However, at least five separate morphosequences are present in the Webatuck Creek lowland between the Hyde Park Moraine and Millerton. Contrary to expectations, an active ice front may have been maintained adjacent to the New England Uplands while the remainder of the glacier stagnated and melted in place on the more open and exposed Hudson Lowlands.

Pine Plains Moraine

A fourth ice margin position oriented essentially east-west is here named the Pine Plains Moraine for several exposures of stratified drift in the Town of Pine Plains, south of Roeliff Jansen Kill (Pine Plains). The moraine was traced continuously for 12 km, from northwest and north of Stissing Mountain to the east flank of that mountain, southwest of the village of Pine Plains. An outwash plain south of the village of Pine Plains, in the headwaters of Wappinger Creek, buried any evidence of the ice margin across this valley. However, there is an ice contact deposit at Bethal 4 km east in the valley of Shekomeko Creek. The ice margin was traced eastward from Bethel to Pulvers Corners and tentatively correlated with a morphosequence at Millerton in the Webatuck Creek lowland.

There is a bold recessional position traced from pitted outwash that forms the lowland divide between Shekomeko Creek and Punch Brook to a huge head of outwash with 180 feet of relief at Bartons Corners. This recessional ice margin was traced around Fox Hill (Copake) to the head of outwash at Whitehouse Crossing that forms the northern divide for Webatuck Creek.

There are several very large heads of outwash in the headwaters of Roeliff Jansen Kill, north of Bartons Corners and along the north valley wall of Roeliff Jansen Kill. Most of these deposits have surfaces that are graded to an elevation of 480 to 500 feet where they abut the valley of Roeliff Jansen Kill. They are particularly well displayed on the Ancram 7-1/2' Quadrangle.

Lake Attlebury

When the ice stood at the Pine Plains ice margin south of the village of Pine Plains, it shed outwash southward into the headwaters of Wappinger Creek. The broad outwash fan/valley train was graded to $\pm$ 450 feet at the hamlet of Attlebury. A dead ice or sediment dam must have existed just southwest of Attlebury. Thus, a lake was impounded between the ice and the since eroded dam, that is here named Glacial Lake Attlebury. As the ice margin retreated, first to the recessional position marked by the drift at Bartons Corners, and then to the north edge of the valley of Roeliff Jansen Kill, the lake expanded in contact with the ice. At its maximum extent, Lake Attlebury expanded northward out of the headwaters of Wappinger Creek and then east and west along the valley of Roeliff Jansen Kill (Fig. 6) with a maximum east-west dimension of about 10 km. The eastern dam in the Roeliff Jansen Kill valley was apparently at Ancram (Ancram) and the western dam was immediately west of Jackson Corners.

From Jackson Corners westward to Elizaville, where Roeliff Jansen Kill spills out onto the Hudson Lowlands from its valley in the high Taconics, several terrace remnants are graded to $\pm$ 300 feet. If the ice margin had moved gradually westward until a lower outlet was established for the 300 foot level, an expansion of the 500 foot level of Lake Attlebury would have preceded the 300 foot level. Because Lake Attlebury evidently drained completely, and must have drained westward along the Roeliff Jansen Kill valley, the 300 foot lake from Elizaville to Jackson Corners must represent a rising water level due to later readvance.

Stagnation

When the ice margin retreated northeastward from Ancram, free drainage already had been established in Roeliff Jansen Kill from Ancram westward to the Hudson River. A bedrock threshold at $\pm$ 480 feet at Ancram controlled westward drainage from the headwaters during deglaciation. There are no continuous ice margin positions in the headwaters region but there is much stagnant ice topography and several outwash and/or alluvial valley trains are graded to the Ancram threshold. With the single exception of a valley choker (MacClintock and Apfel, 1944, p. 1153) moraine halfway between Copake Falls and Hillsdale (Hillsdale), the extensive stagnant ice topography suggests that the glacier remnant in the Roeliff Jansen Kill headwaters stagnated in place when the upper surface downwasted below the crest of the Kijk-Uit Mountain range to the north and northwest. The topographic situation is similar to that described by Cadwell (1981) for the Kinderhook-Lebanon lowlands south of the Rensselaer Plateau.

In the high Taconics, east and north of the Roeliff Jansen Kill, there is almost no stratified drift. No recessional positions are recognized between Roeliff Jansen Kill and Taghkanic Creek except along the divide between the two drainage systems. Several kame complexes occur between Chrysler Pond (Copake) and Copake Lake (Hillsdale) and in the tributary valley of Taghkanic Creek 2 km northeast of Copake Lake. This

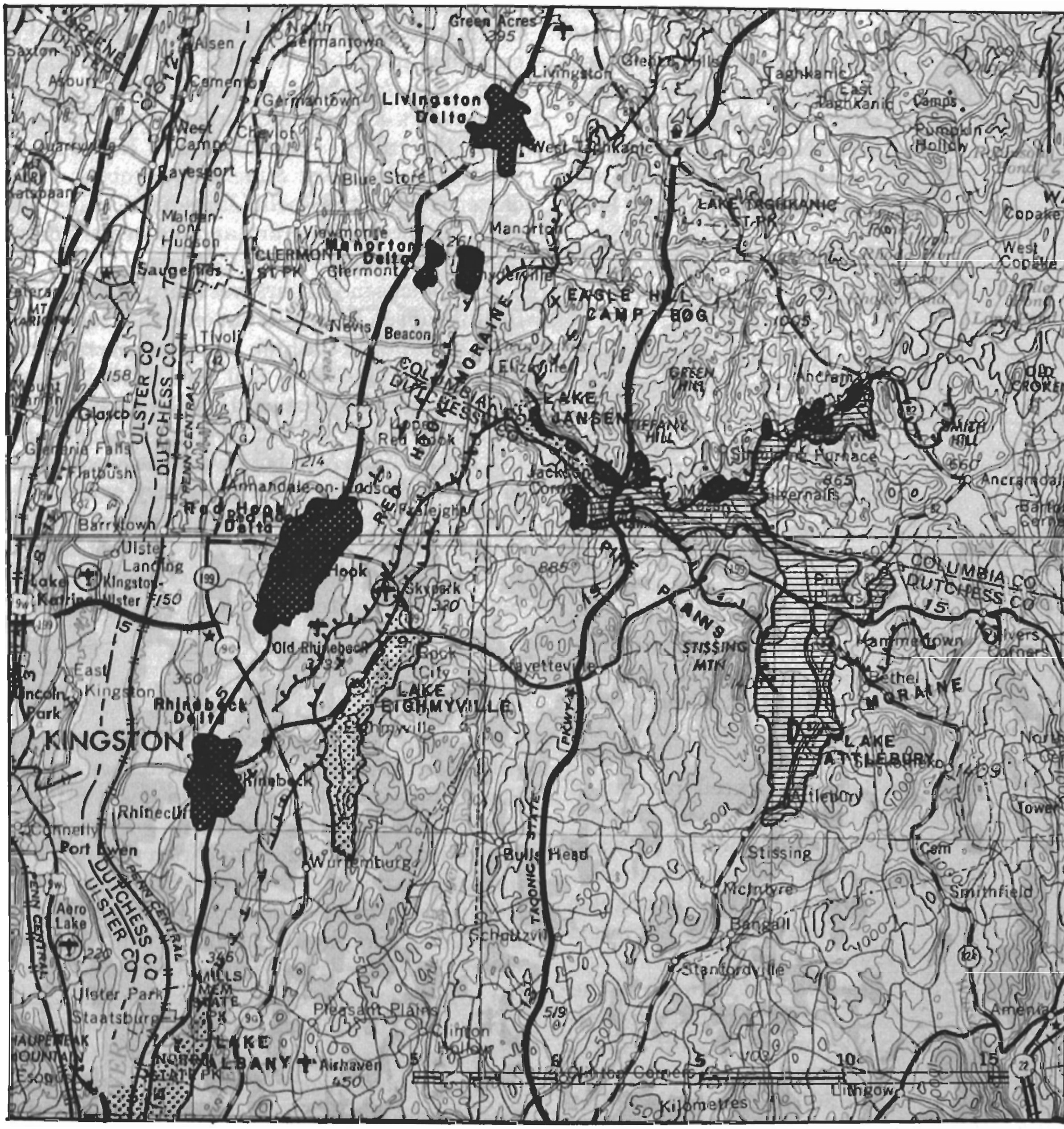

Figure 6 The Pine Plains and Red Hook Moraines. Lake Attlebury was initiated when the ice margin constructed the Pine Plains Moraine and it expanded northward and then east and west at ±500 feet in the valley of Roeliff Jansen Kill. Flat-topped stratified deposits mark the recessional position along the north side of Roeliff Jansen Kill. Later, when the ice readvanced down the Hudson Lowlands, Lake Jansen was redammed at ±300 feet in Roeliff Jansen Kill and Lake Eighmyville at ±330 feet in the valley of Landsman Kill. There is no Lake Albany delta at the outermost Red Hook Moraine but the Rhinebeck and Red Hook deltas mark the medial and innermost positions. The Manorton and Livingston deltas date from disintegration of the ice that constructed the Red Hook Moraine.

ice margin probably is equivalent to the valley choker moraine between Copake Falls and Hillsdale in the headwaters of Roeliff Jansen Kill and may correlate with the Red Hook Moraine, marking a readvance of active ice up the Taghkanic Creek Valley.

Red Hook Moraine

There are three closely related positions that comprise the Red Hook Moraine, here named for their expression in the Town of Red Hook, east of Red Hook village. They are labeled as #1, #2, #3 on the tables and maps in this report. The oldest position may be equivalent, at least in part, to recession from the Pine Plains Moraine. However, the northeast-to-southwest trend is a marked difference from the east-to-west orientation of the ice margins described previously. In addition, the 300 foot lake level between Elizaville and Jackson Corners in the Roeliff Jansen Kill valley implies a redamming of the valley rather than falling water levels associated with recession. Thus, the oldest, or outermost, ice margin position is inferred to mark a readvance that is correlated with the Rosendale readvance (Connally, 1968; Connally and Sirkin, 1973) in the Wallkill Valley, west of the Hudson River. The younger two positions (#2 and #3) are recessional from the outermost (#1).

The oldest ice margin is present at the Red Hook Golf Club (Rock City) where it shed outwash southward from Shooks Pond, down the valley of Landsman Kill. A proglacial lake, here named Glacial Lake Eighmyville, was dammed in that lowland at $\pm$ 330 feet. The ice margin not only dammed the lake at Shooks Pond, but also east of Rhinebeck where it forced drainage farther southward, down Crum Elbow Creek, thus maintaining the 330 foot level. More southerly traces of the outermost position east of the Hudson River were erased or overwhelmed by deposition of the slightly younger Rhinebeck delta of Lake Albany. However, an exposure of diamict over lacustrine sand, just west of the village of Esopus, west of the Hudson River, probably documents the southern limit of the readvance in the Hudson River trench.

We propose a readvance to the outermost Red Hook Moraine by an ice tongue at least 20 km long. This tongue of ice readvanced down the Hudson River trench, and into the Wallkill Valley, after the glacier had receded from the Pine Plains Moraine at least as far north as Roeliff Jansen Kill. This readvance redammed the waters of Roeliff Jansen Kill at $\pm$ 300 feet and forced the new lake to drain southward via the Lakes Kill drainage channel (Clermont). Although this lake is not areally significant, its relationship to the readvance persuades us to propose the formal name Glacial Lake Jansen for this feature.

It is curious that no ice contact delta was deposited in Lake Albany during establishment of the outermost Red Hook Moraine ice margin. Either the readvance was too short-lived or perhaps was accomplished by a thin, overextended surge of unstable glacial ice and debris.

A second ice margin position was traced from the Rhinebeck delta, up the east side of the Rhinebeck Kill valley, and along the Lakes Kill channel. The recession to this position may have been due, in part, to drainage of Lake Jansen as it cut through the distal edge of the readvance to find an outlet southward along the Lakes Kill channel. The second position was traced to the stagnant ice deposits in the Doove Kill valley where it issues from the high Taconics onto the Hudson Lowlands (Clermont), 4 km north of Roeliff Jansen Kill.

A third ice margin, the innermost, was traced from the Red Hook delta to a morainal segment 6 km long from Manorton (Clermont) to Glenco Mills (Claverack) where Taghkanic Creek emptied into Lake Albany from the high Taconics.

When the glacier retreated from the middle position, sedimentation commenced at Eagle Hill Camp (EHC) bog, located just south of Doove Kill at $\pm$ 340 feet. Thus, the base of sedimentation in EHC bog correlates with the beginning of the Red Hook delta of Lake Albany. The Manorton delta (Clermont), and later the Livingston delta (Hudson South), date from disintegration of the ice margin that constructed the youngest, or innermost, position of the Red Hook Moraine. These are the youngest deglacial features in the Mid-Hudson Valley.

The Harlem Valley

The style of deglacial retreat in the Harlem Valley is transitional from the continuous moraines of the Mid-Hudson Valley to the isolated heads of outwash common throughout Connecticut and Massachusetts. There are some morphosequences and some ice margin positions that can be traced to, or correlated with, deposits to the west. However, by and large, the drift of the Harlem Valley gives the impression more of continual stagnation zone retreat by an active sublobe than of major still-stands or oscillatory retreat.

CATSKILL MTNS. (Cadwell, 1984)	MINISINK VALLEY (Connally & others, 1979)	WALLKILL VALLEY (Connally & Sirkin, 1973)	MID-HUDSON VALLEY
Middleburg readvance	?	?	?
↑	?	?	Red Hook Moraine 3
	?	?	Red Hook Moraine 2
Grand Gorge Lake Phase	Rosendale readvance	Rosendale readvance	Red Hook Moraine 1
	?	?	Pine Plains Moraine
↓	?	?	Hyde Park Moraine
Wagon Wheel Gap margin	Wagon Wheel Gap margin	Wallkill Moraine	Poughkeepsie Moraine
?	Cuddebackville margin	Pellets Island Moraine	Shenandoah Moraine

Table 1 Correlation of events between the Mid-Hudson Valley and the Catskill Mountains (Cadwell, this volume), via the Wallkill Valley (Connally and Sirkin, 1973) and Minisink Valley (Connally and others, 1979).

Outwash dominated morphosequences, or ice margin positions, are present at Patterson, just south of the Dutchess County line, Pawling, Wingdale, Dover Plains and Wassaic. Only the ice margin at Wassaic, correlated with the Hyde Park Moraine, was projected westward with any confidence. At least a dozen separate morphosequences were mapped between the morainal position at Wassaic and the northern end of the valley in the vicinity of Copake, New York. The morphosequence at Whitehouse Crossing is correlated westward with the recessional position at Barton Corners.

Outwash south of Dover Furnace documents a south-draining lake at $\pm$ 480 feet during the early recessional history. Once the Tenmile River channel was opened, permitting free drainage to the Housatonic River, a series of outwash shingles resulted. Yet, from South Amenia to the Hyde Park Moraine, prograding foreset beds at $\pm$ 480 feet suggest a second lake, in the middle Harlem Valley, with a surface at 480 feet, probably with a drift dam at South Amenia. From the Hyde Park Moraine north to Copake the outwash shingles attest to free flowing fluvial deposition.

CORRELATIONS

Connally and Sirkin (1967, 1970) named and/or described the moraines in the Wallkill Valley, immediately west of the Mid-Hudson Valley. The Culvers Gap, Augusta, and Sussex moraines all occur well to the south in northern New Jersey. The Pellets Island Moraine, banked against the north slope of the Hudson Highlands west of the Hudson River, almost certainly is the equivalent of the Shenandoah Moraine in the Mid-Hudson Region. The Wallkill Moraine probably correlates with the Poughkeepsie Moraine. There is no apparent equivalent of either the Hyde Park or the Pine Plains Moraine, but the Wallkill Moraine is a compound ice margin, consisting of many recessional ridges, and may be equivalent in part to both the Hyde Park and the Pine Plains ice margins. Alternatively, evidence of equivalent positions west of the Hudson River may have been removed by the Rosendale readvance.

It may be possible to correlate between the Mid-Hudson Valley and the drift of the Catskill Mountains described by Cadwell (1983; this volume). The key to

ICE CONTACT FEATURE	Rebounded LAKE ALBANY*	Stable LAKE ALBANY*	LAKE COVEVILLE (?)*
Livingston Delta	250' (d) 200' (d)	170' (d)	150' (e)
Manorton Delta	240' (d) 200' (e)		
Red Hook Delta	220' (d)	160' (d)	
Rhinebeck Delta	200' (d)	150' (e)	140' (e)
Hyde Park Delta	185' (d)	140' (d)	90' (e)
South Road Delta	175' (d)	135' (e)	100' (e)
Milton Delta	170' (d)		100' (e)
Marlboro Delta	160' (p) 140' (e)		
Beacon-Fishkill Delta	140' (o)	115' (e)	100' (e)
Moodna Creek Delta	140' (d)		80' (d)
Cold Spring Delta		110' (d)	
Peekskill Delta		100' (d)	80' (e)

(d) delta surface, (e) erosional escarpment, (o) outwash, (p) fluvial plain

* Elevations estimated from 1:24,000 maps, 10 or 20 ft. intervals

Table 2 Delta levels in the Mid-Hudson Valley. The first column labeled Lake Albany gives the elevations of the ice margin deltas and defines Rebounded Lake Albany. The second column labeled Lake Albany gives elevations for Stable Lake Albany and the last columns may relate to a level called Lake Coveville by Connally and Sirkin (1973). The symbols indicate delta surface (d), erosional escarpment (e), fluvial plain (p), or outwash (o). Elevations were estimated from 1:24,000 maps with 10 or 20 ft. contour intervals.

such a long distance correlation is the Wallkill Moraine that is correlated with the Wagon Wheel Gap margin of Rich (1935) as remapped by Connally (in preparation). Connally and others (1979) correlated the Wallkill Moraine in the Wallkill Valley with the morphosequence at Summitville in the Minisink Valley. This correlation, though still tentative, agrees with that proposed by Dineen (1983; this volume) when he designates the ice margin at Phillipsport. The deposits at Phillipsport are the proximal, ice contact deposits for this head of outwash, while the distal, outwash deposits occur 2 km south at Summitville. Although Krall (1979, p. 604) found that it was "... difficult to visualize ..." the correlation of the Wallkill Moraine and Wagon Wheel Gap ice margin, via the Phillipsport position, it was much less difficult to establish it through careful mapping in the field than it was to *visualize* it from Union, New Jersey.

In discussing the Wagon Wheel Gap correlation it must be noted that Connally (1983) stated in his abstract that "... readvance . . . near Rosendale . . . is . . . correlated with the Wagon Wheel Gap margin ..." That statement was both incorrect and inadvertent. It was corrected in oral presentation and should stand as here amended maintaining the fundamental disagreement with Krall concerning both the correlation and the necessity for field work as the basis for correlation.

Tentative correlations between the Mid-Hudson Valley and the Catskills, via the Wallkill and Minisink Valleys, are given in Table 1.

GLACIAL LAKE ALBANY

As the Woodfordian glacier retreated northward up the Hudson Valley, interrupted only by the Rosendale readvance, the proglacial lake that began in the lower Hudson Valley expanded northward in contact with the glacier snout. Deltas related to this initial lake level occur at Peekskill, Cold Spring, the Moodna Creek confluence, Beacon, Wappingers Falls (and Marlboro), Milton, South Road (and Vassar and Manchester Bridge), Hyde Park, Rhinebeck, Red Hook, Manorton, and Livingston. From Beacon northward, the deltas were initiated as ice margin deposits (the deposit at Beacon actually is an outwash remnant) and thus all are related to the same sequential recession. The proglacial lake is called Lake Hudson (Reeds, 1927) when confined to the lower Hudson Valley. Ice at the Pellets Island/Shenandoah ice margin marks the northern limit of Lake Hudson, as imaginatively explained by Woodworth (1905, p. 114). As the ice margin retreated northward, permitting expansion of the proglacial lake into the Mid-Hudson Valley, it became Lake Albany in the sense of Connally (1972). Connally (1972; 1978; 1983, and in Woodward- Clyde, 1980), on the basis of deltas within and south of the Hudson Highlands, concluded that Lake Albany extended into the lower Hudson Valley where it was coextensive with Reed's Lake Hudson.

The deltas that define the upper "water plane" of Lake Albany tended to develop where tributaries from the low or high Taconics flow out onto the Hudson Lowlands. In this position, the streams contributed sediment to the glacier, either at the margin or on the downwasting snout. Actual ice contact deposits are seldom visible except when mining operations expose the core of an ice margin delta because when the glacier margin retreated north of an ice contact delta, the streams continued to contribute sediment which prograded and buried the ice contact face or faces. Later, the level of Lake Albany dropped relative to the land surface, or the land rebounded, and the streams readjusted to a lower base level. In some instances the stream redeposited sediments to form a lower delta or terrace, and in others an erosional nip or scarp was developed. The various levels at each position, from Peekskill which is south of the Mid-Hudson Region, to the Livingston delta which is just north of the region, are given in Table 2 and plotted in Figure 7.

In Figure 7 the uppermost deltas related to a regional level in the Hudson Lowlands are connected by an upper "water plane." Historically, this "water plane" was the regional level that defined Lake Albany. Connally (1972) projected this level north to the deltas in the Glens Falls region. A lower "water plane" is projected northward from the Lake Hudson deltas at Peekskill and Cold Spring, to erosional and depositional features present on or near most of the major tributary streams. The lower features are deltaic fragments, terrace remnants, small deltas, fluvial plains, and erosional scarps. In Figure 7A the deltas and "water planes" are shown at elevations determined from 7-1/2' topographic maps with 10 foot or 20 foot contour intervals.

The most obvious aspect of Figure 7A is that if Lake Hudson did indeed expand northward to become Lake Albany, it did so at a level significantly below the "water plane" that connects the ice margin deltas. Connally (1978, 1982) contends that the lower level, projected northward from the lower Hudson Valley, is the true water plane for Lake Albany, referring to this level as Stable Lake Albany. As the ice margin retreated northward, Stable Lake Albany expanded.

A second aspect of Figure 7A is that Stable Lake Albany, which must have been a horizontal surface when it existed, has undergone regional isostatic rebound. In Figure 7B, Stable Lake Albany is rotated back to its original horizontal position. After rotation, the upper "water plane" shows the results of rebound that was active prior to the regional tilting of the stable lake level. This has been proposed to be the result of local popping of the Hudson Lowlands which were overloaded isostatically relative to the surrounding uplands.

Adjacent to the glacier snout, the valley bottom was still depressed so that ice margin deltas were deposited at sites that were depressed relative to the already rebounded land to the south. As the ice margin retreated north of each delta site, the land rebounded locally, elevating the ice margin deltas above the stable, regional level of Lake Albany. The time lag between retreat and rebound of an individual delta might be on the order of 500 years. If this model is correct, the upper "water plane" never existed as a regional lake level and is merely an artifact of local, sequential, postglacial rebound. A simplified sequence of retreat and rebound is sketched in Figure 8, realizing that the actual events were much more gradual and probably more complex than illustrated. Table 3 relates the ice margin deltas in the Mid-Hudson Region with glacial events and moraines in the Mid-Hudson and Wallkill Valleys.

EAGLE HILL CAMP BOG

We here propose the pollen stratigraphic section from Eagle Hill Camp (EHC) bog (Fig. 9) as the standard for the Mid-Hudson Valley. The site was selected using criteria established in prior studies (Connally and Sirkin, 1971; in preparation; Sirkin, 1977; Sirkin and Minard, 1972). Specifically (1) the bog is a closed depositional

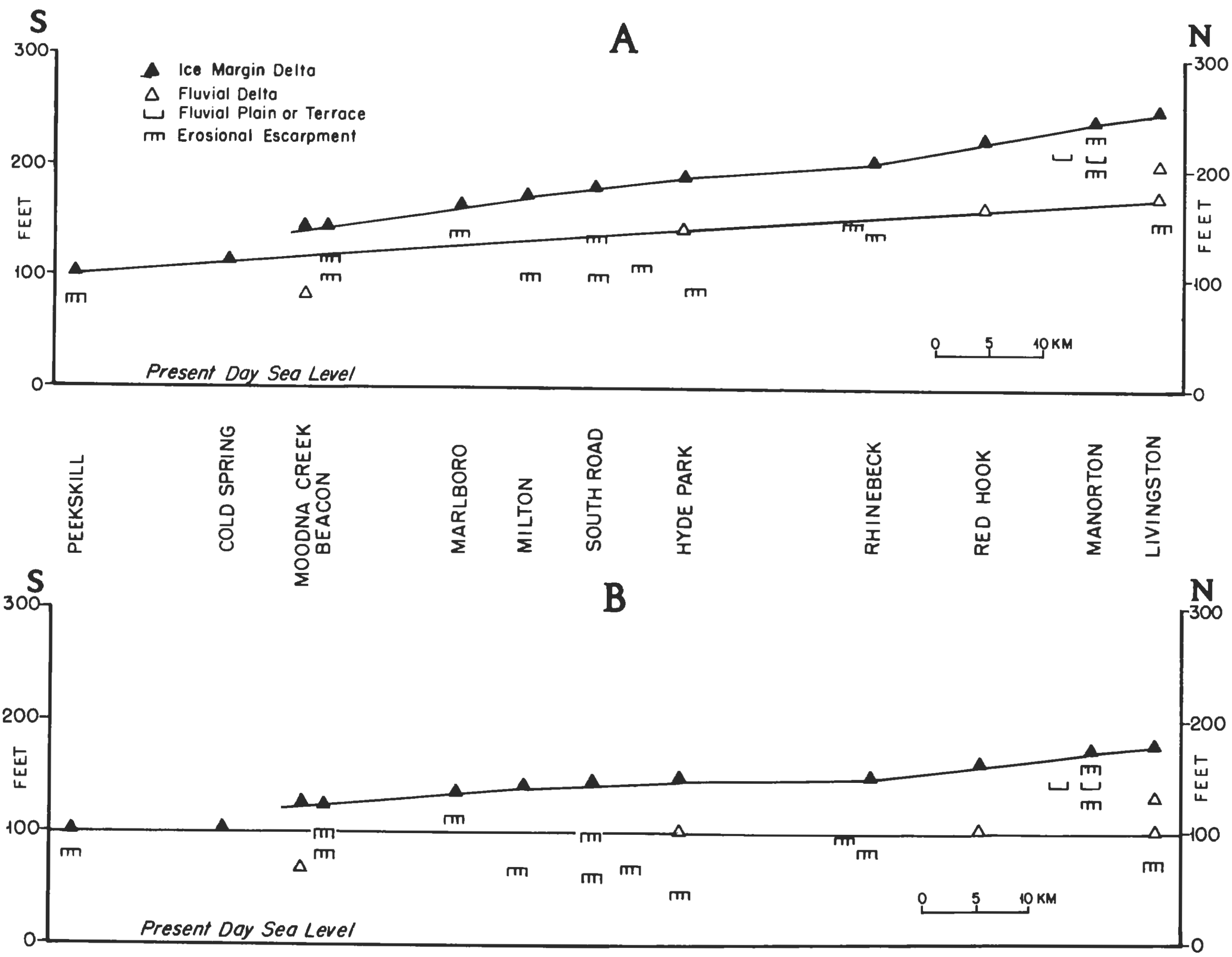

Figure 7 Regional "water planes" in the Mid-Hudson Region. Figure 7A shows deltas and other features as they appear today. Figure 7B shows the features rotated downward about an arbitrary point, eliminating the effects of regional isostatic rebound to make the lower, stable, regional lake level horizontal as it was when Lake Albany was present. Ice margin deltas rebounded sequentially (Figure 8) as the glacier retreated so the upper "water plane" is an artifact of the local rebound of the Hudson Lowlands and never existed as a regional lake.

WALLKILL VALLEY	LAKE ALBANY	MID-HUDSON VALLEY
	Livingston Delta	disintegration
	Manorton Delta	
	Red Hook Delta	Red Hook Moraine 3
	Rhinebeck Delta	Red Hook Moraine 2
Rosendale readvance	----------	Red Hook Moraine 1
	----------	Pine Plains Moraine
	Hyde Park Delta	Hyde Park Moraine
Wallkill Moraine	South Road Delta	Poughkeepsie Moraine
	Milton Delta	recession
	Marlboro Delta	
Pellets Island Moraine	Beacon Outwash	Shenandoah Moraine
	Moodna Creek Delta	?
	Cold Spring Delta	?
	Peekskill Delta	?

Table 3 Correlation of moraines in the Mid-Hudson Valley, east of the Hudson River; Lake Albany features in the Hudson River Lowland; and the events in the Wallkill Valley, west of the Hudson River.

system with exclusively internal drainage, (2) it has a deep section, (3) its base rests on the deglacial surface, and (4) it can be related directly to a geomorphic event – in this example, the deposition of the Red Hook delta. The age of the base of EHC bog dates the establishment of the innermost ice margin position of the Red Hook Moraine.

The bog was sampled with sequential core segments until coarse sediment was encountered at –11.15 m. We concluded that the basal sediment is the same as the glaciofluvial sediment in the rim of the bog. The bog section consists of an upper, dominantly organic (organic-rich) portion and a lower, dominantly inorganic (sub-organic) portion. Accumulation of the sub-organic portion began when the Woodfordian glacier receded from the middle ice margin of the Red Hook Moraine. The sub-organic portion, from -11.15 m to -10.40 m, contains only the herb pollen (T) zone. The organic-rich portion, from -10.40 m to -1.00 m, comprises the spruce pollen (A) through oak pollen (C) zones. No samples were retrieved from the upper 1 m which was composed mainly of water saturated peat. A radiocarbon date of 13,670 $\pm$ 170 yrs BP (SI 4,082) comes from a pooled sample of five increments retrieved from –10.25 m in five closely spaced cores.

A sediment accumulation rate of 0.068 cm/yr is calculated as an average rate for the organic-rich portion from -10.25 to -1.00 m. Applying this rate to the lowest 15 cm of the organic-rich portion, from -10.40 m to -10.25 m, we calculate an age of 13,890 yrs BP (13,670 + 220) for the base of the organic-rich portion.

The time necessary to accumulate the suborganic portion was calculated from two separate accumulation rates. Davis and Deevey (1964) calculated an accumulation rate of 0.036 cm/yr for the basal, suborganic sediments from Rogers Lake in south-central Connecticut.

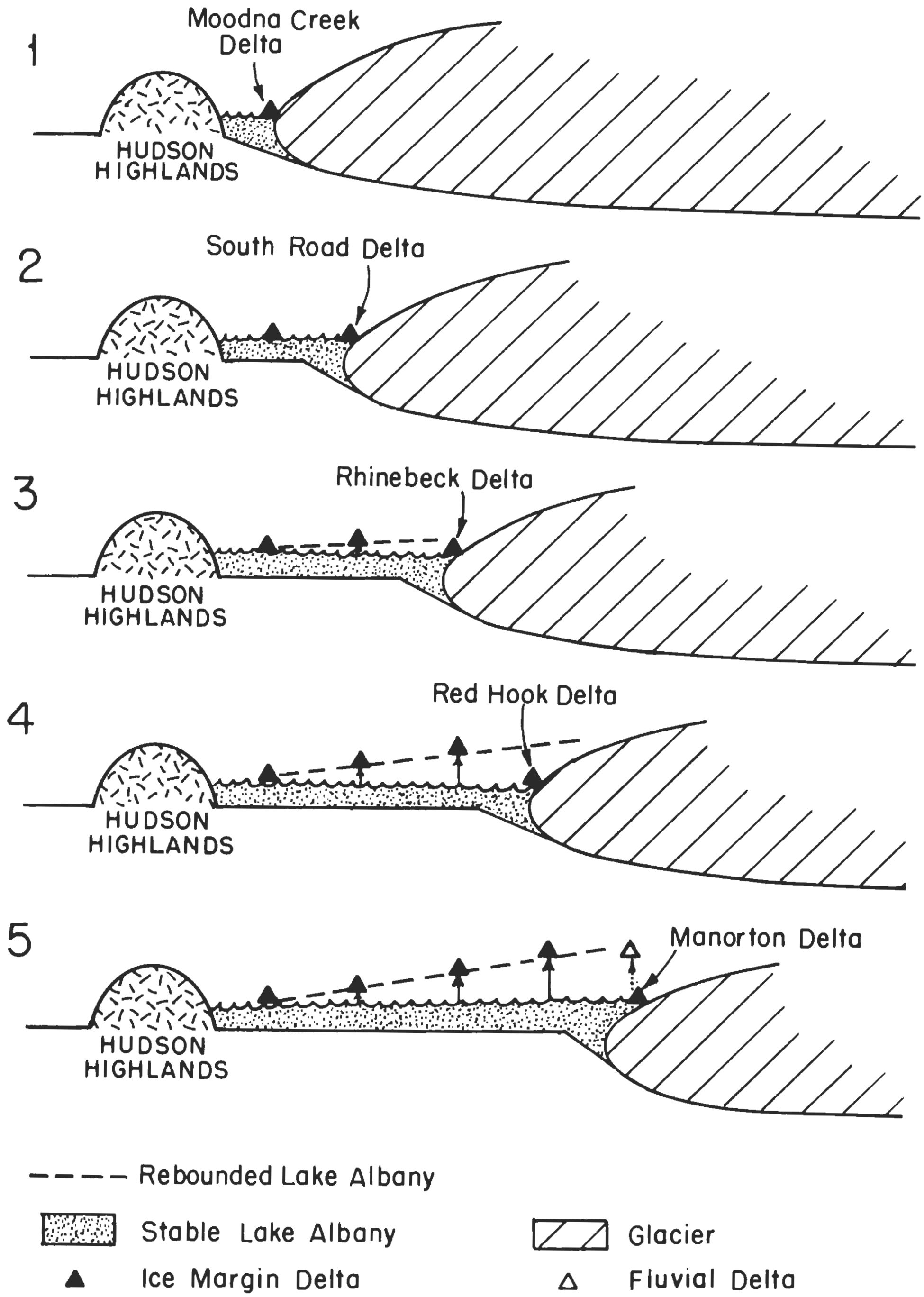

Figure 8 A cartoon representation of glacial recession in the Mid-Hudson Valley. Each ice margin delta rebounds above the water plane of Stable Lake Albany, after an unknown time lag, creating a false "water plane" at a higher elevation that is an artifact of rebound. The ultimate height of Rebounded Lake Albany probably is a function of valley width as well as depth.

Connally and Sirkin (in preparation) calculated an accumulation rate of 0.033 cm/yr from the suborganic portion of the Pine Log Camp bog core (Connally and Sirkin, 1971). The close agreement of these two rates is interpreted as confirmation of calculated ages for suborganic sediments of the herb pollen zone, when bogs meet the stringent criteria we described above. It took a minimum of 2,083 years and a maximum of 2,272 years to deposit the basal 75 cm of the EHC bog section. Thus, the calculated age for the base of the bog, and for the innermost ice margin of the Red Hook Moraine is between 15,970 and 16,160 yrs BP. We suggest an average age of 16,070 yrs BP.

The herb pollen (T) zone at EHC bog is characterized by high values for spruce and lesser amounts of pine, birch, alder, and Rosaceae (Fig. 9). In Figure 9, alder, fir, poplar, willow, as well as the Cedar Group (TJC) and ash, beech, elm, hickory, and maple are lumped together as Miscallaneous Arboreal Pollen (Misc AP). Nonarboreal pollen (NAP) are relatively high in the T zone, up to 22 percent, with grass, sedge, composites, and chenopods represented. Fir in the upper part of the T zone might be a basis for subdividing the zone. However, the pollen spectra indicate a spruce park, or spruce shrub vegetation rather than a park tundra or shrub/herb tundra.

The spruce pollen (A) zone is divided into the A1, A2, and A3-A4 subzones. The A1 subzone has higher spruce representation, along with a fir and pine increase. Subzone A2 reflects a decrease in spruce and NAP. The A3-A4 subzone has a spruce, TJC, birch, and pine assemblage and low NAP.

The pine pollen (B) zone is divided into the B1 and B2 subzones. The B1 subzone has a pine maximum and a general increase in the miscellaneous arboreal pollen (AP), particularly TJC, poplar, and willow, along with a decrease in birch. The B2 subzone, with declining pine values, shows increases in such hardwood species as elm, ash, and maple as well as rises in oak and hemlock. Spruce and fir are gradually replaced upward in the B2 interval. In the NAP, the percentage of grass is significant.

The oak pollen (C) zone is divided into the C1, C2, and C3 subzones. The C1 subzone is dominated by oak and miscellaneous hardwoods, accompanied by birch and TJC, and the first beech rise. The hemlock decline, with values dropping to about 2 percent, occurs late in subzone C1. It is followed by a rise of this taxon to nearly 12 percent of the pollen in subzone C2 where pine decreases while oak, birch, hickory, and beech are prominent. Grass and several species of aquatics are important NAP. The C3 subzone incorporates the oak maxima and a corresponding drop in other hardwoods. Birch diminishes somewhat, but hemlock increases. NAP rises to 30 percent of the pollen in the upper part of the zone; mainly grass, sedge, and aquatics. The rise in composites that usually signals colonial land clearing was not seen in the record due to a lack of samples from the upper 1 m of the section.

The pollen data are comparable to other pollen records in the Hudson Valley, New Jersey, and on Long Island. The standard pollen zonation is recognized at EHC bog and it is correlated with the record previously established (Connally and Sirkin, in preparation). From regional correlation, it becomes apparent that only the latter part of the herb pollen zone, perhaps subzones T2 and T3, or only T3, were sampled at the base of the cored section. This is particularly evident when correlating with the New Hampton (NH) bog, the standard section for the Wallkill Valley, where a somewhat longer and more complete herb pollen zone was described by Connally and Sirkin (1970). At NH bog, more tundra indicators in the pollen spectra of the basal sediments underlie evidence of park tundra. At EHC bog, either the basal tundra vegetation was not significantly developed or it was not sampled and the base of the core is somewhat older than the 16,070 age that we have assigned.

The basal age for NH bog has been recalculated (Connally and Sirkin, in preparation) using 0.036 and 0.033 cm/yr as maximum and minimum accumulation rates for the suborganic portion. The calculated age for the base of the NH bog is between 17,020 and 17,400 yrs BP. We suggest here an average age of 17,210 yrs BP. Sedimentation began at New Hampton when the outermost Wallkill Moraine was established. Thus, the age for its proposed correlate in the Mid-Hudson Valley, the Poughkeepsie Moraine, is suggested as 17,210 yrs BP. Sedimentation began at New Hampton when the outermost Wallkill Moraine was established. Thus, the age for its proposed correlate in the Mid-Hudson Valley, the Poughkeepsie Moraine, is suggested as 17,210 yrs BP which is consistent with our suggested age for the Red Hook Moraine. As further confirmation, Weiss (1971, Table 4) reports a date of 17,950 $\pm$ 620 yrs BP (I-4935) from varved material in his Core 2A from the Hudson River channel, south of the Hudson Highlands. Table 4 shows a correlation for all events in the region, with their respective ages, based on the pollen stratigraphy from the standard sections for the Wallkill Valley (NH) bog and the Mid-Hudson Valley (EHC bog).

SUMMARY

The Woodfordian glacier reached its maximum stand on Long Island about 21,750 yrs BP and recession was underway by about 21,200 yrs BP (Sirkin, 1982; 1983; this volume). By 17,950 yrs BP the ice margin had re-

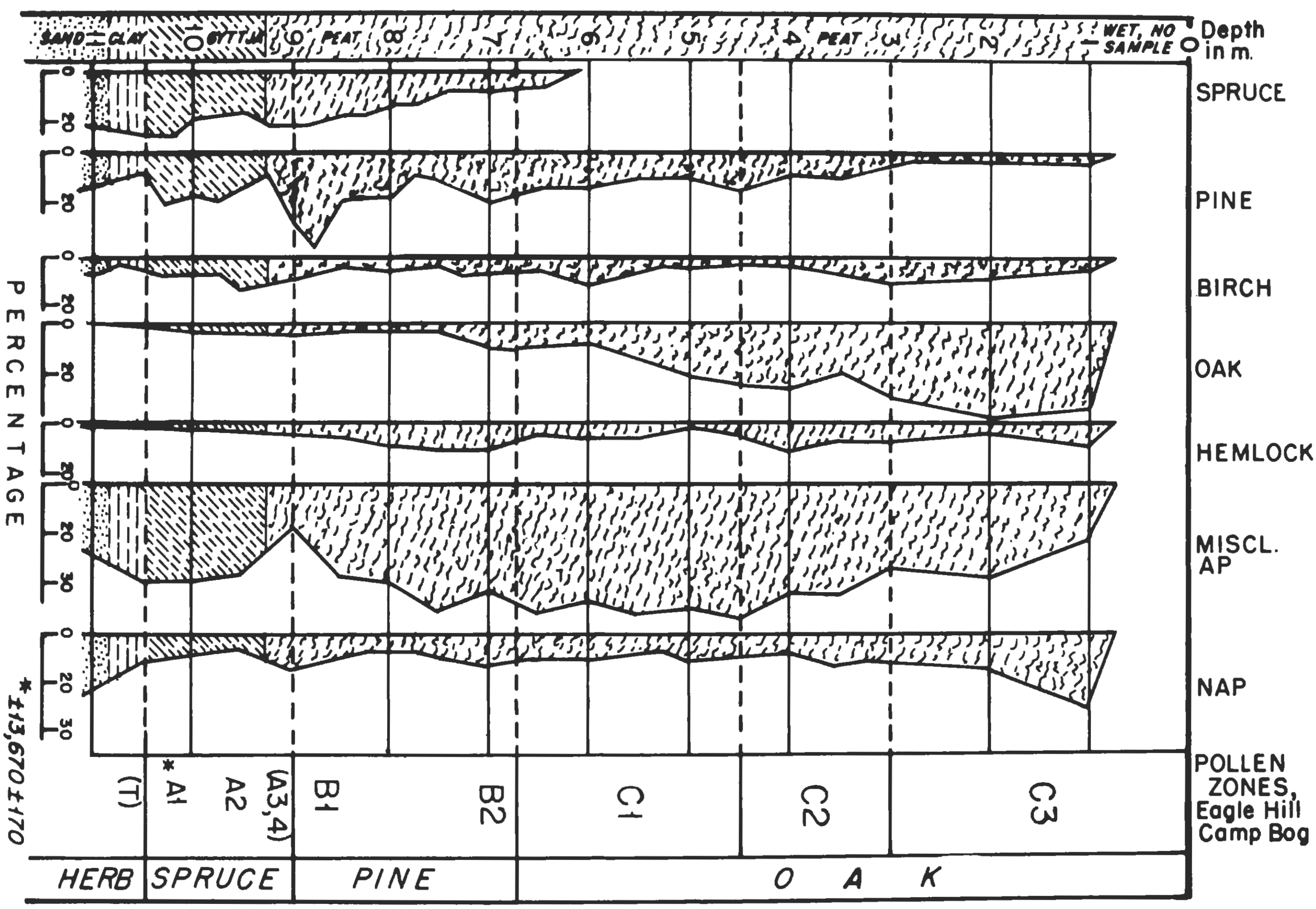

Figure 9 **Summary diagram of the pollen stratigraphy for Eagle Hill Camp bog, Manorton, New York. The bog is located south of Doove Kill (Figure 6) between the medial and innermost Red Hook Moraines.**

CATSKILL MTNS.	MINISINK VALLEY	WALLKILL VALLEY	LAKE ALBANY	MID-HUDSON VALLEY	AGE, B.P.
Middleburg readvance					
			Livingston Delta		
			Manorton Delta		
			Red Hook Delta	Red Hook Moraine 3	16,070
			Rhinebeck Delta	Red Hook Moraine 2	
Lake Grand Gorge	Rosendale readvance	Rosendale readvance	--------	Red Hook Moraine 1	16,100 (?)
	?	?	--------	Pine Plains Moraine	
	?	?	Hyde Park Delta	Hyde Park Moraine	
Wagon Wheel Gap margin	Wagon Wheel Gap margin	Wallkill Moraine	South Road Delta	Poughkeepsie Moraine	17,210
			Milton Delta		
			Marlboro Delta		
	Cuddebackville margin	Pellets Island Moraine	Beacon Outwash	Shenandoah Moraine	
			Moodna Creek Delta		
			Cold Spring Delta		
	Port Jervis margin	?	Peekskill Delta		17,950
	?	Sussex Moraine	?		18,350
	?	Augusta Moraine	?		21,200
	Dingman's Ferry margin	Culvers Gap Moraine	?	Harbor Hill Moraine	21,750

Table 4 Suggested ages for morphostratigraphic features and events in the Mid-Hudson Region and correlation with features and events in the Wallkill Valley, Minisink Valley, and Catskill Mountains.

treated north of the Hudson Highlands, standing at the ice margin position marked by the Pellets Island Moraine west of the Hudson River and the Shenandoah Moraine in the Mid-Hudson Region. During this stillstand, deltas were deposited at Peekskill, Cold Spring, and the confluence of Moodna Creek with Lake Hudson/Albany, south of Newburgh. At the same time, outwash was deposited at Beacon. All document the extension of Glacial Lake Hudson northward to the Mid-Hudson Valley.

By 17,210 yrs BP, the glacier began to construct the Wallkill Moraine west of the Hudson River and the Poughkeepsie Moraine to the east. During retreat to the Wallkill/Poughkeepsie ice margin, ice margin deltas were deposited at Wappingers Falls, Marlboro, and Milton and Lake Fishkill was initiated in the valleys of Fishkill Creek, Sprout Creek and the Whortlekill. The ice contact deltas at South Road, Vassar College, and Manchester Bridge document the presence of Lake Albany during occupation of this ice margin.

For the next few hundred years, the ice margin continued to retreat northward, pausing to construct the Hyde Park and Pine Plains Moraines in the Mid-Hudson Valley. During deposition of the Hyde Park Moraine, the Hyde Park delta was deposited in Lake Albany. The Pine Plains Moraine dammed Lake Attlebury in the headwaters of Wappinger Creek and permitted it to expand east and west into the Roeliff Jansen Kill valley during recession.

About 16,200(?) years ago, the ice margin readvanced at least 20 km down the Hudson River trench and into the Wallkill Valley. During this event, the Rosendale readvance, the Red Hook Moraine was constructed in the Mid-Hudson Valley. There is no Lake Albany feature associated with the outermost Red Hook margin, but the Rhinebeck delta correlates with the middle position and the Red Hook delta with the innermost position of the Red Hook Moraine. The initial readvance dammed Lake Eighmyville in the valley of Landsman Kill and redammed Lake Jansen in Roeliff Jansen Kill. By 16,070 yrs BP the ice had receded to the innermost ice margin position and sedimentation began in Eagle Hill Camp bog. Following disintegration of the glacier at the innermost moraine, the Manorton and Livingston deltas were deposited in Lake Albany where it bordered the high Taconics.

During the approximately 2,000 years it took the glacier margin to retreat from the Hudson Highlands through the Mid-Hudson Valley, the Hudson River trench apparently began localized postglacial rebound. The effect was to elevate the initial ice margin deltas sequentially, from south to north, above the stable level of Glacial Lake Albany.

Correlation of events between the Mid-Hudson Valley and the Catskill Mountains is based on the equivalence of the Poughkeepsie Moraine and the Wallkill Moraine, which in turn is correlated with the Wagon Wheel Gap margin in the Minisink Valley and the eastern and central Catskills. Correlation with events in New England probably depends on finding equivalents to the Hyde Park Moraine in the vicinity of Sharon, Connecticut.

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LATE WISCONSINAN STRATIGRAPHY OF THE CATSKILL MOUNTAINS

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ABSTRACT

Four major deglacial events have been recognized in the eastern Catskill Mountains, Schoharie Valley and Eastern Mohawk Valley. Recent field mapping south of the Middleburg readvance ice margin of LaFleur (1969) strongly suggests that the readvance at Prattsville, in the Schoharie Valley, is equivalent to the Tannersville margin, near the Catskill Front. This correlation reestablishes the Wagon Wheel Gap Margin proposed by Rich in 1935. The Wagon Wheel Gap ice margin now is recognized as one of the Woodfordian ice margin positions in the Schoharie Valley and eastern Catskills. Retreat from this position permitted the development of Glacial Lake Grand Gorge, a proglacial lake that drained through the Grand Gorge channel at 488 m (1600 feet), which reaffirms Rich's Grand Gorge Lake "stage." Kame deltas in the Keyser Kill valley attest to the lowering of Glacial Lake Grand Gorge to 366 m (1200 feet) as meltwater escaped through the Franklinton channel. A readvance to the Middleburg position followed glacier recession and reestablished the 366 m lake and drainage through the Franklinton channel.

INTRODUCTION

Purpose

This paper presents a summary of the Wisconsinan history of the northern Catskill Mountains with emphasis on the deglacial history of the Woodfordian Substage. Knowledge of this area is critical to an understanding of the glacial history of the Susquehanna, Delaware and Hudson River valleys. The study area includes the upper Schoharie Creek and its tributaries as well as parts of the East and West Branch of the Delaware River and Esopus Creek (Fig. 1).

General Geology

The Schoharie Creek valley and vicinity is entirely within the Appalachian Uplands Physiographic Province of New York State (Broughton and others, 1966 p. 34). The 1600 km^2 study area is situated between escarpments developed on gently dipping Devonian clastic rocks of both marine and nonmarine origin. These sandstones, shales and conglomerates are Middle and Late Devonian in age and were derived from the erosion of an eastern land-mass during and following the Acadian Orogeny. The sediments were deposited in an epeiric sea that flooded a major part of eastern North America about 395 million years ago (Rickard, 1981). They are classified as follows (Fisher and others, 1970):

Upper Devonian	– West Falls Group (sandstone, shale, conglomerate)
	– Sonyea Group (sandstone, shale, conglomerate)
	– Genesee Group (sandstone, shale, conglomerate)
Middle Devonian	– Hamilton Group (sandstone, shale)

The Catskill Mountains were glaciated at least once during the Pleistocene Epoch – multiple glaciations are suggested by interbedded tills recorded in subsurface data. A northward gradient, as exemplified by the northward flowing Schoharie Creek, prohibited free drainage of meltwater away from the ice margin and resulted in the development of proglacial lakes during both the advance and retreat of the glacier. Recognition of this drainage pattern is critical to deciphering the glacial history of the Catskill Mountains.

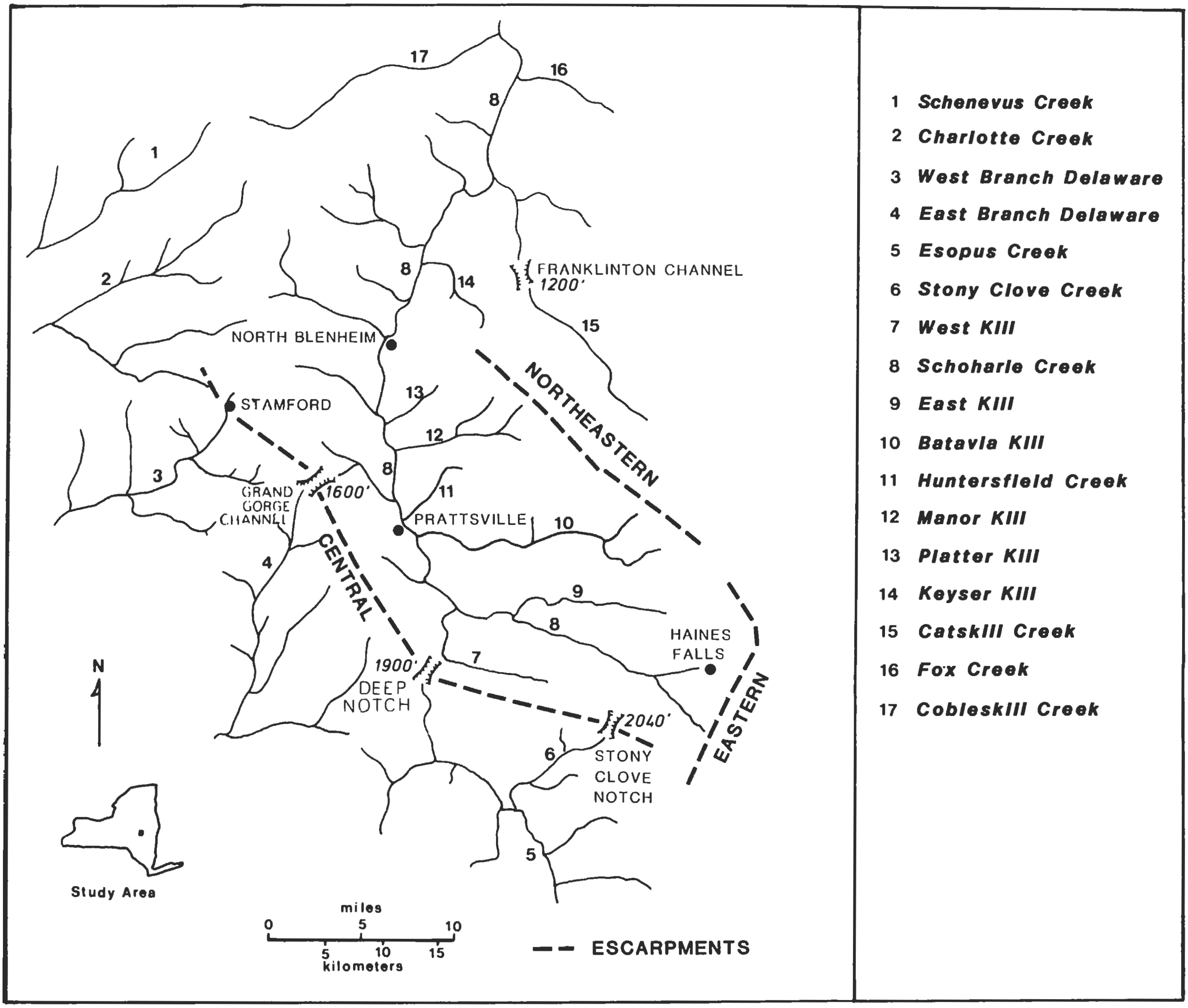

Figure 1. Location map for the central Catskill Mountain region.

Previous Work

Chamberlin (1881-82) suggested that there were local glaciers in the Catskill Mountains during the ice ages. He also recognized that at least one moraine was deposited near Stamford and Grand Gorge during retreat of the ice sheet. Smock (1883) suggested that local detached glaciers were present in the Catskills in his study of the thickness of the continental glacier. Darton (1896) first recognized the outlet for glacier meltwater at Grand Gorge. Chadwick (1928) observed striae in a foundation excavation on top of Slide Mountain (elevation 1282 m) and suggested that glaciers overrode the Catskill Mountains. Rich (1914) discussed drumlins and glacier flow directions on the plateau northeast of the Catskills. Brigham (1908, 1929) and Fairchild (1912), in their studies of the Mohawk Valley, inferred receding ice margins and sequential ice margin lakes that extended into the lower Schoharie Valley.

The first detailed study of local glacier features in the Catskills was presented by Rich (1906). He described a remarkable group of moraines at the head of Fly Brook

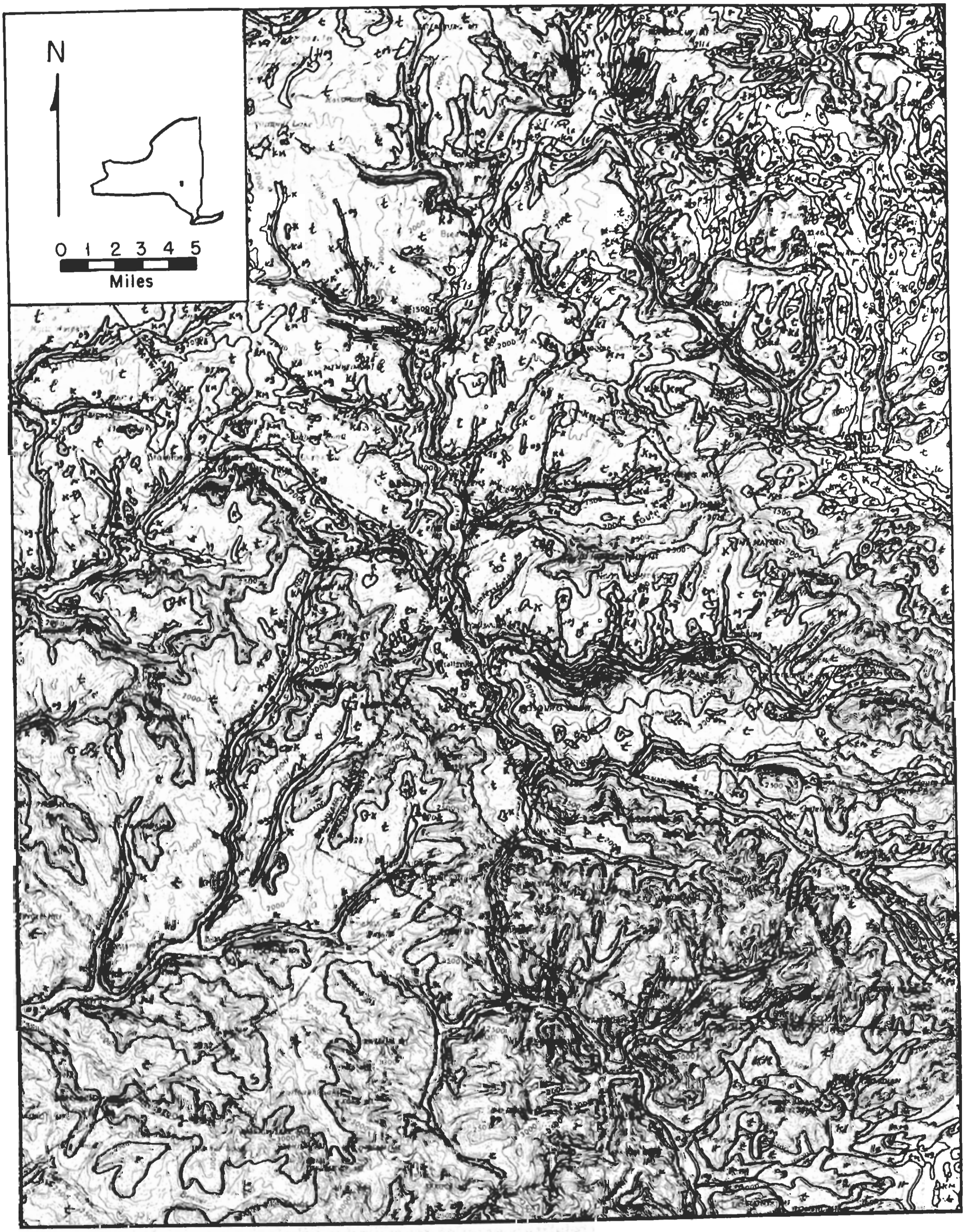

Figure 2. Surficial geologic map of the central Catskill Mountain region. The map symbols and unit descriptions are summarized in Table 1.

Table 1. Surficial Mapping Units and Descriptions

Map Unit	Symbol	General Sediment Characteristics and Environment of Deposition
Alluvium	al	recent deposits of fine sand to gravel generally confined to floodplains within a valley oxidized noncalcareous in larger valleys may be overlain by a veneer of silty overbank deposits subject to frequent flooding
Swamp Deposits	pm	peat-marl, organic silt and sand in poorly drained areas unoxidized may overlie marl and lake silts thickness generally 2-10 meters
Kame Deposits	k	including kames, kame terraces, kame deltas, eskers coarse to fine gravel and/or sand deposition adjacent to ice lateral variability in sorting, coarseness and thickness generally calcareous, locally firmly cemented with calcareous cement
Kame Moraine	km	extremely variable texture (size and sorting) from boulders to sand result of deposition on and around an ice margin during glaciation generally calcareous locally cemented with calcareous cement
Outwash Sand & Gravel	og	coarse to fine gravel with sand proglacial fluvial deposition well rounded and stratified generally thinner, with finer texture away from ice border
Lacustrine Sand	ls	sand deposits associated with deposition in large bodies of water generally a near-shore deposit well sorted stratified generally quartz sand thickness variable (2-20 meters)
Lacustrine Clay	lc	generally laminated clay and/or silt deposited in proglacial lakes usually calcareous
Till Diamicton	t	variable texture (silt, silt-clay, clay, boulder clay, etc.) uniformly unsorted, although there may be local stratified lenses heterogeneous throughout deposited beneath the glacier, relatively impermeable (loamy matrix) variable clast content – ranging from abundant well-rounded diverse lithologies in valley tills to relatively angular, more limited lithologies in upland tills usually is lodgement or melt-out till; compact, firm and consolidated

Table 1 (continued)

Map Unit	Symbol	General Sediment Characteristics and Environment of Deposition
Supraglacial Diamicton	tm	much like till diamicton, but more variable in sorting (coarser than associated till diamicton) generally more permeable than till diamicton deposition adjacent to ice
Bedrock	r	bedrock exposed or within 1 meter of surface may be sandstone, shale or conglomerate

(now known as Johnson Hollow Brook), west of Prattsville, and suggested that local and independent mountain glaciers formed the moraines. Continuing his studies of the glacial features in the Catskills, Rich (1915) examined kame deltas in the Schoharie drainage basin and related them to a glacial lake whose outlet at Grand Gorge permitted meltwaters to flow into the East Branch of the Delaware River. He also noted the absence of deltas in the upper Schoharie Creek valley. Rich's theory of local glaciation was supported by Johnson (1917) who compiled and compared the evidence of local glaciation in the White, Adirondack and Catskill Mountains.

In 1935, Rich presented a detailed description of the glacial deposits and stratigraphy of the Catskill Mountains. His reconstruction of the history of glacial retreat from the Catskills includes six ice margin positions, marginal drainage and glacial lakes. LaFleur (1965), during his study of the lower Schoharie Valley, concentrated on ice-defended (expanding) Lake Schoharie with sequential outlets at 262 m, 213 m and 128 m (860, 700 and 420 feet, respectively). Using subsurface data, LaFleur (1969) suggested four glacial readvances in the Schoharie Valley. Cadwell (1983) suggested a revision of the Woodfordian history of the Catskill Mountains and vicinity to include only three ice margin positions.

Other workers have conducted research in areas peripheral to the Catskill Mountains. Cook (1924, 1942) described the glacial history of the Catskill quadrangle and the disappearance of glacier ice in eastern New York. Connally (1966, 1968, 1983) and Connally and Sirkin (1967, 1970, 1973) described the Wisconsinan moraines in the Hudson and Wallkill Valleys. Dineen (1983) mapped surficial geology along the western Hudson Valley between Schenectady and New Paltz while Fleisher (1983) identified retreating ice margin positions in the upper Susquehanna Valley.

GLACIAL HISTORY

Glacial Deposits

Reconnaissance field mapping on 20 U.S.G.S. 7-1/2 minute topographic quadrangles was compiled on 7-1/2 minute maps reduced to a scale of 1:62,500. Figure 2 is a generalized map of the surficial geology of the study area, at a scale of 1:250,000. The legend is adapted from the one used by the New York State Surficial Mapping Program. Table 1 contains a summary of the mapping units, map symbols, general sediment characteristics and the probable environment of deposition.

The glacial deposits of the Catskill Mountain area can be classified into four general categories: diamicton, ice-contact stratified drift, outwash and glacio-lacustrine deposits. A diamicton is a poorly sorted deposit of variable grain size and may have been deposited beneath a glacier. A diamicton also may be ablation till, formed as inactive glacier ice melts, releasing the glacial debris to be partially washed by meltwater. Ice-contact stratified drift refers to sand and gravel that is deposited by meltwater streams on or adjacent to glacier ice. These deposits generally occur near an ice margin, such as the terminous of a valley ice tongue. Outwash is sand and gravel deposited by meltwater streams flowing away from the ice margin. Glacio-lacustrine deposits are stratified accumulations of sand, silt and clay deposited in a lake marginal to the glacier.

Multiple Glaciation

Exposures that document multiple glaciation in the Catskill Mountain area are scarce. Rich (1935) described some exposures in the Schoharie Valley that suggest a multiple glaciation. However, he emphasized that the most recent glacier had the greatest influence and had a more or less uniform retreat. LaFleur (1969) proposed that four readvances occurred during the retreat of the last glacier based on an analysis of subsurface well data. I have located only one surface exposure with two different tills. This two-till exposure, west of Prattsville, has a clay-rich, clast-deficient diamicton above a silty-clay cobble diamicton.

Records from 15 wells within the Schoharie drainage basin suggest 2 till diamictons separated by sand, gravel and clay; four wells suggest three separate till diamictons; and two wells suggest 4 till diamictons. The absolute age of the tills is not known. It is assumed that the uppermost till diamicton was deposited during the Woodfordian glaciation, and it is possible that all the diamictons are Woodfordian.

Glaciation Models

Rich Model

Rich (1935) recognized 6 "stages" of ice margin positions in the Catskill Mountain area in addition to Early(?) Wisconsin ice retreat positions (Fig. 3).

(1) The Late Wisconsin Terminal(?) Moraine position is marked by a series of moraines south of the central escarpment.

(2) The Peekamoose Gorge-Summitville Lake position also was located south of the central escarpment. He envisioned ice banked against the north side of the escarpment with tongues extending southward through the major cols at Grand Gorge, Deep Notch and Stony Clove Notch.

(3) The ice margin for the Wagon Wheel Gap position extended northward from Wagon Wheel Gap (on the east side of High Point) to the south side of the central escarpment and Plattekill Mountain. In the upper Schoharie Valley the ice margin followed the north side of the central escarpment at an elevation of 762 m (2500 feet), with a lobe extending into the West Kill valley and through Deep Notch.

(4) The Schoharie Valley position actually is an extension of the Wagon Wheel Gap position into the Schoharie Valley. Glacier ice at this time was in the upper Schoharie Creek and Batavia Kill valleys (Fig. 1). Its margin extended westward along the central escarpment at an elevation of about 610 m (2000 feet) to Grand Gorge and South Gilboa.

(5) The Grand Gorge Lake position became established when the lobe of ice occupying the Schoharie Valley retreated far enough to permit drainage of the lake through the Grand Gorge channel. Elevation of the floor of this channel is about 479 m (1570 feet). Grand Gorge Lake was long-lived; it was necessary for the glacier to retreat about 29 km (18 miles) to Middleburg and the Franklinton channel, before a lower outlet was uncovered. The size of the lake is indicated by numerous deltas at an elevation of $\pm$ 488 m (1600 feet) located in the Schoharie, Batavia Kill and Manor Kill valleys. A large delta in the Platter Kill valley has a wave-cut terrace 9 m below the top of the delta, implying a lowering of lake level. Rich noted the absence of Grand Gorge Lake "stage" deltas in the upper Schoharie valley and suggested the area was abandoned by ice at that time.

(6) The Franklinton channel "stage" succeeded the Grand Gorge Lake position with the abandonment of the outlet at Grand Gorge. Meltwater then flowed through this lower outlet at 366 m (1200 feet) to the Catskill Creek and the Hudson Valley.

LaFleur Model

LaFleur (1969) hypothesized four glacial readvance positions in the Schoharie Valley during the retreat of the last ice sheet (Fig. 4).

(1) The Tannersville readvance position, located along the eastern edge of the central escarpment, resulted from a glacier readvance south and eastward up the Schoharie Valley. In the upper Schoharie Valley the margin merged with the main lobe of Hudson Valley ice that crossed the divide near Haines Falls. Glacial meltwater drained through the Grand Gorge channel while the channel was still occupied by an ice tongue. He referred to the lake that subsequently developed as Glacial Lake Schoharie. The glacier then retreated at least to Middleburg, with drainage through the Franklinton channel, prior to the Prattsville readvance.

(2) The Prattsville readvance had side lobes extending into the Grand Gorge, Manor Kill and Platter Kill valleys. Glacial Lake Schoharie was restored, again with the Grand Gorge channel as the exit for all meltwater. As Lake Schoharie increased in size, stagnating ice blocks remained in the Schoharie Valley. Meltwater drainage from the Manor Kill, Platter Kill and Keyser Kill passed over the ice to the Grand Gorge outlet. Near Prattsville, lacustrine clays were deposited in Lake Schoharie at an elevation of 427 m (1400 feet). Again the

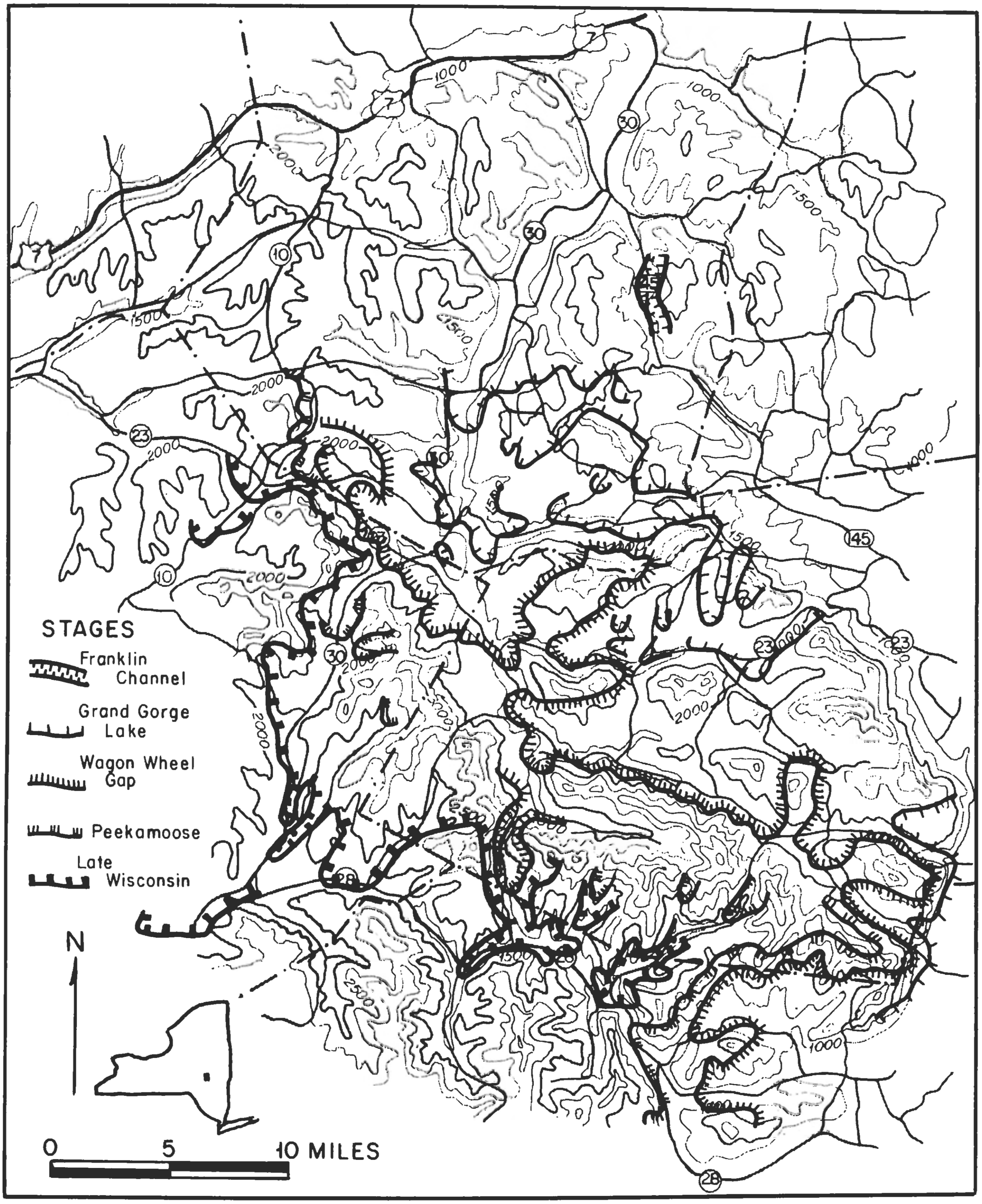

Figure 3. Ice margin positions summarized from Rich (1935).

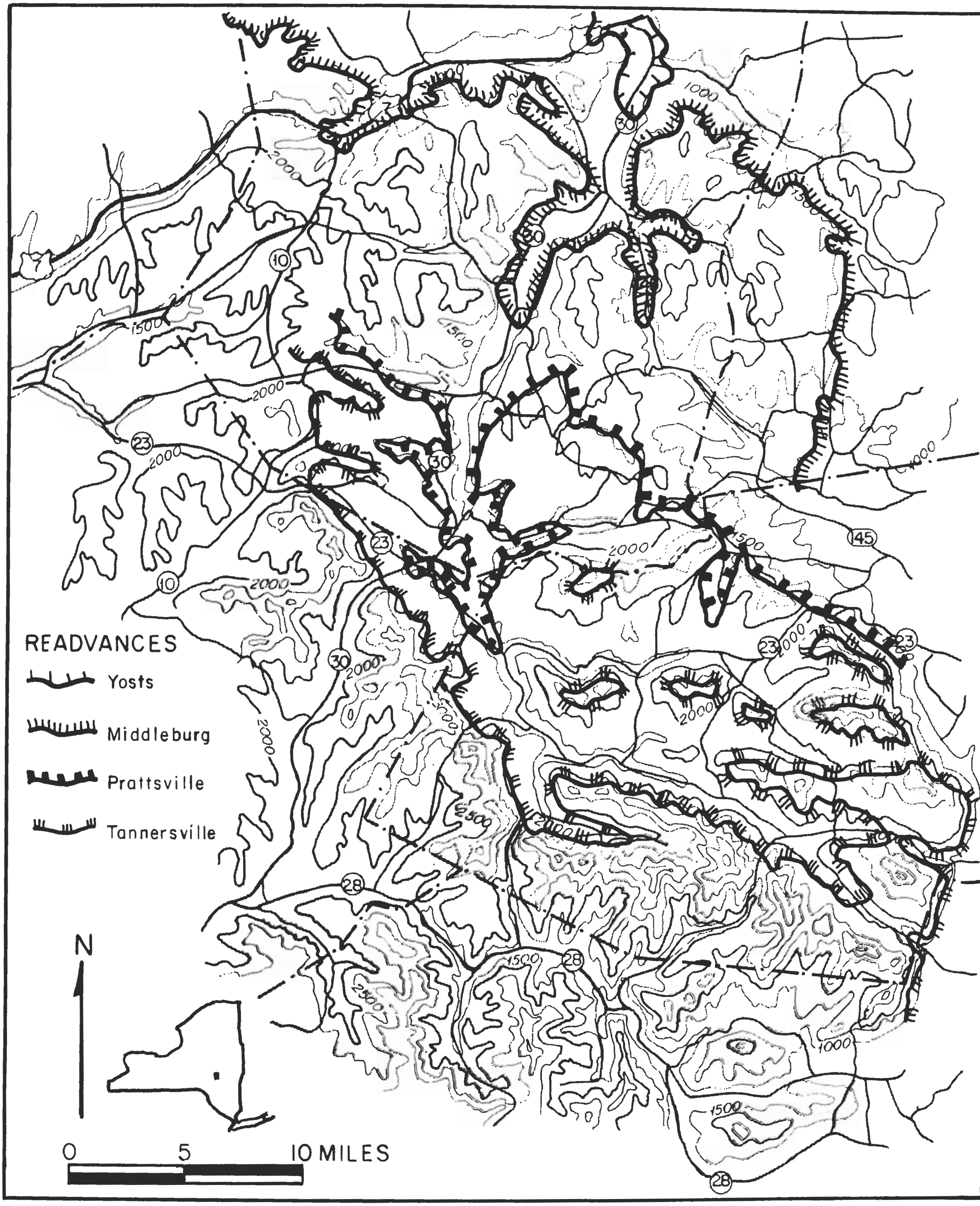

Figure 4. Ice margin positions summarized from LaFleur (1969).

glacier retreated, and again to at least the vicinity of Middleburg, permitting the lowering of Lake Schoharie from 488 m to 366 m and reestablishing meltwater flow through the Franklinton channel into the upper Catskill Valley.

(3) A third readvance of the glacier in the Schoharie Valley, the Middleburg readvance, caused the construction of morainal features near Huntersland, Breakabeen, Richmondville and in the Franklinton channel. Lake Schoharie was reestablished at 366 m (1200 feet) permitting deposition of lake clays.

(4) The glacier then retreated to the north of Schoharie and readvanced to the Yosts position in the lower Schoharie Valley.

Proposed New Model

I suggest that the Catskill Mountains were overtopped by glacier ice at least once during the Pleistocene Epoch. The Woodfordian glacial advance reached a maximum about 21,750 yrs BP (Sirkin and Stuckenrath, 1980) when the terminus of the glacier was at Long Island. The Catskills probably were covered with ice at this time. Three Woodfordian recessional ice margins currently are identified in the study area (Fig. 5). The southernmost continuous position of glacier ice within the Schoharie Valley is similar to those mapped by both Rich and LaFleur; however, it cannot be demonstrated that this margin is a result of a glacier readvance. For this reason, I have retained the terminology of Rich (1935) rather than that of LaFleur (1969). Likewise, LaFleur's Prattsville readvance was not confirmed during this study. I interpret the mapped glacial deposits to represent a progressive retreat of the Woodfordian glacier with the creation of a large proglacial lake that drained through the Grand Gorge channel. I therefore retain Rich's (1935) nomenclature – Glacial Lake Grand Gorge – for that lake.

(1) The Wagon Wheel Gap ice margin is identified along the north side of the central escarpment at an elevation of 762 m (2500 feet). This is similar, in part, to the Wagon Wheel Gap position of Rich and the Tannersville position of LaFleur (compare Figs. 3, 4 and 5). The Delaware River valleys were essentially ice-free, permitting the exit of meltwater through channels at Stamford, Grand Gorge, Deep Notch and Stony Clove. A sublobe of inactive Hudson Valley ice probably remained in the Esopus Creek valley.

(2) The Glacial Lake Grand Gorge ice margin represents the interval during glacier retreat when the glacier-dammed Lake Grand Gorge drained through the meltwater channel at Grand Gorge. This was not an open-water lake. Glacier ice remained in the main Schoharie Valley as ice contact kames and kame deltas were deposited in the Batavia Kill valley at Ashland and Windham; the Platterkill valley 3 km northeast of Gilboa; and at Mosquito Point, 6 km south of Prattsville. Meltwater flowed around and across the ice toward the Grand Gorge outlet. This ice-choked lake remained during glacier retreat to the North Blenheim-Haines Falls position of ice retreat.

(3) The North Blenheim-Haines Falls ice margin represents the final position of the retreating glacier prior to the draining of Lake Grand Gorge. The lake level lowered as the Franklinton channel became ice-free. A 366 m (1200 feet) ice-dammed lake formed in front of the retreating glacier as meltwater flowed through the Franklinton channel. Ice-contact deposits at Mosquito Point suggest that large blocks of glacier ice may have remained within the upper Schoharie Valley as the glacier retreated prior to the Middleburg readvance of LaFleur (1969).

Woodfordian Glaciation

The character of the till deposited by the advancing Woodfordian glacier was a function of the material available to the ice. In the Schoharie Valley west of Prattsville two tills were exposed in a foundation excavation at an elevation of 402 m (1320 feet). Here a clay-rich, clast-deficient diamicton overlies a silty-clay cobble diamicton. The lower diamicton may have been deposited during a pre-Woodfordian glacial advance. The advancing Woodfordian glacier may have then deposited the upper clay-rich diamicton after over-riding lacustrine clays deposited in a proglacial lake in the Schoharie Valley.

The earliest Woodfordian stratified glacial deposits found within the study area are located within the valley of the East Branch of the Delaware River. Kame moraines and kame terraces near Roxbury suggests a sequential retreat of a tongue of ice in the valley. Temporary marginal lakes were created adjacent to the ice tongue, as shown by kame deltas constructed along the valley walls.

Wagon Wheel Gap Ice Margin

The Wagon Wheel Gap ice margin position is located along the north side of the central escarpment at an elevation of 762 m (Fig. 5). Glacier ice probably overtopped the mountain peaks of the eastern escarpment at this time. The Wagon Wheel Gap position ice margin/has been traced northwestward from the eastern escarp-

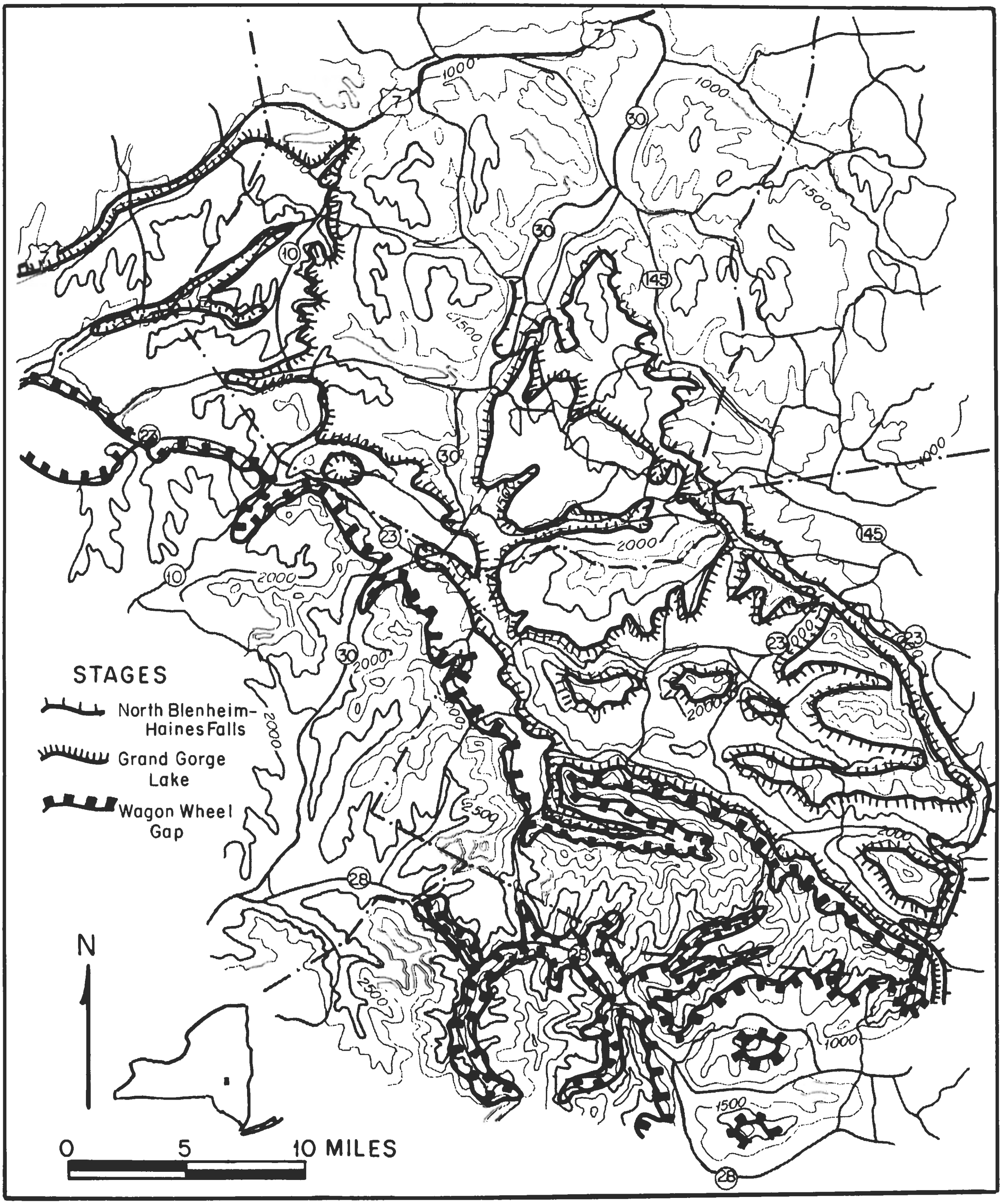

Figure 5. Ice margin positions currently identified in the study area.

ment at Platterkill Mtn. to Stamford. Beyond here, kame moraines suggest that the margin can be traced farther west to the Charlotte Creek valley.

A lobe of ice remained in the Esopus Creek valley as a remnant during the thinning of the glacier at the central escarpment and was attached to the main glacier in the Hudson Valley. A large kame delta at Bushnellsville (see Fig. 2) was deposited in an ice dammed lake as meltwater flowed from the active ice margin in the West Kill valley and Deep Notch. Meltwater also flowed through Stony Clove Notch to the Esopus Creek ice lobe in Stony Clove Creek, where stratified drift was deposited in contact with the disintegrating ice. However, most of the glacial meltwater drained to the Delaware Valley at Grand Gorge and Stamford as tongues of ice extended through gaps in the central escarpment.

Glacial Lake Grand Gorge Ice Margin

The Glacial Lake Grand Gorge ice margin is based upon the development of Lake Grand Gorge whose spillway was at Grand Gorge. This lake developed because the spillway at Grand Gorge was the only exit for meltwater in the upper Schoharie Valley. Lake Grand Gorge should not be confused with Glacial Lake Schoharie (Brigham, 1908) which was confined to the lower Schoharie Valley. Glacier retreat from the Wagon Wheel Gap ice margin was accomplished primarily by downwasting. The glacier thinned first in the vicinity of Prattsville, while more massive stagnant ice remained in the upper Schoharie Valley and active ice occupied the lower Schoharie Valley. The initial stage of Lake Grand Gorge, therefore, was as a supraglacial lake, confined to the Schoharie Valley near Prattsville, whose outflow was through the Grand Gorge channel at an elevation of 488 m (1600 feet). With continued thinning of the glacier the lake became ice-choked, with meltwater flow around and over disintegrating blocks of ice.

Meltwater streams flowed into this lake and constructed deltas in the Platter Kill and Manor Kill valleys at 488-497 m (see Fig. 2). The large delta in the Platter Kill valley was constructed as meltwater flowed from the active ice margin near Broome Center. Active ice adjacent to the eastern escarpment supplied meltwater to the Manor Kill valley. The accumulation of meltwater in these two valleys may have hastened the melting of glacier ice near Prattsville and the development of Lake Grand Gorge.

Sand and gravel deposits in the Batavia Kill and upper Schoharie Valley suggest deposition associated with a stagnant ice mass that was separated from the active retreating Schoharie ice sublobe and the lobe in the Hudson Valley. This lingering inactive ice mass was a major meltwater source as Lake Grand Gorge gradually increased in size. The 482-488 m deltas in the Manor Kill valley were deposited adjacent to a lobe of dead ice by meltwater flowing through a col from the active ice margin at the northeastern escarpment. As the ice in the Batavia Kill valley thinned, this valley became part of the ice-choked Lake Grand Gorge and the 475-488 m deltas between Ashland and Windham were deposited. The absence of deltas in the upper Schoharie Valley at an elevation of 488 m suggests that dead ice remained in this valley throughout the life of Lake Grand Gorge.

Concurrent with the development of Lake Grand Gorge glacier retreat was active in the western part of the lower Schoharie Valley. This part of the ice margin position was traced northward from Bald Mountain (northeast of Stamford) and then generally southwestward in the Schenevus Valley. Meltwater from the retreating ice margin flowed past Stamford into the Delaware Valley; from the ice tongue at Jefferson toward the Charlotte Creek valley; and adjacent to the ice margin in the Schenevus Creek valley. A large mass of dead ice remained in the Charlotte Valley, detached from the retreating margin.

Active glacier ice remained on the eastern flank of the northeastern escarpment at an elevation of 610 m (2000 feet), contributing meltwater through cols to the ice-choked Lake Grand Gorge.

During the Glacial Lake Grand Gorge ice margin local cirque glaciers remained in protected valleys. Meltwater from the local glacier in Schoendorf cirque at the head of Johnson Hollow Brook (formerly called Fly Brook by Rich, 1935), constructed a delta into Lake Grand Gorge at an elevation of 488 m (Fig. 6). Within this cirque is a small lake dammed by a moraine with 12 m (40 feet) of relief. The cirque headwall, with 457 m (1500 feet) of relief, has a complex talus deposit, that suggests the glacier persisted for a relatively long period (Fig. 7). Glacier ice gradually melted from the headwall, permitting the development of six successive talus deposits.

North Blenheim-Haines Falls Ice Margin

The North Blenheim-Haines Falls ice margin represents the last position of active glacier ice prior to the draining of Lake Grand Gorge. Dead ice still choked the Schoharie Valley as lake clays were deposited between the ice blocks. The deposits of clay between North Blenheim and Gilboa, and those near Prattsville, formed at this time. Surface elevations of the clays range from 427 m at Prattsville to 381 m at North Blenheim.

As the glacier retreated to a position north of Broome Center, a series of deltas was constructed in the Keyser

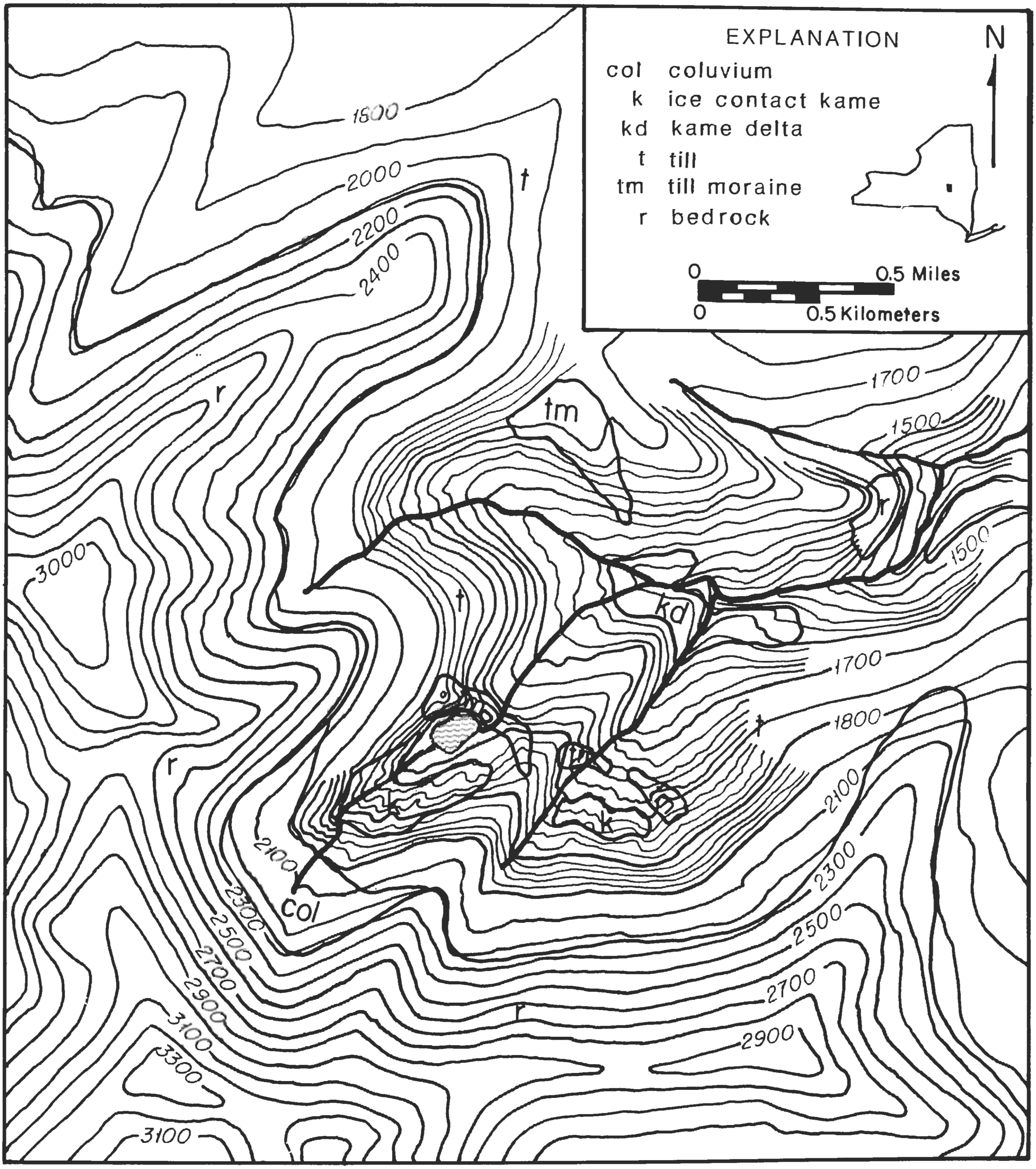

Figure 6. Surficial geologic map of the Schoendorf cirque (located west of Prattsville on Figure 1).

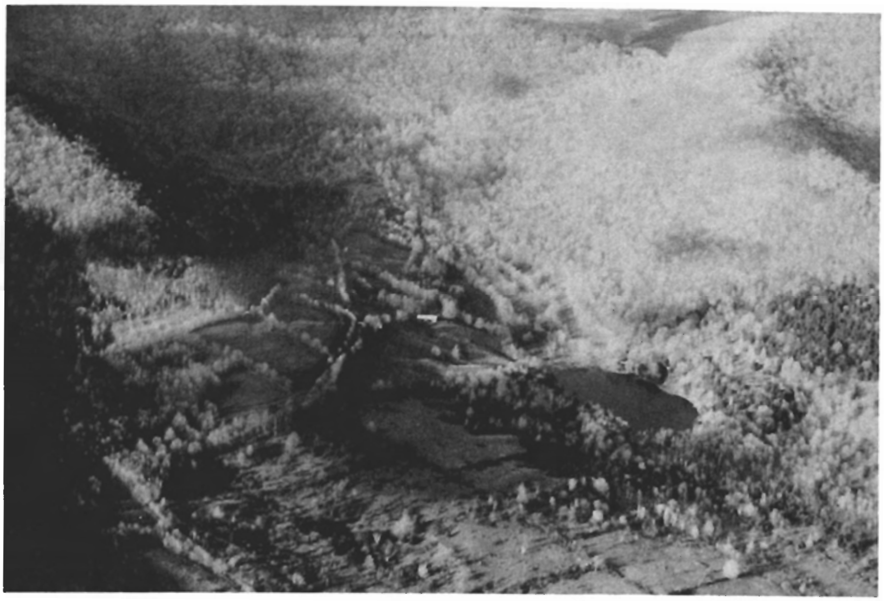

Figure 7. Photographs of the Schoendorf cirque.
7A, above, is a view to the west, showing the cirque headwall and basin features, including the moraine and moraine dammed lake. 7B, below, is a view to the east, showing the moraine, lake, and ice-contact kame (esker) features near the buildings in the foreground. Compare these photographs with Figure 6, the surficial geologic map of the cirque.

SUSQUEHANNA VALLEY	CATSKILL MOUNTAINS	HUDSON VALLEY
?	North Blenheim–Haines Falls	?
New Berlin	Glacial Lake Grand Gorge	Rosendale
Oneonta	Wagon Wheel Gap	Wallkill

Table 2. Correlation of ice margin positions in the Susquehanna River drainage basin, Catskill Mountains and the Hudson Valley.

Kill valley. The 597 m (1960 feet) delta was deposited as meltwater flowed into the Platter Kill valley. With continued ice retreat in the Keyser Kill valley a delta was deposited at 572 m (1875 feet), 1-1/2 miles north of Broome Center. Meltwater flowed adjacent to the ice toward the col northwest of Safford Mountain, and then into Lake Grand Gorge at North Blenheim where a delta was deposited at 488 m. As the glacier ice retreated from the Keyser Kill valley, deltas were deposited in the expanding Lake Grand Gorge at elevations of 494, 488 and 475 m.

The waters of Lake Grand Gorge adjusted from 488 m to 366 m when glacier ice melted from the Franklinton channel. The change of lake level may have been rapid, as suggested by the absence of transition deltas in the Schoharie Valley and the presence of gravel under clay in the Catskill Valley. Two transition deltas are present, however, in the Keyser Kill valley at 427 m and 384 m. Lake clays preserved in the Schoharie Valley at 366 m and 305 m might have been deposited during this transition. The glacier retreated, and this 366 m lake may have drained prior to the Middleburg readvance of LaFleur (1969).

CORRELATION

Tentative correlations between the Catskill Mountains and the adjacent Susquehanna, Delaware and Hudson River Valleys are given in Table 2.

Susquehanna and Delaware River Valleys

Westward, the Wagon Wheel Gap Ice Margin in the Catskill Mountains is correlated with the Oneonta Ice Margin of Fleisher (this volume). At this time active glacier ice was completely removed from the Delaware Valley. Glacial meltwater, however, was flowing through gaps in the central escarpment at Stamford and Grand Gorge, producing the outwash valley train along the West and East Branch of the Delaware River.

The Glacial Lake Grand Gorge ice margin is correlated to the west with the New Berlin Ice Margin of Fleisher (1983; this volume). Meltwater flowed from that margin to the Susquehanna drainage basin at North Harpersfield and to the Delaware drainage basin at Grand Gorge channel. The North Blenheim-Haines Falls Ice Margin has not been traced into the Susquehanna Valley.

Hudson River Valley

The Wagon Wheel Gap ice margin in the Catskills is correlated with the Wallkill Moraine of Connally (1983; this volume), Connally and Sirkin (1970) and Dineen (1983; this volume) in the Wallkill Valley. The Glacial Lake Grand Gorge ice margin is correlated with the Rosendale readvance, south of Kingston, in the Hudson Valley.

CONCLUSIONS

Subsurface data suggest pre-Woodfordian glacier advances. The Woodfordian glacier advance deposited a diamicton on the north side of the central escarpment as the glacier overrode the Catskill Mountains. This glacier began retreat from its terminal position on Long Island by 21,750 yrs BP. The earliest Woodfordian stratified drift in the Catskills was deposited during deglaciation of the East Branch of the Delaware River. The age of this deposit is, however, not known.

The Wagon Wheel Gap ice margin of glacier retreat occurred when glacier ice was located near the crest of the central escarpment. At this time, an ice tongue occupied the gap in the escarpment at Grand Gorge with meltwater flow to the East Branch of the Delaware River. Meltwater also flowed through Deep Notch and Stony Clove Notch where it deposited sediment in contact with disintegrating ice in the Esopus Creek valley. The ice-contact features in Bushnellsville and Stony Clove Creek valleys were formed at this time. Meltwater then flowed to Wagon Wheel Gap on the eastern side of High Point.

Glacial Lake Grand Gorge occurred when glacier ice thinned to an elevation lower than the threshold of Grand Gorge channel (488 m). Kame deltas with an elevation of $\pm$ 488 m suggest that this lake gradually expanded in the central Schoharie Valley and into the Manor Kill, Batavia Kill, and Keyser Kill valleys. Concurrent with the development of Lake Grand Gorge were local cirque glaciers in protected valleys on the high escarpments. Meltwater from the glacier in Schoendorf cirque constructed a delta in Lake Grand Gorge west of Prattsville.

The North Blenheim-Haines Falls ice margin was the last glacial ice position prior to drainage of Lake Grand Gorge through Franklinton channel. During subsequent ice retreat, the glacier established the Middleburg and Yosts(?) readvance positions.

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DEGLACIATION OF THE HUDSON VALLEY BETWEEN HYDE PARK AND ALBANY, NY

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ABSTRACT

Deglaciation in the Hudson Valley involved both stagnant and active glacial retreat. Recession of the Wallkill-Phillipsport-Wagon Wheel Gap ice margin was characterized by a large stagnant area in the Wallkill and Minisink Valleys. A moraine was deposited adjacent to stagnant ice in the Wallkill Valley near New Paltz, while contemporaneously lacustrine silt and clay was deposited in an ice-choked lake in the Minisink Valley. The glacier retreated rapidly from Poughkeepsie to Stuyvesant, concurrent with the northward expansion of Glacial Lake Albany in the Hudson Valley. The Rosendale readvance interrupted glacial retreat and deposited a moraine from Whitfield to West Park. The subsequent retreat from the Rosendale margin was discontinuous, so that recessional moraines were deposited at Stone Ridge, Cottekill, and Ulster Park. Later the glacier stagnated and ice blocks remained in the Hudson Valley between Ravena and Kingston. The Middleburg Readvance glacier overrode stagnant ice in the Hudson Valley and constructed moraines at Middleburg-Franklinton, Oak Hill, Lawrenceville, and Red Hook. Glacial ice in the Schoharie Valley and on the Helderberg Plateau became isolated from the active ice between Ravena and Cairo. In the Hudson Valley, Glacial Lake Albany expanded as the glacier retreated from Ravena to Round Lake. The Delmar readvance then overrode the Lake Albany deposits, and deposited an end moraine from Rotterdam to Altamont, Voorheesville, and Feura Bush. Final recession of the glacier was accompanied by further growth of Lake Albany.

INTRODUCTION

The deglaciation of the Hudson Valley from Poughkeepsie to Albany has been hotly debated since Woodworth's 1905 publication on the late glacial history of the valley. Subsequent workers have not been able to agree on the age and mode of retreat of the glacial ice.

Study Area and Methods

The study area in eastern New York State is bordered by the Schoharie Creek, Mohawk River, Hudson River, and the lower Wallkill Valley (Fig. 1). It includes part of the Helderberg Plateau, Minisink, Schoharie, and Wallkill Valleys, and the Catskill, Shawangunk, and Marlboro Mountains. The Helderberg Plateau is underlain by Devonian limestone and shale with subordinate sandstone. The valleys follow belts of Ordovician shale, while ridges are composed of Ordovician and Silurian sandstone with subordinate shale.

The ice margin positions, style of glacial retreat, and relative glacial stratigraphy of the west side of the Mid-Hudson Valley were studied using 28 reconnaissance-level and 10 detailed 7-1/2 minute quadrangle geologic maps. These maps were compiled for the data base for the Hudson-Mohawk and Lower Hudson sheets of the State Surficial Map. The maps show the areal extent of various glacial materials (till, lacustrine clay, kame terraces, etc.), based on field reconnaissance and airphoto interpretation by the author. Subsurface data from a bedrock topography study by the author also was used. These data include water well and test boring logs, geophysical studies, and stratigraphic sections in gravel pits. The maps and subsurface data are available for inspection at the New York State Geological Survey.

Lake clays or shoreline deposits were used to define the limits of proglacial lakes. Moraine segments, kame complexes, heads-of-outwash, and ice-marginal meltwater channels were used to delineate ice margins.

Previous Studies

Woodworth (1905) suggested that active glacial retreat in the Mid-Hudson Valley was punctuated by periodic readvances. Researchers soon divided into groups that defended two models of glacial retreat: the "stagna-

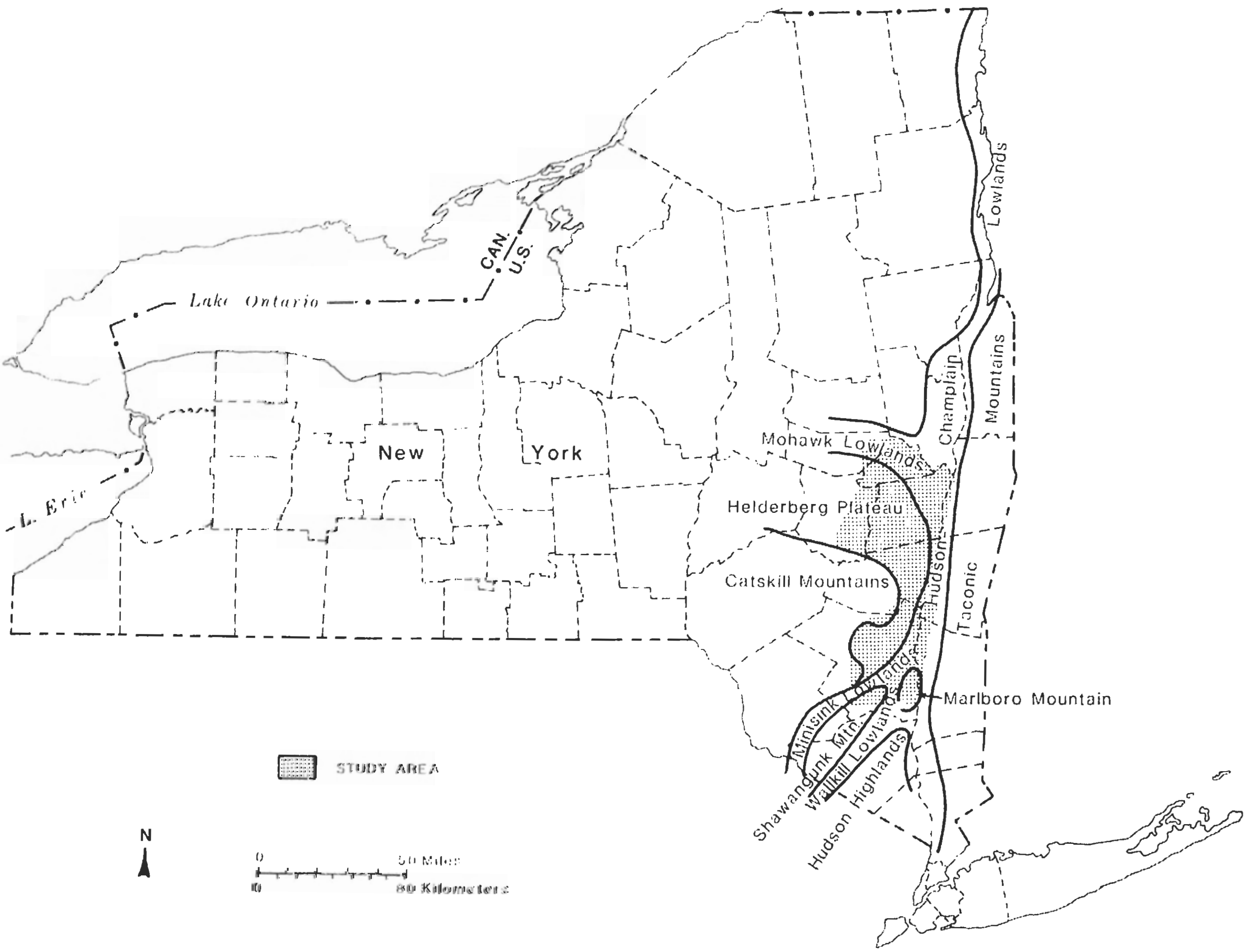


Figure 1. Location of the study area, with the physiographic provinces of eastern New York State.

tors," championed by Cook (1924, 1930, 1942), and the "activists," represented by Chadwick (1910, 1928, 1944), Stoller (1919, 1920, 1922), and Rich (1914, 1935). The "stagnators" gave numerous examples of sediments deposited on or next to stagnant ice. The "activists" countered with examples of deposits from active glaciers, sometimes in the same areas! Cook (1924, 1935) found evidence for stagnant ice on the Helderberg Plateau and for discontinuous glacial lakes between Ravena and Catskill. Stoller (1920, 1922) and Chadwick (1910, 1928), however, showed that an active glacier constructed recessional moraines in the northward-expanding proglacial Lake Albany. Woodworth (1905) and Stoller (1920), in their studies of the Mid-Hudson Valley, emphasized that stagnant ice blocks remained in the inner Hudson Valley during Glacial Lake Albany time.

Later workers found that both stagnant ice and active glacial retreat were necessary to explain the origin of the glacial deposits in the Hudson Valley. Connally and Sirkin (1970) suggested that the Hudson-Champlain glacial lobe stagnated between the Wallkill moraine and New Paltz. They later attributed these sediments to deposition by active ice (Connally and Sirkin, 1971, 1973). Interpretation of the style of glacial recession changed again when Connally (1980, 1982) suggested oscillatory glacial retreat, using the Rosendale and later readvances as evidence for glacier oscillations (Connally and Sirkin, 1970).

LaFleur (1961, 1965) also found deposits emplaced during both active and stagnant glacial retreat. He emphasized active glacial retreat for the main ice mass (LaFleur, 1969a, 1979), but proposed stagnant ice in contact with Lake Albany in the inner Hudson Valley (LaFleur, 1969b).

Rich (1935), later supported by Cadwell (1983), mapped the Wagon Wheel Gap ice margin in the Catskill Mountains and projected it into the Minisink Valley. Happ (1938) and Heroy (1974) mapped an ice margin in the Minisink Valley at Phillipsport and believed that Glacial Lake Wawarsing expanded between the Phillipsport moraine and the retreating Delaware-Minisink sublobe.

Connally and Sirkin (1970) mapped ice margins in the Wallkill Valley at Wallkill and Rosendale. Connally (1980) correlated the Wallkill moraine with the Shenendoah moraine east of Beacon; the Rosendale readvance with the Red Hook moraine, and the Middleburg readvance with an ice margin from Kinderhook Lake to Brainard. In this volume Connally and Sirkin used a projected date from the New Hampton Bog to place the Wallkill lobe at the Wallkill Moraine at 17,210 years ago; they projected withdrawal of the glacier from the Albany area at 13,800 years ago (Fig. 6 in Connally and Sirkin, 1973).

In the Schoharie Valley, LaFleur (1965, 1969a) mapped readvance margins at Middleburg and Schoharie. He correlated the southernmost margin with the Middleburg readvance, and extended it from Middleburg to Rensselaerville (LaFleur, 1979). He traced the northern margin of the Middleburg readvance to Yosts and mapped several recessional ice margins between Schoharie and Niskayuna (LaFleur, 1979). Dineen (1981, 1983) traced the Middleburg ice margin across the Helderberg Plateau from Rensselaerville to Lawrenceville. Dineen and Rogers (1979) found tills and folded clay of a readvance in Glacial Lake Albany sediments between Albany and Delmar. They named it the Delmar readvance.

GLACIAL MOVEMENT

Striae orientations from Brigham (1929), Rich (1935), and the present study are combined with average values of drumlin orientations in each 7-1/2 minute quadrangle to define directions of the glacial movement (Fig. 2). These ice movement indicators delineate the south-moving Hudson-Champlain and west-moving Mohawk Glacial Lobes. Each major lobe was split into sublobes by the underlying bedrock topography, especially during deglaciation when the ice sheet was thin. The late glacial ice movement was funnelled around uplands and into valleys, creating several sublobes with a strong south-west movement, especially across the Helderberg Plateau into the Catskill Valley, and around the southeastern Catskills into the Minisink and Wallkill Valleys.

ICE MARGINS

Nine distinct glacial margins were mapped in the Mid-Hudson Valley between Albany and Hyde Park. These margins record the complex glacial recession caused in part by the influence of bedrock relief on active and stagnant receding ice. The Hudson Lobe readvanced three times during late Woodfordian time; each was associated with multiple recessional margins. In southward-narrowing valleys the advancing glacial lobes were funnelled into long tongues of ice, whereas in southward-widening valleys the glacial lobes spread laterally and advanced only modest distances. At the end of each readvance there was a widespread stagnation, especially in rugged valleys or on cuestas with north-facing escarpments that cut off the glacier from its source of ice.

Large areas of stagnation moraines were deposited. These moraines are composed of meltwater-sorted, poorly stratified sand, gravel, and diamictons. They are characterized by a mosaic of "ablation tills" (loose, unstratified, clay-poor diamictons) and ice-contact stratified drift cut by criss-crossing outwash channels.

Wagon Wheel Gap Ice Margin

The Wagon Wheel Gap ice margin extends from the southeastern Catskill Mountains to the Phillipsport moraine in the Minisink Valley (Fig. 3). The margin wraps around the Shawangunk Mountains into the Wallkill Valley to the Wallkill moraine. The ice margin was named for the major meltwater channel that was carved into a bedrock spur near Cherrytown by drainage from the eastern Catskill Mountains (Rich, 1935). Meltwater from the gap entered Glacial Lake Wawarsing. Cadwell (1983) traced this margin into the Catskill Mountains, and found evidence that peaks in the northeastern Catskill Mountains formed nunataks. Folded clays in the Hudson Valley south of Hyde Park (N.Y.S. Department of Transportation and E. Hanson, Dunn Geoscience Corporation, personal communication, 1982) suggest that this margin was established by a re-

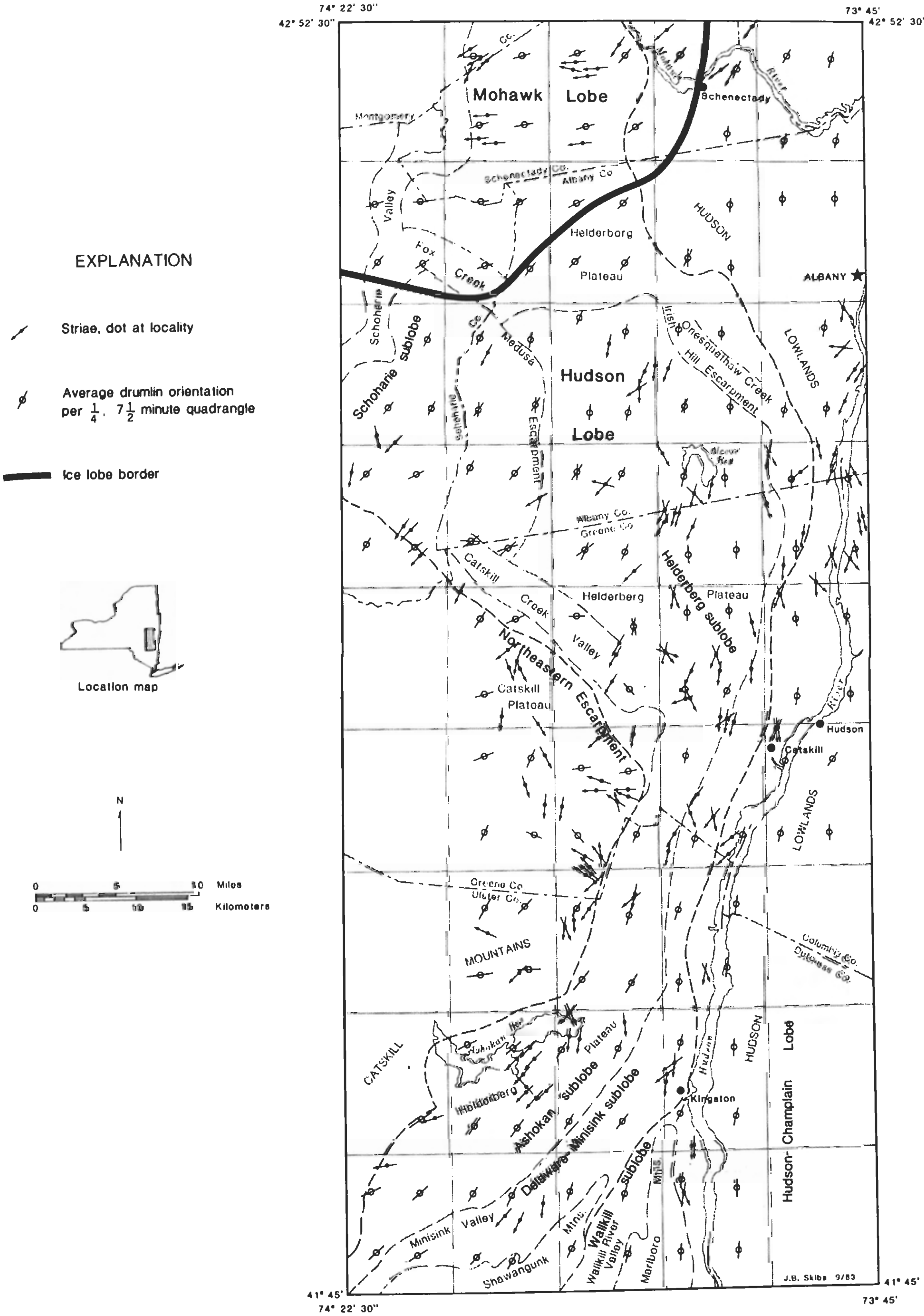


Figure 2. Glacial movement indicators including local physiographic features, glacial lobes and sublobes, and cultural features.

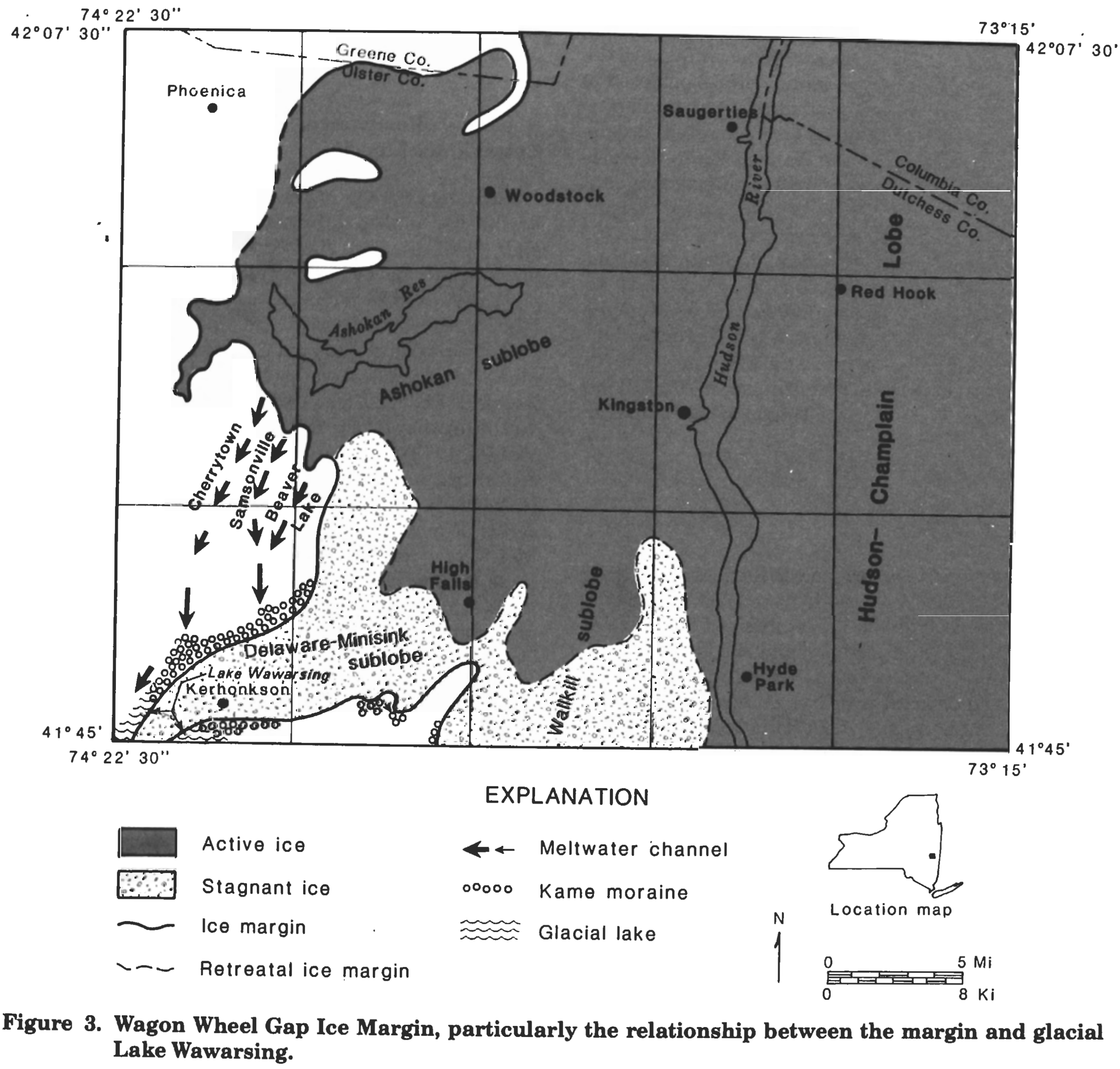


Figure 3. Wagon Wheel Gap Ice Margin, particularly the relationship between the margin and glacial Lake Wawarsing.

advance in the Hudson Valley. After the maximum advance of the glacier, the ice stagnated in the Minisink and Wallkill Valleys (Happ, 1938; Connally and Sirkin, 1970).

Glacial Lake Wawarsing was contained by the Phillipsport moraine at an elevation from 600 to 580 feet (180 to 175 m) above sea level (Heroy, 1974). The lake became choked with stagnating ice and a kame moraine (a ridge composed of stratified drift, deposited next to the glacier) was deposited against the Delaware-Minisink sublobe by meltwater from the active Ashokan sublobe (Fig. 3). Several proglacial channels were carved successively by sediment-laden meltwater streams flowing through Wagon Wheel Gap. As the Delaware-Minisink and

Ashokan sublobes downwasted and retreated eastward, lower lake spillways were exposed, and a 560 foot (170 m) lake was established in the Rondout and Ashokan Valleys. Subsequently, spillways were exposed at 440, 400, 300, and 240 feet (135, 120, 90, and 75 m) for lakes in the Minisink Valley. Flat-topped kame deltas composed of sand, silt, and clay were deposited adjacent to disintegrating blocks of ice as the lake levels lowered. Proglacial lakes also developed in the Wallkill Valley at elevations of 300 and 200 feet (90 and 60 m) during stagnant glacial retreat from the Wagon Wheel Gap ice margin. The wastage of the ice allowed a continuous 180 foot (55 m) lake to coalesce in the Minisink, Wallkill, and Hudson Valleys, while a 600 foot (180 m) proglacial lake developed west of Woodstock. The 240 foot (75 m) Minisink lake and 180 foot (55 m) Minisink-Wallkill-Hudson lake are suggested by subsurface well data (Frimpter, 1970). The glacier retreated from the Phillipsport-Wallkill moraines to Kingston during late Wagon Wheel Gap time (Figs. 3 and 4).

Rosendale Readvance: Whitfield Ice Margin

A moraine complex records the Whitfield stage of the Rosendale readvance ice margin from Whitfield to West Park; proglacial lakes were reestablished in the Minisink and Wallkill Valleys (Figs. 4 and 12). The Delaware-Minisink sublobe overrode and drumlinized lake clay and sand and deposited a thin (1 m) layer of sandy till across the area (J. Brown, USDA Soils Conservation Service, personal communication, 1982; R. Waines, SUNY at New Paltz, personal communication, 1983). The Delaware-Minisink sublobe also deposited a moraine and outwash fan at Whitfield, and held a 390 foot (120 m) lake in the Minisink Valley. At the same time, the Wallkill sublobe dammed 250 foot (75 m) Lake Tillson in the Wallkill Valley; the lake drained into the Hudson Valley through The Hell (Connally and Sirkin, 1970). The Hudson-Champlain Lobe built a moraine and outwash delta into Lake Albany at West Park. Dead ice persisted on the east side of the Wallkill Valley, northeast of New Paltz, depositing a large stagnation moraine north of New Paltz. To the north, the Ashokan sublobe built a kame moraine at Krumville and meltwater again flowed in the Beaver Lake channel (Fig. 4). Several proglacial lakes were dammed by the northeastern edge of the Ashokan sublobe (Fig. 4). Cirque glaciers advanced out of the upper Sawkill and Colgate Valleys.

The Rosendale readvance extruded glacial tongues through gaps in the Northeastern Escarpment of the Catskill Mountains (Fig. 4). Several of these sublobes fed meltwater southwest into Lake Grand Gorge in the upper Schoharie Valley (Cadwell, 1983; this volume).

Rosendale Readvance: Cottekill Ice Margin

Oscillatory retreat of the Rosendale ice subsequently established multiple recessional margins in the Minisink, Wallkill, and Hudson Valleys. The recessional margins are marked by moraine segments in the Hudson Valley from West Hurley to Hyde Park and by falling lake levels in the Minisink and Wallkill Valleys (Fig. 5).

The recession of Rosendale ice opened spillways that allowed the Rondout Valley lake to fall from 390 feet to 240 feet and 220 feet (120, 75, and 65 m), and the Wallkill Valley lake to fall to 220 feet (65 m) (Baker, 1969; Waines, 1979). The receding margin deposited a series of concentric moraine-outwash systems across the mouths of the Minisink and Wallkill Valleys (Fig. 5). A similar series of moraines was formed in the Hudson Valley. Meltwater from The Hell entered the 160 foot (50 m) Lake Albany, building an outwash delta along the edge of the Ulster Park moraine. These moraines probably are related to the Netherwood moraines described by Connally (1980). The Woodstock sublobe flowed west across Lake Woodstock and emplaced a kame moraine at 900 feet (275 m).

Further north, recessional moraines were deposited along the Northeastern Escarpment (Fig. 5). Only the High Knob Gap carried meltwater into the upper Schoharie Valley, and the ice stagnated at Broome Center. During North Blenheim-Haines Falls time, meltwater from the Hudson-Champlain Lobe ceased to flow into Lake Grand Gorge and began to flow southeast, along the face of the Northeastern Escarpment, and into the lower Hudson Valley (Cadwell, 1983, this volume). Lakes were dammed in the Schoharie Valley, near Woodstock, and in the Hudson Valley by receding ice.

Woodstock Ice Margin

The style of glacial retreat of the Hudson-Champlain Lobe changed as the glacier retreated from moraines that were built at Woodstock and West Hurley (Figs. 5 and 6). These moraines were deposited as the ice flow

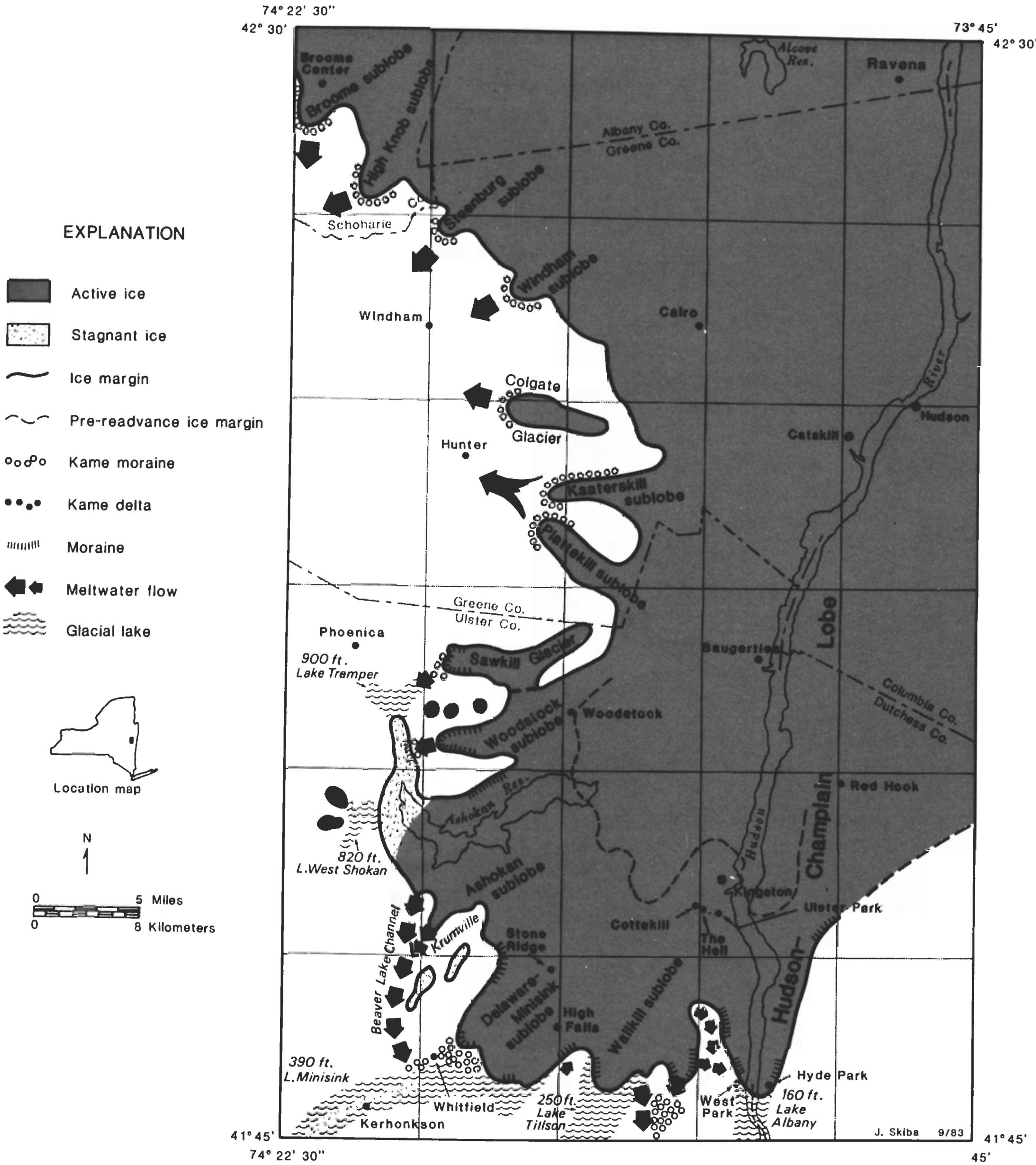


Figure 4. **Rosendale readvance: the Whitfield ice margin, which was the maximum extent of the readvance. Major proglacial lakes were dammed in the tributary valleys. The readvance flowed through gaps in the Northeastern Escarpment.**

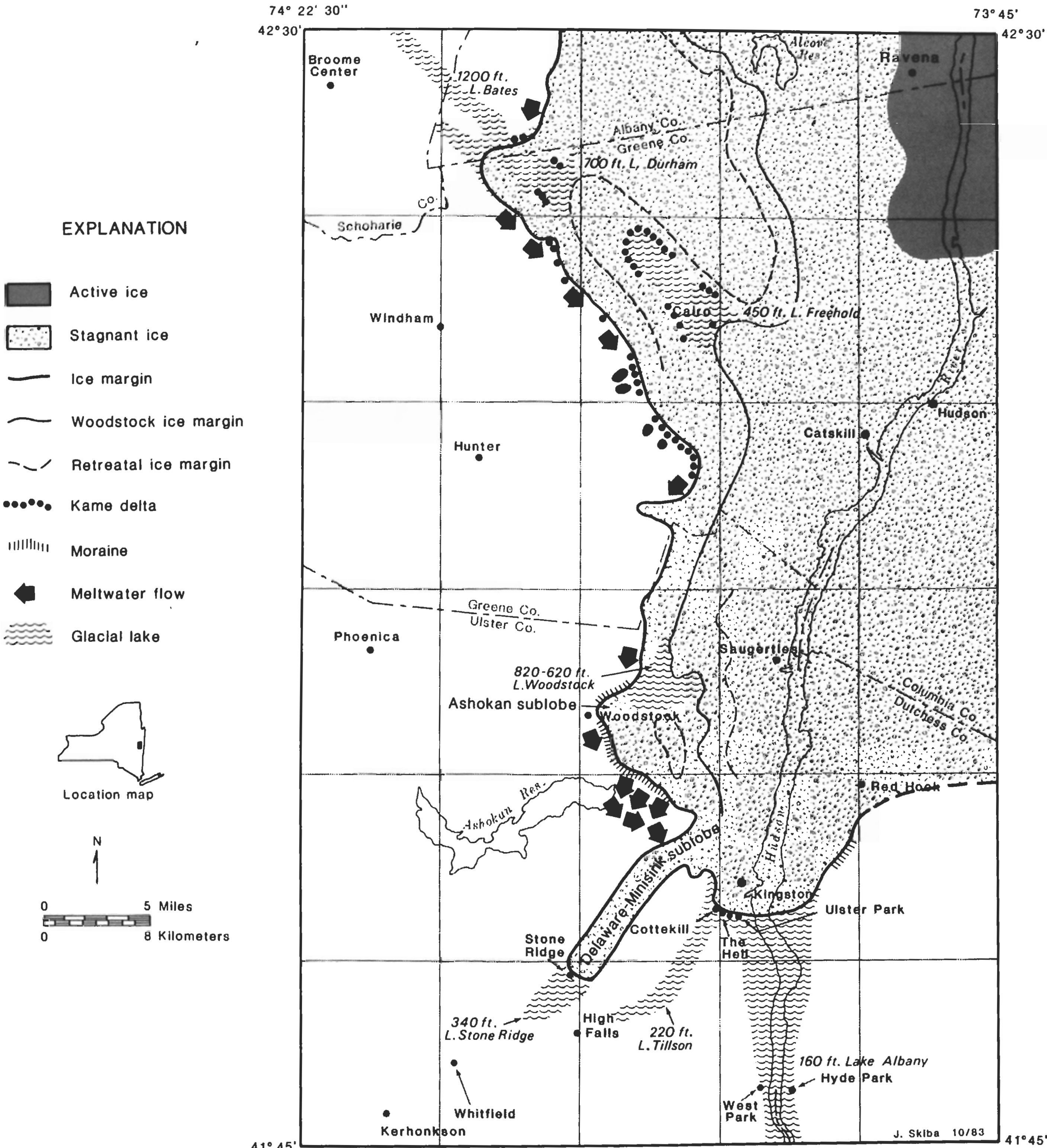


Figure 5. Rosendale readvance: the Cottekill and Woodstock ice margins were established as the glacier retreated from the Whitfield ice margin. Many ephemeral proglacial lakes developed along the receding glacier's edge.

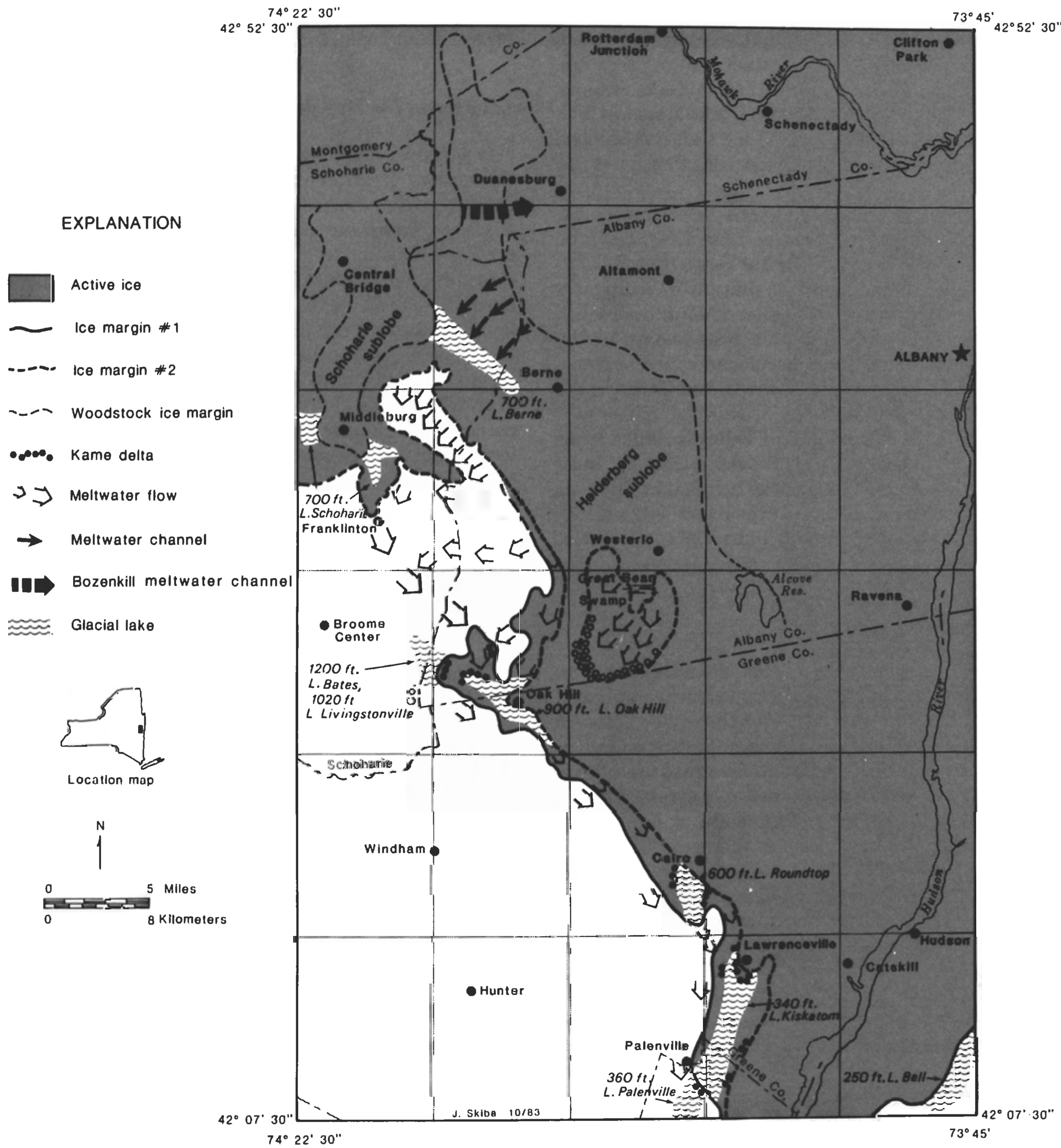


Figure 6. Middleburg readvance and Woodstock ice margin, the glacier's retreat from the Woodstock ice margin was interrupted by the Middleburg readvance. The advancing ice overrode many proglacial lakes and established a margin along the Northeastern Escarpment.

decreased. The glacier stagnated from Kingston to Catskill in the Hudson Valley, on the Helderberg Plateau, and in the Schoharie Valley.

As stagnant ice melted, lower spillways were exposed, and the water levels of the tributary valley lakes fell. These lakes drained into 160 foot (50 m) Lake Albany. The Wallkill Valley contained falling lake stages at 230, 210, and 160 feet (70, 65, and 50 m). Lakes developed next to the ice margin on the Helderberg Plateau (Figs. 5 and 6).

An ice margin developed along the Northeastern Escarpment that ranged in elevation from 1,500 feet (455 m) to the west to 900 feet (275 m) at Cairo (Fig. 6). Outwash was deposited against this margin by meltwater from cirque glaciers in the Blackhead Mountains south of Cairo. As the ice melted from west-to-east in the Catskill Valley, Lake Bates developed contemporaneously with the termination of Lake Grand Gorge in the Schoharie Valley (Cadwell, 1983, this volume). Later ice-marginal lakes in the Catskill Valley included lakes Livingstonville, Durham, and Freehold (Fig. 5). Lake levels fell to 700 feet (215 m) in the Schoharie Valley as the glacier retreated to the northeastern edge of the Helderberg Plateau and opened the Delanson outlet (Fig. 6).

Middleburg Readvance

The glacier readvanced at least 30 miles (48 km) across the Helderberg Plateau (Fig. 12) and established an ice margin from Franklinton, at the divide between the Schoharie and Catskill drainage basins, to Palenville in the Hudson Valley (Fig. 6). This readvance is recorded by a moraine along the Northeastern Escarpment and by till-over-lake sediments in exposures and wells on the Helderberg Plateau (Dineen, 1983). The Middleburg readvance crossed stagnant ice and proglacial lake deposits in the Schoharie, Fox, and Catskill Creek valleys. A stagnation moraine comprised of a network of crevasse fillings was deposited on the stagnant ice. The ice margin wrapped around the base of the Medusa and Helderberg Escarpments (Dineen, 1981), and forced meltwater into a series of ice-marginal lakes that drained south of Cairo (Fig. 6). The advancing ice also drumlinized recessional moraines in the Fox Creek and Switz Kill Valleys (Figs. 2 and 6).

The Hudson-Champlain Lobe readvanced south to Saugerties (Fig. 5), and blocked Glacial Lakes Palenville and Kiskatom on the Helderberg Plateau (Fig. 6). The lakes drained into Lake Albany. The advancing glacier drumlinized kames near Hudson, and subsequently deposited the Red Hook moraine (Connally, 1980). A series of ice-marginal lakes developed in the Catskill Valley as the ice began to retreat in an irregular manner.

Schoharie Ice Margin

The glacier stagnated as the ice thinned over the Helderberg Escarpment. Stagnation spread eastward from the Schoharie to the Helderberg and lower Hudson-Champlain lobes. Meltwater from dead ice in the Fox Creek and from 1,160 foot (355 m) Lake Schoharie poured into an 800 foot (245 m) ice-marginal lake at Oak Hill (Fig. 7). As the stagnant ice in the Catskill Valley melted, several proglacial lakes formed on the Helderberg Plateau and large areas of poorly stratified ice marginal deposits were deposited from Rensselaerville to Freehold (T. Trevail and J. Brown, USDA-SCS, personal communication, 1983). The overflow from the ice-marginal lakes scoured channels along the edge of the Helderberg sublobe and entered stagnant ice in the Hudson Valley. The stagnant ice was riddled with tunnels that drained into Lake Albany. Silty gravel ridges, covered with lake clay, were deposited in the tunnels.

Alcove Ice Margin

Dying ice covered the Helderberg Plateau and the Catskill and Schoharie Valleys, while an active ice margin was established along the Helderberg Escarpment and in the Hudson Valley at Ravena (Fig. 8). The Alcove ice margin was established as active ice deposited moraines near Duanesburg that cross-cut Schoharie moraines (Fig. 8). A large stagnation kame moraine was deposited over dead ice in the Grapeville-Earlton-Leeds area north of Cairo, as meltwater flowed south from active ice in the Alcove area (Fig. 8, Cook, 1942; Chadwick, 1944). As the ice melted out of the Catskill Creek Valley, lake levels fell to the level of Lake Albany.

The Fox Creek and Schoharie lakes lowered as the spillways at Delanson deepened (LaFleur, 1965) and a kame delta was deposited against ice at Rotterdam (Fig. 8). The receding glacier left a recessional moraine at McKownville as Lake Albany flooded the area from Ravena to Round Lake (Dineen and Rogers, 1979; Dineen and others, 1983). Many subaqueous gravel cones formed in deep lake water along the receding ice margin, while dead ice choked the Hudson Valley from Ravena to Kingston.

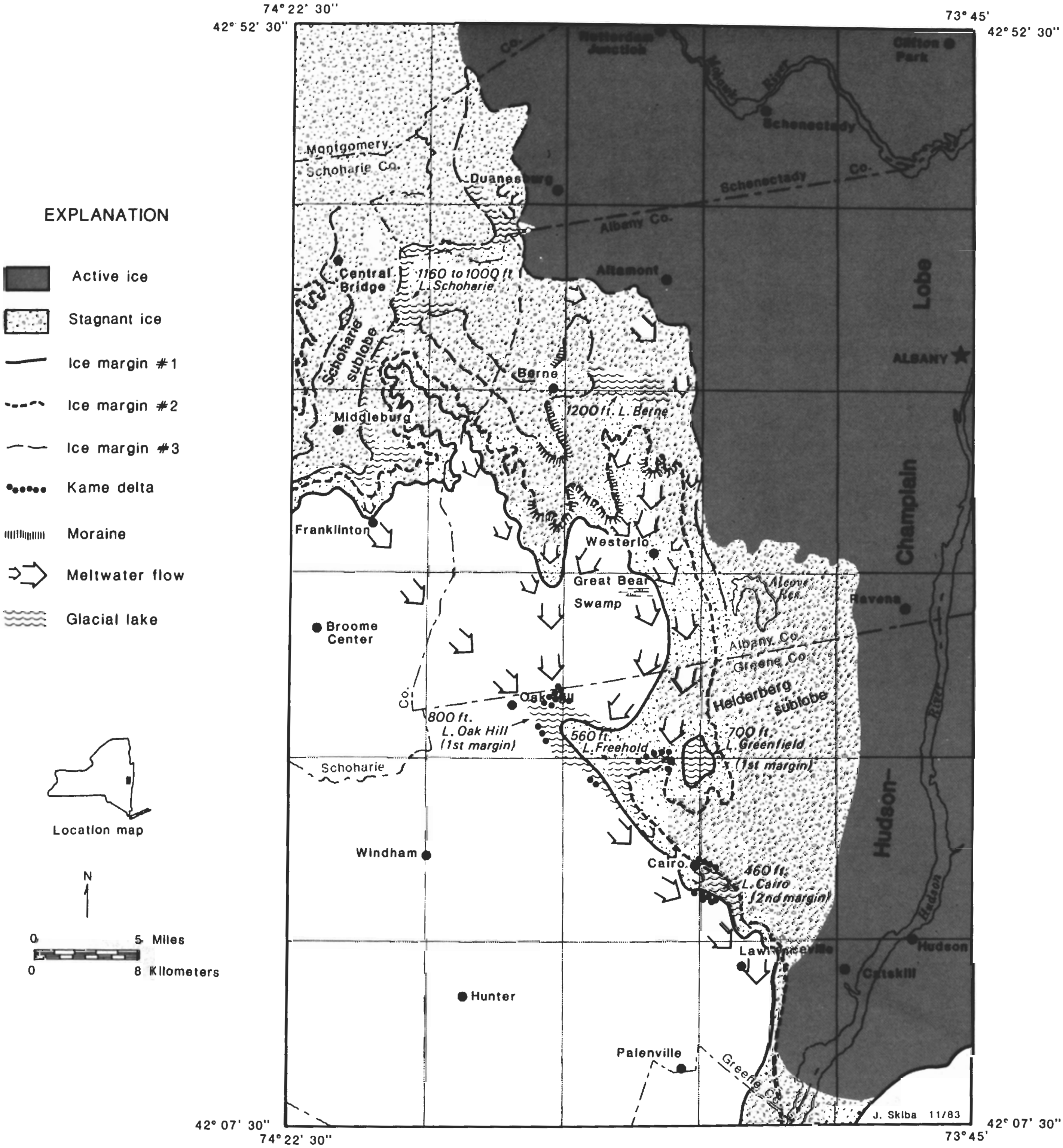


Figure 7. Schoharie ice margin includes several margins that formed as the ice retreated from the Middleburg readvance. These margins are associated with a complex array of proglacial lakes and meltwater streams, with deltas and outwash trains. Many of these lakes formed in areas covered by earlier lakes of the Woodstock ice margin.

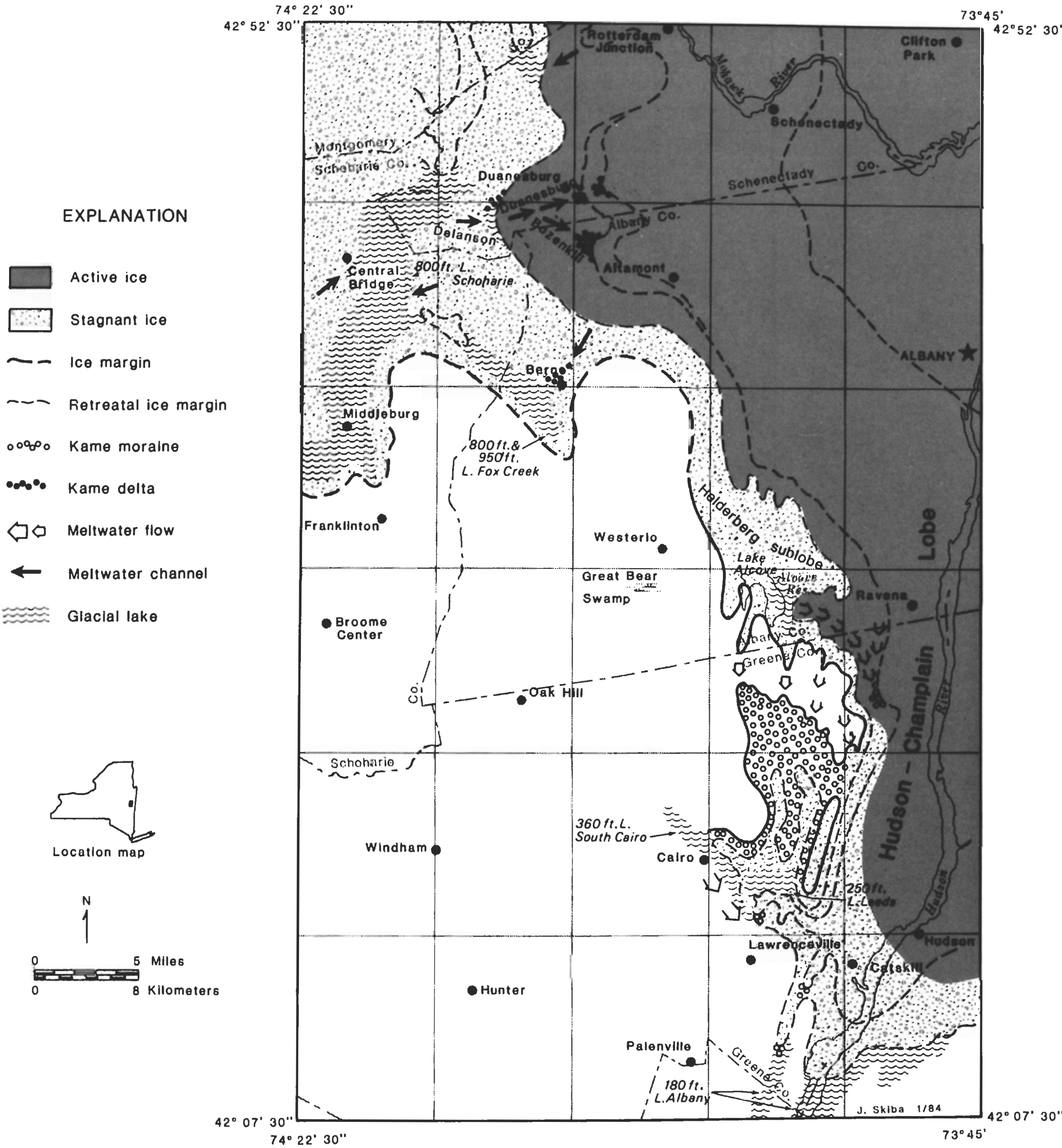


Figure 8. Alcove ice margin marks a reorganization of ice flow in the thinning glacier. The moraines of this margin cross-cut the Schoharie margin's moraines. Many detached ice blocks remained on the Helderberg Plateau.

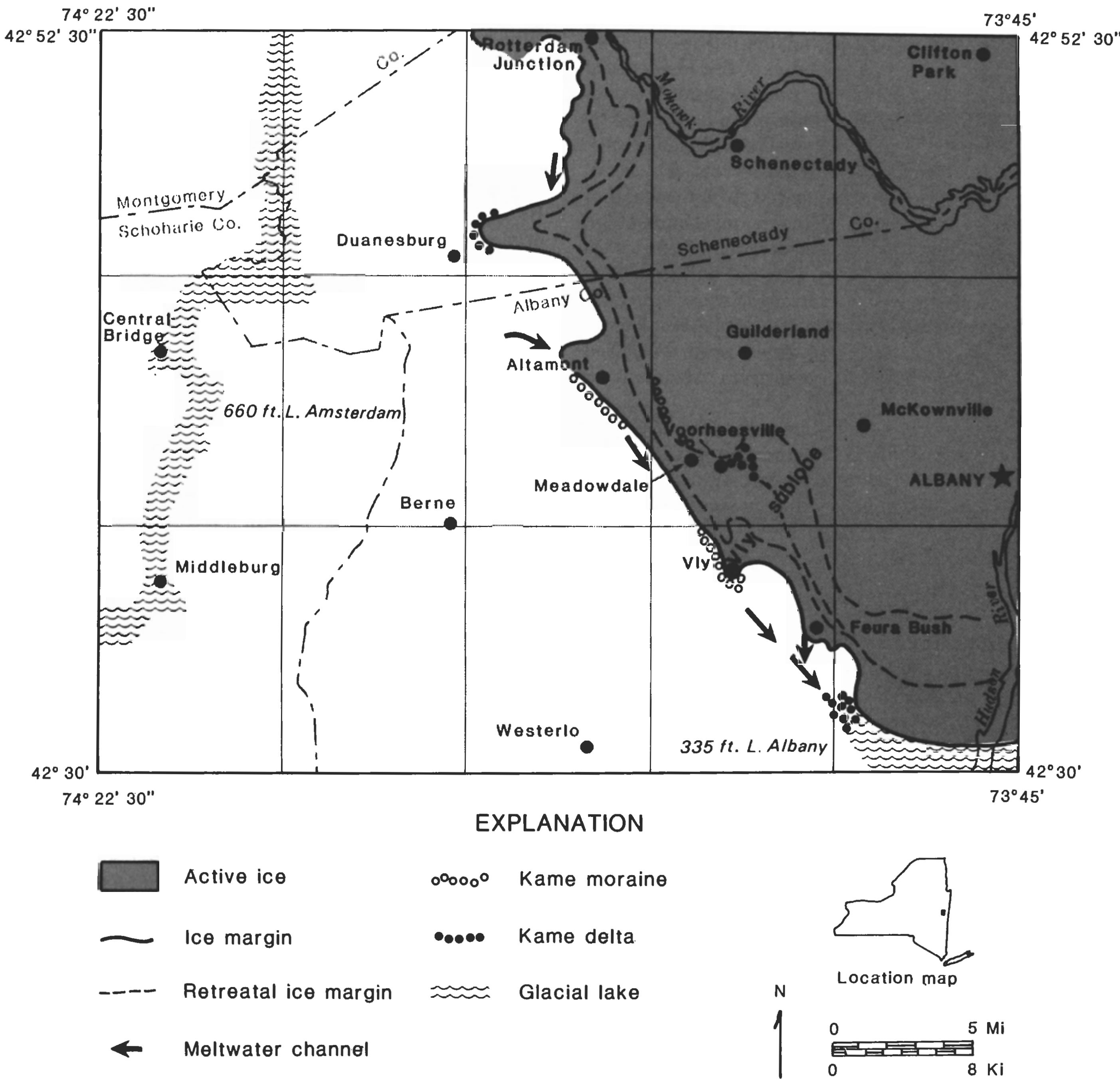


Figure 9. **Delmar readvance involved the Hudson Lobe, which readvanced over Lake Albany deposits. The margin lay along the base of the Helderberg Escarpment. Several recessional moraines were built as the glacier retreated from the Delmar margin.**

Delmar Readvance

The Delmar readvance of the Hudson Lobe surged south 20 miles (32 km) from Round Lake to South Bethlehem (Figs. 9 and 13; Dineen and others, 1983). The readvance molded kame-cored drumlins north of the study area at Schaghticoke and drumlins-over-lake clays at Easton. It reached the edge of the Helderberg Plateau. Like the Middleburg and Rosendale readvances, the Delmar readvance maximum is marked by indistinct moraines, whereas well-developed recessional moraines were formed during retreat. The readvance folded lake clays, and deposited thin, discontinuous till over lake sand and silt (Dineen and others, 1983; Dineen and Rogers, 1979). Outwash spread from an ice margin that extended from the Vly and Onesquethaw divide to Feura Bush, and boulder beds were deposited in the Lake Albany deltas at South Bethlehem (Fig. 9; Dineen and Rogers, 1979). Moraines were deposited along the Helderberg Escarpment from Altamont to Ravena (Fig. 9).

As the Delmar ice retreated, it deposited the Meadowdale moraine at Voorheesville and the Guilderland kame terrace. The ice front might have retreated rapidly because of extensive calving of the ice front into Lake Albany.

Schenectady Ice Margin

Lake Albany fell from 330 feet to 310 feet (100 to 95 m) at the same time as the glacier retreated to the Schenectady ice margin. The Schenectady ice margin was characterized by large masses of dead ice that remained in the Hudson Valley until the 250 foot (75 m) Lake Quaker Springs stage of Lake Albany (Fig. 10). Ice blocks also survived at Niskayuna (Hanson, 1977), Coeymans, and New Baltimore. A large ice block persisted at Glenmont until the Lake Fort Ann stage of Lake Albany (Dineen and Rogers, 1979). Many of the clay-covered kames from Castleton to Stuyvesant were deposited next to dead ice, and subsequently were buried by Lake Quaker Springs clay and silt.

AGES OF READVANCES

The ages of the various readvances are still uncertain. The deposits associated with these events show a paucity of *in situ* carbon-rich material. Most ages in the Hudson Valley are "bog bottom" dates. The bog bottoms are dated by extrapolating the accumulation rate for organic matter-poor sediment below the well-dated base-of-organic-sediment. The latter dates are suspect because they are based on bog sediment accumulation rates that are known to vary from 0.036 cm/yr (Sirkin, 1977, p. 210) to 0.08 cm/yr (Coates, 1976, p. 83). The bog sediment accumulation rates probably vary even more than these figures because of local variations in organic sediment supply, intensity of periglacial sedimentation, availability of fine sediment, aeolian sediment transport, etc.

For this study, the Great Bear Swamp and Meadowdale Bog were sampled. The basal, relatively inorganic silt and clay samples of these bogs were dated by Krueger Enterprises of Cambridge, Mass.; base-of-organic-sediment samples were dated by Krueger Enterprises and the Queen's College Radiocarbon Laboratories (Fig. 11).

Great Bear Swamp is a large kettle 5 miles west of the Alcove Reservoir in the Township of Westerlo, Albany County (Fig. 8). It lies at an elevation of 1210 feet (370 m) on the Helderberg Plateau. It is in stagnant ice deposits of the Schoharie ice margin. Donald Lewis (New York State Museum, Science Service – Biological Survey) examined the pollen from a 4.3 m long bog core. Its lithologic log and pollen stratigraphy are summarized on Fig. 11. The radiocarbon age of the base-of-organic sediment is 11,590 $\pm$ 980 yrs BP, while the radiocarbon age of the base-of-core is 19,875 $\pm$ 980 yrs BP. The extrapolated bog-bottom date is 15,060 yrs BP (Fig. 11). Connally and Sirkin (1973, fig. 6) previously estimated the date of deglaciation for this latitude 14,000 yr. B.P.

The Meadowdale moraine has been dated using sediment from the Meadowdale (D'Hommeadeau) bog, a small, closed kettle bog in Voorheesville Township, Albany County (Fig. 9). The Meadowdale bog is on the north (ice-proximal) slope of the Meadowdale moraine, at an elevation of 350 feet (105 m). The moraine was deposited during the recession of the Delmar readvance. The base-of-organic-sediment sample yielded an age of 10,490 $\pm$ 320 yrs BP, a base of core age of 16,650 $\pm$ 660 yrs BP, and an extrapolated bog bottom date of 14,930 yrs BP (Fig. 11). Connally and Sirkin (1973, fig. 6) estimated an age of 13,800 yrs BP for deglaciation of the Albany area.

Based on my dates, and the dates of Connally and Sirkin (1973, this volume), I conclude: 1) that the glacier retreated from the Helderberg Plateau between 16,000 and 14,000 years ago; 2) that the Middleburg readvance occurred about 15,500 years ago; 3) that the Delmar readvance might have occurred approximately 15,000 years ago; and 4) Lake Albany is at least 14,000 years old in the Albany area.

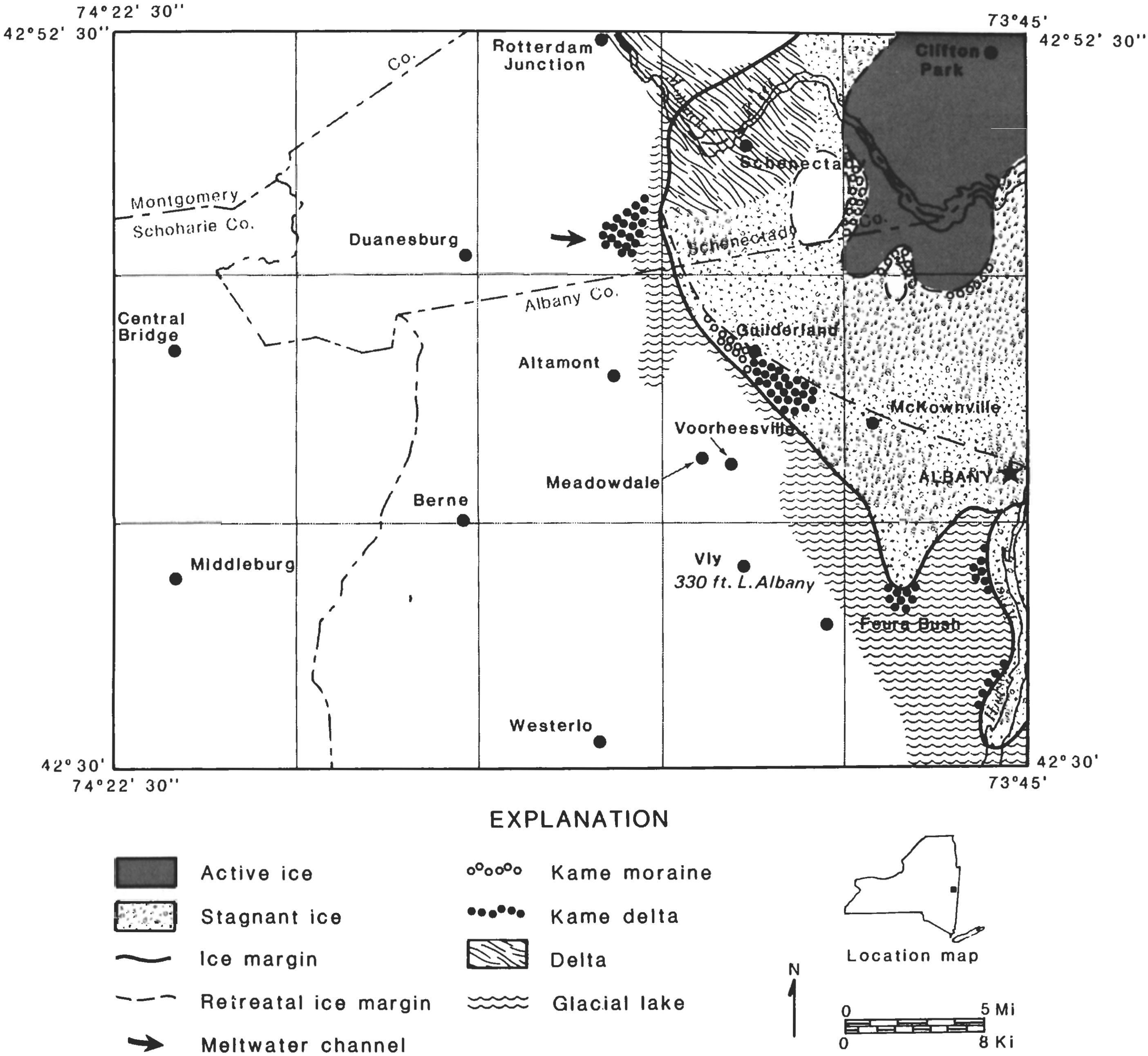
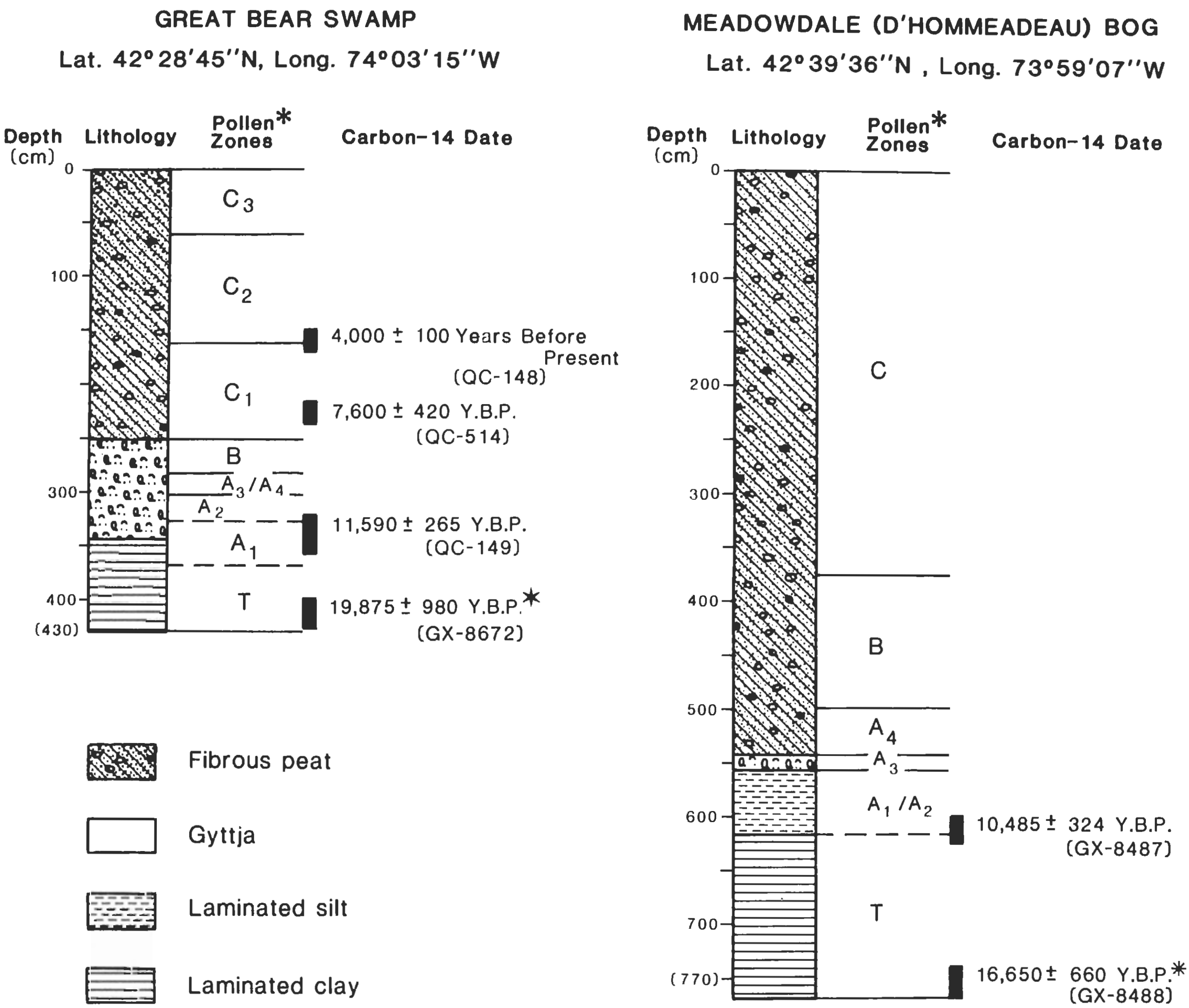


Figure 10. Schenectady ice margin began as the Schenectady delta of the Mohawk River built across stagnant ice. A recessional moraine was deposited near Guilderland.



*From Donald M. Lewis, Biological Survey, N.Y.S. Museum & Science Service

*Extrapolated age is 15,060 Y.B.P. at 0.039 cm/yr , or 13,152 Y.B.P. at 0.08 cm/yr

*Extrapolated age is 14,930 Y.B.P. at 0.039 cm/yr , or 12,485 Y.B.P. at 0.08 cm/yr

Figure 11. Pollen core logs and ages of pollen zones and glacial recessions. See Figures 8 and 9 for the pollen core locations.

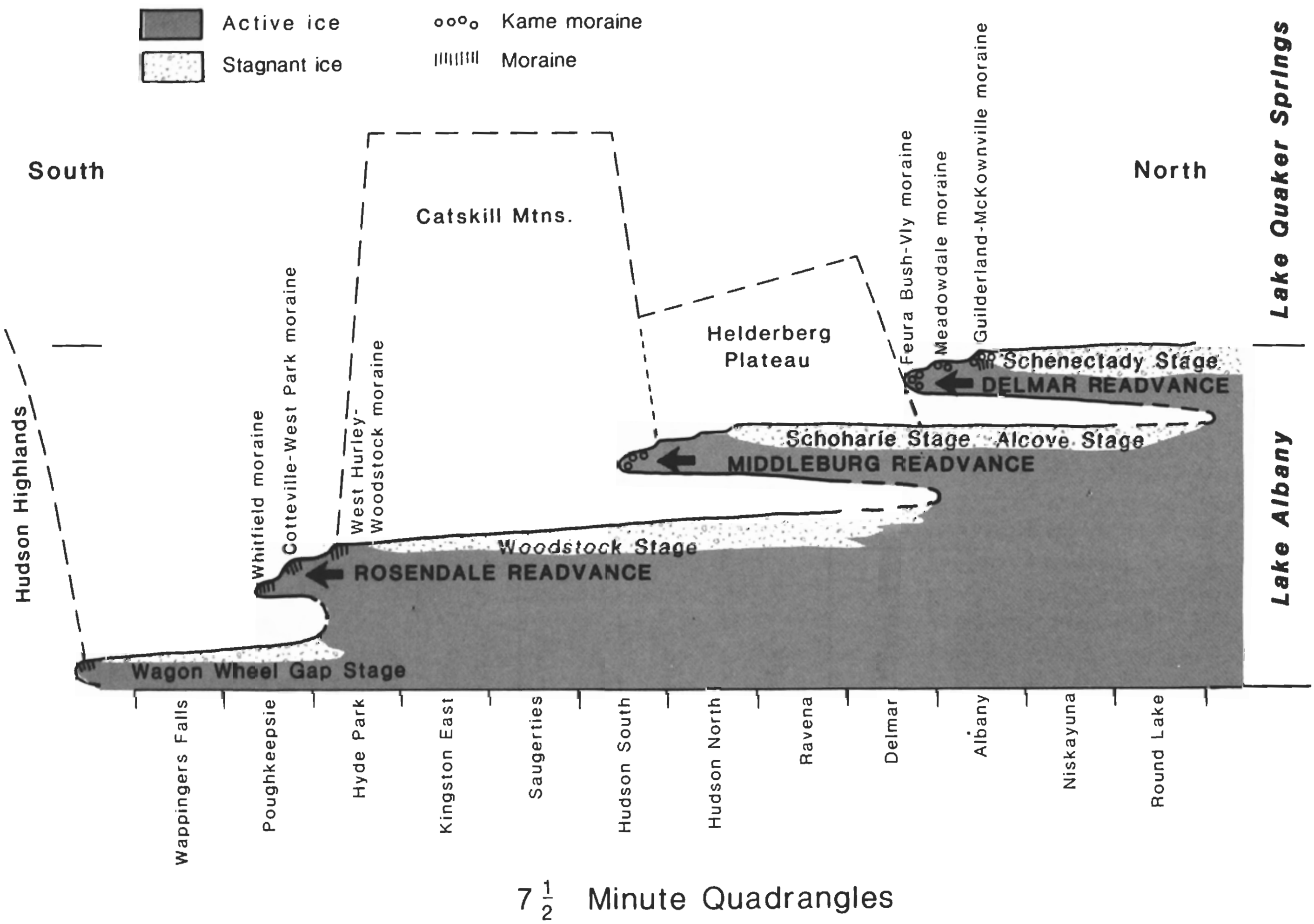


Figure 12. Magnitude of glacial retreats and readvances, stages have no formal stratigraphic meaning, they refer to phases of glacial retreat.

SUMMARY AND CONCLUSIONS

Three readvances occurred as the Hudson-Champlain Lobe retreated from New Paltz to Schenectady. Nine ice margins can be recognized in this area, including the readvances of 10 to 40 miles (16 to 64 km, Fig. 12, Table 1). Each readvance is associated with rising lake levels in the Hudson Valley tributaries and characterized by pulsations during their initial retreat. The six recessional ice margins are characterized by falling lake levels and broad areas of stagnant ice.

Bedrock topography influenced both the advance and retreat of the Hudson-Champlain Lobe. During readvances, glacial motion was impeded by areas of rugged relief and uplands; consequently, ice was diverted away from uplands into lowlands. Thus, glacial lobes tended to advance farther in the valleys than in the uplands.

During a recession, south-dipping plateaus, areas of rugged relief, and narrow transverse valleys caused the glacier to stagnate, but smooth areas of low relief and deep, wide valleys encouraged active retreat. Ice stagnated in portions of valleys that were distal to the main

TABLE 1. Correlation Chart

ICE MARGINS	HUDSON VALLEY	WALLKILL AND MINISINK VALLEYS	SCHOHARIE VALLEY CATSKILL MOUNTAINS HELDERBERG PLATEAU
Schenectady	Lake Quaker Springs 310 ft. Lake Albany	150 ft. Lake Albany	
Delmar readvance	330 ft. Lake Albany	180 ft. Lake Albany	660 ft. Lakes Amsterdam & Schoharie
Alcove	330 ft. Lake Albany	180 ft. Lake Albany	Delanson Outlet: 800 ft. Lake Schoharie
Schoharie	180 ft. Lake Albany	180 ft. Lake Albany	1160 ft. Lake Schoharie, Schoharie or Yosts Margin (La Fleur, 1969a, 1979)
Middleburg readvance	180 ft. Lake Albany Lake Kiskatom Red Hook Moraine (Connally, 1980)	180 ft. Lake Albany	Middleburg readvance: Franklinton Notch (La Fleur, 1969a)
Woodstock	Woodstock Moraine 160 ft. Lake Albany	230 ft. Lake Tillson	1200 ft. Lake Schoharie (La Fleur, 1969a)
Rosendale readvance: Cottekill	Ulster Park Moraine 160 ft. Lake Albany Netherwood Moraines (Connally, 1980)	Cottekill Moraine: 340 ft. Lake Stone Ridge 220 ft. Lake Tillson	North Blenheim-Haines Falls Margin (Cadwell, 1983) Drainage into Hudson Valley
Rosendale readvance: Whitfield	West Park Moraine 160 ft. Lake Albany Poughkeepsie Moraines (Connally, 1980)	Whitfield Moraine: 390 ft. Lake Pataukunk 250 ft. Lake Tillson	Lake Grand Gorge (Cadwell, 1983) Drainage into Schoharie Basin
Wagon Wheel Gap	Shenendoah Moraine (Connally, 1980)	Phillipsport Moraine- Lake Wawarsing (Happ, 1938 & Heroy, 1974) Wallkill Moraine (Connally & Sirkin,1970)	Wagon Wheel Gap Margin (Rich, 1935)

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ice lobes, such as the upper Schoharie, Minisink and Wallkill Valleys, and in tributary valleys at right angles to the main glacial sublobes, such as the Schoharie, Fox, Catskill, and upper Esopus Valleys.

The width of the Hudson-Champlain Lowland also influenced glacial lobe behavior. The Hudson Lowland between Ravena and Kingston is narrow and bounded by adjacent highlands, and contains rough topography, locally with 150 to 200 feet (45 to 60 m) of relief. This area retained large quantities of dead ice until the end of Lake Albany time. Wide lowlands, such as between Ravena and Glens Falls, contained only a few large blocks of stagnant ice left behind by an actively retreating ice margin. Readvances into lowlands that narrowed to the south tended to extend farther than readvances into southward-widening lowlands. The more significant readvances, such as the Middleburg (30 ± miles, 50 ± km) and Delmar (20 ± miles, 32 ± km), spread into southward-narrowing, lake-filled lowlands, where the Catskill Mountains and Helderberg Plateau caused convergent flow in the ice (Figs. 1, 2, and 13). During the Rosendale readvance, ice flow diverged around the southeastern edge of the Catskill Mountains into the widening Hudson Lowlands, and only advanced 5 ± mi. (8 ± km) (Figs. 1, 2, and 12).

Readvancing glacial lobes deposited only discontinuous and inconspicuous moraines at the line of maximum advance. A series of recessional moraines were deposited during the oscillatory retreat. The maximum readvance ice margin was 2 to 5 miles (3 to 8 km) beyond the first strong recessional moraine. Thus, both stagnant and active glacial retreat occurred in the mid-Hudson Valley, suggesting that the debate about stagnant vs. active glacial ice was due to an over-extension of models that worked well on a local scale.

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WOODFORDIAN STRATIGRAPHY IN THE WESTERN CATSKILL MOUNTAINS

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ABSTRACT

Deglaciation in the western Catskill Mountains during Woodfordian time involved both active ice retreat and the stagnation of detached ice masses. The extent of ice detachment was related to the degree of topographic isolation, which generally increases to the south in this region. Thus, different valley systems were characterized by contrasting styles of deglaciation.

Ice remained active longest in the West Branch Delaware River valley where topography was the most conducive to a replenishment of ice flow. Glaciofluvial deposits in this valley can be grouped into six zones of ice-contact morphosequences, demonstrating that deglaciation took place by a process of stagnation-zone retreat. High local relief and valley sinuosity induced larger stagnation zones in the lower East Branch drainage. Numerous kame deltas along valley walls record the development of three successive glacial lake levels in the East Branch during deglaciation. Farther to the south, topographic conditions favored the stagnation of a large ice mass in the Beaver Kill and Willowemoc Creek valleys. These valley systems contain an abundance of drift left by the ablation of stagnant ice masses (i.e. resedimented supraglacial and englacial debris).

Former active ice positions are identified along upland drainage divides by the presence of ice marginal erosional landforms (cols, notches and spillways). Regional correlations are established by stratigraphic relationships between meltwater deposits at valley confluences and by the application of an ice profile gradient model to the topographic setting. These correlations allow deglaciation events in the Susquehanna River valley to be linked with ice retreat in the upper Delaware drainage system.

INTRODUCTION

Purpose

Glacial geology in the Catskill Mountains has been studied previously, but the complexities of Woodfordian stratigraphy in this region have not been fully explained. Additionally, the correlation of deglaciation events between the Hudson Valley and western New York State has been prevented by the difficulties involved in tracing former ice marginal positions through the Catskill region. This paper presents data from a recent study that addresses these problems in the western Catskill Mountains.

Study Area

The study area encompasses nearly 2,000 square km in the western Catskill Mountains (Fig. 1) which represent a transition from the highest Catskill peaks to the east and the Susquehanna geomorphic section to the west. This region is underlain by Devonian sandstones, siltstones and shales that dip gently to the southwest (Fletcher, 1962). The low structural dip and differences in erosional resistance between sandstones and shales have contributed to the formation of a series of cuestas with north to northeast-facing scarps (Thornbury, 1965). Rich (1935) identified three major cuesta forms in the Catskill region: the Northeastern, Central and Southern Escarpments. The western Catskills lie to the southwest of the Central Escarpment, which is the drainage divide for the upper Delaware River system (Fig. 1).

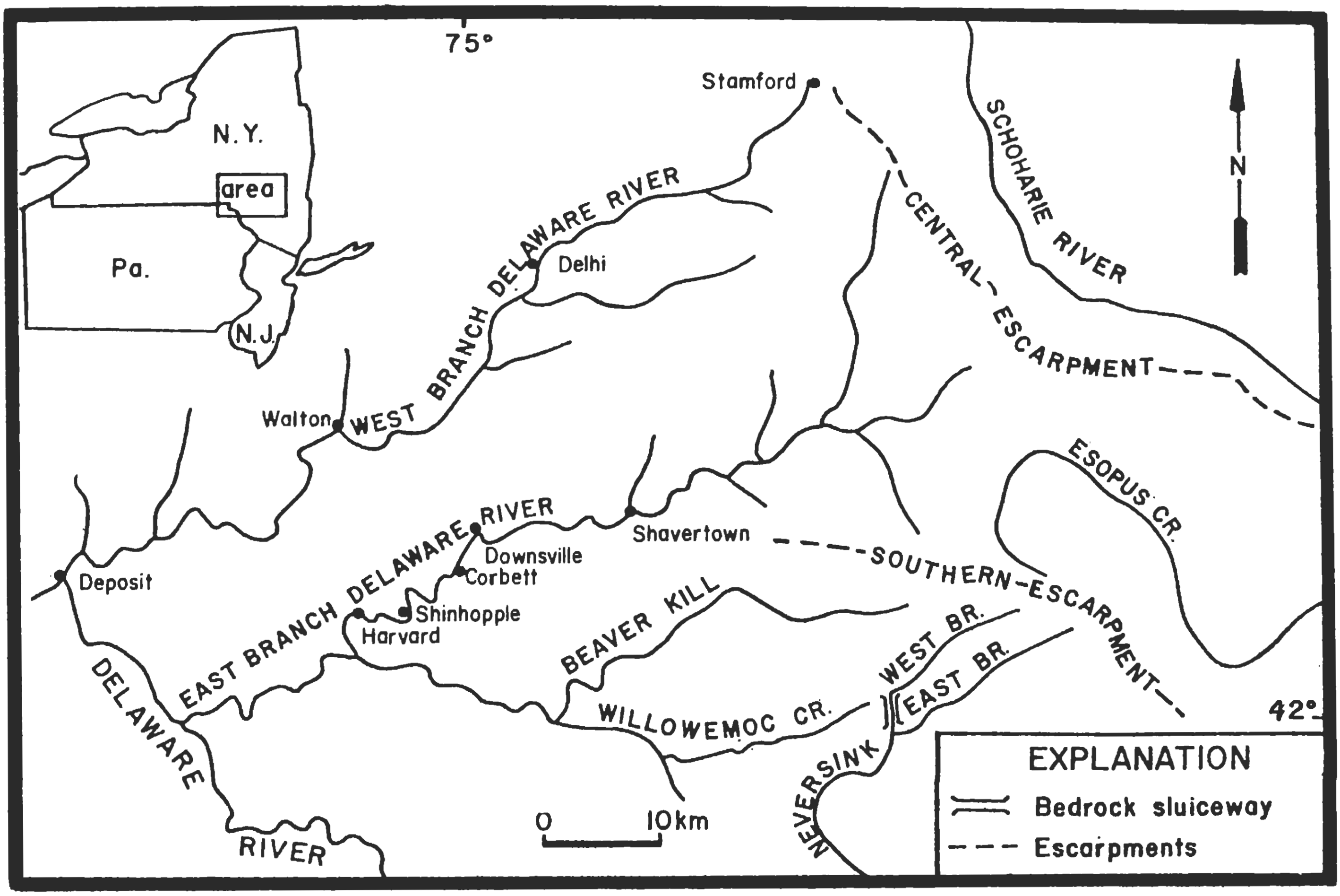


Figure 1. Map of the western Catskill region showing locations and the relationships between escarpments and drainage systems (from Kirkland, 1979). Study area encompasses western half of map.

The western Catskill Mountains form a rugged terrain, with steep slopes and high relief. Summit elevations reach a maximum of 1,110 m, and local relief ranges from 250 m to over 500 m. Relief would be even greater were it not for thick glacial deposits in the valley bottoms.

This landscape is dominated by the entrenched East and West Branch Delaware River valleys which are up to 1 km wide as a result of glacial modifications. Upland tributary valleys are much narrower, and generally have interlocking spurs. Original drainage directions are thought to have been structurally controlled (Soren, 1963) and then modified by stream capture during Tertiary time (see Coates, 1963) for a review of drainage evolution theories).

Pleistocene deposits appear to be entirely Late Wisconsinan (Woodfordian) in age (Kirkland, 1979), although this region was undoubtedly covered by earlier glaciations (Coates, 1976). Ice advanced from the northeast during Woodfordian time and its flow was strongly influenced by the rugged topography (Rich, 1935; Kirkland, 1973; 1979).

Previous Work

Rich (1935) distinguished two drift sheets in the Catskill region on the basis of landform appearance, drift color, and the weathering of clasts. Thus he concluded that there are two ages of glacial deposits preserved here: "Early and Late Wisconsin." Rich also reconstructed five stages of "Late Wisconsin" deglaciation, showing that ice retreated to the northeast. The first stage marks what Rich thought was the position of farthest advance by "Late Wisconsin" ice into the western and central Catskills. The four subsequent stages represent ice marginal positions during withdrawal from the Central Escarpment.

Coates (1963) and Kirkland (1973; 1979) rejected Rich's (1935) hypothesis of two drift sheets in the Catskill region. Instead, they attributed apparent differences in weathering and color to lithological variations within a single-aged drift sheet. Kirkland considered all glacial deposits in the Catskills to be Late Wisconsinan in age.

Kirkland's (1973; 1979) study of stratified valley deposits in the western Catskills allowed him to document "three widely different" styles of deglaciation in the upper Delaware drainage basin. Glaciofluvial deposits in the West Branch valley can be grouped into zones of fluvial ice-contact morphosequences, suggesting that ice retreated by a series of ice tongue detachments. Kame deltas in the East Branch indicate that three successive glacial lakes developed in this valley during ice retreat. In the Beaver Kill and Willowemoc Creek valleys to the south, deglaciation was accomplished by mass stagnation.

Krall (1977) mapped the Cassville-Cooperstown Moraine which occurs north of the Catskill region, and attempted to correlate this ice position with deglaciation events in the Catskill Mountains and the Hudson Valley. By linking the Cassville-Cooperstown Moraine with the Wagon Wheel Gap Stage of Rich (1935) and the Rosendale Readvance of Connally and Sirkin (1970) farther to the southeast, Krall was able to trace a line of ice retreat which he believed had parallel analogs through the western and central Catskills.

Connally (1979) took strong exception to many of Krall's (1977) correlations and refuted the idea that discernable ice margins can be identified in the western Catskills. Connally cited the studies of Kirkland (1973), who was unable to trace active ice positions across the rugged western Catskills, and Coates and Kirkland (1974), who suggested that areas to the south of the Catskills may have experienced regional stagnation. The problems of correlation in the western Catskills also are mentioned by Kirkland (1979).

Two factors make the correlation and age assignment of glacial sediments in the western Catskills a largely conjectural task. First is the absence of informative stratigraphic sections. Secondly, no investigators have provided radiocarbon dates for Catskill localities. Thus, age assignments have been based on the presence of Woodfordian drift to the south in Pennsylvania (Crowl, 1980; Crowl and Sevon, 1980) and New Jersey (Connally and Sirkin, 1970; 1973; Cotter and others, this volume). No data have been discovered in this study to lessen the speculative nature of correlation and age assignment.

GLACIAL SEDIMENTS IN THE STUDY AREA

The model of deglaciation developed for the western Catskills is closely related to the distribution of glacial sediments in this area. The purpose here is to review the descriptions and interpretations of mapping units employed in the western Catskills. These sediments are described more fully in Ozsvath (1985).

Uplands

Level areas in the uplands are covered with a relatively thin (0.25 to 2.0 m thick) diamicton unit that rests directly on bedrock. Well logs show that thicknesses may exceed 20 m on lee side slopes (Soren, 1963). Its distribution generally is continuous, with the exception of upland passes where bedrock is exposed. On most valley walls this material has undergone slope colluviation.

Exposures of this upland mapping unit reveal a massive, matrix-supported diamicton that is dense and compacted. The matrix material ranges from sandy silt to silty loam with minor amounts of clay. Randomly dispersed clasts commonly are pebble to cobble-sized and boulders are rare. The composition of clasts which are predominantly angular to subrounded, always reflects the immediately local bedrock. In some locations, clasts have a weak fabric oriented roughly parallel to the land surface.

Five properties of this mapping unit suggest that it was transported and deposited subglacially: (1) it is composed of sediments that are of very local origin; (2) it rests directly on bedrock in upland areas throughout the region; (3) the material is dense and compacted; (4) its textural characteristics are relatively uniform; and (5) there is an absence of either associated or interbedded stratified sediments (Marcussen, 1975; Kruger, 1979; Haldorsen, 1982). The abundance of angular, locally derived clasts suggests short transport distances prior to deposition. This mapping unit has been interpreted as lodgment till, derived from the thin basal debris layer of an ice sheet (Boulton, 1972). Such an interpretation is consistent with what other workers have mapped on upland areas in both the eastern Catskill region (Cadwell, this volume) and adjacent areas of New York State (Coates, 1976; Caprio, 1980; Gillespie, 1980; Gubitosa, 1984; Fleisher, this volume).

Major Valleys

Major valleys served as the main passageways for meltwater flow during deglaciation as demonstrated by the abundance of glaciofluvial deposits in these areas.

The greatest amount of stratified drift is found in the East and West Branch valleys, where depths to bedrock can exceed 40 m (Soren, 1963).

Five mapping units were used to classify stratified drift, the majority of which are ice-contact deposits: (1) undifferentiated kame (including lone kames, kame fields and poorly exposed ice-contact deposits); (2) kame terrace; (3) kame delta; (4) kame moraine; and (5) outwash/alluvium.

Ice-contact landforms commonly are found along valley walls with the exception of kame moraines, which may occur as cross-valley forms. Exposures into these deposits reveal features characteristic of ice-contact deposition including: soft sediment deformation structures (faults, oversteepened, overturned and folded beds, convolutions, diapirs, etc.), diamicton inclusions, overall poor sorting and stratification, the presence of angular clasts, and abrupt changes in grain size and sorting. The term "kame terrace" is used in a genetic sense, meaning that such landforms are composed of sediments deposited by braided meltwater streams flowing between stagnant ice and the valley wall. Facies of braided stream deposition were identified according to descriptions given by authors such as Boothroyd and Ashley (1975), Church and Gilbert (1975), and Miall (1977). Kame deltas are characterized by the presence of sedimentation units typical of the deltaic environment: horizontally to cross-stratified sand and gravel topsets, steeply dipping parallel to cross-stratified sand, silt and gravel foresets, and horizontally stratified fine-grained bottomset beds (Gustavson et al., 1975; Thomas, 1984). Kame moraines are essentially "heads of outwash" (Koteff, 1974), formed in front of stagnant valley ice margins (Kirkland, 1973). They contain sediment facies indicative of rapid deposition in the proximal proglacial environment: concave upwards sequences of imbricate, crudely bedded, very coarse-grained sand and gravel dipping downstream (Boothroyd and Ashley, 1975; Church and Gilbert, 1975).

In a topographic setting such as the western Catskills, valley bottoms usually are covered with proglacial outwash sediments forming a flat-surfaced "valley train" (Rich, 1935; Flint, 1971). Valley bottom deposits are poorly exposed in the study area; however well logs (Soren, 1963) confirm that subsurface sediments along the bottoms of major valleys are predominantly sand, silt and gravel. These deposits were mapped as "outwash/alluvium" because of the difficulty in distinguishing between glaciofluvial and Holocene river sediments (Rich, 1935; Kirkland, 1973). Wells penetrate thick clay beds in some locations (particularly in the East Branch valley), suggesting the presence of local lakes in those areas during deglaciation. However, lacustrine sediments are not exposed at the surface in sufficient quantity to warrant a separate mapping unit.

As was suggested by Kirkland (1973; 1979), suites of glaciofluvial deposits (particularly in the West Branch valley) can often be grouped into ice-contact morphosequences (Koteff, 1974) that mark the former positions of stagnant valley ice tongues. The distribution of these sequences indicates that their positions in the valley frequently are controlled by the junctions of tributary streams that contributed additional sediments to the glaciofluvial system (Kirkland, 1979).

Tributary Valleys

Sediments in tributary valleys generally are less stratified and sorted as the valley size decreases. Most of the smaller tributary valleys contain little stratified sediments and what does exist occurs as very poorly sorted ice-contact drift. The predominant sediment type in these tributary valleys is a diamicton unit.

Valley diamicton sediments form a gently undulating to hummocky deposit along the lower walls and valley bottom. In contrast to the upland diamicton (lodgment till), this mapping unit is characterized by lateral and vertical variability in both texture and composition. The matrix material generally is coarser-grained than that of the upland diamicton, ranging from loamy silt to silty sand. Clasts, which may be up to boulder-sized, vary from angular to well-rounded. Deposits are predominantly massive, but inclusions of sorted and poorly stratified sediments are common. These inclusions range from isolated stringers or lenses of sand and silt to graded beds of fine gravel and sand. In some examples, diamicton sediments grade almost imperceptibly into beds of very poorly stratified sand and gravel. The composition of the valley diamicton does not necessarily reflect the local bedrock, but may contain exotic lithologies derived from bedrock to the north and east (Ozsvath, 1985).

Four properties of the valley diamicton indicate that it was derived from the resedimentation of englacial and supraglacial debris during ice ablation: (1) there is a broad spectrum of sediment size, sorting and stratification; (2) inclusions of stratified sediments are common; (3) some clast lithologies indicate long transport distances; and (4) its hummocky surface morphology is not typical of subglacial lodgment or melt-out till (Marcussen, 1975; Eyles, 1979; Lawson, 1979; Haldorsen, 1982).

Within the valley diamicton mapping unit are discrete landforms ranging from small hummocks to finger-like, crossvalley forms. These types of landforms

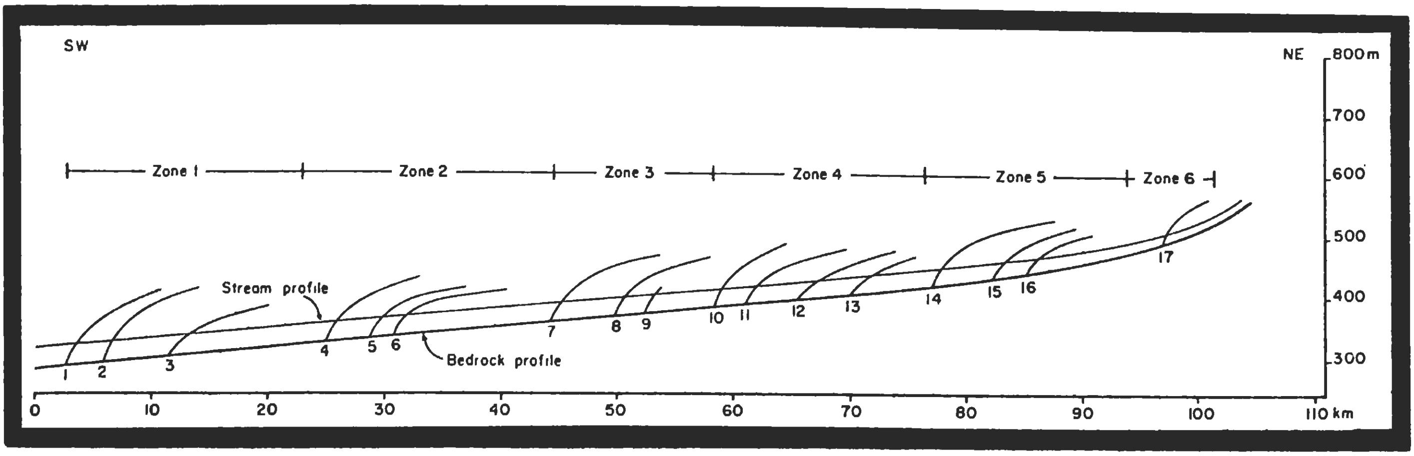


Figure 2. Longitudinal profile of the West Branch Delaware River valley showing stagnation zones and valley ice profile gradients reconstructed through individual morphosequences (from Kirkland, 1979).

had previously been attributed to active ice deposition on the basis of their morphology (Rich, 1935; Kirkland, 1973). However, the sedimentology and topographic setting of these features indicates that they were formed in a dead ice environment (Ozsvath, 1985). This is consistent with the conclusion that valley diamicton sediments were deposited in association with the ablation of stagnant ice masses.

Summary of Glacial Sediments

The distribution of glacial deposits in the western Catskills follows a regional pattern. Stratified glaciofluvial sediments are found in the major valleys where meltwaters were concentrated during deglaciation. The majority of these are ice-contact deposits formed adjacent to stagnant valley ice tongues. The degree of stratification and sorting decreases with decreasing valley size and most of the smaller tributary valleys contain no stratified drift. Instead, these smaller valleys are blanketed with diamicton sediments that were deposited as stagnant ice masses withered away. The uplands are covered with a relatively thin mantle of lodgment till derived from the basal debris layer of the Laurentide ice sheet.

DEGLACIATION MODEL

Figures 2 and 3 illustrate the styles of deglaciation in the East and West Branch valleys as presented by Kirkland (1979). A series of ice tongue detachments created six zones of stagnant ice in the West Branch valley (Fig. 2). These zones, ranging from 16 to 24 km in length, contain three or four fluvial ice-contact morphosequences. A complete sequence grades upstream from outwash to pitted outwash to kame moraine and kame terrace. Deglaciation of the East Branch valley created a succession of three glacial lakes (Fig. 3). These lakes are substantiated by the presence of numerous kame deltas preserved along the East Branch valley walls.

This study has confirmed most of Kirkland's interpretations and accepts his deglaciation models for the East and West Branch valleys as a starting point for regional correlations. The evidence for six ice tongue detachments in the West Branch and at least three in the East Branch eliminates the possibility of regional stagnation in the upper Delaware drainage basin. Although no recessional moraines can be identified in this rugged terrain, the absence of such features does not necessarily mean that active ice margins were nonexistent during deglaciation. In fact it is unlikely that ice borders traversing maturely dissected uplands would leave morainal ridges (Johnson, 1941). There are few level areas in the uplands upon which these features could be preserved, and valley segments would be removed by glaciofluvial streams. Thus, other types of evidence must be used to reconstruct former ice positions.

IDENTIFICATION OF ICE POSITIONS

Meltwater Pathways

As discussed by Coates and Kirkland (1974) and Cadwell (1978), former ice marginal positions in the Appalachian Plateau have left erosional imprints in the landscape. In the western Catskills these imprints are

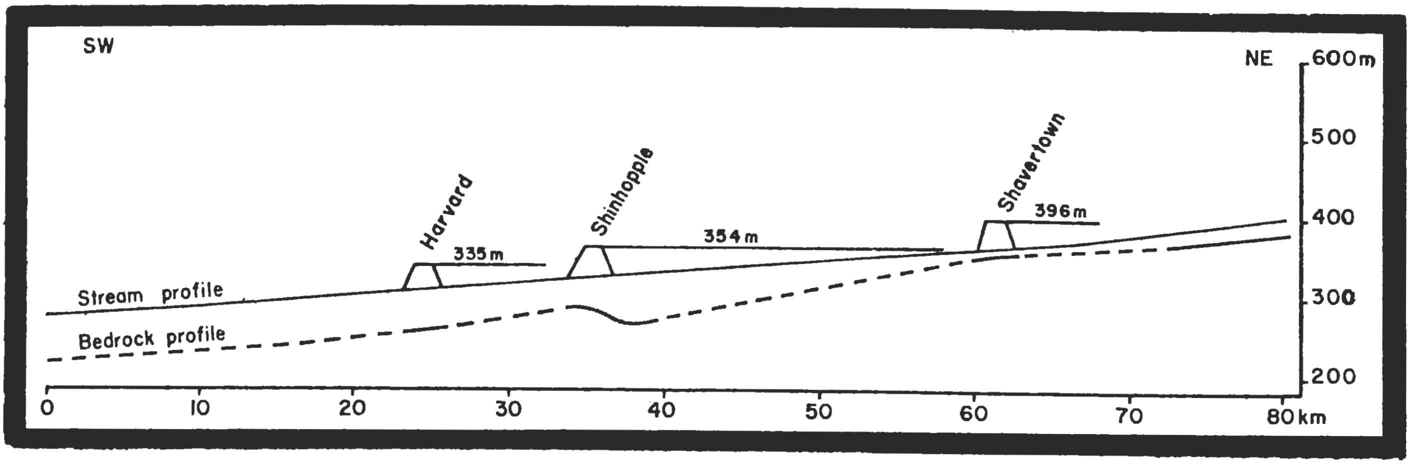


Figure 3. Longitudinal profile of the East Branch Delaware River valley showing lake levels during deglaciation (from Kirkland, 1979).

in the form of cols, notches and spillways in the upland divides (Ozsvath, 1985, Plate 1). Thus, these features can be used to reconstruct former ice marginal positions (Fig. 4). Large channels (greater than 50 m in depth) that were not created by the drainage of proglacial lakes are particularly useful. The age of these landforms is difficult to determine in a terrain that has undergone several cycles of glaciation. However, the presence of meltwater deposits downgradient from a channel demonstrates that it was utilized during the most recent deglaciation. The grading of outwash at valley confluences indicates the sequence in which meltwater pathways were used.

Topography and Ice Gradients

It is possible to establish upper and lower limits on ice marginal configurations by evaluating the constraints placed upon a theoretical ice profile gradient by topography. Thus, former ice positions can be reconstructed where the elevations of cols, notches and spillways line up along a theoretical gradient. The ice sheet profile gradient used in this study was taken from the 1.0 bar basal shear stress model of Nye (1952) used by Kirkland (1973; 1979) and Coates and Kirkland (1974) (Fig. 5). This gradient compares favorably with empirically derived gradients for retreating ice in the east-central New York region (Denny and Lyford, 1963; Mathews, 1974; Krall, 1977).

EAST AND WEST BRANCH CORRELATIONS

Evidence for correlations between these two valleys is found in the south-flowing tributaries that carried meltwaters from ice in the West Branch across the divide into the East Branch drainage. Three such tributaries are of particular importance and merit discussion.

Two of these south-flowing tributaries have similar histories: Baxter Brook, which joins the East Branch at Harvard, and Trout Brook, which joins the East Branch at Shinhopple (Fig. 1). Meltwater deposits in both tributary valleys are graded to outwash in the East Branch that is approximately 7 m above the present floodplain. It is apparent that the last meltwaters to flow through these valleys postdate the drainage of Harvard Lake (Fig. 3), which was 13 to 20 m above the Harvard-Shinhopple floodplain. The dam for the Shinhopple Lake (Fig. 3) must also have been breached to allow outwash to come through from upstream in the East Branch. Because the Shinhopple Lake (Fig. 3) had drained by this time, kame deltas deposited into this lake had already formed. Therefore, active ice in the East branch valley must have retreated to at least Shavertown (Fig. 1) when active ice in the West Branch was discharging meltwaters into Baxter and Trout Brook valleys.

The third south-flowing tributary is Cadosia Creek valley that joins the East Branch approximately 17 km southwest of Harvard (Fig. 1). Meltwater deposits in this

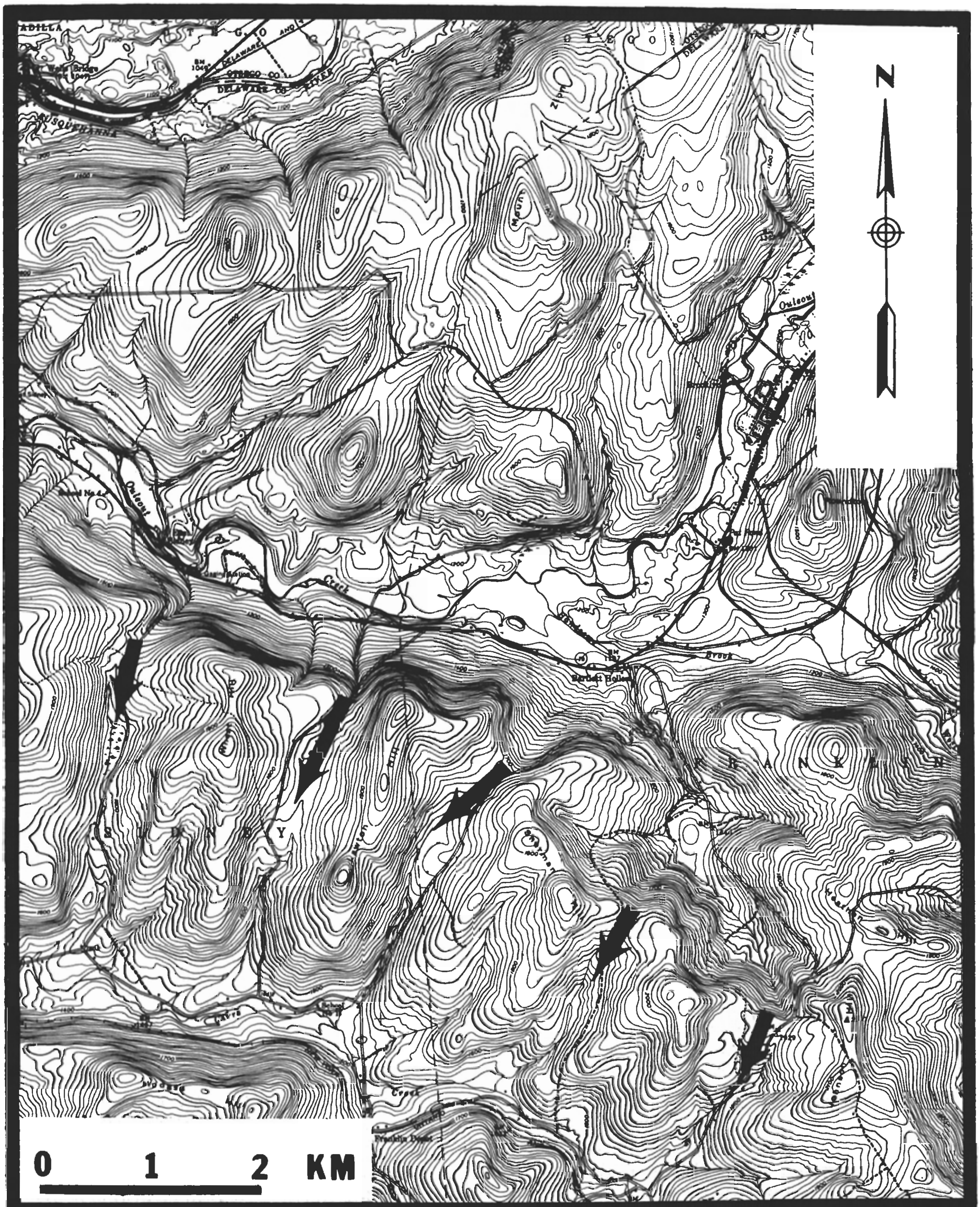


Figure 4. A reduced portion of the Franklin, NY 7.5 minute quadrangle showing the location (with arrows) of meltwater pathways (cols, notches and spillways) along a drainage divide. These features suggest the former presence of an ice margin to the northeast of this divide. Contour interval is 20 feet.

tributary valley grade to deposits in the East Branch that are approximately 30 m above the present floodplain. Thus, the use of Cadosia Creek as a meltwater channel predates the last meltwater drainage through Baxter and Trout Brooks.

DEGLACIATION CHRONOLOGY

The distribution of ice marginal erosional landforms in the western Catskills suggests that drainage divides were oriented so as to favor the formation of topographically controlled ice positions. These positions were used in conjunction with the correlations between the East and West Branch valleys to reconstruct regional correlations. Such reconstructions are by nature speculative, but they provide a framework for discussing the deglaciation chronology of this region.

Figure 6 shows the regional correlations deduced from this study. The lines drawn on this map represent active margins but each position also is associated with areas of stagnant ice in valleys to the south or southwest. These correlations support Krall's (1977) contentions that ice retreat through the western Catskills did produce discernible margins. However, Krall's specific correlations (Krall, 1977, fig. 9, Lines 2b, 3a, and 4a) have not been confirmed.

It is noteworthy that six positions have been reconstructed in the West Branch valley between Deposit and Stamford (Positions 1-6) where Kirkland (1973; 1979) identified six zones of ice detachment. Thus, each active ice position in the West Branch valley had a corresponding detached ice zone downgradient from it. The same was true for the East Branch, but the stagnant ice zones were generally more extensive. Ice also became detached into tributary drainage basins with high topographic relief and into valleys oriented perpendicular to the ice flow direction. Such valley systems are characterized by deposits formed in contact with ablating ice masses (i.e., valley diamicton and very poorly sorted ice-contact stratified drift) (Ozsvath, 1985).

The first stage of deglaciation was mass stagnation of ice into the Beaver Kill and Willowemoc Creek valleys as the ice sheet thinned over the Southern Escarpment (Fig. 6, southernmost ice position). The East and West Branch valleys had not undergone deglaciation at this point, although ice may have already begun to stagnant in the Small Lake Section of Pennsylvania to the southwest (Coates and Kirkland, 1974).

Recession from the Southern Escarpment to the East Branch/West Branch divide left large masses of stagnant ice in the East Branch valley southeast of Shinhopple (Fig. 6, letter "S," just north of the Southern Escarpment). It was during this time that Harvard Lake was impounded in the lower reaches of the East Branch valley (Fig. 3). This predates the last flow of meltwaters through Baxter and Trout Brooks, as explained earlier. The exact configuration of the upland ice margin is uncertain but it is likely that active ice existed along the East Branch/West Branch divide at Apex (Fig. 6), discharging meltwaters into Cadosia Creek (south of Apex).

Following the drainage of Harvard Lake, active ice remained along the East Branch/West Branch divide (Fig. 6, position extending from northeast of Apex towards Andes), discharging meltwaters into Baxter and Trout Brooks. Before ice retreated from this position, Shinhopple Lake also drained and stagnant ice existed in the East Branch valley from near Shinhopple to at least Shavertown (Fig. 6, "S-town"). It is likely that Shavertown Lake (Fig. 3) was in existence at this time. The point at which active ice crossed over into the East Branch drainage cannot be determined from this study but the East Branch valley ice margin must have been at or above Shavertown for reasons already discussed. Thus, the margin may have extended eastward towards Andes and then southward to the East Branch valley. The westward extension of this margin also is unknown, although cols and topographic considerations suggest that it may have been equivalent to either Positions 1 or 2 (Fig. 6). It is possible that active ice remained along the East Branch/West Branch divide during both Positions 1 and 2.

Six ice positions have been defined in the West Branch drainage. Position 1 extended from the spillway at Howes (Fig. 6) eastward along the northern drainage divides of adjacent tributary basins. This trend was repeated five times as ice retreated to the Central Esccarpment (Position 6). Two segments of ice positions are shown along the eastern side of the West Branch valley but their correlations with other positions is uncertain. The ice margin along the Central Escarpment has been recognized by other workers (Rich, 1935; LaFleur, 1969; Cadwell, 1983; this volume), and it represents a major stage in the deglaciation of the Catskill region.

The marginal positions depicted on Figure 6 suggest that active ice retreated through the western Catskills in a roughly clockwise fashion, pivoting around higher peaks to the east. Thus, the rates of active ice recession from the Southern to the Central Escarpment were not equal in the East and West Branch valleys because of the differences in topographic conditions. There are more barriers to flow in the East Branch drainage and these caused ice to stagnate more quickly and in larger zones than ice in the West Branch. This explains why Kirkland (1973) found evidence of mass ice stagnation in the area around Andes and in the Little Delaware River valley to the north of Andes (Fig. 6).

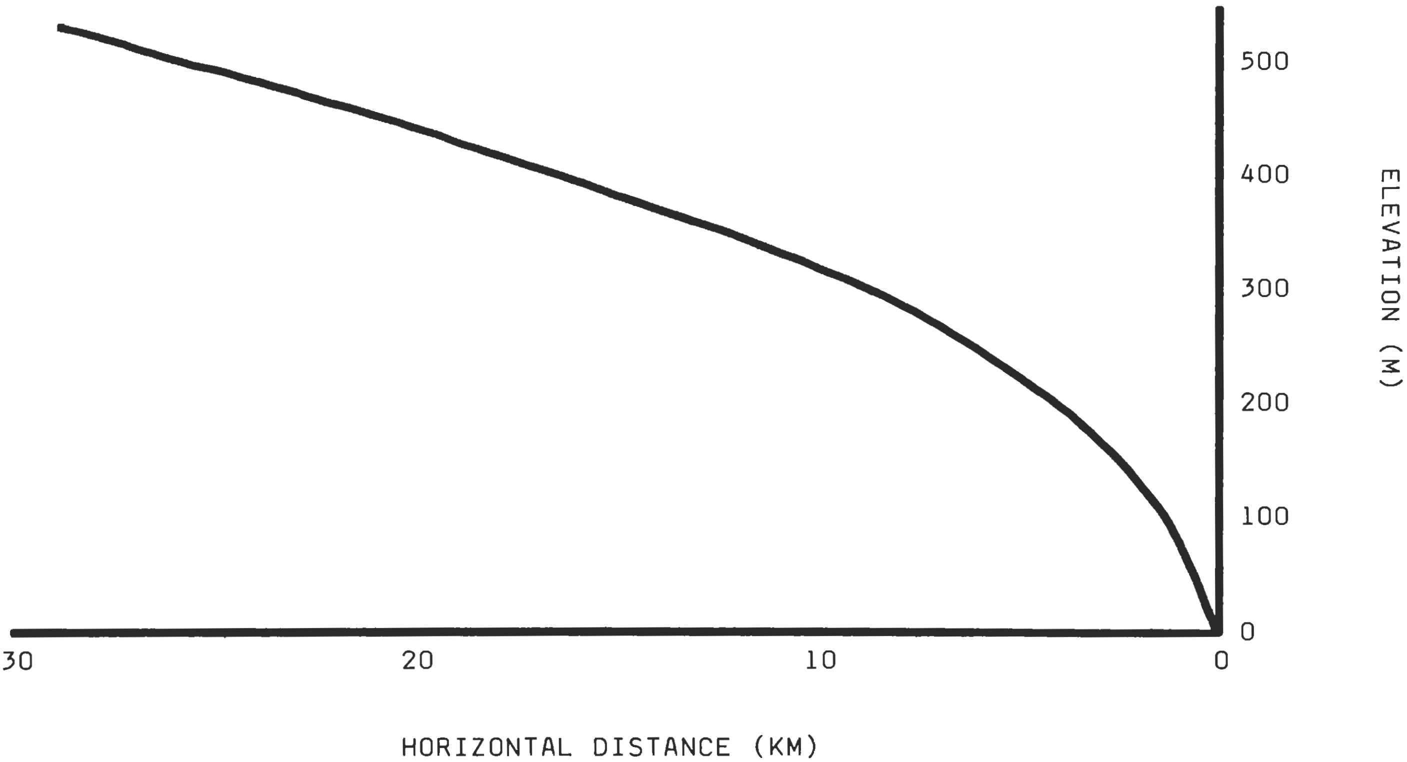


Figure 5. Theoretical ice sheet profile gradient based on a 1.0 bar basal shear stress model (after Kirkland, 1973).

CORRELATIONS WITH OTHER REGIONS

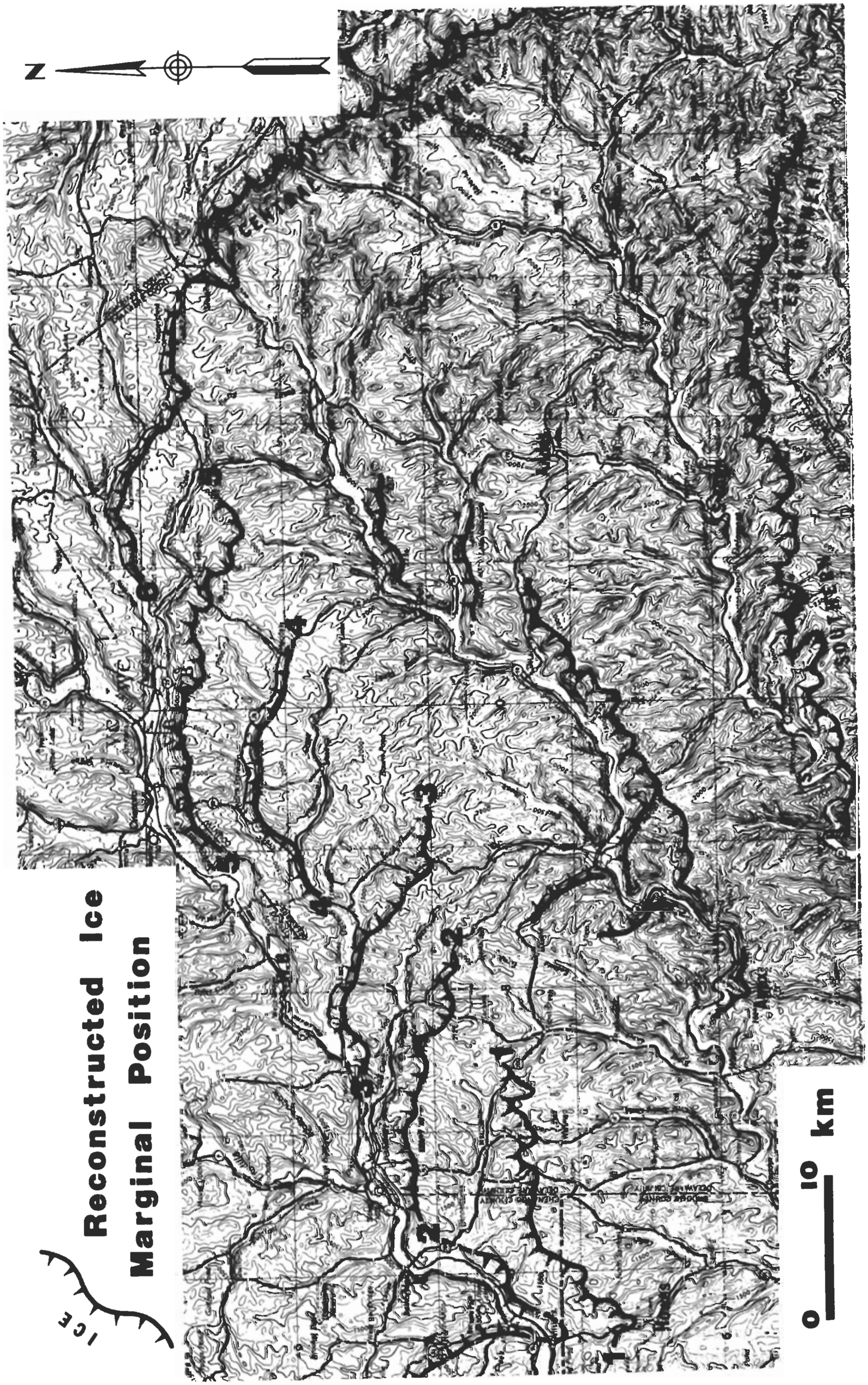
Position 2 on Figure 6 corresponds with the Chenango Forks- Sidney margin in the Susquehanna River valley mapped by Fleisher (this volume). Fleisher correlated this margin with a position defined in the Chenango River valley by Cadwell (1972) (see also Cadwell, 1981, p. 101, Margin 3). If this correlation is correct, then a minimum date for the withdrawal of ice from Position 2 is 16,650 $\pm$ 1,800 yrs BP (based on a C^{14} analysis of wood taken from a kettle hole bog along the margin mapped by Cadwell, 1972).

Position 4 on Figure 6 also corresponds with a margin mapped by Fleisher (this volume). Position 4 has been extended into the Susquehanna valley and linked to the Wells Bridge Moraine (Fig. 6, letters "W.B.," northwest of Position 4). Fleisher estimated that the retreat of ice from this position occurred 14,860 $\pm$ 800 years ago based on a C^{14} date obtained from lake sediments deposited behind the Wells Bridge Moraine (allowing 1,000 years for the lake duration). This age assignment for Position 4 is consistent with that assigned to Position 2 in terms of progressive retreat to the northeast.

Using these ages to correlate with events in the Hudson drainage is difficult because glacial stratigraphy in that region is undergoing reevaluation. It is now generally agreed that the Wagon Wheel Gap Position along the Central Escarpment (Fig. 6, Position 6) is time-equivalent to the Wallkill Moraine in the Wallkill Valley (Dineen, 1983; Cadwell, this volume; Connally and Sirkin, this volume) and the Poughkeepsie Moraine in the Hudson Valley (Connally and Sirkin, this volume). The Wallkill Moraine originally was dated at about 15,000 yrs BP (Connally and Sirkin, 1970) which is conformable with the Delaware/Susquehanna correlations (Fig. 6, Positions 2 and 4). However, new evidence indicates that this ice position may be as old as 17,210 yrs BP (Connally and Sirkin, this volume). This older date is difficult to integrate with the proposed Delaware/ Susquehanna stratigraphy, suggesting that deglaciation events may not have been synchronous in the Hudson and Susquehanna drainage systems.

SUMMARY

Glacial sediments in the western Catskills reflect a depositional landscape formed by ice disintegration. Such landscapes are characterized by an abundance of ice-contact stratified drift and "ablation till" (Flint, 1971). Conditions favorable for ice disintegration in-


N
Reconstructed Ice Marginal Position
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clude high relief, maturely dissected topography, and thin, slow-moving ice (Flint, 1971; Sugden and John, 1976). These conditions are likely to have prevailed in the Catskill region during Late Wisconsinan time (King and Coates, 1973).

Woodfordian ice retreat was greatly influenced by topography throughout the western Catskill Moutains. Progressive thinning of the Laurentide ice sheet over the Central Escarpment caused detachment of marginal ice masses in topographically controlled zones. Thus, the style of deglaciation involved both stagnation-zone retreat (in valleys oriented parallel to the ice flow direction) and mass stagnation (in valleys oriented transverse to ice flow). Six stagnation zones have been defined in the West Branch valley, and at least three in the East Branch (Kirkland, 1973). The area of largest stagnation occurred south of the Southern Escarpment in the Beaver Kill and Willowemoc Creek valleys.

Six ice marginal positions in the West Branch valley have been identified, perhaps corresponding with the six zones of ice stagnation defined by Kirkland (1973; 1979). These positions show progressive retreat of Woodfordian ice to the northeast, roughly paralleling the trend hypothesized by Krall (1977). Two of the six positions have been assigned ages on the basis of correlation with work done in the Susquehanna drainage basin: Margin 2 (16,650 $\pm$ 1,800 yrs BP) and Margin 4 (14,860 $\pm$ 800 yrs BP). These deglaciation events may not have been synchronous with ice retreat in the Hudson Valley.

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GLACIAL GEOLOGY AND LATE WISCONSINAN STRATIGRAPHY, UPPER SUSQUEHANNA DRAINAGE BASIN, NEW YORK

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ABSTRACT

The Late Wisconsinan stratigraphic units in the upper Susquehanna drainage, eastern Appalachian Plateau, New York, consist mainly of thick stratified drift within valleys and lodgement till on valley slopes and divides. Landforms created by downwasting include kames, kame-fields, and a few small eskers. Backwasting in active-ice retreat formed kame-moraines and associated valley trains of thick outwash. Features of a glaciofluvial environment are partial valley trains of gravelly outwash with associated kame terraces. Glacier retreat from an ice-contact glaciolacustrine environment left hanging deltas, delta kames, deltaic outwash terraces, and a few strandlines above lacustrine plains, all of which are related to lacustrine stratigraphic units encountered in numerous wells and test borings.

These stratigraphic units and the landforms they constitute suggest general active ice retreat (backwasting), with local stagnation (downwasting) caused by topographically controlled ice-margins within a general environment of ice-contact lakes. Moraine dams, valley trains and ice dams served to impound lakes.

Well data indicates a common stratigraphic association beneath the valley floors. A gravel cap (up to 50 feet, 15 m) usually mantles silt and sand (generally 100 to 200 feet, 30 to 61 m thick) that overlies coarse gravel of undetermined thickness. Wells that contain this sequence seldom penetrate bedrock.

Owing to limited local radiocarbon dates, the deglacial chronology is based mainly upon correlation of morphostratigraphic units with the Schoharie drainage to the east and Chenango to the west. General deglaciation and subsequent readvances occurred between approximately 15,500 and 14,000 yrs BP.

INTRODUCTION

The effects of glaciation on the Appalachian Plateau are clarly represented in the upper eastern Susquehanna drainage basin. While glacier ice repeatedly advanced and retreated across New York State during the Pleistocene, only the chronology of the Late Wisconsinan events can be interpreted.

I examined the glacial geology within an area that included most of Otsego County, and adjacent parts of Delaware, Schoharie and Chenango Counties, along the upper eastern reaches of the Susquehanna River and its major tributaries (Fig. 1). It extends from the village of Sidney on the southwest to the Susquehanna headwaters at Cooperstown to the northeast. Included are major portions of the valleys drained by the Butternut, Charlotte, Cherry Valley, Oaks, Otego, Ouleout, Schenevus, Susquehanna, and Unadilla Creeks. The study area encompasses the twenty-nine 7.5' quadrangles shown on Figure 1.

The variable influences of glacier pulses, glacier flow regimes, and ice-marginal activities across New York State has resulted in challenging problems related to correlation and chronology. In spite of this, a comprehensive picture is developing through the combined efforts of many contributors who have concentrated on specific areas, drainages, and problems. In addition to the general overview treatment given by Fairchild (1925) and Rich (1935), Coates (1974) and Coates and Kirkland (1974) considered the regional significance of main drainage ways and the general distribution of ice-marginal deposits. Krall's work (1972) included the drumlins near Richfield Springs and the occurrence and correlation of various moraines, notably the Cassville-Cooperstown Moraine. The work of Whipple (1969) also

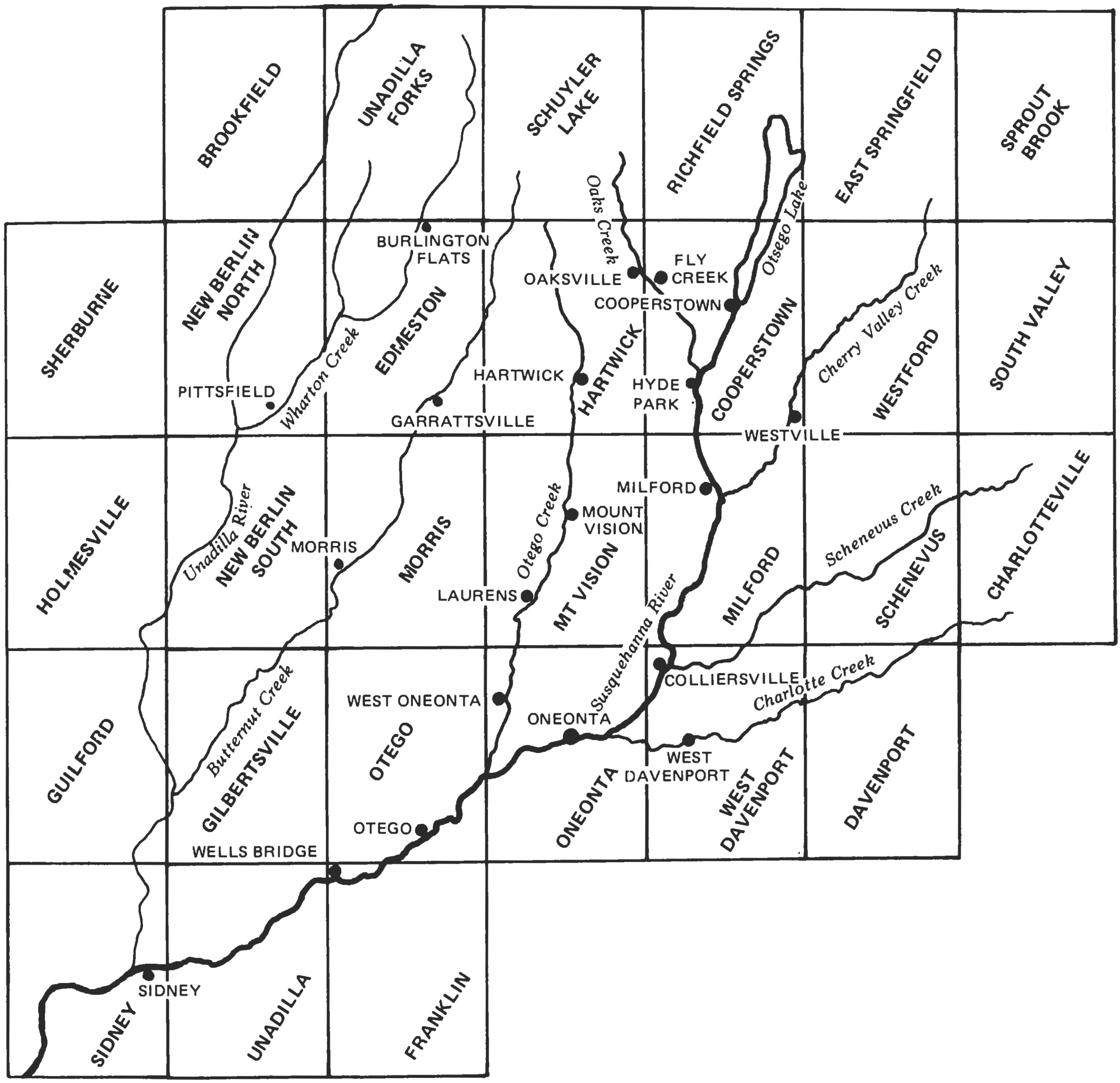


Figure 1. Index map

contributed to an understanding of the glacial geology north of Cooperstown. The chenango drainage to the west was studied by Cadwell (1972), who suggested a conceptual model for stages of deglacial events, proposed a retreating sequence of ice-marginal positions, and also provided a radiocarbon date that has helped to establish ice-marginal chronology. To the east in the Schoharie drainage basin, LaFleur (1969) and Cadwell (1983) considered the unique aspects of glacial pulses southward into a north flowing river system and the northern Catskill slopes. The glacial events of the western Catskills have been investigated by Kirkland (1973) and Fleisher (1977a) but only recently has the deglacial chronology in the West Branch of the Delaware River drainage basin received detailed consideration (Ozsvath and Coates, 1983). A general treatment of the upper Susquehanna glacial geomorphology was reported by Fleisher (1977b), in which the deglacial environments of deposition were discussed.

This report emphasizes two main topics: 1) the stratigraphic record of ice-contact environments of deposition and 2) the definition of criteria by which progressively retreating ice margins can be identified and correlated. Attention is given to the association of deglacial landforms with subsurface stratigraphy. A Late Wisconsinan chronology is suggested based on the physical correlation of ice-marginal positions with those in adjacent areas for which radiocarbon dates have been reported.

GEOLOGIC SETTING

This portion of the Appalachian Plateau is characterized by deeply dissected middle to upper Devonian clastic strata that include interfingered beds and lenses of sandstone, siltstone, shale and sparse conglomerates of the Hamilton and Genesee Groups. The regional dip is to the south-southwest at angles typically less than 10 degrees. In general, the bedrock of the region is well expressed by the topography. Some divides and broad, arcuate cuestas are capped by massive, resistant beds, allowing the subtle structural configuration to be seen on a regional scale.

The topography shows the compound erosional influence of a fluvioglacial origin with the Susquehanna River as the main trunk stream. The present drainage consists of glacially-enlarged valley troughs that reveal an inherited ingrown meander pattern of preglacial origin. Ice-scoured bedrock mantled by thin lodgement till characterizes the gentle sloping uplands; however, till shadows and isolated stratified drift exist there. Plucking of large bedrock blocks formed some basins in which upland lakes and bogs have formed.

A significant and noteworthy aspect of the physiography in this part of the Appalachian Plateau is the great antiquity of the drainage basin. Long before the Mohawk River eroded headward to capture the Adirondack headwaters of the Susquehanna, a pattern of deeply ingrown river meanders was developed and can still be recognized in the topography. The ancestral Susquehanna, which flowed southwesterly, developed major south-flowing tributaries. Much smaller tributaries enter from the south and southeast. This is primarily the result of two major factors: 1) the south and southwesterly flowing streams originated resequent to bedrock controls, whereas smaller subsequent streams followed northwesterly trending bedrock cuestas and 2) the main direction of ice movement from the north-northeast enlarged valleys oriented parallel to glacial flow, whereas the transverse trending valleys gained little in size as the result of glacier erosion.

Local relief typically reaches 600 to 700 feet (183 to 213 m). However, if one consideres the drift that chokes the valley floors to thicknesses that generally range between 200 and 300 feet (61 and 91 m) (Randall, 1972; Gieschen, 1974), the erosional relief is seen to be considerably greater. This drift is sorted to some degree, even in moraines, and consists of glaciofluvial and glaciolacustrine gravel, sand, silt, and clay.

Two large lakes (Canadarago and Otsego) dominate the headwater valleys of the Susquehanna River. Both are vestiges of ice-contact lakes that occupied moraine-dammed valleys. Otsego Lake, the larger of the two, currently occupies a basin with a maximum depth of 166 feet (51 m), which contrasts with only 44 feet (13 m) in Canadarago Lake (Harman, 1974; Weir and Harman, 1974). Sometime following deglaciation the spillways of both lakes breached their impounding moraines and the lakes were lowered to their present elevations. A similar geomorphic history can be interpreted for other valleys that are completely free of lakes today but contain thick lacustrine sediments.

GLACIAL MODIFICATION

The overall effects of glacial erosion are represented in the landscape by the oversteepened flanks of broad glacial troughs and truncated spurs. Because the general grain of the topography consists of drainageways and interfluves oriented to the south-southwest and parallel to the direction of general ice flow, the most common upland character is that of a smoothed and streamlined terrain. However, cols and meltwater channels commonly are found on divides and uplands formed by structurally controlled cuestas oriented against the general direction of the ice flow. In addition, umlaufbergs in various stages of development can be found in Sidney, Unadilla, and Cooperstown and are evidence of the large-scale erosion that influenced selected portions of the terrain.

In their study of umlaufbergs, Coates and Kirkland (1974) found a progressive sequence of development from which they interpreted multiple stages of erosion, and therefore, multiple glaciations. A similar line of reasoning may be applied to the spectacular bedrock gorges that commonly indent many of the oversteepened valley walls and obsequent slopes of cuestas. They appear joint controlled but lack general alignment with published joint trends. They typically have steep to vertical walls and a bedrock floor that serves as the channel for short streams of exceptionally steep gradient. In some cases these streams flow through abrupt first-order drainageways, whereas others form short steep and narrow segments along the main drainageway of 3rd and 4th order streams. In all cases, diamictons interpreted as lodgement till is being eroded, exhuming and enlarging the bedrock gorge that must have existed prior to the last glacier advance across these slopes. Small alluvial fans commonly are found at the mouths of many gorges and consist of channery gravels characteristic of their origin.

The greatest variety of glacial landforms can be found along the lower valley walls and floor where diamictons interpreted as super and englacial stratified drift are concentrated. Landforms consist mainly of kame moraines, dissected and pitted outwash, kames and kame terraces. These are found in association with hanging deltas and lacustrine plains. A form of meltwater deposit referred to as *deltaic outwash terraces* typically stand 60 to 80 feet (18-24 m) above the modern flood plain and contain gravel in massive foreset beds. In addition, delta kames and short eskers suggest that ice-contact lakes existed and some local stagnation occurred. Cadwell (1972) suggested a deglacial ice margin defined by prominent valley ice lobes to account for similar landforms in adjacent valleys to the west. This applies equally well for the upper Susquehanna.

An additional noteworthy characteristic of the ice is that it was apparently charged with sufficient meltwater to produce sorted and stratified diamictons as valley-floor landforms. In contrast to this is the unsorted, unstratified, clay-rich, highly compact lodgement till that mantles the valley walls and divides and which formed subglacially in an environment that was influenced by meltwater but to a lesser degree.

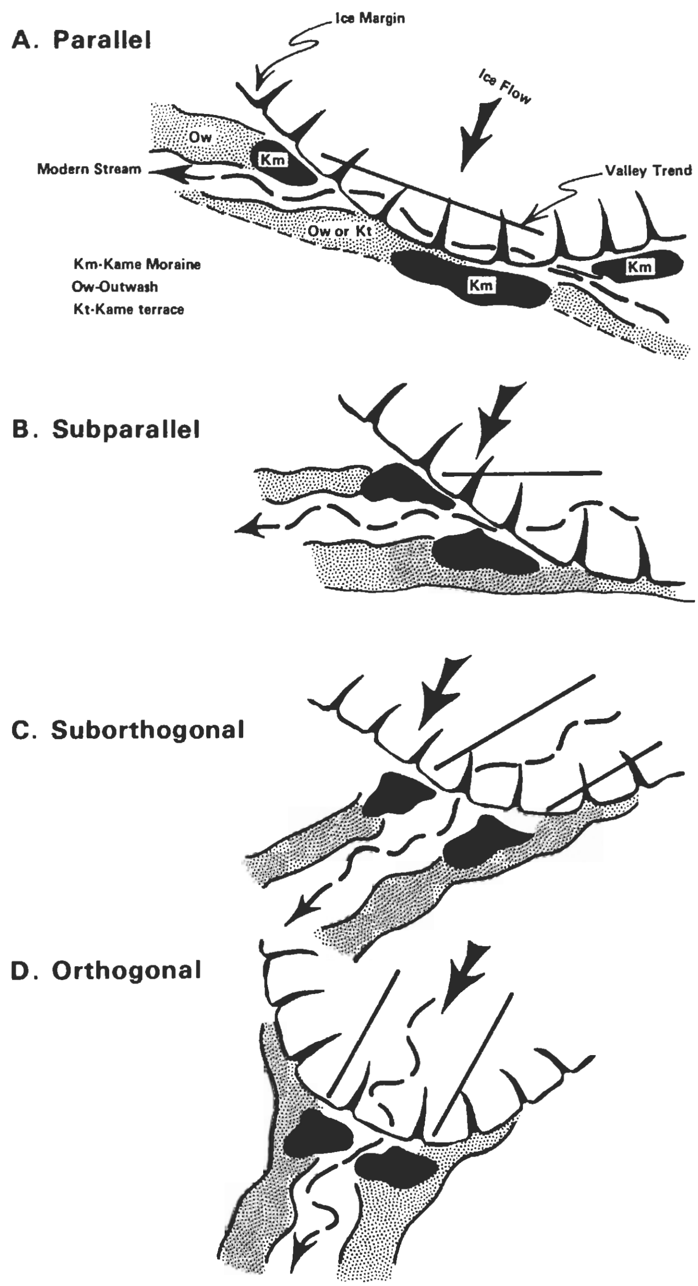


Figure 2. Ice marginal landforms and valley trend. The distribution of landforms is determined by the orientation of the ice margin relative to the valley trend.

CONDITIONS OF DEGLACIATION IN HIGH RELIEF TERRAIN

Topographic Influence – Ice-marginal Trend and Valley Orientation

Local relief on the Appalachian Plateau is sufficient to have influenced the ice-marginal configuration during retreat. Cadwell (1972) described a valley lobe formed by an ice tongue extending down-valley from an ice margin oriented perpendicular to the valley trend. The position of the lobe would be defined by its marginal ice-contact deposits. Similar deposits would depict the ice-marginal position where its orientation paralleled the valley trend. Both alternatives, with intermediate variations and resulting landforms, are illustrated in Figure 2 (Fleisher and Cadwell, 1984).

Kame-moraines and associated outwash are the primary landforms of an active ice-marginal position in the upper Susquehanna drainage basin. They are distributed discontinuously along a valley in which the ice margin and valley trend were parallel (Fig. 2A). With a subparallel orientation (Figure 2B), the depositional landforms are less widely spread along the valley.

Similarly, a cross-valley distribution of kame-moraines is approached (Figure 2C) and finally attained (Figure 2D) as the ice margin trend approaches a direction perpendicular to the valley. Here, a valley ice-lobe dominates the depositional environment, with the kame-moraine in the most favorable position to form a dam across the valley as the ice recedes.

Topographic Influence – Drainage Pattern and Incised Meanders

An asymmetric drainage pattern and system of incised valley meanders served to control ice flow during the thinning phase of glacier retreat. Consequently, the locations of characteristic ice-marginal landforms can be related to large-scale drainage features.

During retreat, major south-flowing tributaries would have continued to supply actively flowing ice streams along marginal positions oriented parallel to the trends of major valleys. These streams of active ice would provide a depositional environment in which kame-moraines would form at the mouths of tributaries. Such ice streams would be separated by zones of passive ice, in which stagnation would ultimately occur and ablation moraines form (Figure 3A). The major depositional distinction between active and passive ice is the occurrence of thick and widespread outwash deposits which suggest continuous contact with active ice and its meltwater system. This is in contrast with the detached, stagnant ice that downwasted to form ablation landforms. This is best illustrated by the occurrence of kame fields (on the valley floor and adjacent valley walls), which appear as isolated patches of hummocky terrain lacking an association with large-scale outwash. Presumably, the amount of available meltwater, from which outwash would originate, was limited to the volume of the downwasted ice.

Figure 3B illustrates the influence of an ingrown meander on the location of a kame-moraine and pitted outwash. The valley relief is sufficiently high (800 feet, 244 m) and curvature tight enough to inhibit the main thread of flow within an active ice lobe, thereby creating a zone of reduced activity. As a result, static flow leads to stagnation and ice burial, with associated dead-ice landforms, one of which is pitted outwash. Where sufficiently large blocks of ice become detached and entirely buried by outwash, a large valley floor kettle ("dead-ice sink") forms (Fleisher, 1986). The concept of a dead-ice sink as an environment of deposition is further developed *in the following discussion of valley floor landforms.*

Topographic Influence – Through and Non-through Valleys

The south-flowing tributaries to the Susquehanna River occupy one of two valley types: 1) through valleys that do not climb headward to upland elevations but instead have low longitudinal profiles within open-ended glacial troughs and 2) non-through valleys that were enlarged less by glacial erosion and rise longitudinally to divide elevations well above those in adjacent through valleys. Normal backwasting within a non-through valley leads to thinning ice over the divide which impedes continued flow to the ice lobe down valley and promotes general stagnation. It is within non-through valleys that landforms produced by downwasting (eskers and kame fields) are most commonly found (Fleisher and Cadwell, 1984).

A MODEL FOR DEGLACIAL ENVIRONMENTS

Figure 4 illustrates the environments of deposition associated with the most common model for glacier retreat in this region (Fleisher, 1977a). A moderately extensive ice lobe is illustrated terminating in an ice-contact lake

Part A

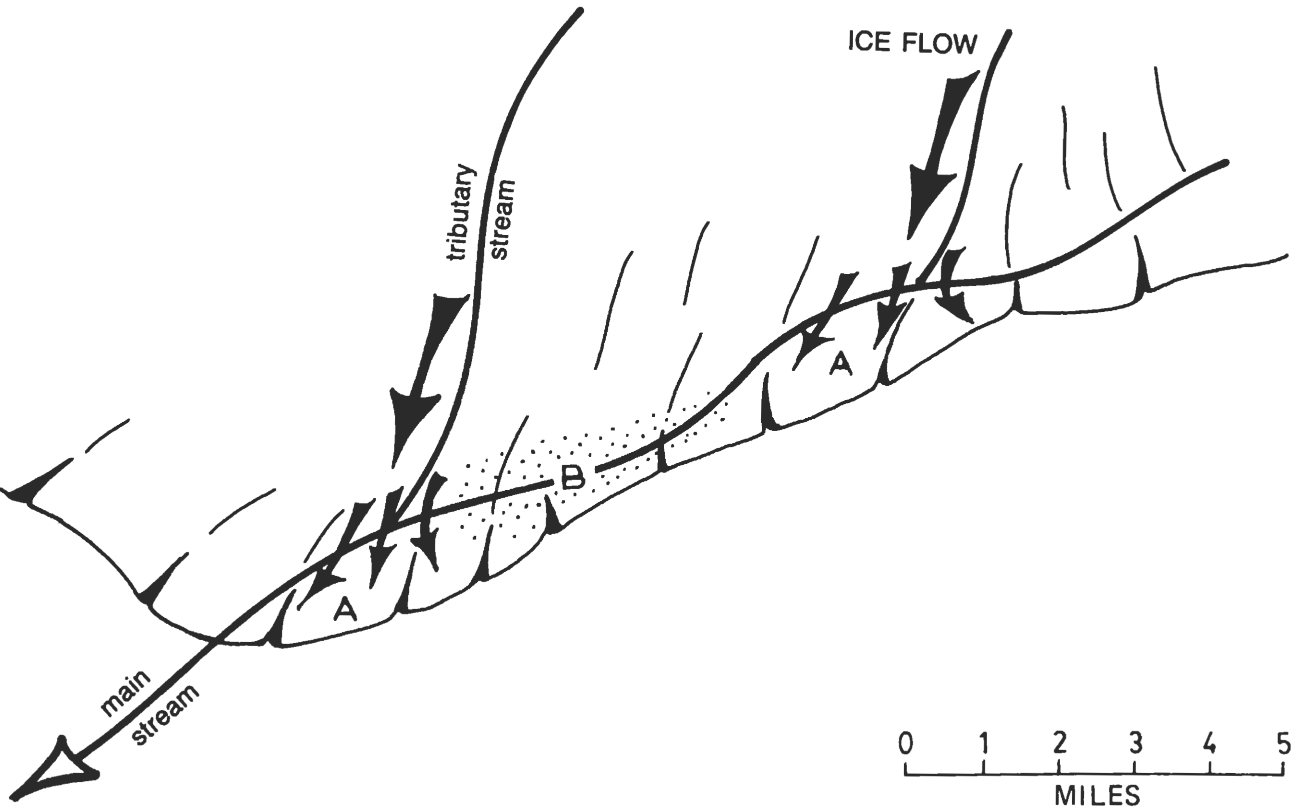


Figure 3. Topographic control of landform occurrence.

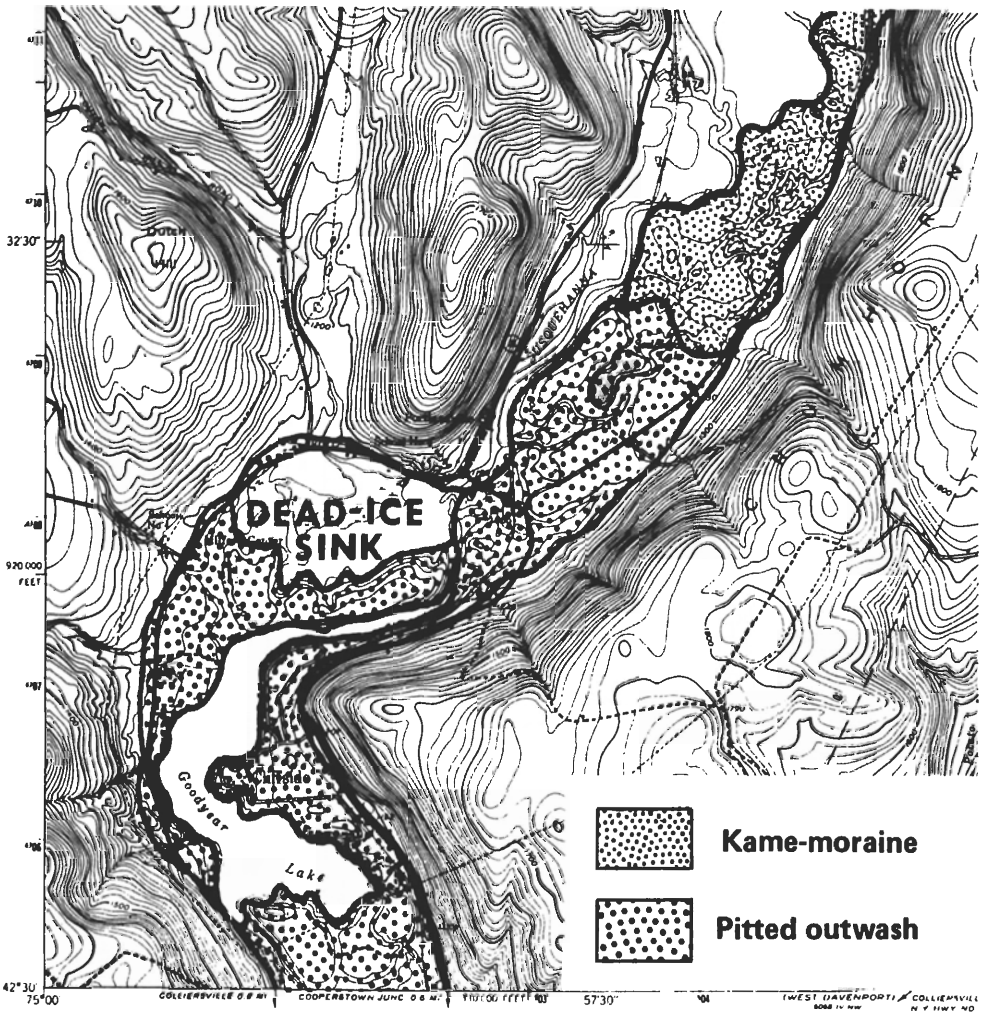
Part A – Tributary valleys aligned parallel to the active ice flow direction feed active ice streams along an ice margin that follows the course of the main valley trend. Active flow (A) separate zones of passive flow (B). Kame-moraines and associated outwash are indicative of active flow, whereas kame fields (lacking significant outwash) mark locations of passive ice. The Schenevus Creek Valley is diagrammatrically represented here.

whose dam is located some distance down-valley. At any given time a variety of deposits will accumulate in different, yet associated environments. While lodgement till is deposited subglacially on valley walls and divides, lateral meltwater drainage deposits sand and gravel in kame terraces. Where the same meltwater streams enter the ice-contact lake, deltas form. With continuous retreat from previous delta positions, new space is created but quickly filled by sediments in the form of massive foreset beds as the delta migrates in contact with the ice margin. Upon drainage of the lake, the dominant landforms consist of continuous deltaic outwash terraces (in the form of paired terraces) and a lacustrine plain across which the modern flood plain develops (cf. Figure 4).

Deltaic outwash terraces formed in this way have been referred to in earlier literature as kame terraces or delta kame terraces (Cadwell, 1972). Although they are the product of deposition by the same streams that form kame terraces, their environment of deposition is not between the ice lobe and the adjacent valley wall, but rather at the lobe terminus in an ice-contact lacustrine environment. It is suggested that the term *deltaic outwash terrace* be used instead of kame terrace for features with this topographic expression and internal deltaic structure.

The flat valley floor of the lacustrine plain is underlain by deposits of fine sediment that are in part the bottomset facies of prograding deltas but mainly consist of laminated lacustrine silt (Fig. 4). Lacustrine deposits often attain thicknesses in excess of 400 feet (122 m), indicating a deep lake basin of significant duration during one or more glacial events.

Part B



Part B – A tight ingrown meander in which active flow is inhibited during retreat resulting in dead ice burial and the formation of pitted outwash and large valley floor kettle (dead-ice sink). This example is taken from the Susquehanna Valley at Portlandville (Milford quadrangle). Dead-ice sink is about 4500 ft. across, with north to the top of the map.

Glacial Lake Otego

Landforms and subsurface stratigraphy indicate that the upper Susquehanna drainage basin was the site of several large ice-contact and proglacial lakes during deglaciation (Fleisher, 1977a). A good example lies along the valley of the Susquehanna River between the hamlet of Wells Bridge and the City of Oneonta. A large moraine blocked the valley at Wells Bridge creating a body of water 12 miles (19.2 km) long, almost 1 mile wide and more than 400 feet (122 m) deep, referred to by Melia (1975) as Glacial Lake Otego.

Sorted and stratified sandy gravel exposed in surface excavations and recorded in the logs of water wells (Randall, 1972) constitutes the moraine. Depth to bedrock is in excess of 170 feet (52 m) and possibly as deep as 266 feet (81 m), as seen in a well 2.5 miles (4 km) down-valley. The moraine fills the valley to an elevation of 1120 to 1140 feet and is associated with a broadly dissected valley train. Figure 5 illustrates the hummocky, rolling valley-floor topography of a breached valley plug referred to here as the "Wells Bridge moraine." The cross section includes landforms and deposits associated with active ice retreat during backwasting.

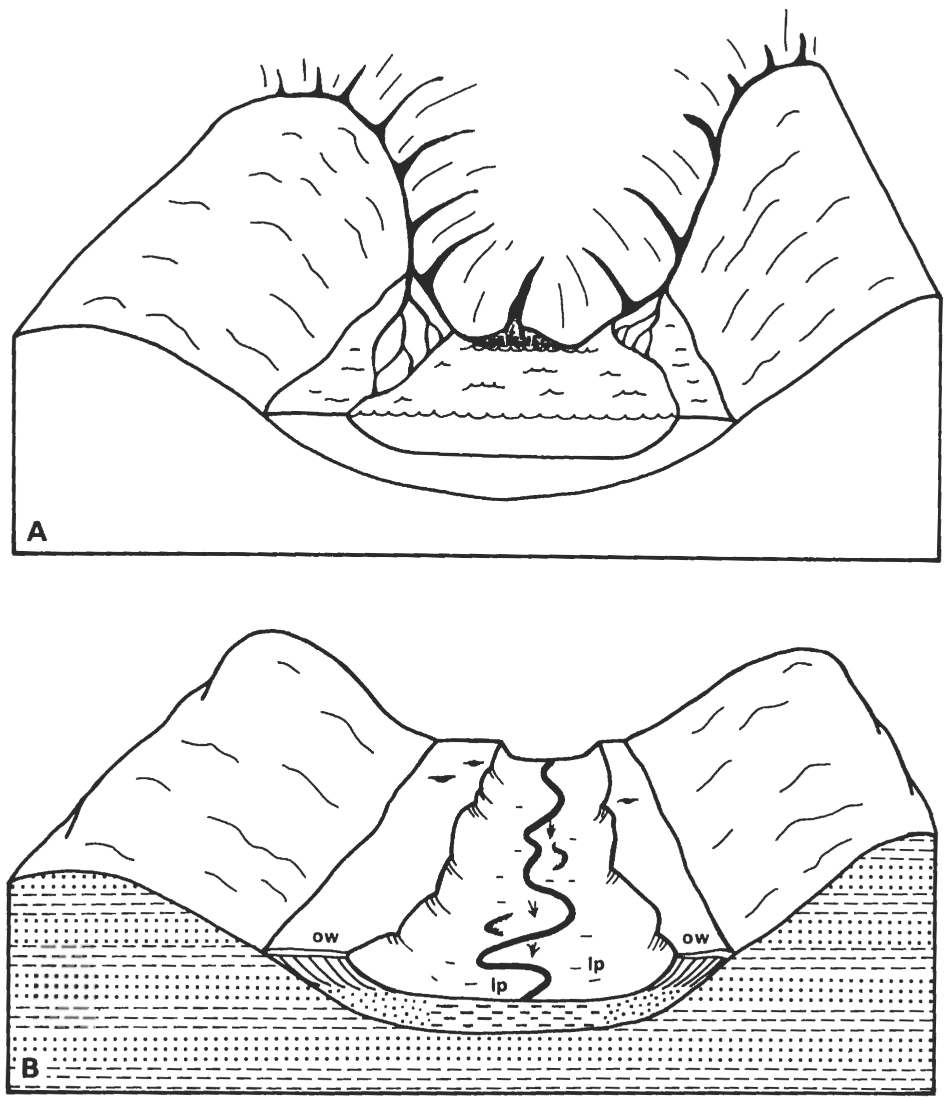


Figure 4. Ice lobe model for glacier retreat.

Part A – During glacier retreat valleys were typically occupied by an ice lobe, with associated lateral drainage into an ice contact lake.

Part B – After glacier retreated and lake drained, the valley is shown to contain deltaic outwash terraces (ow) separated by a lacustrine plain (lp) across which the modern flood plain has developed.

As the ice retreated, a lake began to form behind the moraine and in front of the ice. This ice-contact lake was fed primarily by lateral and englacial drainage from the ice lobe margins. Initially, the lake was small and collected a heterogeneous assemblage of collapsed till, debris flows, glaciofluvial sand and gravel, and glaciolacustrine sand and silt. Such an association is typical of the ice-contact environment and can be anticipated elsewhere in the upper Susquehanna. As the ice withdrew, the moraine persisted and Glacial Lake Otego expanded in size.

An interpretation of rapid retreat is supported by discontinuous deltaic outwash terraces. Gravels interbedded occasionally with thick silt and clay indicate

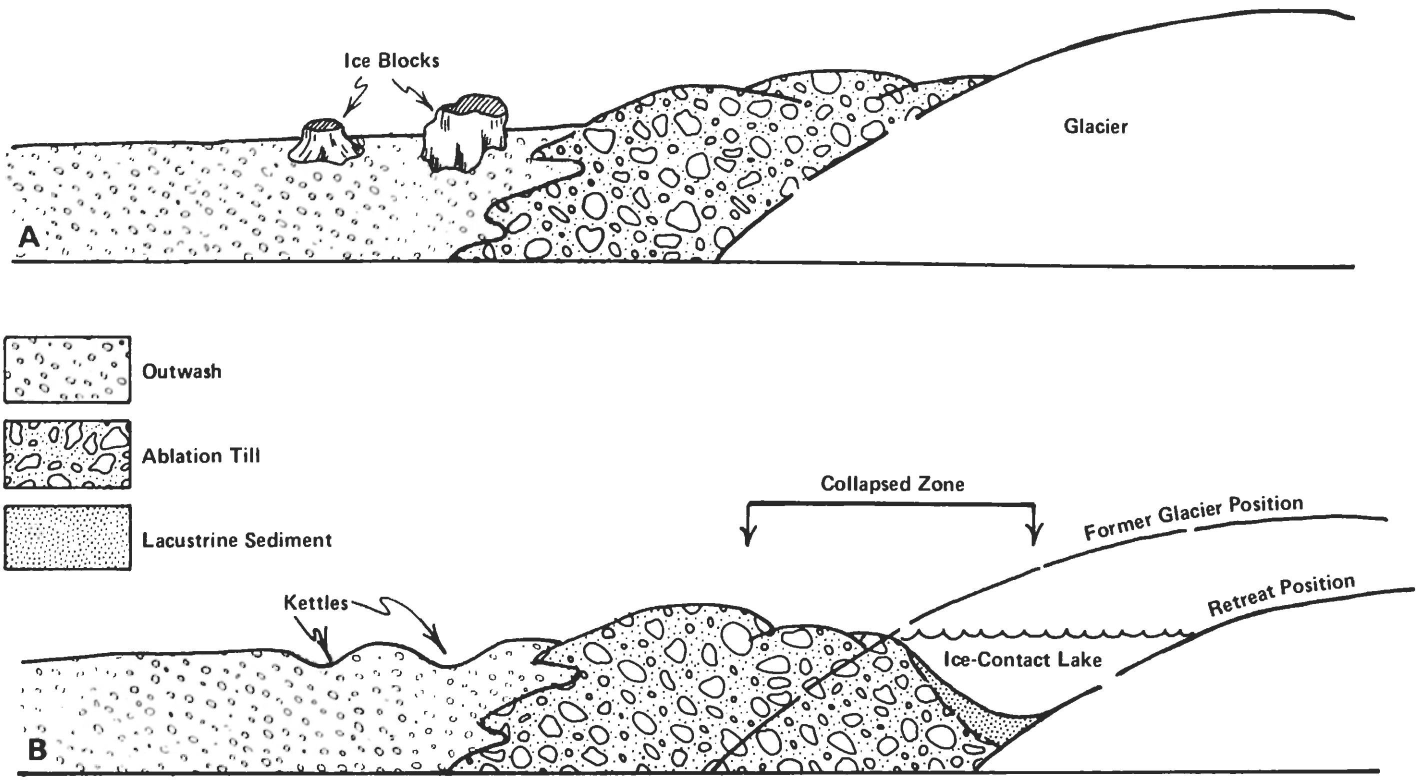


Figure 5. Origin of Wells Bridge moraine. Diagrammatic illustration of ice retreat from moraine which dammed the Susquehanna Valley to form Glacial Lake Otego.

Part A – Ice margin and associated deposits.

Part B – Retreat and subsequent formation of the Wells Bridge moraine.

periodic accumulation of highly turbid flows derived from ice, ice-rafted gravels or possible valley-wall mass wasting.

The topography of a modern flood plain covering an ancient lacustrine surface in the vicinity of Otego is shown in Figure 6. Well number 20 is located near the middle of the former lake and about 5 miles (8 km) from the nearest up-valley ice-marginal position. At this location, 434 feet (145 m) of valley fill was penetrated, almost all of which is silt (although it is referred to as quicksand in driller's logs). Assuming an average sedimentation rate of 1 foot (.3 m) per 2 years for silt (Ashley, 1975) and a minimum thickness of 420 feet (140 m), an estimate of at least 840 years of uninterrupted lacustrine sedimentation would have been required to produce the deposits of Glacial Lake Otego.

The Johnson-Russ archaeological site within the breach of the Wells Bridge moraine indicates when the lake may have drained. Funk and Dineen (personal communication, 1983) developed the relationships illustrated in Figure 7 at this site. Although some questions still remain on the source and correlation of deposits at the site, a radiocarbon date of 13,860 ± 800 yrs BP from a depth of 2.5 feet (0.8 m) below the surface of a sand and gravel terrace that stands approximately 12 feet (4 m) above the modern flood plain is interpreted to mean the moraine was breached 13,060 to 14,660 yrs BP. Adding 840 years for the minimum duration of the lake places the minimum age for ice retreat to be on the order of 13,900 to 15,500 yrs BP. This establishes a minimum age for the Wells Bridge moraine and any landforms graded to Glacial Lake Otego.

A well-defined strandline in the vicinity of Otego marks the high water level of Glacial Lake Otego at an elevation of 1140 ± 10 feet. It is best expressed 2 miles (3.2 km) east of Otego along the eastern and southern valley wall. There the east-west valley trend makes an abrupt right angle turn to the north. The resulting west facing slope trapped sand moved eastward by currents developed over an uninterrupted 5 mile (8 km) fetch. Ev-

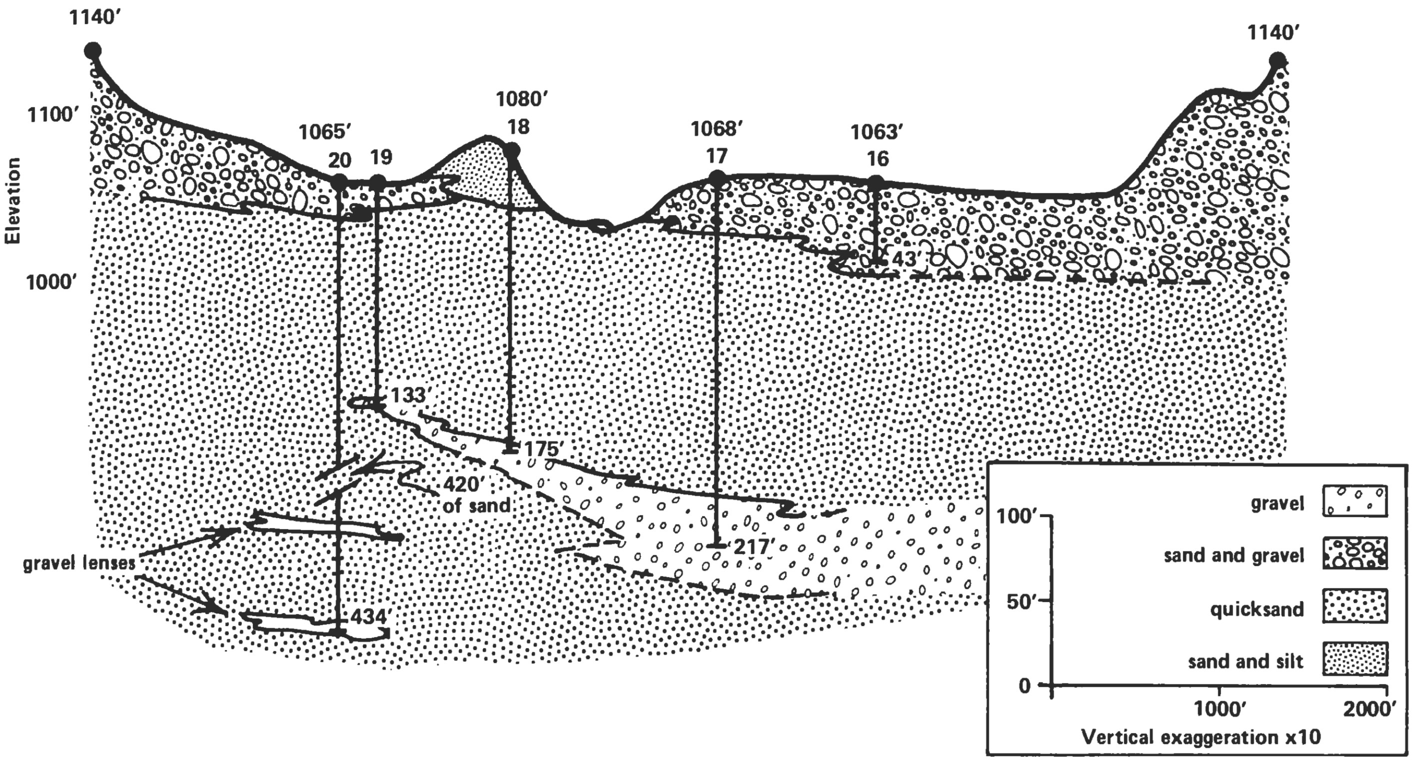


Figure 6. Cross-sectional diagram of Glacial Lake Otego stratigraphy. The logs of five water wells along a general north-south azimuth approximately one mile (1.6 km) east of Otego indicate an exceptionally thick accumulation of quicksand. With a vertical exaggeration of X10, the log of well number 20 was compressed to illustrate a total depth of 434 ft (132 m). Note that 420 ft (128 m) of quicksand is represented in one break.

idence for Glacial Lake Otego as a continuous body of water can be traced up the Susquehanna Valley to the Otego Creek tributary 12 miles (19.2 km) east of Wells Bridge. Here, at West Oneonta, surface excavation of a kame delta reveals the topset/foreset contact at approximately 1140 feet and establishes the continuity of lake level.

A variety of well-developed glaciolacustrine features that formed in association with Glacial Lake Otego also exist in the vicinity of Oneonta. The city itself is situated on a large dissected hanging delta complex that marks the entrance of Oneonta Creek and Silver Creek to the Susquehanna valley.

The landforms and valley-floor stratigraphy similar to those of Glacial Lake Otego can be seen elsewhere in the upper Susquehanna and its major tributaries. This suggests other lakes, possibly an entire chain of lakes, were an important part of the deglacial history.

Valley Floor Landforms

There are four distinctly different types of valley floor topography, each consisting of a characteristic assemblage of landforms related to some phase of the glacial history. They are well preserved, with only minor erosional modification.

1. Kame-moraines – these are recognized by their rolling, hummocky, kame and kettle topography that commonly spans the full width of the valley floor and plugs the valley for 1/2 mile (0.8 km) or more. Typically they denote an active ice-lobe margin and are associated with thick outwash deposits. Relief above the valley floor varies, but 80 to 100 feet (24 to 30 m) is not unusual. An example from the Charlotte Creek valley at West Davenport can be seen in Figure 8.
2. Outwash terraces and pitted outwash – these landforms constitute the highest planar features of the valley floor by standing 60 to 80 feet (18 to 24 m) above the modern flood plain. Their gravelly texture permits high infiltration and very little runoff. These features occur along all major drainages and typically can be traced headward to kame-moraine positions. Figures 8 and 9 illustrate pitted outwash associated with kame-moraines.

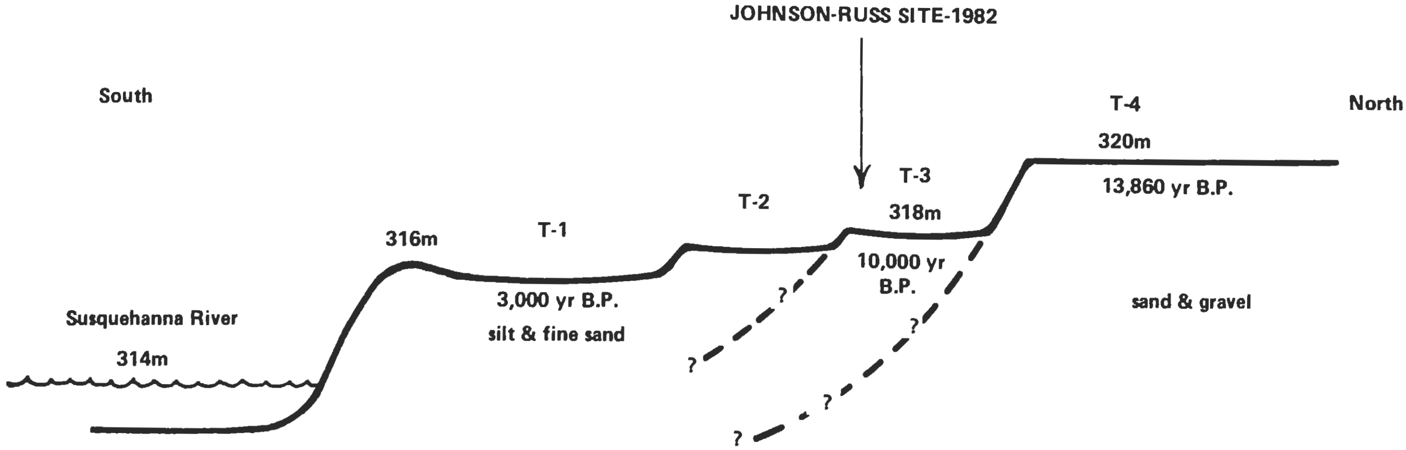


Figure 7. Diagrammatic cross section of Johnson-Russ archeological site (Funk and Dineen, personal communication, 1983). T-1 through T-4 are low relief terraces at progressively higher elevations, with associated radiocarbon dates. This cross section is located adjacent to the Susquehanna River and within the breached Wells Bridge moraine.

3. Dead-ice sinks – these features have topographic expression similar to a lacustrine plain, but with limited longitudinal extent along the valley floor. They are typically associated with extensive outwash terraces that restrict the modern flood plain width both up-valley and down (Fleisher, 1984, 1986). Because they are often traversed by tightly meandering streams, they may be misinterpreted as the product of lateral erosion. However, a dead-ice sink is anomalously wide when compared to lateral dissection in the outwash of adjacent parts of the valley. Such features represent sites at which large stagnant ice blocks were detached during retreat, then buried by stratified drift. In a sense, they are enormously large kettles that dominate the valley floor. However, they differ from kettles in that they occupy the entire valley floor, occur on the flood plain and are sites of continuous sedimentation. The topographic expression of a dead-ice sink is shown in Figures 9 and 10.

4. Lacustrine plains – two different types of lacustrine environments are interpreted from landforms and associated well-log stratigraphy, 1) ice-contact and 2) late glacial or post-glacial lake. Within the ice-conact lake environment, meltwater discharge from a slowly retreating active ice lobe formed deltaic outwash terraces and isolated hanging deltas. Beneath the lacustrine plains are very thick silts that are typically capped by sand and gravel, whose origin remains to be determined.

   The late glacial or post-glacial environment is characterized by such diagnostic features as a broad and extensive, low relief flood plain (less than 10 feet, 3 m) with swampy areas and exaggerated meanders that may traverse the entire valley width. Typically, the breached moraine that once served as a dam can be found down-valley. The valley floor usually is mantled by a veneer of overbank silt that covers clay-rich lacustrine deposits. Both surface expression and subsurface stratigraphy indicate a lacustrine environment free of high-discharge inflow. Numerous examples can be cited, as in the Cherry Creek Valley (Westford quadrangle) in the vicinity of Middlefield (Figure 11).

WELL LOG STRATIGRAPHY

Well and Test Boring Data

Water well data and test boring stratigraphy for the entire upper Susquehanna is summarized in Randall (1972). Almost all of the data come from the stratified drift of the main valleys with occasional information on the divides and lesser tributaries. Of particular interest is the physical stratigraphy from more than 200 wells and borings associated with mapped glacial landforms. In order to obtain a visual perspective of the stratigraphic units and their areal distribution, I summarized Randall's written subsurface logs on topographic maps with their locations specifically identified. A sample map is shown in Figure 12. Through this technique a vast quantity of detailed stratigraphy can be viewed at a glance, quadrangle by quadrangle. As a result, several general observations were made.

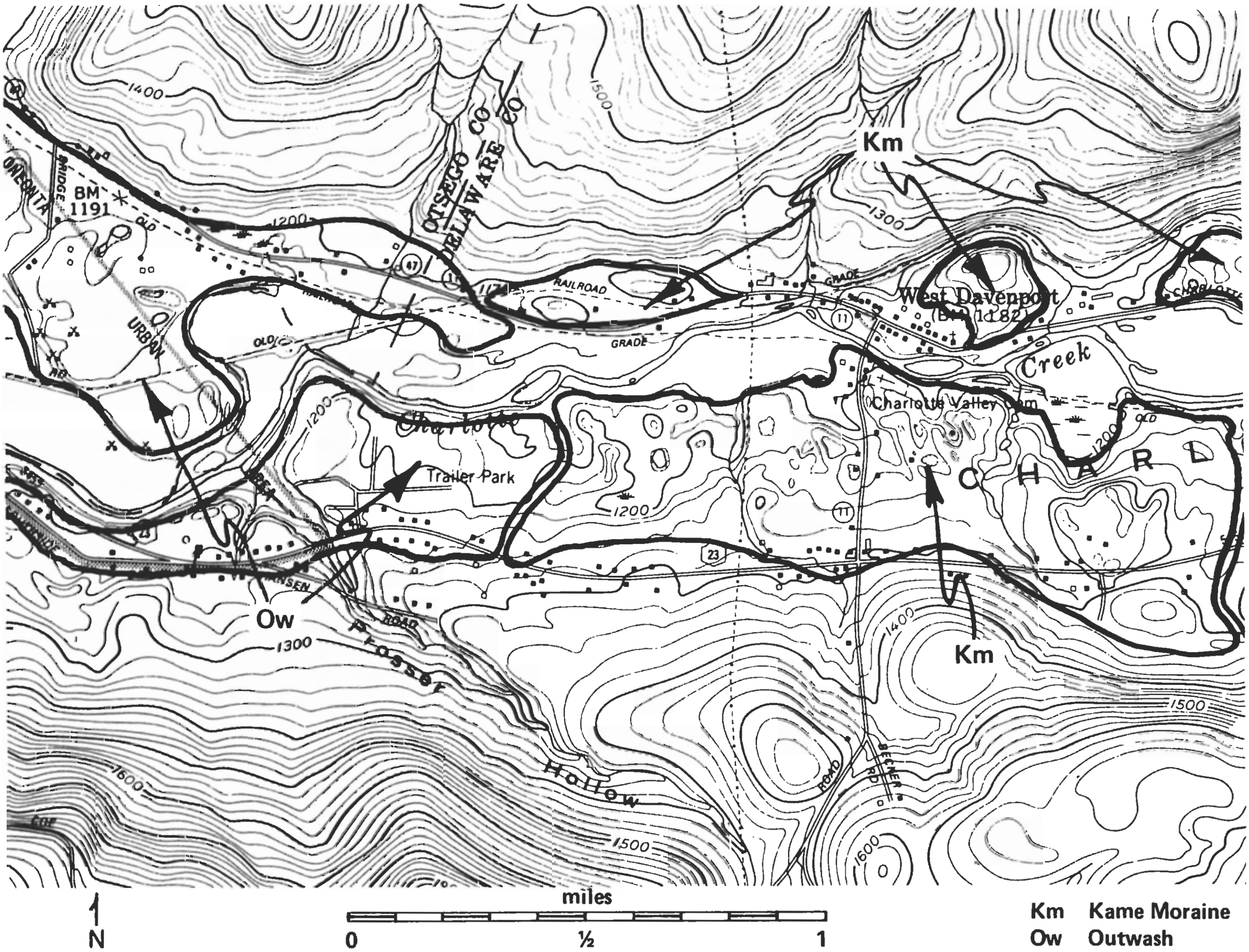


Figure 8. Kame-moraine and associated outwash, West Davenport, West Davenport quadrangle.

Because the origin of the data is primarily logs of domestic water wells, they generally stop in the first suitable aquifer. In some examples an aquifer was encountered within 100 feet (33 m) of the surface. However, the majority of the wells are considerably deeper, extending to depths in excess of 400 feet (133 m) in several locations and usually terminating in gravel. Wells drilled near the valley sides generally penetrated bedrock, as did those on valley walls and divides. Test borings sunk by the Department of Transportation along the route of Interstate 88 also yield useful stratigraphic information, although seldom below 100 feet (33 m) of the surface. From this information a regionally uniform association between several stratigraphic units can be recognized. A gravel cap usually mantles a thick accumulation of silt and sand that in turn overlies coarse gravel. The upper gravel varies in thickness from 10 to 50 feet (3 to 15 m) (occasionally up to 100 feet, 30 m) and is part of either fluvial gravel, outwash, deltaic outwash terraces or kames. The finer sediment below consists of silt and clay, silt, silt and sand, quicksand or sand. It is not uncommon for this unit to reach thicknesses between 100 and 200 feet (30 and 61 m). In several wells, 300 to 400 feet (91 to 122 m) of quicksand is reported! This thick, fine grained unit is interpreted to be glaciolacustrine in origin. Where the overlying gravel is lacking, a broad lacustrine plain forms the valley floor and is veneered by overbank silts. The underlying lower gravel unit usually has a very coarse texture and is often associated with black sand. Because this gravel serves as a very

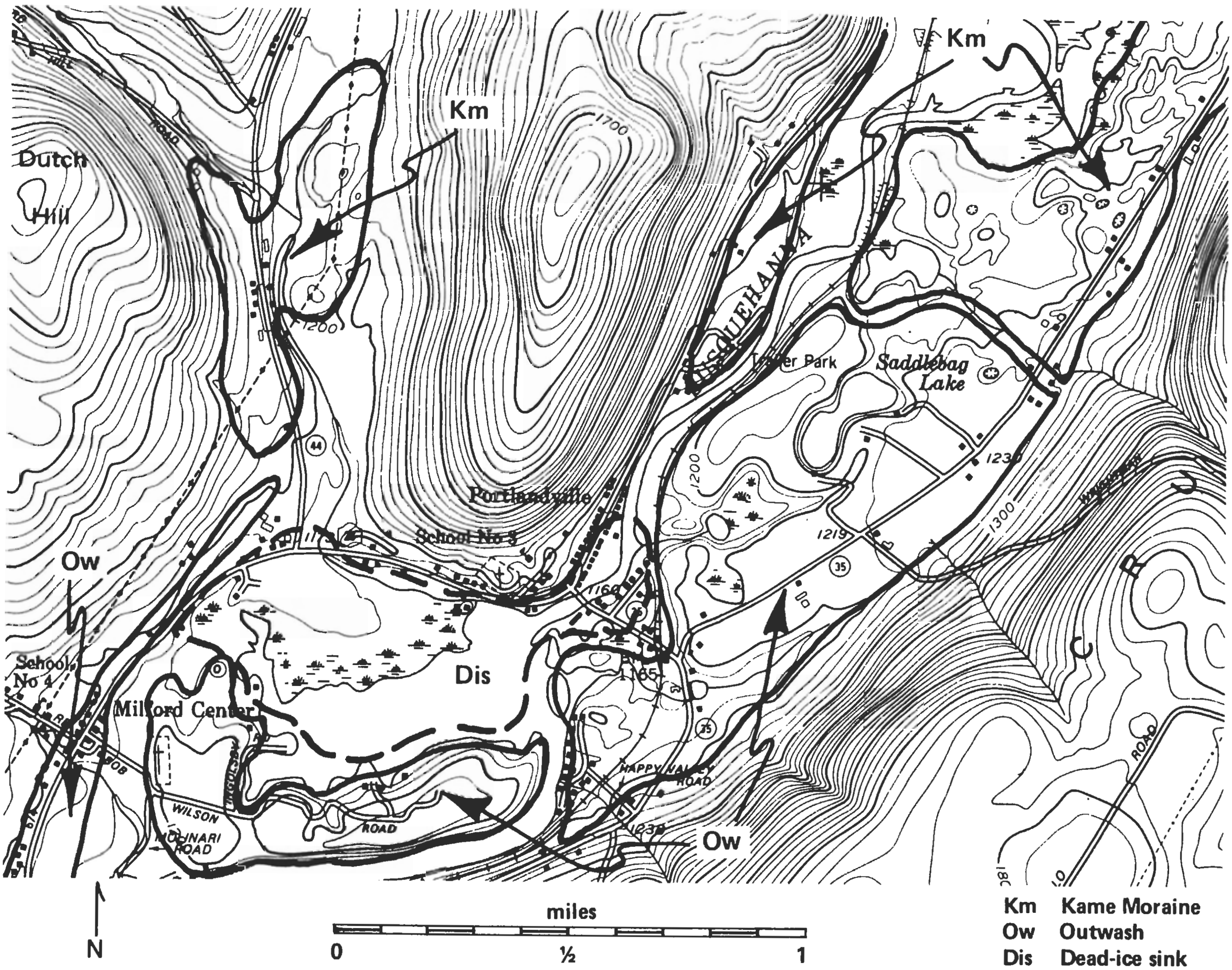


Figure 9. Kame-moraine, pitted outwash and dead-ice sink, Portlandville, Milford quadrangle.

substantial aquifer, it is seldom penetrated. Thus, wells that contain this three-fold stratigraphic sequence seldom encounter bedrock.

Application of Stratigraphic Data for Interpretation of Glacial Events

The ice lobe model for deglaciation and accompanying ice-contact lakes is supported by the common stratigraphic sequence described above. The fact that this sequence is so widely observed throughout the upper Susquehanna indicates that this deglacial model has regional application.

In several critical locations the well stratigraphy is useful as a criterion for distinguishing between recessional and readvance moraines and thereby is very helpful for interpretations of deglacial events. Two examples are selected to illustrate this. The first is the Oneonta kame-moraine located immediately east of Oneonta along the main Susquehanna valley and the second is the Cassville-Cooperstown (Kame) Moraine 2 miles (3.2 km) south of Cooperstown near the Susquehanna headwaters.

The topographic expression of the moraines generally is similar except that the Cassville-Cooperstown Moraine is a more prominent landform (wider, greater internal relief and laterally more distinct) and breached less by the modern Susquehanna. They both are associated with thick, extensively developed outwash that yield downvalley to lacustrine landforms and stratigraphy.

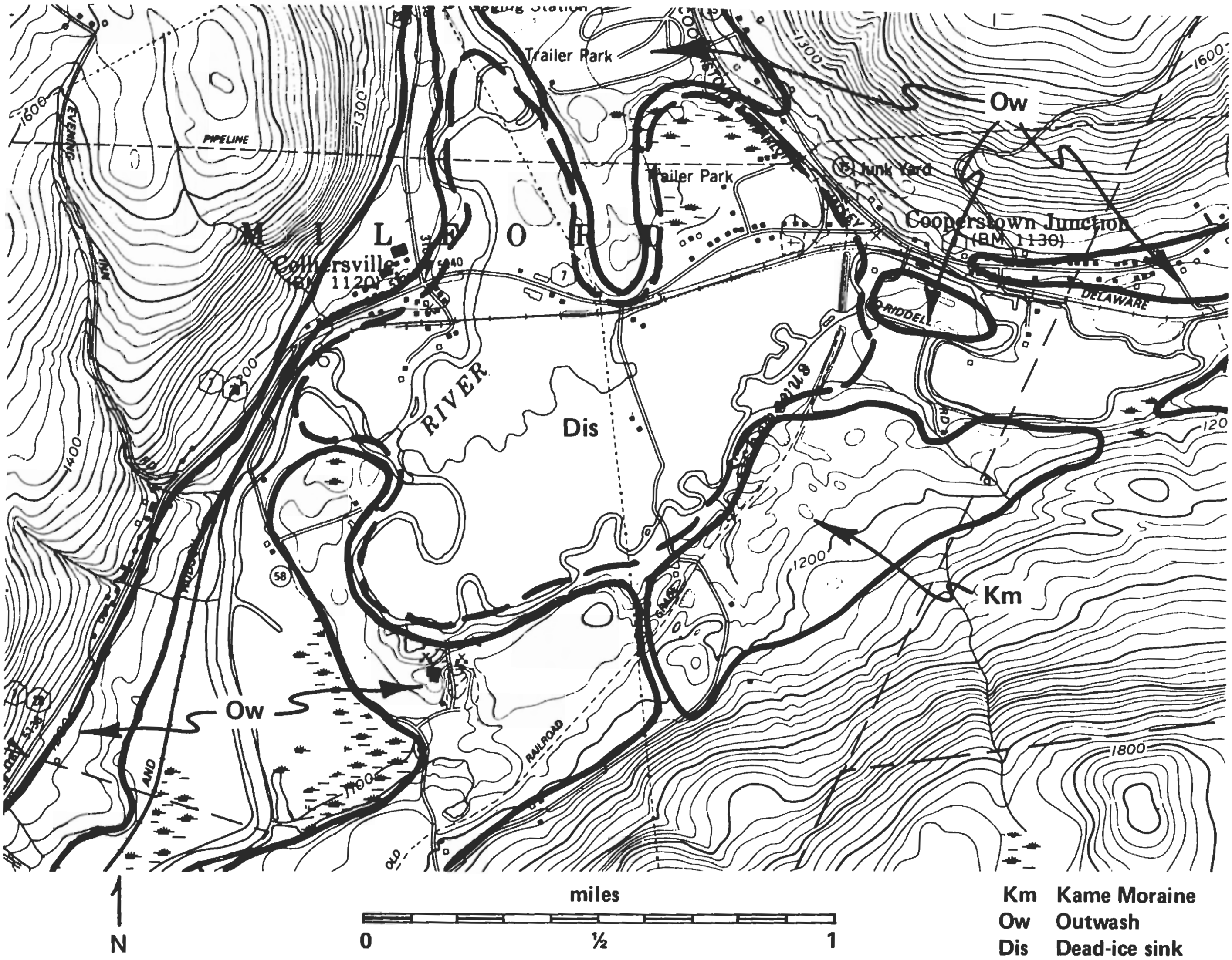


Figure 10. Kame-moraine, deltaic outwash terrace, and dead-ice sink, Emmons to Colliersville, Oneonta and West Davenport quadrangles.

The stratigraphy beneath the moraines is illustrated by longitudinal profiles in Fig. 13 that utilize well data from Randall (1972) and personal communication with local residents. Two distinctly different stratigraphic units are recognized in both locations; sand and gravel of the moraines and outwash and finer sediments beneath the lake plains. The sand and gravel within the Oneonta Moraine are continuous with depth and interfinger down-valley with lacustrine silt. This is interpreted to mean synchronous deposition of both units and that the moraine and its associated deltaic outwash were deposited when Glacial Lake Otego still occupied the valley. Because a stadial readvance to this position would not account for the interfingering of the two stratigraphic units, the Oneonta Kame-Moraine is interpreted to be recessional in origin.

Figure 13B illustrates the valley stratigraphy associated with the Cassville-Cooperstown Moraine. Outwash sand and gravel in close proximity to the moraine overlie more than 200 feet (61 m) of silt and clay that is continuous with silts beneath the lacustrine plain down-valley. The uniformly thick lacustrine sediments beneath the outwash are interpreted to extend northward beneath the adjacent moraine. Provided the lacustrine sediments are continuous under the moraine, a readvance to this position across earlier lake deposits occurred. Therefore, this moraine is interpreted to mark the limit of readvance and is an end moraine. Because

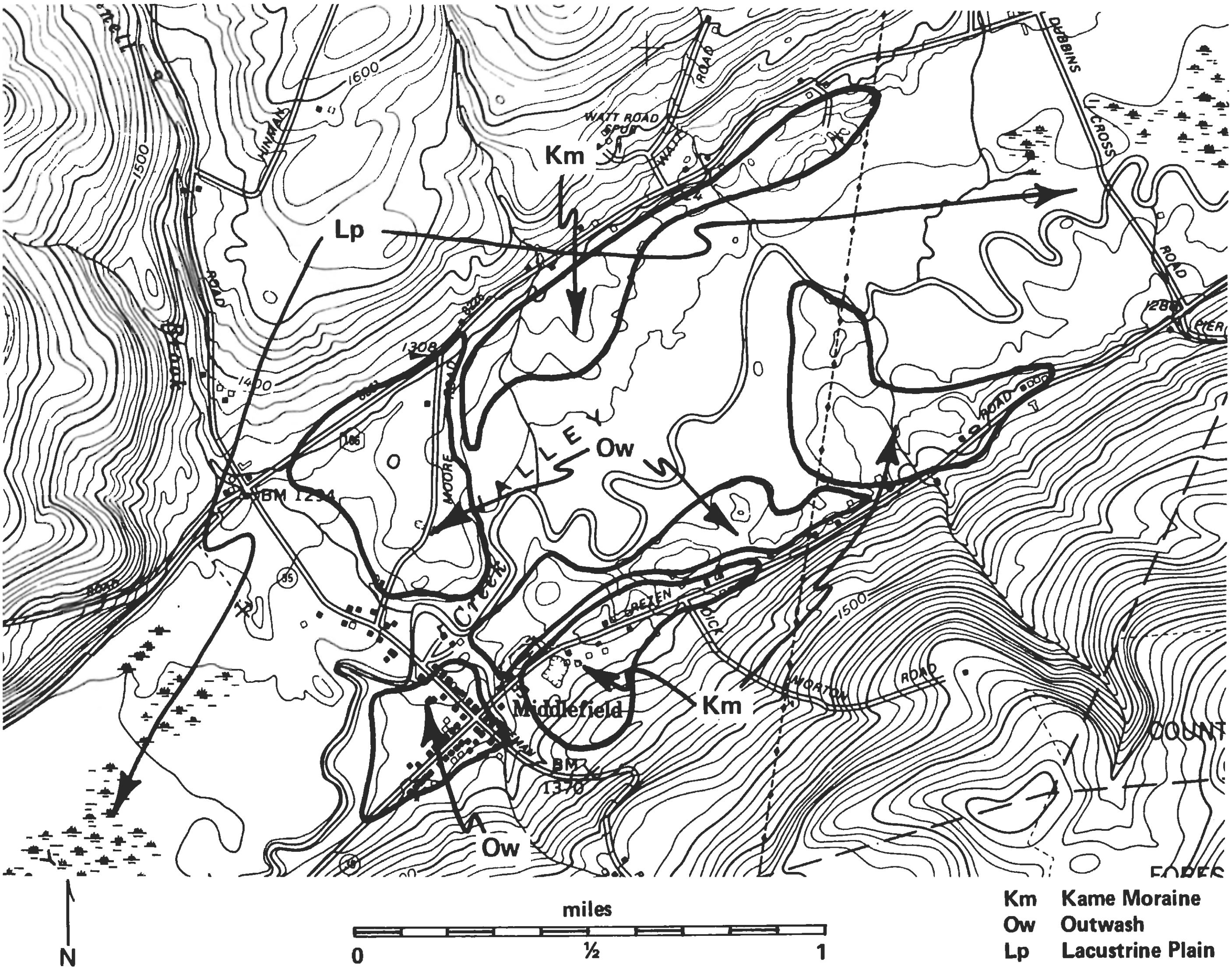


Figure 11. Lacustrine plain and small moraine-outwash complex, Middlefield, Westford quadrangle.

the distal extent of outwash is graded to the projected level of Glacial Lake Milford (Fleisher, 1977b) that was dammed by a kame-moraine at Portlandville at an elevation of 1220 to 1230 feet, it is suggested that the lake was still in existence when readvance occurred.

GLACIAL LAKES IN THE UPPER SUSQUEHANNA DRAINAGE BASIN

Landforms and stratigraphy indicate that the upper Susquehanna and many of its tributaries contained ice-contact and proglacial lakes. The characteristic pattern consists of a moraine dam and associated deltaic outwash, ice-contact outwash behind the moraine, hanging deltas, thick sequences of lake sediments and broad lacustrine plains across which the modern drainage flows. In each case strandline features are graded to the average moraine elevation and each moraine is successively higher up-valley.

The locations and suggested names of four moraine-dammed lakes at the head of the Susquehanna drainage basin are as follows:

1. Glacial Lake Milford (1220-1230 feet) – along the main Susquehanna Valley in the vicinity of Milford, dammed by the moraine at Portlandville.
2. Glacial Lake Middlefield (1260-1280 feet) – dammed by the Cassville-Cooperstown Moraine in Cherry Creek Valley.

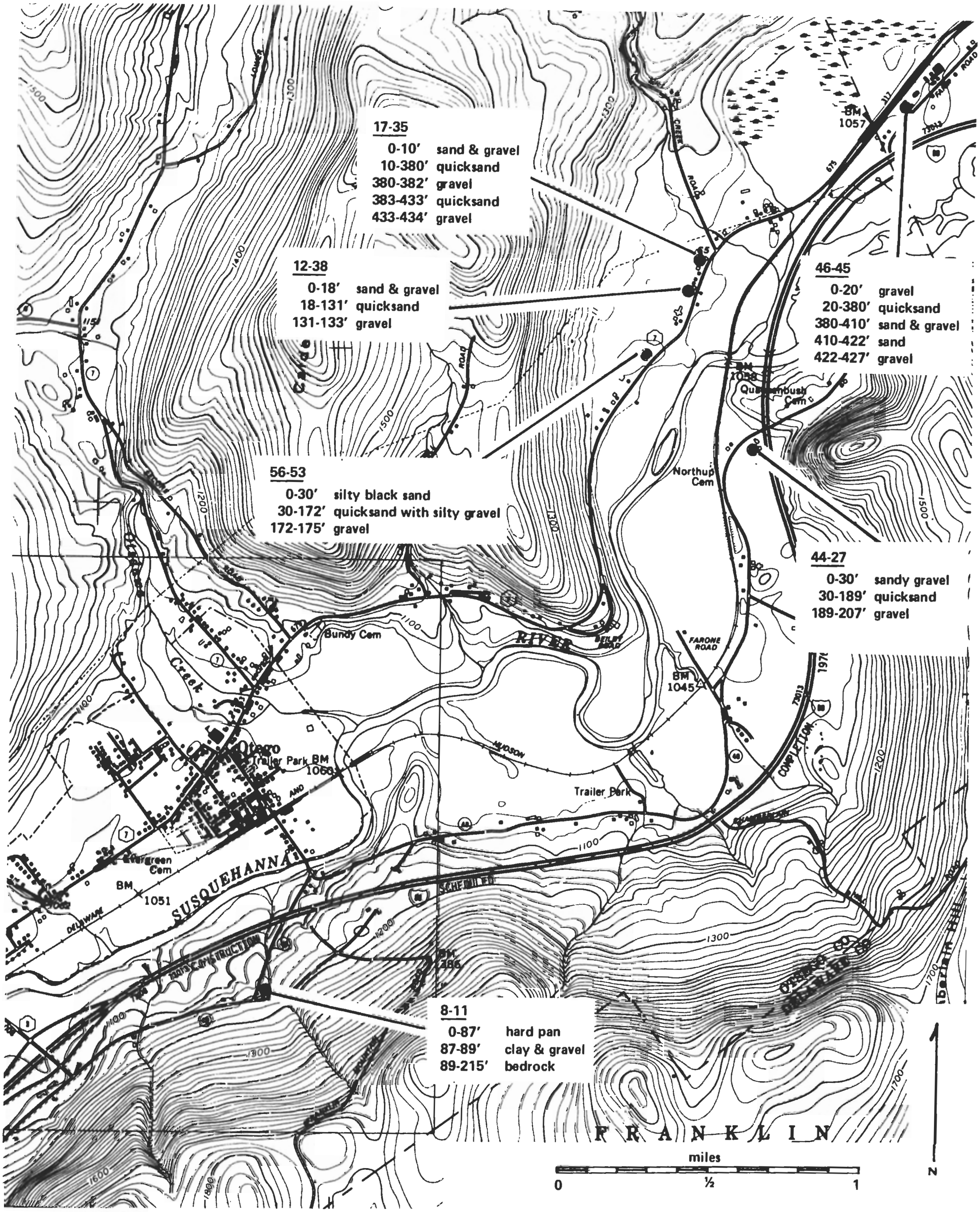


Figure 12. Sample well log map display.

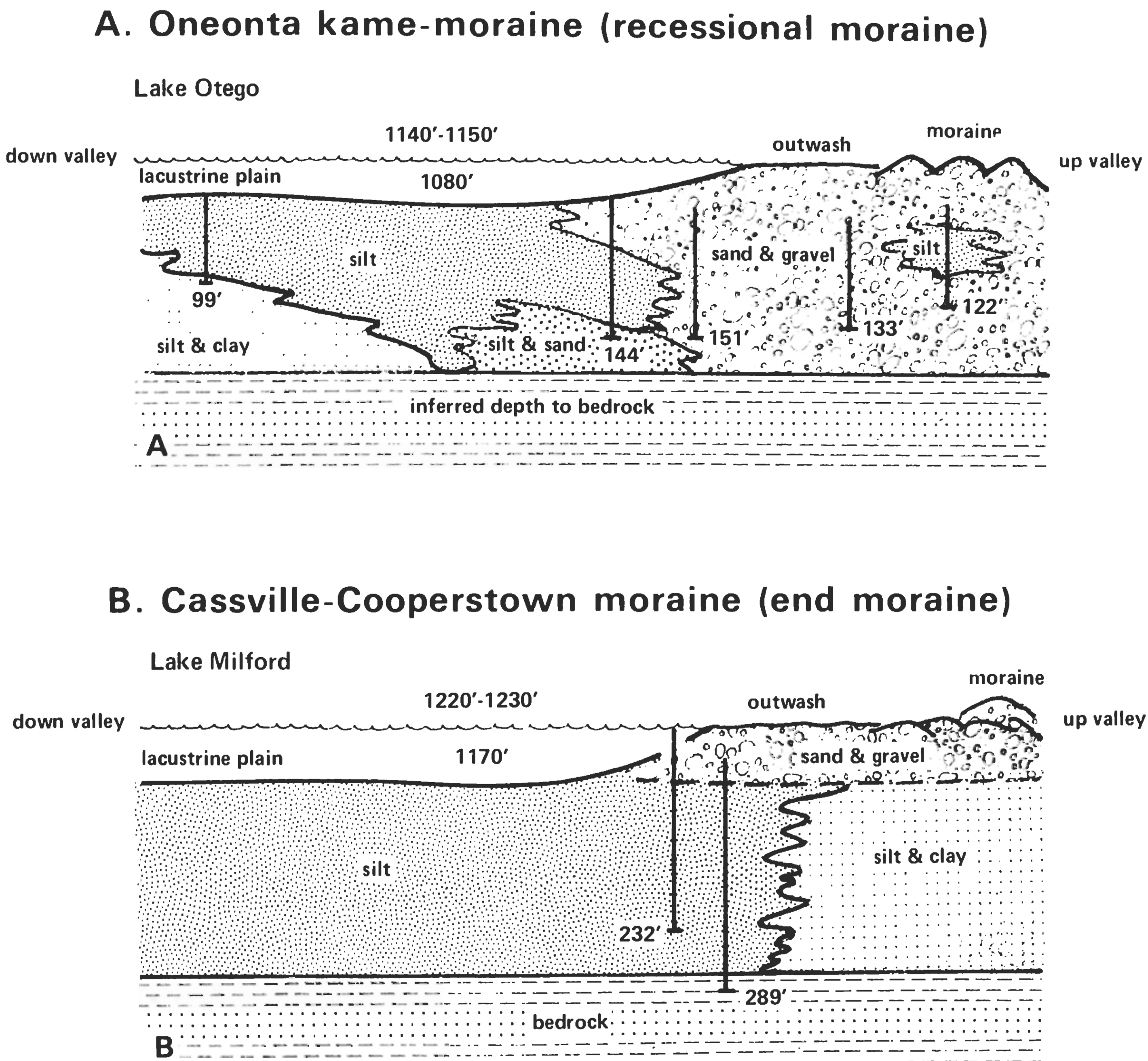


Figure 13. Stratigraphic distinction between recessional and end moraine.
Part A – Outwash sand and gravel interfinger with lake silt.
Part B – Outwash sand and gravel overlie lake sediments.

3. Glacial Lake Oaksville (1300-1320 feet) – within Oaks Creek Valley and confined by the Cassville-Cooperstown Moraine; includes the current position of Canadarago Lake.

4. Glacial Lake Cooperstown (1250-1260 feet) – held behind the Cassville-Cooperstown Moraine; a higher lake level for Otsego Lake at Cooperstown.

Comparable lakes existed in other major valleys. Thick deposits of sand, silt, and clay behind correlative moraines are found in the Chenango, Unadilla and Wharton Creek valleys (Randall, 1972). Various types of deltaic landforms also indicate large bodies of water occupied these valleys. Similar evidence for smaller lakes associated with stagnant ice masses of non-through valleys is found in the Butternut, Otego Creek, Schenevus

Creek and Charlotte Creek valleys. Ample field documentation exists for a rather complete chain of lakes in the upper Susquehanna during late glacial and early post-glacial time.

RECOGNITION AND CORRELATION OF ICE MARGINAL POSITIONS

The glacial history of the eastern Susquehanna region is recorded primarily by the landforms and stratigraphy developed along ice-marginal positions. Little is known about the advance of the ice across the dissected Appalachian Plateau but much can be interpreted about deglaciation during phases of oscillatory retreat and readvance. Topographic criteria are used for the recognition of ice-marginal positions (Fleisher, 1984) and stratigraphic differences distinguish recessional moraines from end moraines. The criteria proposed for the recognition of ice-marginal positions fall into two categories, which are:

A. Valley floor facies
   1. kame-moraines and associated outwash
   2. thick, pitted outwash in the form of partially dissected valley trains
   3. dead-ice sinks

B. Valley wall and divide facies
   1. kames, kame fields and/or ablation moraines (not associated with extensive outwash) on valley walls and divides between valleys containing kame moraines
   2. upland meltwater channels (currently occupied by under fit streams)

Plate I illustrates the correlation of ice-marginal positions in a general northwest to southeast trend across the upper Susquehanna drainage. In general, it appears that the ice remained active and retreated by normal backwasting, except where various local topographic controls interfered with active flow causing local stagnation and downwasting.

Five major ice-margin positions have been recognized and mapped (see Plate I). The location of the Wells Bridge margin is defined by an ice-lobe position near Rockdale in the Unadilla River valley. From there it crosses the divide and continues on a southeastward trend to Wells Bridge where it effectively dammed the Susquehanna during ice retreat creating Glacial Lake Otego. From Wells Bridge it crosses the divide and continues into the Ouleout drainage at Franklin.

The Oneonta margin is best defined at Oneonta by a dissected kame-moraine that extends into the mouth of Charlotte Creek valley at West Davenport. The moraine loops across the Susquehanna where it is stratigraphically defined as a recessional moraine (cf. Figure 13). From here the ice margin trend is controlled by the relief of the Charlotte Creek valley where a variety of ice-contact stratified drift deposits and landforms mark its location.

Along the southernmost part of the dual New Berlin margin are smaller kame-moraines and pitted outwash surfaces that typify ice-lobe retreat perpendicular to the main drainage from the Unadilla valley southeastward to the Susquehanna at Portlandville. At Portlandville thinning ice stagnated due to its inability to maintain active flow through the incised valley meander as shown in Figure 9. Buried ice formed a dead-ice sink here, as well as farther down-valley at the mouth of Schenevus Creek. From here the ice-margin was topographically controlled along the northeasterly trend of the Schenevus Creek valley.

Along this trend the effects of topographic control (Figure 2) can best be seen. Schenevus Creek follows a valley of considerable relief (900 + feet, 274 m) that is relatively narrow and confined. During glacier retreat this valley served to limit the position of the paralleling ice margin, thereby creating highly variable depositional conditions of the ice-contact environment along the trend of a single valley. As a result, the ice margin is representd by various combinations of kame-moraines, massive outwash, dead-ice sinks, ablation moraines, evidence of former proglacial and ice-contact lakes, and glaciofluvial deposits are distributed along this valley.

Secondary, and possibly tertiary, margins parallel to the New Berlin position converge to the southeast, almost meeting beyond the Susquehanna Valley. Considering the short distance that separates them, their general parallel trends, and the lack of a clearly defined eastward correlation, additional field work may reveal these to be subphases of a single ice-margin position.

One of the most prominent and clearly marked margins is represented by the Cassville-Cooperstown moraine (Krall, 1977), from which it takes its name. This is the only margin which stratigraphically and topographically can be demonstrated as a position to which the ice readvanced, as opposed to retreated.

The final ice-margin occurs in the headwaters of Cherry Valley, where it can be mapped across the valley in lobate fashion. Although massive constructional landforms are for the most part lacking along the valley floor, it is well represented by hummocky terrain ascending the valley walls. To the east, the margin can be traced along the base of the Appalachian Plateau escarpment where it appears to join with the Middleburg readvance margin.

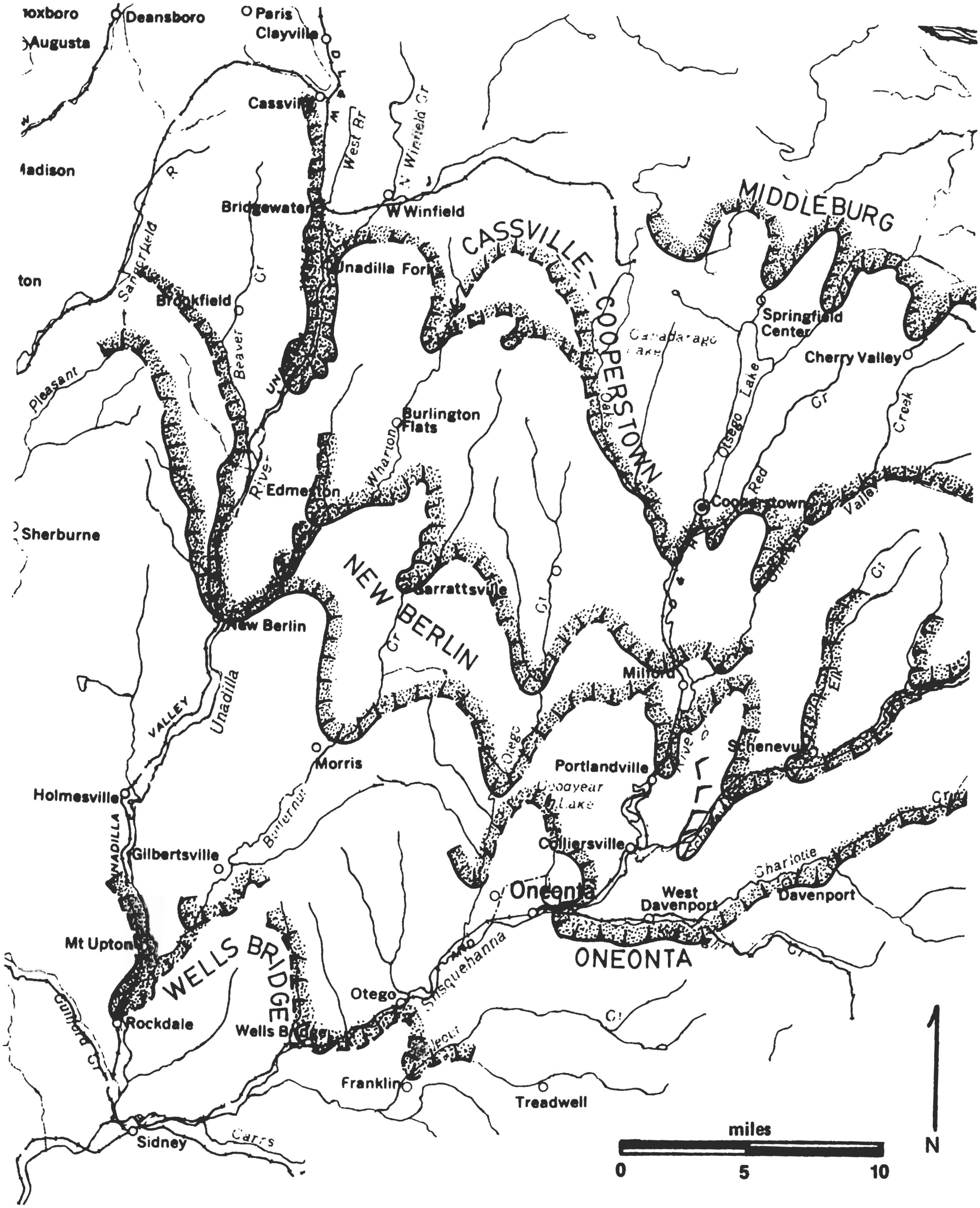


Plate I Ice Margin Positions, Upper Susquehanna Drainage basin.

TABLE I CORRELATION CHART

Susquehanna Drainage (Fleisher, 1983)	Schoharie Drainage (LaFleur, 1969)	Catskill Mountains (Cadwell, 1983)
Middleburg margin	Middleburg readvance	Middleburg readvance
Cassville-Cooperstown margin	Middleburg readvance	Middleburg readvance
New Berlin margin	Prattsville readvance	North Blenheim-Haines Falls Grand Gorge Lake Phase
Oneonta margin	Tannersville readvance	Wagon Wheel Gap margin
Wells Bridge margin		

DEGLACIAL CHRONOLOGY AND REGIONAL CORRELATION

By all indications the upper Susquehanna drainage basin is mantled by a single drift of a common stade. Through correlation with stratigraphic units from which radiocarbon dates have been obtained within the drainage basin and beyond, it is possible to infer the deglacial history.

A date of 16,650 ± 1,800 yrs BP was reported by Cadwell (1972) for a kettlehole bog in a kame-moraine in the Chenango Forks moraine near the village of Chenango Forks. This generally is consistent with Flint's suggestion (1971) that deglaciation began 17,000 to 18,000 yrs BP on the Appalachian Plateau. An additional date of 13,860 ± 800 yrs BP was obtained from terrace gravel deposited within the breached Wells Bridge moraine along the Wells Bridge ice margin. This date establishes a minimum age for that ice margin and Glacial Lake Otego which was impounded behind the moraine. Glacial Lake Otego sedimentation is estimated to have required a minimum of 840 years to account for the exceptionally thick silts and sands that accumulated during its existence.

Melia (1975) sampled four separate bogs within the study area (at Wells Bridge, Laurens, Milford Center, and Maryland) and correlated them with the established pollen stratigraphy for New York State. Only the Maryland Site (near the mouth of Schenevus Creek and within the New Berlin ice margin) was sufficiently deep to provide a relatively complete early post-glacial record that suggests an age of about 14,000 yrs BP (Sirkin, 1967; Miller, 1973). The upper Susquehanna was ice-free by that time.

It appears as though this area was uncovered by deglaciation from the Chenango Forks area (16,650 ± 1,800 yrs BP) to the New Berlin margin (minimum age 14,000 yrs BP) and progressively to the northeast. A readvance to the Cassville-Cooperstown margin at a still undetermined time is estimated to be older than 12,000 to 13,000 yrs BP. The lack of definitive evidence within the Susquehanna drainage basin precludes an interpretation of retreat or readvance for the Middleburg margin at this time, although a recessional position to the Cassville-Cooperstown readvance is favored.

The correlation chart of Table I lists the ice margins recognized within the upper Susquehanna drainage basin and their proposed correlation eastward with established nomenclature of the Schoharie Valley and the newly proposed positions suggested by Cadwell (1983) for the Catskill Mountains.

SUMMARY

1. The deglacial ice margin orientation relative to valley trend influenced the formation of valley ice lobes and the subsequent distribution of ice-marginal landforms.
2. The high-relief terrain controlled flow patterns at the ice margin causing local areas of stagnation and resulted in the isolation of large detached ice blocks. The downwasting that followed created environments of dead-ice deposition called dead-ice sinks.
3. Ice-contact and proglacial lakes were common throughout the upper Susquehanna drainage basin. Glacial Lake Otego is a good example of the resulting landforms and stratigraphy.

4. The stratigraphy of water well logs has been applied to distinguish the recessional Oneonta moraine from the readvance represented by the Cassville-Cooperstown moraine.

5. Criteria for recognition of ice-marginal positions include a valley facies of stratified drift landforms (kame-moraines and associated outwash, pitted outwash and dead-ice sinks) and less distinct divide facies features (kame fields and meltwater channels).

6. The upper Susquehanna was still ice covered circa 16,650 ± 1,800 yrs BP when retreat from the Chenango Forks area to the southwest occurred. Retreat from the Wells Bridge margin could not have occurred more recently than about 15,000 yrs BP. The New Berlin margin is estimated at about 14,000 yrs BP. The upper Susquehanna was ice free prior to 12,000 to 13,000 yrs BP, a maximum age for the Cherry Valley margin.

7. Retreat was from a series of correlative ice margins oriented in a general northwest-southeast trend, with ice lobes occupying all major valleys.

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PLEISTOCENE GEOLOGY OF THE WESTERN MOHAWK VALLEY, NEW YORK

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ABSTRACT

Late Wisconsinan thinning of the continental ice sheet over the Adirondack Mountains and adjacent Appalachian and Tug Hill uplands channeled lobate ice tongues into intervening lowlands. Outcrops along West Canada Creek and its tributaries expose tills of several provenances interbedded with sediments deposited in meltwater lakes that were impounded between the Ontario and Mohawk glacier lobes. The ponded waters drained at various times across the ice, south into the through-valley system of the plateau, or eastward along the ice margin to the Hudson Valley.

Erosional contacts and intercalation of fluvial gravels suggest that the glacier dam presented by Mohawk Valley ice was lowered or removed at least once during Late Wisconsinan deglaciation. An episode of fluvial deposition is inferred to have occurred between emplacement of West Canada and Hawthorne tills. In the absence of absolute dating criteria, this interval is tentatively ascribed to the Erie Interstadial when the Great Lakes drained east by the Mohawk River. The Indian Castle limit described by Fullerton (1971) delineates parts of the margins of two till sheets here distinguished as the Norway and Holland Patent tills. Fluvial transport from an Ontario lobe ice margin occurred during deposition of the Rome till.

While responding to the same climatic impulses, the Mohawk and Ontario sublobes attained their maximum extents asynchronously, whether because of different distances from accumulation areas, shifting outflow centers, or changing contribution by obstructed throughflow across the southeastern Adirondacks.

INTRODUCTION

The late glacial record preserved in the western Mohawk Valley is among the most complete and detailed in central New York. Exposures along West Canada Creek and its tributaries contain particularly valuable evidence of the complex changes in ice flow and drift provenance that resulted during Late Wisconsinan deglaciation (Fig. 1).

Deglacial thinning of ice over the Adirondack Mountains and later over the Tug Hill Uplands led to differentiation of the ice margin into distinct lobes channeled into the adjacent lowlands. The late glacial stratigraphy in the western Mohawk Valley records the interplay of ice from the north and west (Black River and Oneida sublobes of the Ontario lobe respectively), from the east (Mohawk sublobe of the Champlain-Hudson lobe) and from partly obstructed Adirondack throughflow. The principal evidence involves the interbedding of till sheets derived from diverse provenances with sediments of glacially impounded meltwaters. In the Mohawk Valley, too, must be sought the evidence of eastward drainage episodes long inferred from relationships in the Great Lakes Basins.

Although the western Mohawk Valley figured in investigations of Quaternary geology from the earliest arrival of the Glacial Theory in North America, only recently has it begun to receive the detailed attention that its stratigraphic record deserves.

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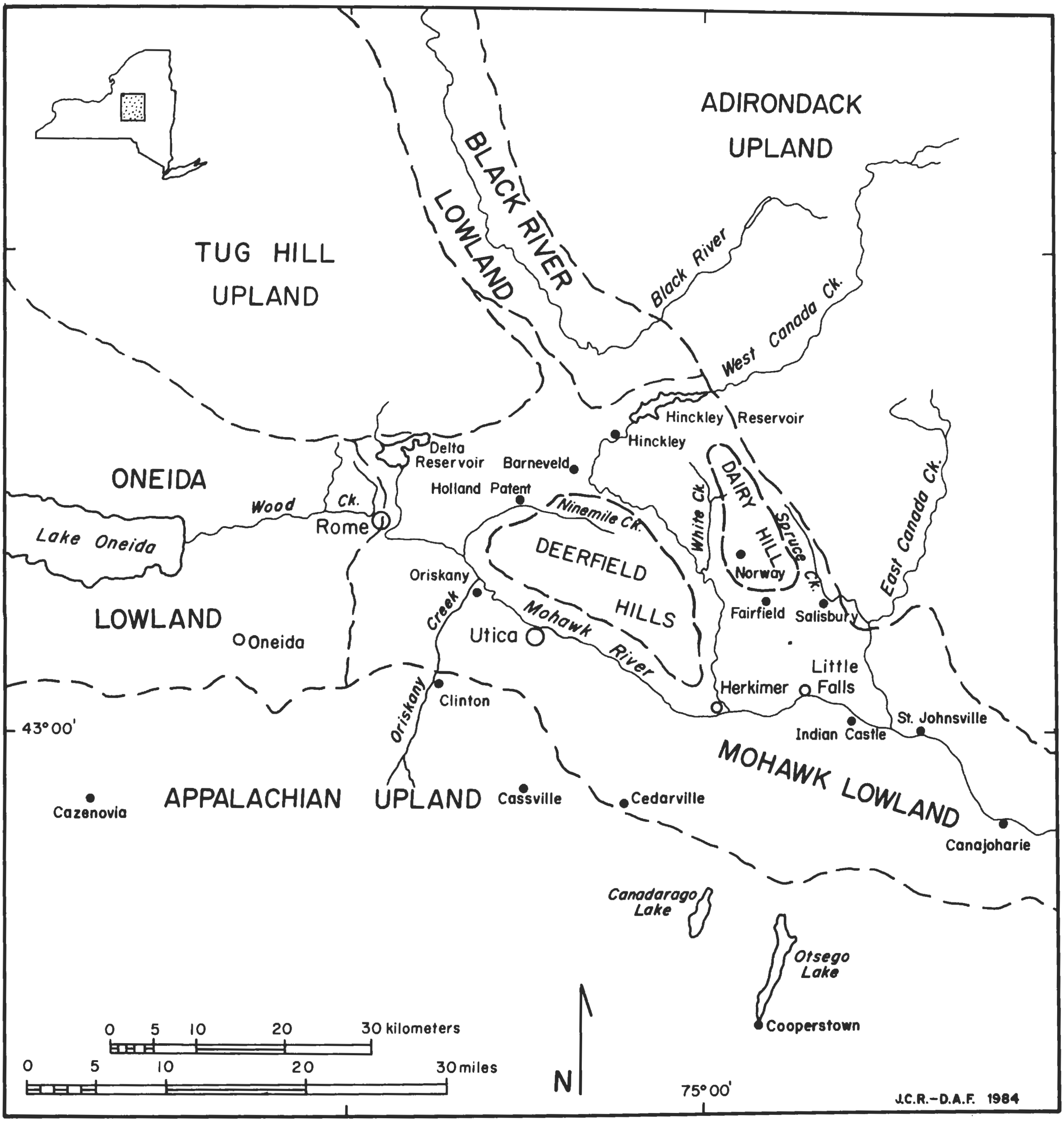

Figure 1 Physiographic relationships and location of study area.

PREVIOUS INVESTIGATION

In the Survey of the Third Geological District, Lardner Vanuxem (1842) recognized scratches on polished rock in locations near Utica as being of probable glacial origin. They were too widely distributed, at such diverse elevations and with such orientation as to argue against origin as iceberg scratches. On the basis of striae, too, J.D. Dana (1863) inferred the past existence of glacier flow constrained along the axis of the Mohawk Valley subsequent to regional north-south flow of continental ice. T.C. Chamberlain (1883) concluded that "massive ice currents swept around the Adirondacks and entered the Mohawk Valley at either extremity, while a feebler current at the height of glaciation, probably passed over the Adirondacks and gave to the whole a southerly trend." Additional data led W.J. Miller (1909b) to adopt a similar view.

A.P. Brigham (1893) provided perceptive analysis of the topography and glacial deposits of the Mohawk Valley, supplemented later (1929) by more detailed coverage of quadrangles in the lower Mohawk Valley. H.L. Fairchild (1912) sketched, in broad-brush outline, the regional history of glacial impoundment and meltwater drainage in the Black River and Mohawk Valleys, but added little in the way of detailed stratigraphic or morphologic documentation. Surficial deposits received only incidental mention in geologic reports on adjacent quadrangles by Cushing (1905), Miller (1909a), Kay (1953), Dale (1953) and Nelson (1968).

Glacial geological studies by Street (1966), Wright (1972), Jordan (1978), Chambers (1978) and Muller (1978) in the adjacent Tug Hill Uplands provide descriptive detail and interpret recessional history, but develop minimal supporting stratigraphic documentation. Fullerton (1971, 1980) was the first to provide detailed descriptions of western Mohawk stratigraphic units and to propose regional correlations, defining the Indian Castle Moraine as partly correlative with the Valley Heads Moraine complex of the Finger Lakes region. Krall (1972, 1977) recognized till of Mohawk provenance as far west as Oriskany Creek south of Clinton. On the plateau south of the Mohawk Valley he described and traced a moraine which he named the Cassville-Cooperstown Moraine, tentatively relating it to the Wagon Wheel Gap substage of Rich (1935) in the Catskills and the Rosendale readvance in the Wallkill Valley (Connally and Sirkin, 1973).

This report is a statement of progress in investigations undertaken in the western Mohawk Valley by a research group from the Geology Department at Syracuse University. In summarizing findings of our associates mapping in contiguous quadrangles, we acknowledge with appreciation our indebtedness to M.D. Antonetti (1982), J.M. Loewy (1983), C.A. Lykens (1984), R.J. Foresti (1984) and to G.R. Flick for freely sharing the results of their work while it was still in progress.

REGIONAL SETTING

The western Mohawk Valley region includes parts of the Black River, Oneida and Mohawk lowlands, and of the Tug Hill, Deerfield Hills and Adirondack uplands (Fig. 1).

Preglacial and glacial erosion along the strike of nonresistant Cambro-Ordovician strata dipping off the Adirondack dome produced the Mohawk and Black River Lowlands. Although developed on stratigraphically higher beds, the Oneida Lowland is continuous with the Mohawk Lowland across the low divide near Rome. The Oneida and Mohawk Lowlands isolate the Tug Hill and Deerfield Hills from the Appalachian Plateau to which they are physically related.

During deglaciation, emergence of uplands was the primary control on development of glacial lobes and ice deployment. The stratigraphic record which we describe documents this aspect of Late Wisconsinan history. Four distinct ice-flow trajectories developed during deglaciation.

1. The Mohawk sublobe spread west as an offshoot of the Hudson-Champlain lobe, confluent at times with ice from eastern Adirondack sources.
2. The Oneida sublobe spread east from the Oneida Lowlands around the southern flank of the Tug Hill as an offshoot of the Ontario lobe.
3. The Black River sublobe spread south-southeast between the Tug Hill and the western Adirondacks as an offshoot of the Ontario lobe.
4. Outflow across the southern Adirondacks, though considerably obstructed, contributed variously both to the Mohawk and Black River sublobes, and acted alone in depositing the Kuyahoora till.

Interplay of the several ice lobes and of the meltwater lakes which they impounded is recorded in Quaternary drift exposures on West Canada Creek and its tributaries. Because successive glacial advances reached their maximum extents in proglacial lakes ponded in the western Mohawk Valley, the preservation of glacial stratigraphy is unusually complete. Individual sections expose more than 60 m of continuous sedimentary record and contain deposits from as many as four glacial advances. It is our objective in this paper to formulate a correlation of stratigraphic units, to interpret the his-

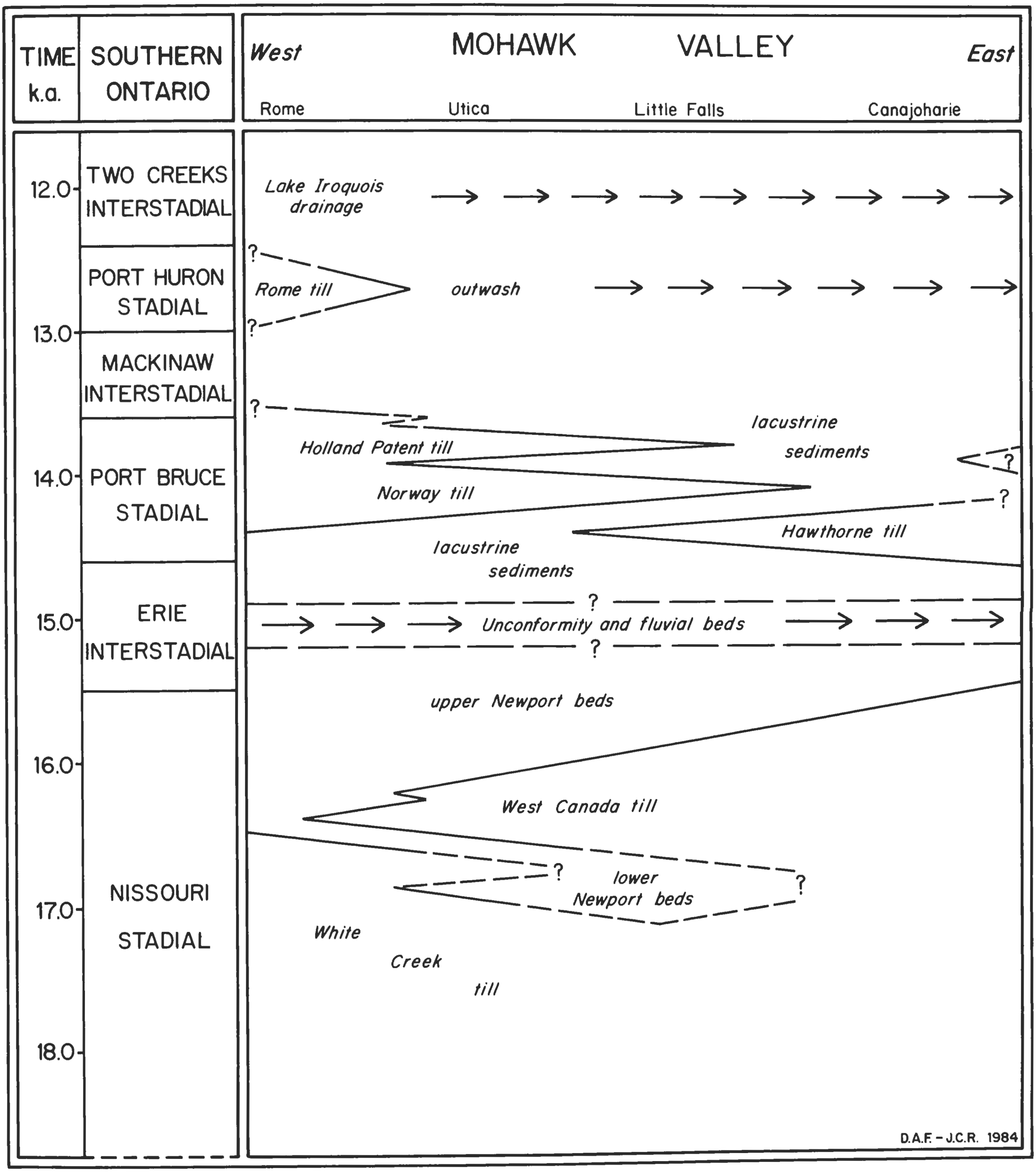

Figure 2 Time-distance diagram showing lithologic relationships of Late Wisconsinan stratigraphic units.

tory which these units record and to suggest a tentative chronology and correlation with units and events in adjacent areas.

In the discussion that follows, the word "till" is used with full genetic implication of deposition by glacial ice. At practical mapping scales, resedimented materials intimately associated with processes at the ice margin are included as well, in the sense of a till sheet, but the immediate physical presence of the ice margin at time of deposition is implicit. The nongenetic term "diamicton" is used in description and where the origin of the unit is in question. In stratigraphic description, where scale and sedimentologic relationships permit, associated resedimented materials are distinguished by an appropriate term such as flow-till.

A generalized time-distance diagram showing names of important stratigraphic units for the western Mohawk Valley (Fig. 2) is tentatively correlated with chronostratigraphic units of Dreimanis and Karrow (1972) for the Erie and Ontario Basins. This nomenclature presently affords an appropriate time-stratigraphic framework for this study. The proposed correlations are based, however, on inferred regional relationships without independent chronologic control. As yet, no material suitable for radiocarbon dating has been found. Furthermore, the nature of Late Wisconsinan environments within the region makes it unlikely that datable material exists in the portion of the stratigraphic record with which we are concerned. For this reason, one of us (Ridge and others, 1984; Ridge, in prep.) is investigating secular variation of paleomagnetic declination in lake sediments of the region to test correlation with curves developed for the Genesee Valley of western New York (Brennan and Gruendike, 1981; Brennan and others, 1984; Braun and others, 1984).

A time-distance diagram (Fig. 2) has been formulated to show schematically the duration and extent of deposition of seven diamict units and the associated stratified deposits, all of which are of Late Wisconsinan age. Flow directions for ice that deposited the diamictons are inferred on the basis of till fabrics, striae, end moraine configuration, till sheet margins and provenance data that include pebble counts, total carbonate content, calcite/dolomite ratios, heavy mineral counts and geochemical analyses (Antonetti, 1982; Franzi, 1984a; 1984b). On this evidence, the several till sheets are identified as representing Oneida, Mohawk or Black River provenance. Sediments deposited when glacier cover was lacking represent either lacustrine impoundment in the Mohawk Valley or intervals of outwash deposition and subaerial drainage.

Significant unconformities or gaps in the glacial record, representing intervals of erosion and free drainage eastward through the Mohawk Valley, are interpreted as representing the Erie and Mackinaw Interstadials. Recognition of these hiatuses supports previous interpretation of eastward drainage of the Great Lakes (Hough, 1966; Dreimanis, 1969; Moerner and Dreimanis, 1973).

GLACIAL HISTORY

Nissouri Stadial

White Creek Till

The oldest stratigraphic unit recognized in the western Mohawk Valley is here identified as the White Creek till, a typically medium to dark blue-gray, calcareous clast-rich diamicton. Where the basal contact is not concealed, this unit lies directly on bedrock. The type area of the White Creek till is along White Creek from the confluence of White Creek and Factory Brook south to West Canada Creek.

The combination of dominant limestone, shale and dolostone with many metamorphic clasts and subordinate sandstone fragments suggests a northerly provenance. This orientation parallels pebble fabrics in the White Creek till as well as regional streamlining in the central Tug Hill (Street, 1966; Chambers, 1978). The ice sheet was apparently thick enough to override all topographic features of the western Mohawk Valley drainage basin without significant deflection. The White Creek till probably represents deposition during advance and retreat of ice which reached its Woodfordian maximum extent far to the south.

Division of the ice sheet into distinct lobes is clearly apparent in later stages of deposition of the White Creek till. In stream banks along the east side of Woodchuck Hill, 2.4 km northeast of the confluence of White and West Canada Creeks, the White Creek till is interbedded with lacustrine sediment (Section 5, Fig. 3). Striations, pebble fabrics and grooving on the till surface trend S70W, suggesting ice flow from the Adirondacks which passed southeast of Dairy Hill into the West Canada Creek Valley. Striations and pebble fabrics to the northwest on White Creek, in other exposures of White Creek till, reflect Adirondack ice flow from the north around the northern end of Dairy Hill into West Canada Valley. Irregularity in thickness and elevation on the upper surface of the till suggests buried morainal topography of Adirondack provenance at the margin of ice that flowed north of the Deerfield Hills.

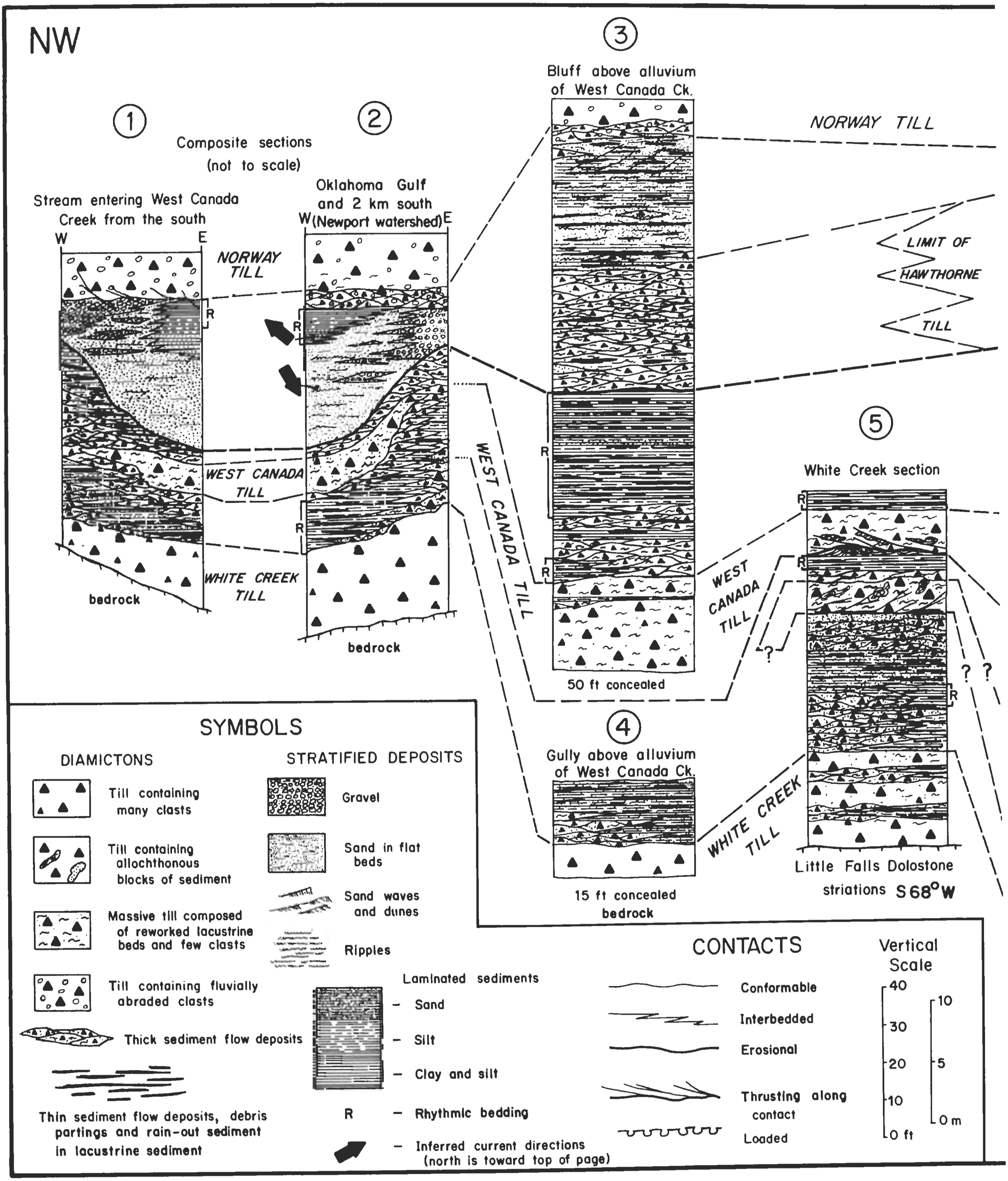

Figure 3A Representative stratigraphic sections showing inferred correlations in West Canada Creek area.

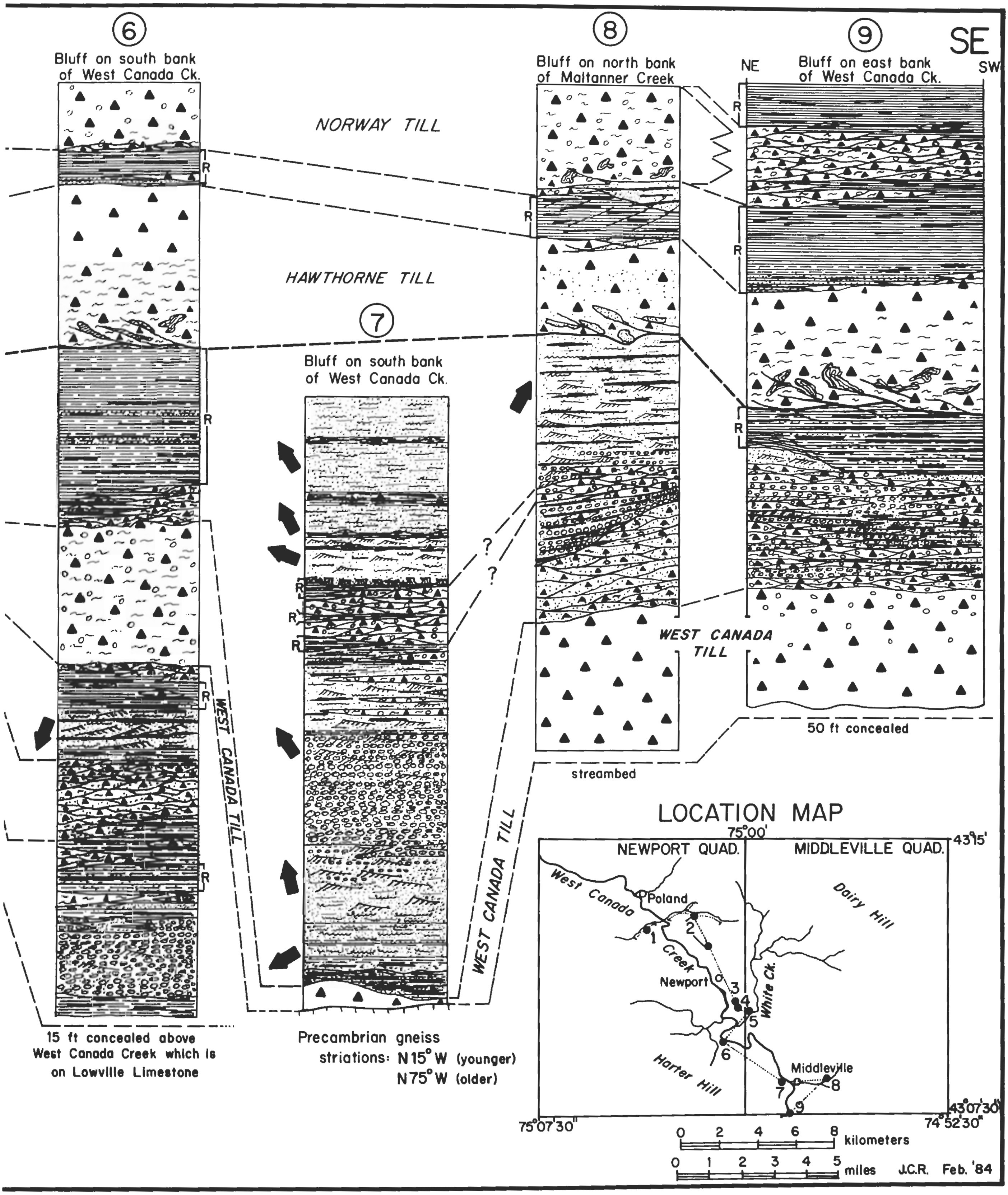

Figure 3B Representative stratigraphic sections showing inferred correlations in West Canada Creek area.

LOWER WEST CANADA AND MOHAWK VALLEYS	UPPER WEST CANADA VALLEY	BLACK RIVER VALLEY
valley train		
ROME TILL (O) and outwash	valley train (inwash ?)	
ice-marginal deltas and lacustrine beds	valley train	
HOLLAND PATENT TILL (O) and ice-marginal deltas	valley train and ice-marginal deltas	
lacustrine sediment	lacustrine sediment	lacustrine sediment
NORWAY TILL (O)	NORWAY TILL (O) Hinckley Moraine	ALDER TILL (B)
lacustrine sediment		lacustrine sediment
	lacustrine sediment	
HAWTHORNE TILL (M)		
lacustrine sediment		
Unconformity and fluvial gravels	Unconformity	Unconformity or lacustrine sediment (?)
upper Newport beds (lacustrine sediment)		
	lacustrine sediment	lacustrine sediment
WEST CANADA TILL (M)		
lower Newport beds (lacustrine sediment)	KUYAHOORA TILL (A)	ice-marginal gravels
	lacustrine sediment	TILL (B) ?
WHITE CREEK TILL (O-B)		
		D.A.F.-J.C.R. 1984

Figure 4 Lithostratigraphic equivalence of units in the western Mohawk Valley, upper West Canada Creek and Black River headwater. Primary provenance is indicated as follows: A = Adirondacks; B = Black River Sublobe; M = Mohawk Sublobe; O = Oneida Sublobe.

Kuyahoora Till

Loose, gray to yellow-brown, sandy, stony diamicton in the Hinckley, Ohio, North Wilmurt and northern part of the Middleville Quadrangles is here named the Kuyahoora till. The suggested relationship of Adirondack and Black River units to stratigraphy exposed along West Canada Creek is shown in Figure 4. Dominance of metamorphic clasts, and only toward the southern limits a presence of subordinate limestone cobbles clearly indicate the Adirondack provenance of the Kuyahoora till. Its margin lies against the north end of Dairy Hill and extends northwest to Sand Hill, a kame complex built into waters impounded at more than 466 m above sea level. Morphostratigraphic evidence suggests deposition during more than one glaciation, making Kuyahoora a lithofacies, rather than the name of a single till sheet. Bailey Hills and Ninety Five Hill (3.2 km northwest of Hinckley Reservoir) represent juxtaposed, partly synchronous deposition at the margins of the Black River, Oneida and Adirondack ice lobes.

White Creek – West Canada Interphase Deposits

At several exposures along White Creek, the White Creek till is overlain by and interbedded with a lacustrine sequence that contains sediment flow deposits, cross-bedded and rippled sand beds, sandy and silty turbidites and silty rhythmites. These sediments grade upward into clayey rhythmites and turbidites which are interbedded with the base of the West Canada till. At exposures along White and West Canada Creeks (Sections 5 and 6, Fig. 3), this sediment package also contains a sandy or silty, sparsely stony diamicton and a package of sediment flow deposits of uncertain provenance. These units may represent a small oscillation of the advancing Mohawk sublobe after deposition of the White Creek till.

West Canada Till

Gray to dark-gray, calcareous, stony to sparsely stony clay diamicton with eastern provenance (Mohawk sublobe) occurring above the White Creek till is here named the West Canada till. Throughout the White Creek and West Canada Valleys, northwest of Middleville, West Canada till is sparsely stony. Clasts are more numerous and the unit is sandier along the southern flank of Dairy Hill and in the West Canada Valley south of Middleville. Dominance of shale and limestone clasts with metamorphic and dolostone clasts, northwest-trending pebble fabrics and increasing interbedding with lacustrine sediments to the west indicate an easterly source.

Sections along Maltanner Creek and at Middleville along West Canada Creek (Sections 7, 8 and 9, Fig. 3) contain a very stony diamicton formerly called the "Maltanner till" (Muller and others, 1983), but now believed to be correlative with the West Canada till to the west. For this reason the term "Maltanner till" has been dropped. The unit is dominated by shale and limestone clasts and contains much dolostone. Vertically, the diamicton contains more shale in its base but has increased abundances of sandstones and fluvially abraded clasts upward. Pebble fabrics indicate northwest-southeast flow. Striations below the diamicton (Section 7, Fig. 3) indicate N15W and N75W trends, the latter being probably from an easterly source.

In Hawthorne Gulf on the southern slope of the Deerfield Hills, the West Canada till is a fissile, silty to clayey, calcareous, moderately stony, black diamicton, containing numerous metamorphic and dolomitic clasts. This till indicates westward penetration by the Mohawk sublobe with its border at an elevation of at least 305 m above sea level against the southern Deerfield Hills.

Subsequent events make it difficult to establish the limits of the glaciation responsible for the West Canada till. On the north flank, the margin of the Mohawk sublobe is marked by a kame delta at Diamond Hill (3.2 km north of Salisbury) (Fig. 1). The well-developed, east-west drumlin alignment across the Millers Mills and Jordanville Quadrangles reflects dominance of the Mohawk sublobe in ice-flow direction that was repeated during the subsequent Port Bruce Stadial. Noting this change from earlier southerly streamlining to an easterly drumlin alignment, Krall (1977) ascribed it to iceflow during building of the Cassville-Cooperstown Moraine. He further demonstrated that till of eastern provenance extends as far west as Clinton in the Oriskany Valley. Although a direct basis for correlation is lacking, it is tempting to relate deposition of the West Canada till to the same event.

Associated with the West Canada till at Middleville (Sections 7, 8 and 9, Fig. 3) is an interbedded lacustrine sequence that includes cobble gravel beds within lacustrine sediments. Several arguments point to a subaqueous, near-glacial origin for these beds. They are coarse, poorly-sorted, and show inverse graded bedding. Although dominantly well-rounded, the clasts include scattered small boulders and angular fragments. Stratification and structures suggesting deposition by a braided stream are lacking. The gravel units occur between crossbedded sand layers that show up-valley current directions, suggesting deposition from an impounding ice mass to the southeast.

Erie Interstadial

During the Erie Interstadial, the ice of the Mohawk sublobe retreated to the east permitting progressively lower levels of impoundment in the western Mohawk Valley. Lacustrine sediments that interfinger with the top of the West Canada till comprise an upper unit of the lake beds which we call the Newport beds. The name is useful, though not very specific. The stratified, generally fine-grained units which it includes are intercalated with several till sheets and represent impounding during oscillation of two distinct ice margins. In large part, they indicate sedimentation into water ponded north of and draining over, through or under ice that occupied the Mohawk Valley (Fig. 5A).

When not ponded north of the Mohawk and Oneida sublobes, the interlobate lake initially drained south into the through-valley network of the plateau, or east toward the Hudson Valley (Fig. 5B). Cedarville Col at 369 $\pm$ 3 m (Millers Mills Quadrangle), which later became the threshold of overflow during deposition of Norway till, may also have controlled lake level for a time during the Erie Interstadial. We cannot, however, rule out the possibility that a lower bedrock col to the west, though now drift-filled, may then have opened into a through-valley on the Appalachian Plateau. Further eastward recession of the Mohawk sublobe first lowered impounded water to the level of Lake Schoharie, controlled by the Delanson Outlet (Gallupville Quadrangle), initially at about 262 m. Inferred elimination of lacustrine conditions in the western Mohawk implies eastward withdrawal of the Mohawk sublobe sufficient to uncover the West Hill Channel that drained Lake Amsterdam at about 162 m above sea level (LaFleur, 1979). On the basis of unconformable relationships within the Newport beds, we infer such an erosional interval. Evidence in Ontario has been interpreted as indicating initiation of eastward drainage of the upper Great Lakes and Lake Erie during the Erie Interstadial (Hough, 1966; Dreimanis, 1969; Moerner and Dreimanis, 1973). Accordingly, we have tentatively correlated this deglacial event in the West Canada sequence with the Erie Interstadial of southern Ontario.

Port Bruce Stadial

The Port Bruce stadial was a time of more varied glacial activity in the western Mohawk Valley than previously recognized. Renewed ponding of meltwater and three glaciations are indicated by stratigraphic evidence along West Canada Creek and its tributaries.

Hawthorne Till

In the West Canada Valley, from North Creek to White Creek, the top of the Newport beds is truncated by a silty to clayey, calcareous, moderately to sparsely stony black diamicton, here named the Hawthorne till (Sections 6, 8 and 9, Fig. 3). This till is dominated by black shale clasts and contains abundant dolomite. From White Creek northwest in the West Canada Valley, an equivalent unit is a thick sequence of sediment flow deposits probably recording proglacial deposition at the maximum extent of ice that deposited the Hawthorne till (Section 3, Fig. 3).

In Hawthorne Gulf, a very clayey package of sediment flow deposits and a black, very clayey and sparsely stony diamicton indicate westward penetration of the Mohawk sublobe. The limit of the Mohawk sublobe at this time is inferred to have been no higher than 274 m above sea level against the southern Deerfield Hills. The kame moraine .4 km north of Salisbury (Fig. 1) is considered to mark the northern limit of the lobe at this time. Cushing (1905) referred to this feature as the "Pinnacle kame moraine" and reported that it overlies laminated clay with northward dip, suggesting readvance over the lake floor to this position.

Norway and Holland Patent Tills

Overlying Hawthorne till or age-equivalent sediments in exposures in Hawthorne Gulf is a moderately stony, silty diamicton with dominant calcareous shale and limestone clasts. This unit is the Norway till, named for exposures in the Town of Norway (Fig. 3) (Middleville Quadrangle). Its clast content reflects Oneida sublobe provenance and its eastward limit reaches almost to St. Johnsville in the Mohawk Valley.

In exposures along West Canada Creek (Sections 6 and 9, Fig. 3), Hawthorne and Norway tills are separated by a sequence of varved sediments which though truncated, indicates that an interval represented by at least 160 varves elapsed between deposition of the two units. It is not possible to prove that the Oneida and Mohawk sublobes were never in contact with each other along the central axis of the Mohawk Valley, but it seems clear that the two lobes did not simultaneously attain their maximum extents.

The Black River and Oneida sublobes, both of them responding to maximum expansion of the Ontario outflow center, coalesced along the western part of the Hinckley Moraine System immediately west of Hinckley (Fig. 1). The Black River sublobe deposited Alder till (Fig. 4), dominated by Trenton and Black River carbonate clasts and rich in metamorphic cobbles, particularly along the eastern margin of the lobe. The Hinckley Moraine System (Fullerton, 1971) is a series of kame ridges deposited by Oneida sublobe ice into Lake Miller, a local ponding in the West Canada Valley. Lake Miller drained south by Spruce Creek at present elevation of 430 m above sea level. The surface elevations of deltas built into Lake Miller suggest northward postglacial rebound of about .85 m/km.

The glacial advance which led to deposition of Norway till and to building of parts of the Hinckley Moraine Sys-

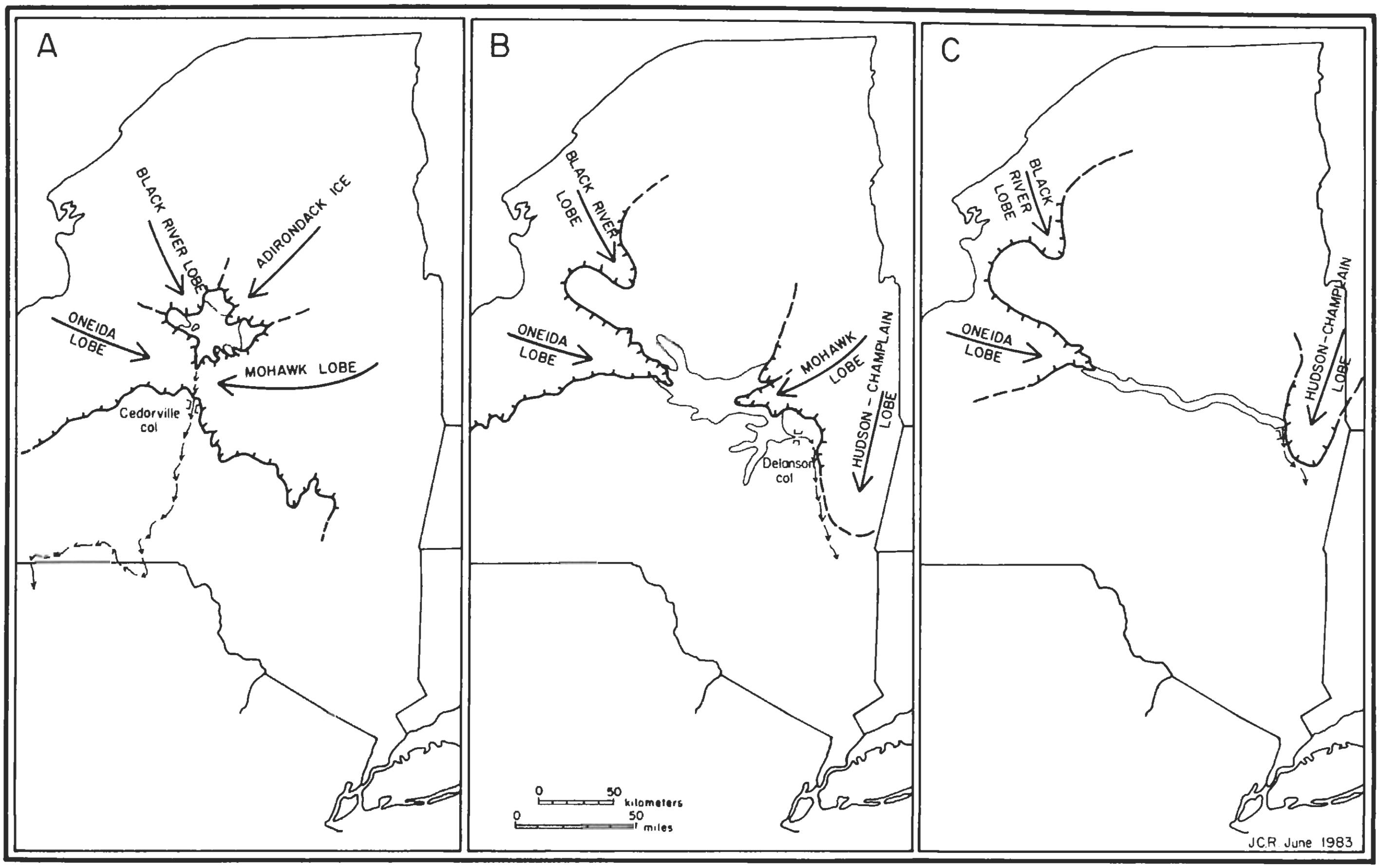

Figure 5 Inferred deglacial lake sequence in the western Mohawk Valley.

tem is essentially the Indian Castle Readvance of Fullerton (1971), a name which we now abandon for reasons discussed below. Correlative ice positions south of the Mohawk correspond, at least in part, to the Valley Heads Moraine System (Muller, 1965) and the Cazenovia moraine (Fullerton, 1971). In most through-valleys of central New York this moraine comprises the present drainage divide. Cedarville Col at 369 $\pm$ 3 m afforded the lowest threshold for outflow of water impounded between the Oneida and Mohawk sublobes.

Recession of the Oneida sublobe is marked by lacustrine deposits overlying Norway till in the western Mohawk Valley. The lake in which these sediments were deposited persisted as long as the western margin of the Mohawk sublobe lay west of Amsterdam. Antonetti (1982) mapped a thin, but rather continuous extension of these lake beds consisting of at least 140 varves westward across South Trenton Quadrangle into the Ninemile Creek basin.

Overlying this rhythmite sequence, Antonetti (1982) described the Holland Patent till which is difficult to distinguish from Norway till other than by its stratigraphic position. In general, the Holland Patent till is finer in texture and clast size. Along its eastern margin, it is marked by a clay-rich facies and end moraines north of Hawthorne Gulf in the Newport Quadrangle. Preliminary mapping in the Herkimer and Little Falls Quadrangles suggests that the Holland Patent till extends east of Little Falls. Southeast of Fairfield, the projected border of Holland Patent till corresponds closely to Fulerton's (1971) Indian Castle border. The close association of the Norway and Holland Patent till sheets permits the conclusion that both relate to glacier fluctuations which built the massive Valley Heads Moraine System in the Finger Lakes area. If this is so, the 40 kms of recession indicated by the intercalated rhythmite sequence in the Ninemile Creek Valley sheds light on the manner in which the large valley-blocking moraines of the Valley Heads System were built.

Mackinaw Interstadial

Mapping in the Remsen and Newport Quadrangles suggests that continued eastward recession of the Mohawk sublobe did not lower the level of ponding in the Mohawk Valley below the Delanson outlet (Fig. 5B).

Contacts between topset and foreset beds in kame deltas in the West Canada Valley from Poland northwest to Barneveld show water levels at elevations of 290 to 302 m. Preliminary information in the Herkimer and Little Falls Quadrangles suggests that the Holland Patent ice did not dam the West Canada Valley, and that water levels were controlled by a spillway in the Mohawk Valley to the east. If isostatic tilting as steep as that inferred for deltas of Lake Miller (.75 m/km in a generally northerly direction) can be assumed, the Delanson Channel at 256 m may conceivably have been the outlet for the lake that occupied West Canada Valley. The Oneida sublobe retreated as far west as Oriskany Creek while impoundment in the Mohawk Valley was at an elevation of at least 162 m. Lacustrine ponding in the Mohawk lowered further, as indicated by topset-foreset contact elevations of 140 m in deltas at Herkimer and west of Frankfort in the Ilion and Herkimer Quadrangles (Fig. 5C). Further retreat west of Rome may have permitted free drainage from the Ontario Basin across the present Oneida-Mohawk saddle. Meltwater channels on the south flank of the Tug Hill Plateau (Muller, 1978), the Syracuse Channels (Hand and Muller, 1972; Hand, 1978) and related channels from Syracuse east to Oneida (Sissons, 1960) were abandoned.

Port Huron Stadial

The last readvance into the Mohawk Valley was the Stanwix Readvance of the Oneida sublobe (Fullerton, 1971), for which a Port Huron age is suggested. This episode is recorded in the Oneida Lowland by red till, here named the Rome till. Eastward expansion of the Stanwix Readvance spread ice across the mouth of Ninemile Creek and deflected the upper Mohawk River at the present site of Delta Reservoir, north of Rome (Chambers, 1978; Halberg and others, 1962).

Rome till can be seen over lacustrine deposits in a few places in the South Trenton Quadrangle and westward along the south margin of the Ontario lake plain. Recession from the Stanwix margin was followed by lowering of "hyper-Iroquois waters" (Fairchild, 1912) and renewal of Lake Iroquois, controlled by the level of alluvial fan construction south of Rome where the Mohawk River, leaving the Tug Hill Plateau, first begins its eastward course.

DISCUSSION

Indian Castle Readvance of Previous Authors

Fullerton (1971) applied the term Indian Castle Readvance to an expansion of the Oneida sublobe during an interval of open drainage in the Mohawk Valley.

Our field data show that the Indian Castle Readvance of Fullerton represents two separate Port Bruce readvances of the Oneida sublobe (Fig. 2). Further, the mapped limit of the Indian Castle Readvance only partly delineates each of these two advances. Rather, it corresponds to the margin of the Norway till (the Hinckley Readvance) from Hinckley Reservoir southeast to Fairfield, but it follows closely the limit of Holland Patent till (the Barneveld Readvance) south of Fairfield to Indian Castle.

We also question Fullerton's suggestion that free drainage existed in the Mohawk Valley during deposition of the Hawthorne and Norway tills. In the Deerfield Hills along West Canada and Hawthorne Creeks, Norway till is found overlying and intercalated with lacustrine sediments which interfinger downward into Hawthorne till. These sections show impounding between the two lobes at least to the elevation of 369 m for overflow south through Cedarville Col. We recognize no evidence of free drainage such as to permit the eastward spread of outwash in the Mohawk Valley during Port Bruce time.

Parallel Response of Separate Lobes

In contrast to other areas discussed in this volume, glacial topography of the western Mohawk Valley is not dominated by deposits of stagnating and disintegrating ice. Except for the Valley Heads ice margin projecting into through-valleys of the Plateau from Cedarville westward, kame and kettle topography is of extremely limited extent. In retrospect, at least, this is a predictable consequence of the low basal shear and marked instability of a floating ice foot. Surging advance and retreat by calving embayment probably characterized the regime. Minimal subglacial erosion and the accumulation of an unusally thick and complete late-glacial stratigraphic succession are the fortunate results.

A significant, though not surprising, conclusion based on stratigraphic relationships in the western Mohawk Valley is that the Oneida and Black River sublobes of the Ontario lobe responded to the same climatic impulses as did the Mohawk sublobe.

Equally interesting is the failure of the lobes to attain their maximum expansions simultaneously in this area where the asynchrony can be rather accurately determined. The stratigraphic evidence in West Canada Creek exposures shows regional flow south-southeastward in this area southwest of the Adirondack Uplands at the time of maximum Woodfordian ice expansion. Subsequent westward spreading of the Mohawk sublobe is marked by the West Canada till that was deposited during retreat of the Oneida sublobe that had deposited the White Creek till (Fig. 2). Renewal of glacigenic conditions following the Erie Interstadial saw the return of the Mohawk sublobe, rapidly reaching its maximum extent and beginning its retreat before the Oneida sublobe attained its maximum expansion. At least 160 years elapsed between deposition of Hawthorne and Norway tills in the West Canada Valley.

A reasonable explanation for the asynchronous culmination of these two advances may lie in unequal response times that reflect different distances from the centers of outflow. Alternatively, it may reflect dominance of a late outflow center of the Ontario sublobe and spreading from the Ontario Basin as shown by radial disposition of drumlins in west central New York (Holmes, 1952).

A third alternative hypothesis for asynchronous culmination of advances by the Mohawk and Oneida sublobes suggests that rates of growth and recession by the Mohawk sublobe responded to changing throughflow across the southeastern part of the Adirondack massif during critical stages in expansion and recession. Although it is generally accepted that the highest peaks of the Adirondacks were overtopped by continental glaciation, evidence in the western Mohawk Valley supports Chamberlin's (1883) assessment that the Adirondacks served rather as a barrier than as a major contributor to flow during most of the Pleistocene.

During deposition of West Canada till, the Mohawk sublobe was augmented by ice flow through the Adirondacks as far west as the headwaters of East Canada Creek. Similar conditions applied during deposition of Hawthorne till. During recession the Adirondack throughflow failed earlier than did axial flow of the Mohawk trunk glacier, as shown by the manner in which the Gloversville moraine in the central Mohawk Valley transects the lateral moraine of the expanded Sacandaga tongue of Adirondack throughflow. Late-glacial, ice sources in the Adirondacks (Craft, 1976) were too small to have contributed ice to the Mohawk Valley.

CONCLUSIONS

The stratigraphy and morphology of the western Mohawk Valley record the interplay of Oneida, Black River, Mohawk and Adirondack glacier lobes. White Creek till represents, in part, deposition by overriding ice independent of local topographic deflection. However, lobation of the Late Wisconsinan ice margin is first apparent in juxtaposition of the upper part of the White Creek till of Oneida provenance and West Canada till of Mohawk provenance.

Intercalation of lacustrine deposits with 6 till units in the West Canada Creek Valley records the impounding of meltwater between the lobes up to levels controlled variously by drainage thresholds south to through-valleys of the plateau, or east to the Hudson Valley.

At some time prior to final withdrawal of the ice sheet, eastward drainage through the Mohawk Valley became subaerial, implying substantial thinning and withdrawal of the Mohawk sublobe. In the absence of other criteria for absolute dating, this development of free drainage affords the best present basis for regional correlation. On this basis, correlation is proposed with geologic/climatic units established for southern Ontario. The White Creek and West Canada tills are assigned to the Nissouri Stadial; the Hawthorne, Norway and Holland Patent tills, to the Port Bruce Stadial, and the Rome till of the Stanwix Readvance, to the Port Huron Stadial (Fig. 2).

Although the several lobes responded to the same regional climatic impulses, they did not simultaneously reach maximum extent. The lag between parallel culminations of Oneida and Mohawk sublobes was at least 160 years. Possible causes for asynchrony include differing distances from the accumulation area, shifting centers of outflow, differences in basin configuration and changing contributions of obstructed Adirondack throughflow.

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PLEISTOCENE STRATIGRAPHY OF NORTHWESTERN CONNECTICUT

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ABSTRACT

In Late Wisconsinan time the expanding Hudson-Champlain Valley glacial lobe spread southeast over the western part of Connecticut. The resultant erosional and depositional features document the characteristics of the eastern edge of this ice lobe. Meltwater channels and glaciofluvial deposits record the retreat of the ice margin as the ice shrank back to the Hudson River valley during its waning phases.

Glacial deposits are widespread but discontinuous in northwestern Connecticut. Bedrock outcrops extensively in upland areas and most of it exhibits little weathering, but there are some localities where saprolite is preserved beneath glacial sediments. Two tills have been described in the study area – a Late Wisconsinan (Woodfordian) upper till and a weathered lower till of probable Early Wisconsinan (Altonian?) age.

Drumlins and bedrock striations indicate that the principal Late Wisconsinan ice-flow direction in western Connecticut was toward the south-southeast. A notable exception is in the northwestern part of the state, where east-southeast ice-flow indicators near Salisbury suggest topographic control of movement as the ice thinned in late-glacial time.

The pattern of ice recession is best defined in the major valleys, where glaciofluvial and glaciolacustrine deposits are locally abundant. The characteristics of these meltwater deposits support the concept of stagnation-zone retreat. Morphosequences and meltwater channels mapped by the authors reveal that the Hudson Valley lobe retreated generally to the north-northwest in the study area. However, thinning of the ice sheet over the Housatonic Highlands and southern end of the Taconic Range resulted in an irregular configuration of the ice margin. Deposition of water-laid glacial sediments occurred chiefly in the larger valleys, and it is difficult to correlate meltwater deposits across the uplands between these valleys.

Numerous glacial lakes and associated deposits formed where meltwater was ponded behind drift barriers or in valleys which sloped toward the ice margin. Prominent end moraines were deposited in one of these lakes. Fluvial ice-contact and outwash deposits typically occur where glacial streams could drain freely away from the ice margin. Major meltwater stream systems from New York flowed eastward into the Housatonic region via the Still River and Tenmile River Valleys and into the Salisbury area from the Taconic Mountains.

Ice-margin positions in the Housatonic River valley northwest of New Milford and southeast of Salisbury are tentatively correlated with the Shenandoah and Poughkeepsie Moraines, respectively, in adjacent New York. A later position in the vicinity of the New York-Connecticut border may be correlative with the Hyde Park Moraine in New York.

INTRODUCTION

The purpose of this paper is to summarize the Pleistocene stratigraphy of the Housatonic River basin and adjacent areas in northwestern Connecticut. The authors' investigation of Late Pleistocene surficial deposits in western Connecticut has determined characteristics of Wisconsinan glaciation and the history of deglaciation in part of the dissected New England Uplands (Kelley, 1975a, 1975b; Thompson, 1975).

The study area lies along the middle reach of the Housatonic River and has local relief exceeding 400 m. Glacial tills, glaciofluvial and glaciolacustrine deposits, and erosional features have been mapped in detail by the authors in the Danbury, New Milford, Kent and Ellsworth 7.5-minute quadrangles, and at reconnaissance level in adjacent areas. W.B. Thompson has compiled the geology of the entire region for the *Surficial*

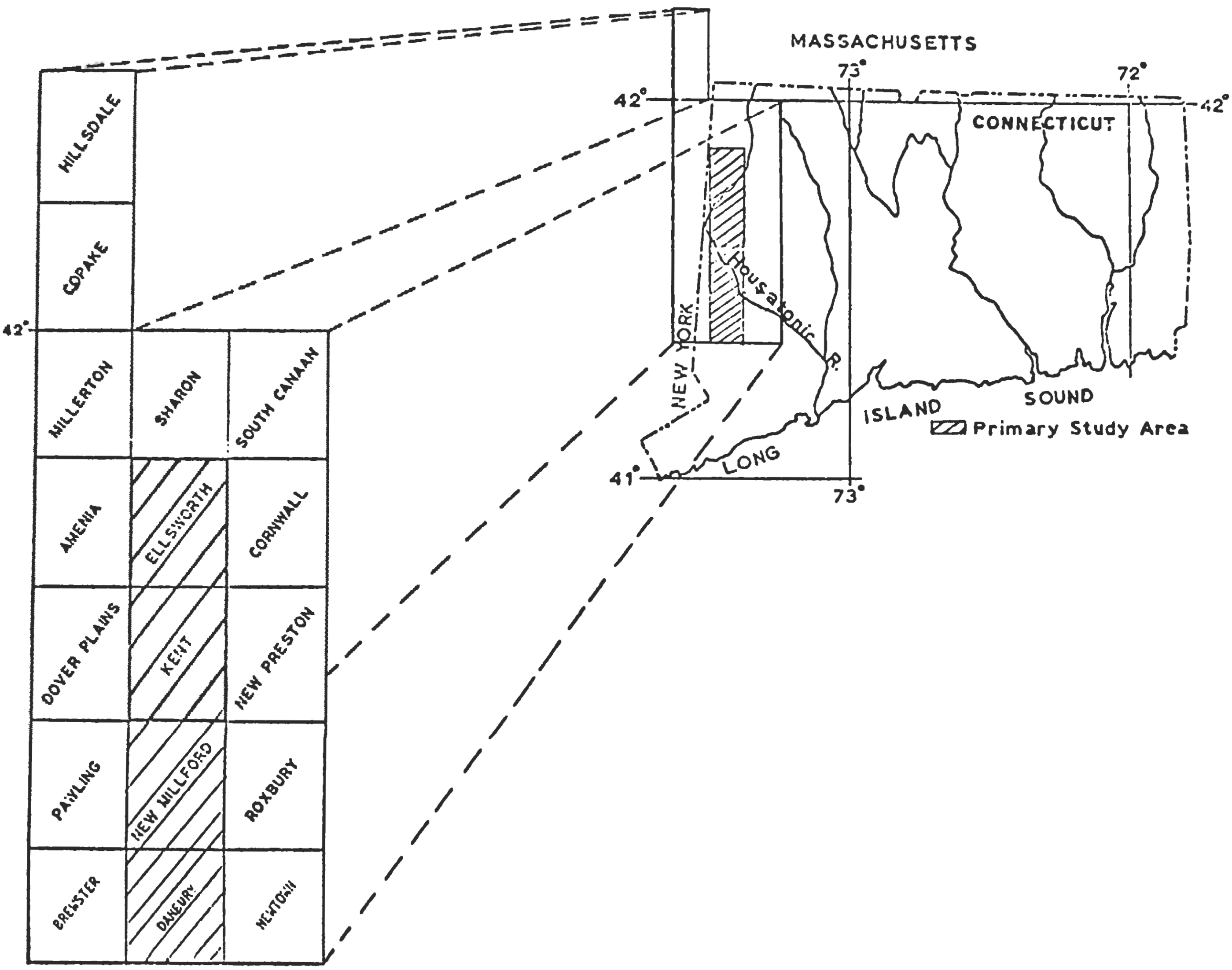

Figure 1 Location map of study area in western Connecticut.

Geologic Map of Connecticut (Schafer and others, In prep.). Figure 1 shows the boundaries of the study area, which includes the 7.5-minute quadrangles between Danbury and the Massachusetts border. The eastern boundary of this area is approximately coincident with the drainage divide between the Housatonic and Naugatuck River basins.

The emphasis of this paper is the recession of the Late Wisconsinan (Woodfordian) ice sheet and the associated series of meltwater deposits. Till stratigraphy is discussed only briefly, because the tills have only partly been investigated in detail. There is, however, sufficient information on this subject to generally characterize the till units and correlate them with till units in other parts of southern New England. The Late Wisconsinan meltwater deposits are better understood because they have been the focus of intensive quadrangle mapping by Federal and State geological surveys.

Mapping in western Connecticut during the last 15 years has been directed toward interpreting and correlating meltwater deposits using the morphosequence concept. Koteff and Pessl (1981) demonstrated the utility of this concept in the reconstruction of deglaciation over much of New England. The present authors have found the morphosequence concept applicable to the major valleys, where glacial retreat resulted in deposition of successive water-laid morphosequences. Former positions of the stagnant ice margin can be recognized by the location of the ice-contact heads of these deposits. Active ice to the north and northwest of the stagnant zone supplied the debris which washed over and through the stagnant ice to form the morphosequence deposits. Correlation of meltwater deposits between north-south valleys is difficult because the uplands are underlain by till and bedrock, and ice-margin position indicators are scarce or absent. In some places, meltwater channels can be used to tentatively correlate upland ice positions.

PREGLACIALLY WEATHERED BEDROCK

The range of elevation in the study area is from 9 m to 724 m (30 to 2375 feet), from the Housatonic River to the northwest corner of the state. The topography of most of the region is of two principal types – areas of knobby, bedrock-controlled terrain and areas dominated by thick till in the form of southeast-trending, glacially-streamlined hills. There are numerous outcrops of Precambrian and Paleozoic crystalline bedrock in the hilly terrain in northwestern Connecticut.

Most bedrock exposures in the Housatonic region consist of solid, unweathered rock. There are, however, places where deeply weathered and disintegrated bedrock is present either at the ground surface or beneath surficial sediments. These remnant weathering profiles are significant to the glacial history of Western Connecticut because they indicate that erosion by Pleistocene glaciers did not extend deep enough in certain areas to completely remove the saprolite cover. Rottenstone, *in-situ* disintegrated rock, is especially common in the Stockbridge Formation (Inwood Marble), that underlies much of the Housatonic Valley. Thickness of the weathered zone exceeds 3 m (10 feet) (Melvin, 1970) and locally reaches 30 m (100 feet). Thompson (1975) concluded that alteration of dolomite marble was the combined result of mechanical weathering and dissolution of intergranular calcite. The thick saprolite probably started to form during preglacial time. A pre-Quaternary age for some of the saprolite is inferred from studies of water-worn clasts of fine-grained black carbonaceous material (probably peat) from gravel pits north of East Canaan village. These pits are situated in a delta that was deposited into glacial Lake Norfolk during the recession of the Late Wisconsinan ice sheet. Clasts collected by W. S. Newman from the O'Connor pit yielded a C-14 age of greater than 40,000 yrs BP (W-2615, U.S.G.S. Radiocarbon Laboratory). Additional samples were collected by Thompson in 1979 from deep in the foreset beds in the proximal part of the delta, north of the O'Connor pit. Examination of the latter clasts by the Paleontology and Stratigraphy Branch of the U.S.G.S. revealed the presence of pollen grains and spores comprising an assemblage of probable Late Cretaceous age (Frederiksen,

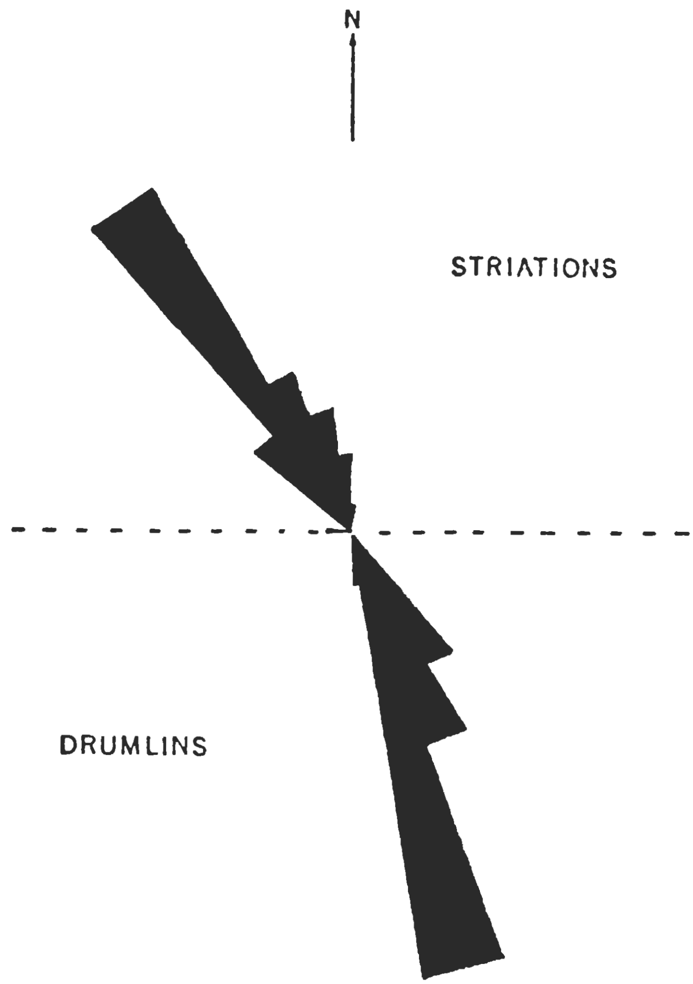

Figure 2 Striation and drumlin axis orientations in northwestern Connecticut.

pers.comm.). Other carbonaceous clasts, the first to be recorded in the area, were discovered in 1967 by Melvin (pers.comm.) in the foreset beds of the glacial Lake Norfolk delta at nearby Norfolk village.

The organic material described above probably was eroded from Cretaceous deposits in the marble-floored lowland areas just northwest of the Lake Norfolk deposits. The fragile carbonaceous clasts and their proximity to possible source areas suggest that they were transported only a short distance by the last ice sheet before deposition in glacial Lake Norfolk. Schafer (pers. comm.) has observed that saprolites of known or probable Cretaceous-Tertiary age are common on the marble belt from Connecticut northward into Massachusetts and Vermont. They include the well-known lignite deposit at Brandon, Vermont, and several iron-oxide deposits which were formerly mined for iron ore.

GLACIATION

The western half of Connecticut was glaciated by the Hudson-Champlain Valley and Connecticut River valley ice lobes. Southeast-oriented directional data (Pessl, 1971; Figures 2 and 3) indicate that the eastern part of the Hudson-Champlain lobe was responsible for glaciation of the study area. Thickening of the glacial lobe forced ice through low-elevation cols in the Hudson, Housatonic and Taconic Highlands, and ultimately caused the mountains and highlands along the western New England border to be overtopped (Taylor, 1903). The ice then overrode successive northeast-trending ridges in western Connecticut. The orientation of the advancing ice margin in northwestern Connecticut was northeast-to-southwest. Late glacial ice-flow indicators reveal that the northeast-to-southwest ice-margin trend persisted during deglaciation, at least across the plane of the uplands, although relief features altered and partly obscured this trend.

Upland ridges and valleys oblique to the direction of glacial movement differentially affected the margin by restricting or favoring ice flow. This, in turn, caused a markedly sinuous ice margin, especially during deglaciation. A high-altitude observation of northwestern Connecticut at that time would probably have revealed a ragged, lobate and serrated ice margin, similar to the margins of ice fields in Alaska and other moderate-to-high relief regions today.

Ice-flow directional indicators are interpreted as sensitively reflecting topographic control of the thinning ice lobes (Kelley, 1975a). The Taconic Mountains, north of the study area, rise to elevations higher than the Housatonic Highlands and the adjacent ridges to the east and west. Ice from the Hudson-Champlain Valley lobe penetrated through cols along the New York-Connecticut border and encountered the western margin of the Connecticut Valley lobe on the uplands between the Hudson and Connecticut Valleys. As the active ice thinned, the Connecticut River lobe dominated ice flow east and southeast of the Taconic Range. This influence caused a southward deflection of the locally weakened Hudson lobe ice-flow in western Massachusetts and Connecticut. However, lower-elevation uplands south of the high Taconic Range, such as the Housatonic Highlands, permitted vigorous southeastward ice flow to be maintained across that area. This flow resisted southward deflections by the Connecticut Valley ice lobe.

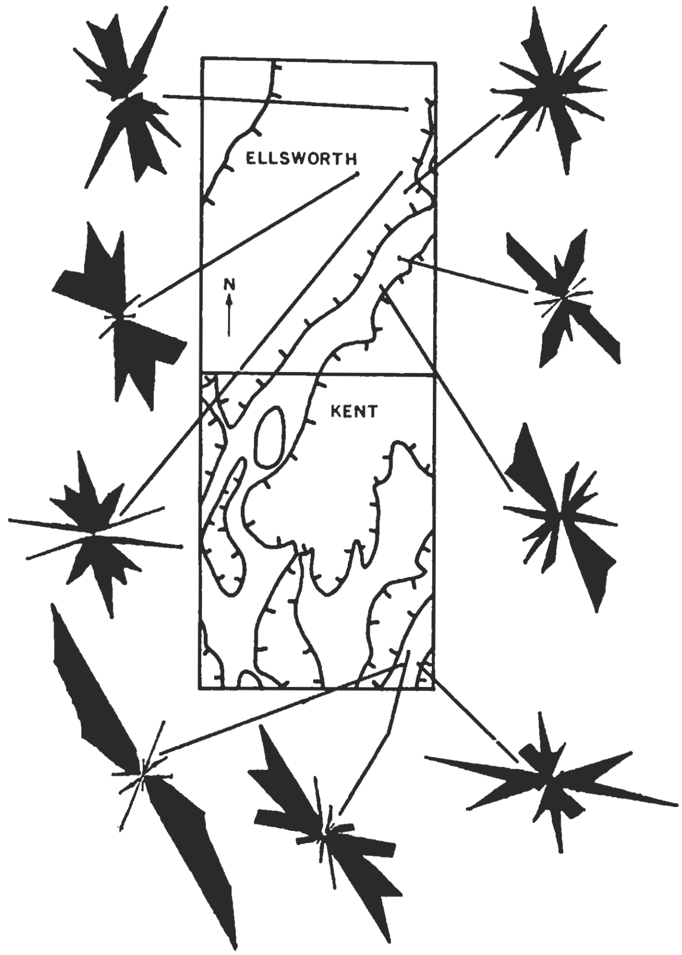

Figure 3 Pebble fabrics in tills of Ellsworth and Kent quadrangles, Connecticut.

Till Stratigraphy

There are two distinct tills in northwestern Connecticut: a dominantly nonoxidized "upper till" overlying a deeply oxidized "lower till." Exposures of these tills near Danbury and New Milford are correlated with the upper and lower tills described in New England by Pessl (1971) and earlier investigators (Thompson, 1975). Newton (1979) designated the upper unit as the "Bakersville Till" and the lower one as the "Thomaston Till."

The lower, olive brown to dark olive gray, deeply oxidized till is a lodgement till and forms the cores of many drumlins. This till is compact, silty, and less stony than the upper till. The lower till also has subhorizontal and subvertical joints, and dark brown iron-manganese-oxide staining on the surfaces of both joints and stones.

The upper till is typically nonoxidized, light olive brown to light olive gray, sandy, and very stony. Pieces of the lower till and lenses of washed sediment are locally included in the upper till.

Regional relationships support the hypothesis that the upper till of western Connecticut was deposited during the Late Wisconsinan glaciation of New England. The authors concur with Pessl and Schafer (1968) that the lower till was deposited by an earlier glaciation. Considering the slight degree of clast weathering and development of secondary clay minerals in the lower till (Thompson, 1975), this deposit may be no older than Early Wisconsinan (Altonian).

DEGLACIATION AND THE DEVELOPMENT OF LATE WISCONSINAN MELTWATER DEPOSITS

During the late phases of deglaciation, local bedrock ridges transverse to ice flow controlled glacial flow and ultimately initiated zones of ice stagnation. Deposits associated with successive ridges record the presence of thick, presumably active ice northwest of each ridge, while related valley features southeast of each ridge were produced in association with stagnant ice.

Field evidence indicates that thick ice northwest of ridges directed meltwater through upland cols and meltwater channels. The meltwater descended the southeast slopes, eroding some materials and depositing others on stagnant, isolated, lower-elevation ice surfaces in the valleys. A typical example of this situation is at Bulls Bridge, south of Kent. An abandoned meltwater channel with a threshold elevation of 184 m (603 feet) crosses Spooner Hill (Kent Quadrangle). Boulder concentrations associated with terraces on the northwest-facing hill slope, and knob-and-kettle topography to the east, record the flow of meltwater along an ice margin, through the channel, and down to the lower-elevation ice beyond. Here the maximum difference in ice-surface elevations on opposite sides of Spooner Hill was 30 m (98 feet) in a distance of 1 km (0.6 mile) (Kelley, 1975a).

Abandoned outwash channels crossing upland ridges indicate temporary ice-margin positions which developed between the relatively active ice on the uplands and the distal stagnant ice. An approximate estimate for the width of the stagnant ice zone can be calculated by determining the difference in elevation that developed between the active and stagnant ice surfaces. Elevation differences have been estimated across several ridges based upon the location of related upland meltwater-channel thresholds and lower-elevation ice-contact deposits. These estimates range from 17 m/km (90 feet/mile) to 40 m/km (210 feet/mile). The maximum local relief in the study area is 397 m (1300 feet). Assuming this is the minimum thickness for active ice northwest of the ridges, and using the calculated gradients discussed above, a stagnation zone between 10.5 km (6.5 mile) and 23 km (14.3 mile) wide probably developed in this region. This stagnation zone would have included stranded, detached ice in two or three valleys in the lee of the nunatak ridges. Thus, stagnant rather than active ice-marginal retreat dominated deglaciation and deposition in the valleys and resulted in the formation of meltwater deposits in the Housatonic River Valley and its tributaries. These deposits include abundant glaciolacustrine sediments. Valley-train outwash deposits are relatively uncommon, however, because the hilly terrain favored meltwater ponding during ice recession, especially in the small valleys that sloped directly or obliquely toward the receding ice margin.

Surficial geologic mapping demonstrates that most of the meltwater deposits can be grouped into morphosequences of the types described by Koteff and Pessl (1981). End moraines are rare in the interior of southern New England but the outwash heads at the proximal end of many ice-contact morphosequences offer an alternative means of identifying former ice-margin positions.

Figure 4 shows successive positions of the Late Wisconsinan ice margin in those areas where they are known from outwash heads and meltwater channels, or can be inferred from topographic considerations or by correlation of closely spaced meltwater deposits. This map indicates an overall pattern of northwestward ice retreat, but with many local variations.

Glaciolacustrine deposits formed in temporary lakes ranging from small ice-marginal ponds to large bodies of open water such as Glacial Lake Danbury. Prominent ice-contact deltas and lake-bottom deposits of sand, silt, and clay record large lakes in the Housatonic, Still, and Hollenbeck Valleys. Sediments also were deposited in small ponds between the ice margin and upland ridges and cols. In those ponds which filled with lacustrine sediments, the early deltaic facies were buried beneath fluvial deposits that graded directly to the spillway. The result was a lacustrine-fluvial morphosequence as defined by Koteff and Pessl (1981).

Glacial lake levels were controlled by several spillway types, including cols on till or bedrock ridges, stratified-drift dams, and plugs of stagnant ice. Lacustrine deposits also formed in the south-draining Housatonic River Valley where meltwater was impounded against earlier ice-contact outwash heads.

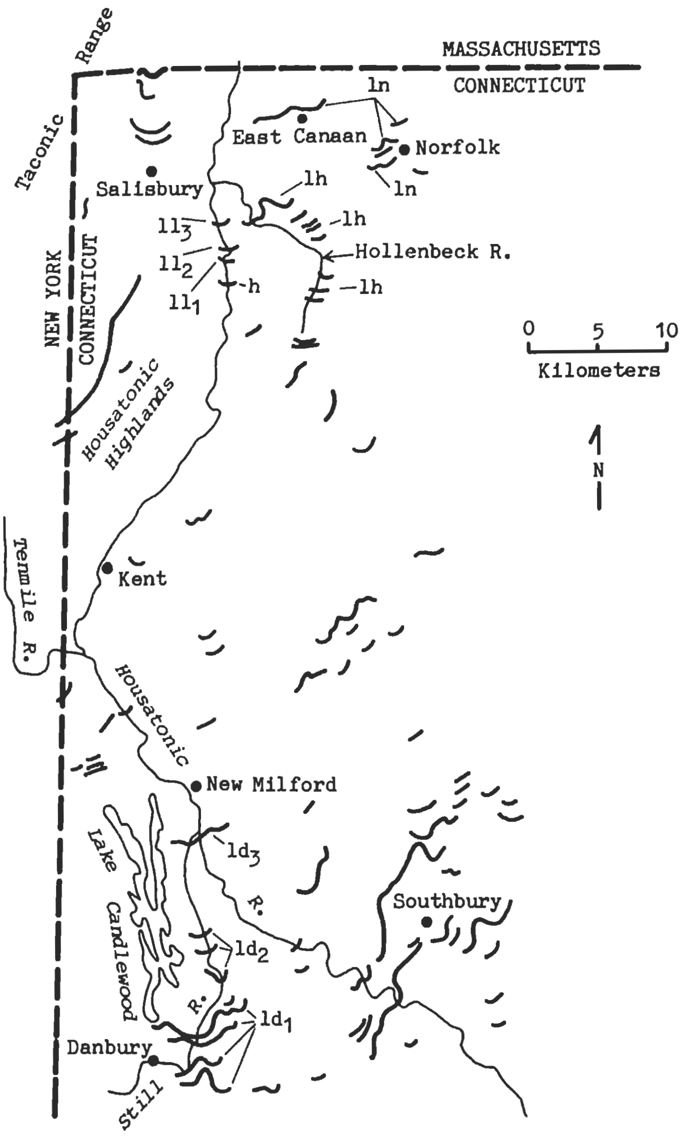

Figure 4 Map showing recessional positions of the late Wisconsinan ice margin in northwestern Connecticut. Lines indicate heads of stratified-drift morphosequences and correlations between deposits believed to be of equivalent age. Labeled ice margin positions were associated with the following glacial lakes: ld, Lake Danbury; lh, Lake Hollenbeck; ll, Lake Lime Rock; ln, Lake Norfolk; h, drift dam that controlled Lake Lime Rock.

Glacial Lakes of the Danbury – New Milford Area

Glacial Lake Danbury developed as the ice margin retreated northward from the divide between the Still River basin and the Saugatuck River basin to the south. The extent of this lake and the spillways for its four successive stages are shown in Figure 5. At first, sand and gravel were deposited among stagnant ice blocks which choked the lowland between the divide and the present location of the Still River. An open lake later developed as the stagnant ice melted. Deltas and lake-bottom sediments were then deposited in the location of the city of Danbury. The Saugatuck River stage of Lake Danbury continued to drain southward across the Saugatuck divide while the ice margin withdrew to the successive positions (ld1 position, Figure 4). Meltwater entered Lake Danbury from the north and west. The western meltwater drainage came from a narrow tongue of Hudson Valley ice that occupied the upper end of the Still River Valley in Danbury. Recession of this ice tongue initiated Glacial Lake Kenosia which was located just west of the present city of Danbury. Thick sediments that accumulated on the bottom of Glacial Lake Kenosia underlie the basin now occupied by Mill Plain Swamp and the Danbury airport (Thompson, 1975).

Continued northward retreat of the ice margin (to ld2 position, Figure 4) allowed Lake Danbury to drain eastward across the divide between the Still River and Pond Brook and thence to the Housatonic River (Figure 5). During the Pond Brook stage, unusual sand deposits were built on the west side of the Still Valley near Brookfield village, consisting of delta foreset beds. Topset beds are lacking and the tops of the deposits are too low to have been graded to any available spillway. These sand deposits do not fit into the existing classification scheme for morphosequences although they are, in fact, ice-marginal features. The authors consider the Brookfield deposits to be subaqueous fans. Their sediment supply was terminated by rapid ice retreat before they could build up to the lake surface.

The Pumpkin Hill stage of Lake Danbury began when the ice margin reached the present Lanesville area in New Milford. A prominent delta was deposited at Lanesville (ld3 position, Figure 4) very close to the associated spillway across Pumpkin Hill. The esker which fed this delta is 2.8 km (2 miles) long, and is indicative of the minimum width of the stagnant zone fringing the active ice.

The Pumpkin Hill stage of Lake Danbury was short-lived and soon developed into the final Housatonic River stage as ice receded from the Housatonic Gorge (south of

New Milford). At this time meltwater flowed directly down the Housatonic Valley (Figure 5). The abundant glaciolacustrine sediments near New Milford indicate that an ice or drift barrier temporarily blocked the gorge and delayed the termination of Lake Danbury. Outwash eventually filled the lake basin at New Milford, ending the Housatonic River stage.

Glacial Lake Candlewood developed in the former Rocky River basin to the north and west of the Still River (Thompson, 1975). Most of the deposits of this lake now are obscured by modern Lake Candlewood, a manmade lake that floods the Rocky River Valley. Glacial Lake Candlewood drained southward across a divide into Lake Danbury.

Deglaciation of the Housatonic River Valley Between New Milford and Cornwall

Along the Housatonic River from New Milford to Cornwall, the northeast-southwest trend of the upland ridges, coupled with the northwestward ice retreat, caused the development of extended stagnant ice tongues in the Housatonic River valley and detached ice blocks in adjacent isolated valleys. Striations and the orientation of meltwater channels across ridges indicate that the receding ice margin was sub-parallel to the trend of these ridges. The scarcity of ice-marginal deposits and the fragmentary nature of drainage channels hinder correlation of ice-marginal positions and supports the concept of a rapid active ice-marginal recession across the upland ridge crests. Ice thinning over the uplands permitted topographic relief to control local deglaciation.

Active ice continued to flow through broad, low-elevation cols after flow across the uplands terminated. This weakly active ice facilitated the development of valley plugs and numerous drainage diversions, linear boulder concentrations, and associate meltwater drainage features (Kelley, 1975a).

Heads of outwash and vertical intervals between terraces along the Housatonic River valley record successive marginal positions of the shrinking valley ice

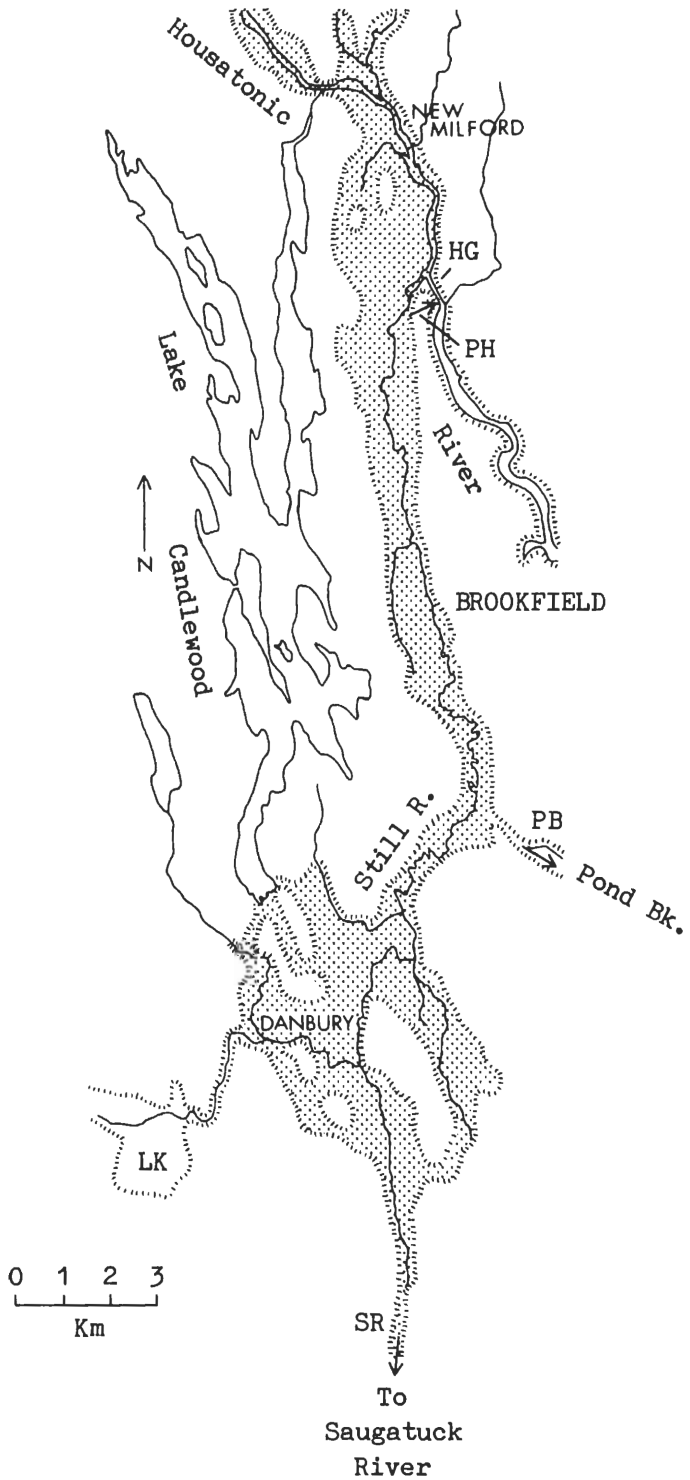

Figure 5 Map of Still River Valley region, showing extent of glacial Lake Danbury deposits (shaded area). Arrows indicate spillways for first three stages of Lake Danbury: SR, Saugatuck River stage; PB, Pond Brook stage; PH, Pumpkin Hill stage. Final (Housatonic River) stage drained through the Housatonic Gorge (HG). LK is glacial Lake Kenosia basin.

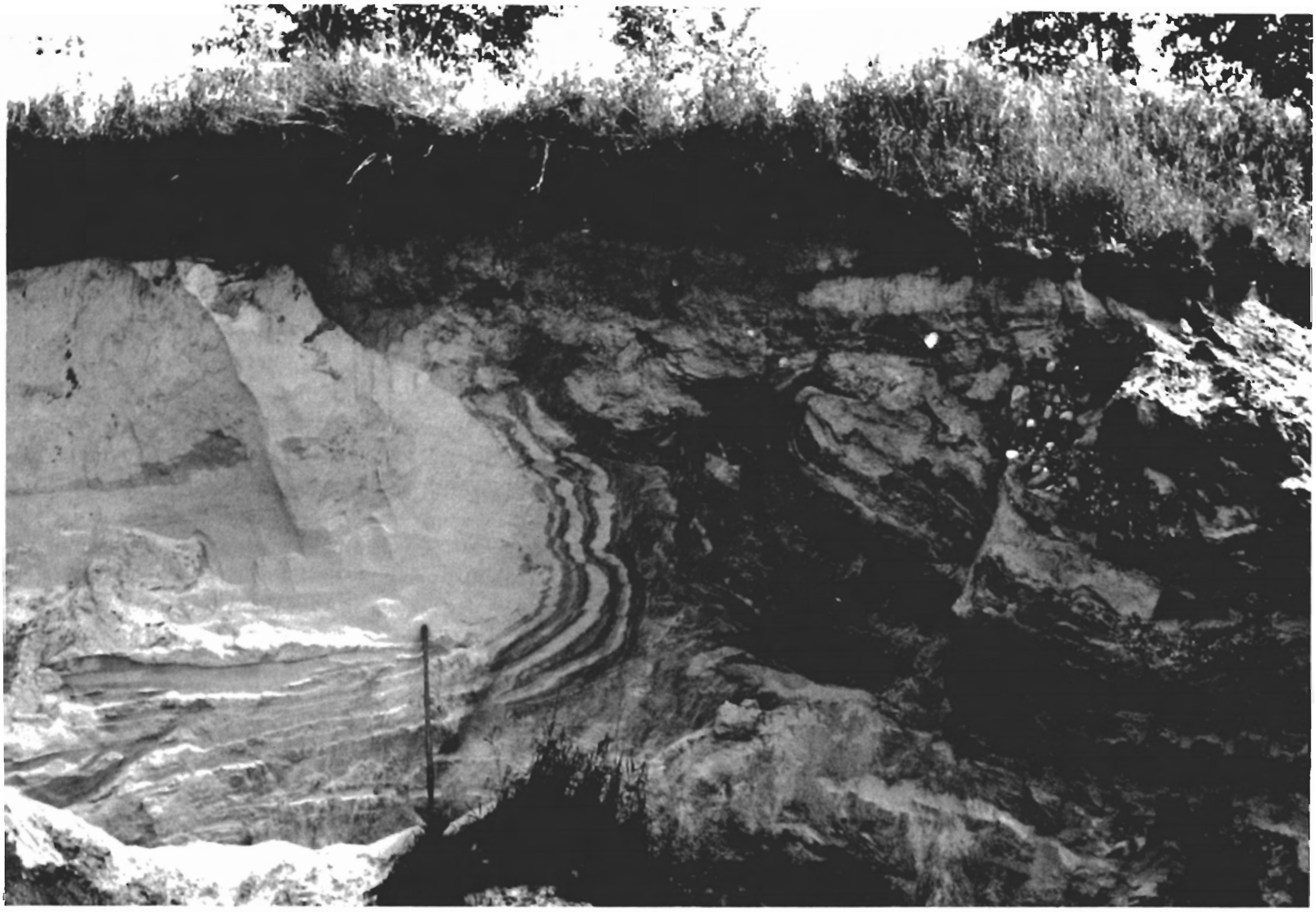

Figure 6 Southward view of deformed sand beds in stratified end moraine, south of Cobble Hill in Canaan. Height of pit face is about 5 m.

tongue. Outwash materials a short distance south of Kent can be traced westward from the Housatonic Valley to a meltwater source in the Tenmile River valley of adjacent New York. These materials suggest that the waning Hudson Valley lobe continued to deliver debris to the Housatonic River valley after active ice retreated from the Housatonic Highlands. Lacustrine silts and varved clays define small, ephemeral ponds that developed during late phases of deglaciation. Areas with northward drainage and larger valleys to the west in New York held larger, more persistent glacial lakes into which prograding outwash extended and buried or partly buried underlying lacustrine materials. The limited areal extent and thickness of ice-contact sediments associated with disintegrating ice blocks in many upland basins and valleys suggest that entrained rock debris was sparse in stagnant ice masses cut off from the active-ice sediment source.

Postglacial upland bogs contain as much as 7 m (23 ft) of organic material mixed with silt and clay. A radiocarbon date and pollen analysis obtained from buried peat indicate that the Housatonic Highlands were free of ice and forest vegetation was reestablished by 12,750 $\pm$ 230 yrs BP (RL-245, Kelley,1975a)

Glacial Lakes of the Canaan Area

The glaciolacustrine deposits of the Canaan area have been described by Holmes and others (1970), Holmes and Newman (1971), and Thompson and others (In prep.). Three closely related lakes formed during the deglaciation of the Housatonic and Hollenbeck River valleys in this part of northwestern Connecticut. Glacial Lake Lime Rock was dammed by a plug of ice-contact sand and gravel that blocked the Housatonic Valley near Cornwall (Position h, Figure 4). A series of three deltas were deposited into Lake Lime Rock as the ice margin retreated northward. The heads of these deltas indicate successive ice-margin positions southeast of Salisbury (Positions ll-1, ll-2, ll-3, Figure 4).

Glacial Lake Hollenbeck which probably was contemporaneous with Glacial Lake Lime Rock and was ponded between the ice margin and the north-draining valleys of the Hollenbeck River and Wangum Lake Brook. Lake Hollenbeck first drained southward across the Hollenbeck River-Birdseye Brook divide. Ice retreat to the north and northwest subsequently opened lower spillways, enabling the lake water to drain westward into the Housatonic Valley, probably into Lake Lime Rock. A pronounced series of sandy ridges and deltas were deposited adjacent to the ice margin in Lake Hollenbeck (Position lh, Figure 4). Some of the deltas have flat tops and grade to their respective lake levels, while other deposits probably formed as subaqueous fans.

At least some of the ice-marginal sand ridges of Lake Hollenbeck probably are end moraines. An exposure in one moraine at the Canaan town landfill displayed complex faults and recumbent folds, indicating ice shove from the northwest (Figure 6). It is inferred from topographic relationships that the ridge was deposited in water at least 15 m (50 feet) deep.

Lake Hollenbeck disappeared as the ice margin withdrew northward, allowing the lake to empty. However, meltwater continued to be ponded in the Housatonic Valley north of the bedrock threshold at Great Falls (east of Salisbury), forming glacial Lake Great Falls. This lake expanded northward from the bedrock barrier and into Massachusetts (Holmes and Newman, 1971).

Deglaciation of the Salisbury Area

Data collected while mapping the Sharon quadrangle (Thompson and others, In prep.) has defined an assemblage of meltwater deposits and other features revealing details about the deglaciation of the northwestern corner of Connecticut. Streamlined hills and bedrock striations indicate east-southeast ice flow around the southern end of the Taconic Mountains in the area between Salisbury village and the New York border. This probably was a deglacial flow trend in response to thinning of the Hudson Valley ice lobe.

A series of kames, meltwater channels and outwash deposits in the Salisbury area supports the existence of an ice mass downwasting in the vicinity of the New York-Connecticut border. The high kame deposits just northwest of Salisbury village do not contain clasts of the local Stockbridge marble. These gravel deposits were derived from the west or northwest. The low-level outwash fans, such as the one that underlies Salisbury village, also were deposited by eastward-flowing streams but the latter deposits contain locally derived marble clasts. Additional deposits were derived from an ice tongue north of Salisbury. The combined meltwater flow from ice to the north and west deposited prograding outwash over earlier lake deposits in the Salmon Creek Valley that extend southeast from Salisbury.

CORRELATIONS

The fragmentary nature of the glacial meltwater deposits in the study area precludes extensive correlation or the tracing of ice margins over long distances. Ice-margin positions between Kent and New Milford in the Housatonic River valley (Figure 4) may be correlative with the Shenandoah Moraine in New York State (Connally, this volume). A long northeast-trending ice-margin position, based on prominent meltwater channels south-southwest of Salisbury (Figure 4), may be correlative with the Hyde Park Moraine (Connally, this volume).

The plug of sand and gravel that impounded Glacial Lake Lime Rock in the Housatonic Valley (position h, Figure 4) marks a distinctive ice-marginal position that may be intermediate in time between the deposition of the Shenandoah and Hyde Park Moraines. If so, this deposit is correlative with the Poughkeepsie Moraine (Connally, this volume).

SUMMARY

Glacial deposits form a widespread but discontinuous cover on the bedrock in the Housatonic River basin and adjacent areas in northwestern Connecticut. Most bedrock outcrops in this region exhibit little weathering, but preglacial saprolite locally is preserved beneath glacial sediments. Two principal types of till are present in the study area. The upper till is Late Wisconsinan (Woodfordian) in age, while the slightly more weathered lower till is believed to represent an earlier Wisconsinan (Altonian?) glaciation. Ice-flow indicators suggest general ice-flow directions toward the south-southeast. However, evidence of late-glacial eastward ice flow from the Hudson Valley ice lobe occurs in the vicinity of the New York border.

During the Late Wisconsinan deglaciation of northwestern Connecticut, the active ice margin receded rapidly northwestward across and oblique to the upland ridge crests in response to backwasting and downwasting. Local relief restricted active ice flow and stagnation occurred in valleys which were cut off from the active ice. Melting and thinning of stagnant ice tongues in larger valleys, such as the Housatonic, caused rapid northward recession of the stagnant ice margin. The zone of stagnant ice distal to the active ice is inferred to have been 10-24 km (6-15 miles) wide. Thinning of the ice sheet over the Housatonic Highlands and southern end of the Taconic Range resulted in an irregular configuration of the ice margin.

The glaciolacustrine and glaciofluvial deposits formed along the receding ice margin have been grouped into morphosequences that indicate successive positions of the stagnant ice margin. The pattern of morphosequences and meltwater channels supports a north-northwestward retreat of the Hudson Valley lobe across the study area. Meltwater ponded in many places, either behind drift barriers or in valleys that sloped northward toward the ice margin. Glaciolacustrine deposits include deltas, subaqueous fans, lake-bottom sediments, and stratified end moraines. Outwash prograded along the Housatonic and other valleys, locally filling glacial lake basins and burying earlier lacustrine sediments beneath fluvial sand and gravel. Meltwater stream systems from New York State flowed eastward into the Housatonic region via the Still River and Tenmile River Valleys, and into the Salisbury area from the Taconic Mountains.

Ice-margin positions in the Housatonic River Valley northwest of New Milford and southeast of Salisbury are tentatively correlated with the Shenandoah and Poughkeepsie Moraines, respectively, in adjacent New York. A later position in the vicinity of the New York-Connecticut border may be correlative with the Hyde Park Moraine in New York.

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DEGLACIATION STRATIGRAPHY, MODE AND TIMING OF THE EASTERN FLANK OF THE HUDSON-CHAMPLAIN LOBE IN WESTERN MASSACHUSETTS

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ABSTRACT

The Hudson-Champlain ice lobe that occupied the middle reaches of the Hudson River lowland during retreat of the Laurentide ice sheet in late Wisconsinan time extended eastward about to 73° W longitude in the Massachusetts uplands. Striations associated with the eastern flank of the lobe trend southeast whereas many striations farther east trend southwest, recording flow on the western flank of the Connecticut Valley lobe. Because large, laterally continuous moraines are not present in western Massachusetts, the stratigraphic relations of lake deposits and the positions of ice-marginal deltas in glacial lakes provide the best available evidence for the chronology of deglaciation which was characterized by stagnation-zone retreat. Glacial lakes that require a northeast trend of the solid, grounded margin of the Hudson-Champlain lobe include Lakes Housatonic, Bascom, Tyringham, Hinsdale, Dalton, and Konkapot. In adjacent New York and Vermont, a northeast trend of the margin of the Hudson-Champlain ice lobe is required by Lakes Berlin and Shaftsbury and by a series of stages of Lake Kinderhook. Successive ice-margin positions inferred from the chronology of deposits and from assumptions used in early work by F. B. Taylor show the approximate extents of ice sublobes in valleys. A ^{14}C date of 14,000 ± 130 yrs BP from the Massachusetts upland is consistent with recent estimates of the minimum age of deglaciation of the region, perhaps as long ago as 15-16,000 yrs BP.

INTRODUCTION

The late Wisconsinan deglaciation of western Massachusetts and adjacent parts of New York and Vermont by the eastern flank of the Hudson-Champlain lobe (Connally and Sirkin, 1973) of the Laurentide ice sheet did not produce distinct, cross-country morainic till ridges or regionally extensive multiple drift sheets (Taylor, 1903). Correlation by the alignment of ice-contact heads of stratified drift from valley to valley generally is uncertain and stratified deposits otherwise traceable in the field are scarce in comparison with other areas of southern New England and with the axial part of the Champlain-Hudson lowland to the west. No clear stratigraphic evidence is known in most of the area to indicate that climatically driven regional readvances of the ice margin occurred, or even that minor or local readvances occurred. Thus no physically traceable units are available that mark coeval retreatal positions of the ice margin across the area.

The evidence that is available for the pattern, mode, and timing of glacier retreat in the region includes indicators of ice-flow direction and ice-marginal and other sediments deposited in the succession of ice-marginal lakes that formed during deglaciation. These features permit a detailed correlation of map units within and among the numerous glacial lake basins in Massachusetts, New York, and Vermont, and are the basis of the deglaciation stratigraphy proposed in this paper.

The purpose of this paper is to present some of the results of field mapping and of compilation of published and unpublished reports carried out as part of the U.S. Geological Survey program of surficial geologic mapping in Massachusetts in cooperation with the Massachusetts Department of Public Works. These results, which build on and update the classical regional study by Frank Bursley Taylor (1903) on the recessional ice borders in Berkshire County, Massachusetts, are related to recent work in adjacent parts of New York (D. H. Cadwell, 1981) and to the published deglaciation stratigraphy of southwestern Vermont (Shilts and Behling, 1967; Stewart and MacClintock, 1969; Stewart and MacClintock, 1970).

The principal results include data on the directions of ice movement and on the distribution, character, and relative ages of stratified-drift deposits. These data specify the positions, or at least the trends, of certain ice margins that acted as dams for glacial lakes or that supplied sediments to ice-marginal and other deposits.

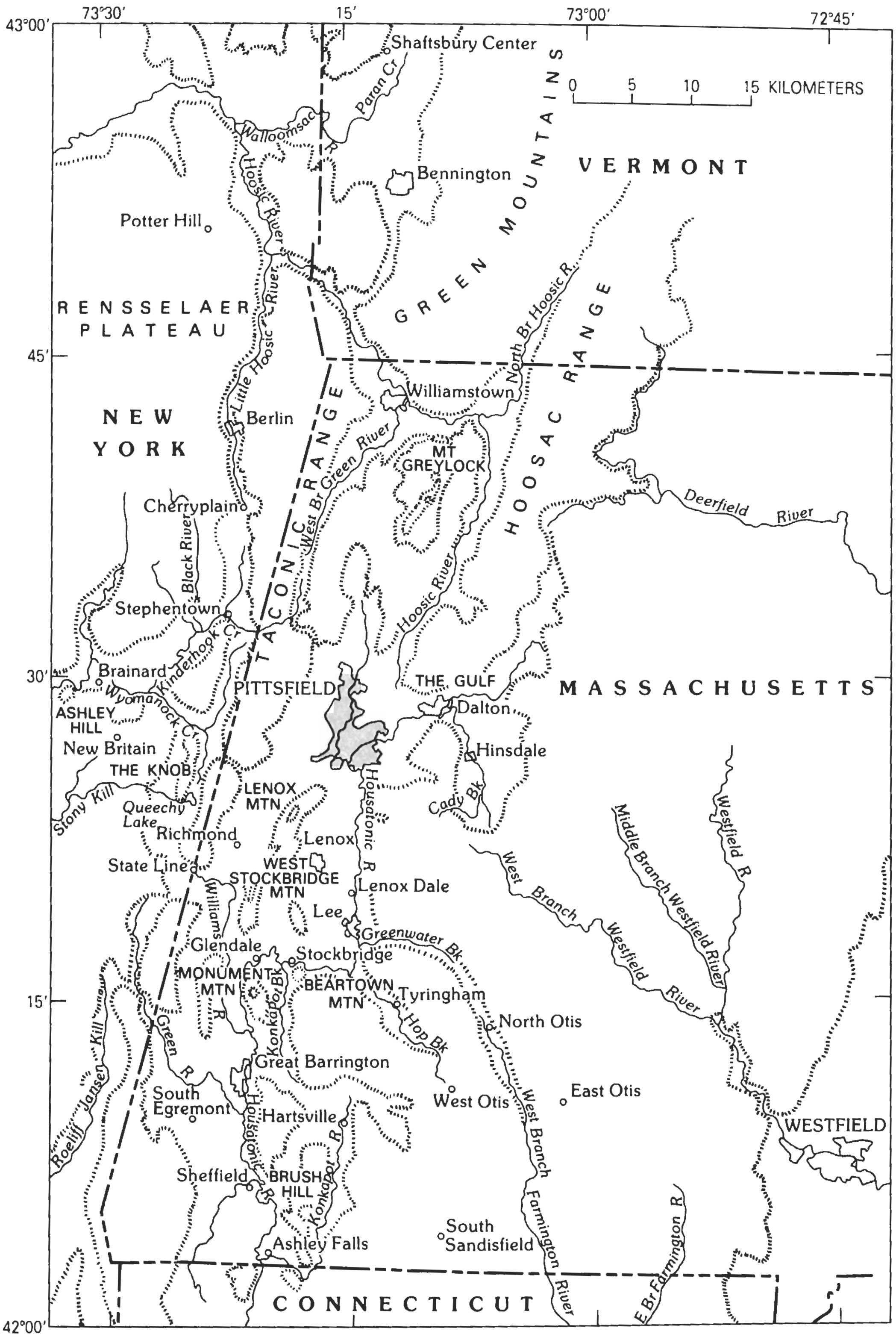

Figure 1 Map of western Massachusetts and adjacent parts of New York, Vermont, and Connecticut showing principal rivers, upland areas and mountains, and localities mentioned in text. The Massachusetts area is part of the Berkshire Hills. Hachured lines, drawn at bases of steep slopes, outline major upland areas and mountains.

These relationships imply approximately coeval ice-margin positions within different basins. The trends of retreatal ice-margin positions indicate a basic chronostratigraphic framework for the deposits that permits locally precise stratigraphic correlation and approximate regional correlation of the principal deposits in western Massachusetts and adjacent parts of New York and Vermont.

The postglacial ^{14}C dates of the region, obtained from the organic-carbon fraction in lacustrine sediments or from peat or organic-rich sediments in bogs, are reported by Connally and Sirkin and by Cotter and others in this volume. Two additional ^{14}C dates from the uplands of western Massachusetts are consistent with these regional estimated ages of the deglaciation.

Regional Setting

Most of western Massachusetts is a hilly upland of moderate relief known as the Berkshire Hills. Mount Greylock rises over 1,060 m (3,480 feet) and highlands in much of the area, especially the Taconic Range and other uplands along the state border in the west (Fig. 1), stand above 600 m (1970 feet). West of this upland, largely underlain by resistant plutonic igneous and metamorphic rocks, the general altitude and relief decrease across the Rensselaer Plateau in New York State that is underlain principally by graywacke, to the Hudson River lowland, underlain by carbonate rocks and shales. To the east, the terrain descends to the Connecticut River Valley that is narrow in the north where it forms the Vermont-New Hampshire boundary but which widens to the south in Massachusetts where it forms a lowland carved in the weak shales and arkosic sandstones of Late Triassic and Early Jurassic age.

The principal streams draining the Berkshire Hills (Fig. 1) include the Housatonic River that flows south in the southwest part of Massachusetts, and the Hoosic River that flows northwest past Williamstown into Vermont and then west to the Hudson River. The Deerfield and Westfield Rivers flow east and southeast, respectively, to the Connecticut River. The West Branch of the Farmington River and several tributary streams drain south into Connecticut before joining and turning east to the Connecticut River. These master streams flow at altitudes usually below 300 m in valleys that are generally 300 to 400 m (985 to 1310 feet) below the upland surface.

The sides of these main valleys generally are steep. Where the valley is floored by marble, the steep slope generally is on the resistant rocks adjacent to the marble; both the Housatonic and Hoosic Rivers flow in open, marble-floored valleys for most of their lengths in Massachusetts. Where the marble outcrop belt is wide, the valleys widen out into basins, as at Pittsfield, Stockbridge, Sheffield, and Williamstown. The headwaters of Kinderhook Creek in Massachusetts also follow a valley cut in carbonate rocks.

Previous Studies

The first recorded observations concerning glacial geologic phenomena in the study area appear to be those made by a Dr. Stephen Reed, of Richmond, Massachusetts, in 1842 (see Taylor, 1910). Reed described a line of erratic boulders that extends southeast from The Knob, in New York, through the town of Richmond, Massachusetts. Reed's "Richmond Boulder Train" was much discussed in the following decades (see, for example, bibliographies in Taylor, 1910, and in Kelley and Newman, 1975) and was a significant factor in the final acceptance of the glacial theory.

The general pattern and relative timing of glaciation and deglaciation of western Massachusetts have been known since F. B. Taylor (1903) published the results of his reconnaissance studies of the area. He described his field evidence for a retreating lobe of ice centered in the Hudson River valley. Taylor inferred that the last ice sheet moved southeast across western Massachusetts from the Champlain-Hudson lowland. He mapped the distribution of hummocky till in recessional moraines, border drainage channels, and deltas deposited in glacial lakes. All of these he considered to be ice-border features related to recession of the glacier margin.

Areas with hummocky topography that did not seem to be caused by relief of the bedrock surface were mapped as moraines by Taylor. Many of these areas are underlain by relatively thick till but some of his moraines include kames and other stratified drift. Taylor interpreted the moraines as sediments deposited under the ice along a glacier front or margin where the ice had thinned and could no longer carry its load. He correlated these moraines from valley to valley on the basis of their occurrence in successive steps in parallel valleys.

Taylor used the term "border drainage channel" to describe shallow valleys, generally cut in till and now dry or carrying grossly underfit streams, that slant diagonally down a slope. Taylor appropriately inferred that such a valley can only have been carved by a stream that had ice forming one of its stream banks, that is, by a meltwater stream flowing along an ice margin.

From the deltas, Taylor inferred the former presence of glacial Lakes Housatonic and Bascom. Other deltas that Taylor mapped record Lakes Hinsdale and Dalton of this paper, and two lakes in New York. Evidence for the lakes included not only the deltas built into them

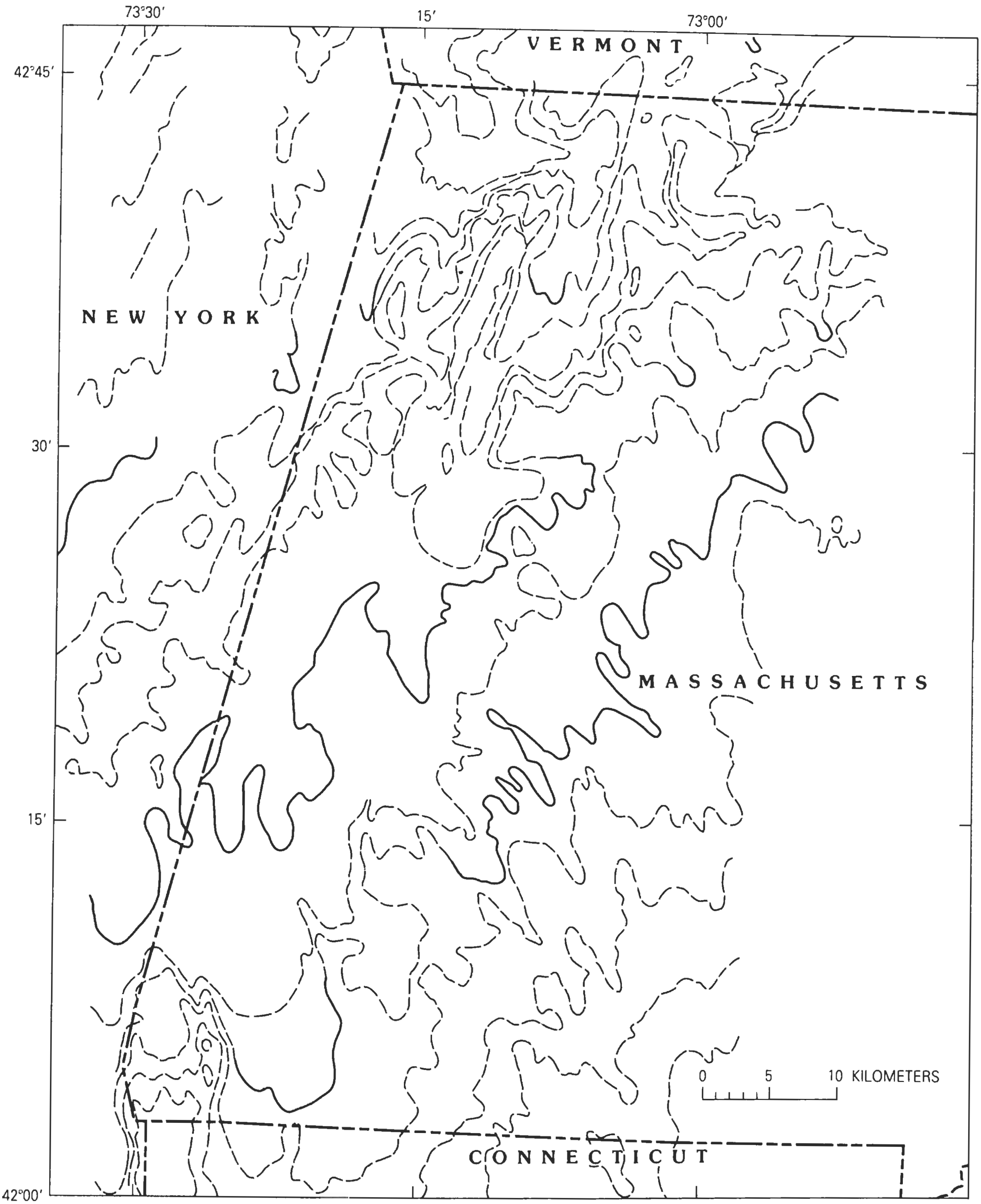

Figure 2 Map showing selected recessional ice borders as mapped by Taylor (modified from Taylor, 1903, fig. 10). Solid lines show correlations strongly inferred from alignment of moraines or stratified drift in upland valleys, or from glacial Lake Housatonic in the Housatonic lowland.

but also the presence of lake spillways carved at elevations corresponding to the general altitudes of the correlated delta plains. These lakes, as Taylor noted, required the presence of ice that was continuous and firmly aground to retain them. Based on the synthesis of all his field evidence, Taylor constructed a map showing the positions of a series of recessional ice borders across Berkshire County, Massachusetts and adjacent parts of Connecticut, New York, and Vermont (Fig. 2).

Taylor supplemented the 1903 paper with unpublished data in 1903 to 1908 and a short published paper (Taylor, 1910) and an abstract (Taylor, 1916). In his later work, as shown by his unpublished maps of the region, Taylor dropped his emphasis on the use of moraines to define the alignments of the ice-margin positions and stressed instead the importance of stratified drift, including kames and fluvial deposits. He also noted the significance of clay deposits that confirm the former presence of glacial lakes. The 1910 paper discussed Reed's Richmond boulder train and described a second "train" or fan, consisting of boulders of the same rock type and provenance that Taylor named the Great Barrington train.

After Taylor, little research in Quaternary geology was done in western Massachusetts for more than half a century, though J. W. Goldthwait (*in* Antevs, 1922) included striation data from this area in his map of the regional pattern of the ice sheet. Then, beginning in 1964, G. William Holmes (for example, Holmes, 1968) made a series of reconnaissance maps at 1:24,000 showing the distribution of surficial materials in 23 7.5-minute quadrangles. Scattered quadrangles were mapped in greater detail for the GQ series of the U.S. Geological Survey beginning with Zen and Hartshorn (1966). In one of these, Holmes and Newman (1971) mapped the general distribution of clays deposited in Glacial Lake Great Falls (their "Falls Village lake"). Norvitch and others (1968) published a map showing the generalized distribution of sand and gravel in the Housatonic River basin. Newton and others (1975) briefly discussed the glacial deposits of the Housatonic River valley. Kelley and Newman (1975) discussed the Richmond boulder fan. Warren and Harwood (1978) inferred the distribution of a series of recessional ice fronts of late Wisconsinan age in the Ashley Falls and South Sandisfield quadrangles.

Shilts and Behling (1967) and Shilts, (*in* Stewart and MacClintock, 1969) described the extension of Glacial Lake Bascom into Vermont and deduced that when the ice margin had retreated far enough to uncover a col at Potter Hill in New York State, the level of Lake Bascom dropped to the level of the Potter Hill spillway. Shilts gave the name Glacial Lake Shaftsbury to the lake thus lowered and shrunken.

GLACIATION

The advance of the Laurentide ice sheet across western Massachusetts occurred before about 20-21,000 yrs BP, which is the approximate age assigned to the Roslyn Till and associated meltwater deposits in the Harbor Hill Moraine of western Long Island (Sirkin and Stuckenrath, 1980; Sirkin, 1982; Connally and Sirkin, this volume). This terminal moraine marks the southernmost advance of the sector of the Laurentide ice sheet that covered western Massachusetts. Peat-ball clasts within meltwater deposits of late Wisconsinan age in the Housatonic River valley in Connecticut have yielded radiocarbon ages of 28,000 $\pm$ 100 yrs BP, >33,000 yrs BP (W-2043 and W-2174; R. L. Melvin, U.S. Geological Survey, Hartford, Conn.), and >40,000 yrs BP (W-2615; W. S. Newman, Queens College, N.Y.). The first of these dates seems to indicate that the ice advanced across the area after 28,000 years ago.

At the climax of the late Wisconsinan glaciation of New England and during the initial phases of deglaciation, ice moved southeast across western Massachusetts, as shown by glacial grooves and striae with azimuths of 130° to 140° on most of the higher summits and in many upland valleys throughout the area (Fig. 3). The long axes of drumlins in the uplands in the central part of the area are parallel with the trends of nearby upland striations and they therefore confirm the inference that the flow of the ice was southeastward.

The direction of ice movement near the time of glacial maximum also is indicated by the Richmond boulder train, a fan of boulders derived from a distinctive amphibolite that crops out in The Knob in New York State. Large, angular, unweathered boulders of this rock were carried southeast across the upland southwest of Pittsfield (Fig. 3), where they are scattered across hills and valleys in complete disregard of the topography. Taylor (1910) suggested that these boulders were plucked from The Knob and carried "on or in the upper part of the ice" while ice was thick across the area.

DEGLACIATION

Topographic Control of Ice Flow

In response to thinning and melting back of the ice sheet margin during the initial phases of deglaciation, major ice lobes developed in the Champlain-Hudson and Connecticut valleys. Radial patterns of striae record the lobation (Taylor, 1903, p. 328; Goldthwait, *in* Antevs, 1922; Dineen, this volume; Larsen and Hartshorn, 1982). Thus most striae near the western side of the Connecticut Valley trend southwest to west (Fig. 3), and one

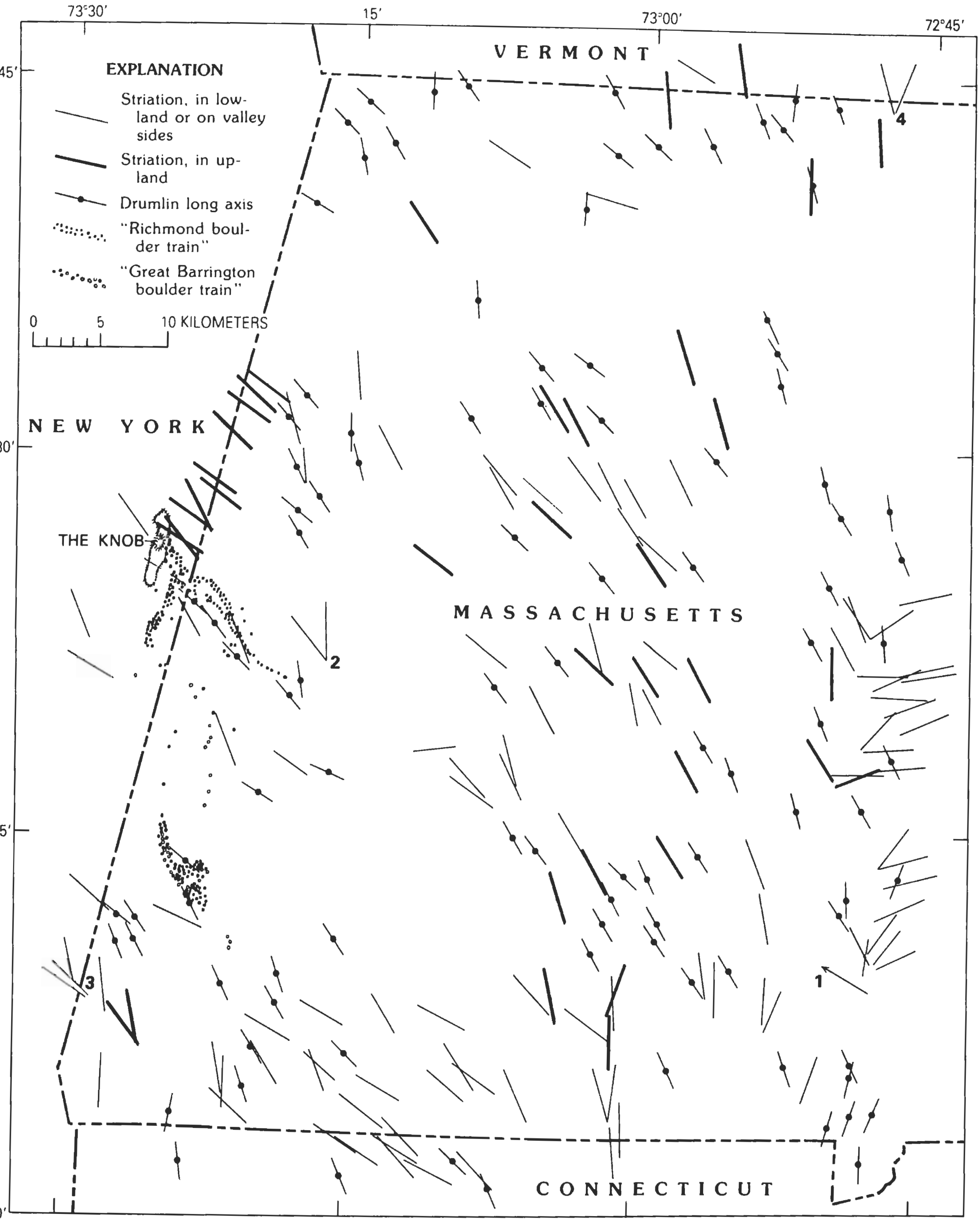

Figure 3 Map showing ice-flow indicators of western Massachusetts and adjacent parts of New York, Vermont, and Connecticut. Data from published quadrangle reports and unpublished data of F.B. Taylor and C.R. Warren. Distribution of Richmond and Great Barrington boulder trains modified from Taylor (1910, fig. 1). Numbers indicate localities cited in the text.

striation with a unique azimuth in the Westfield River valley even trends north of west (locality 1, Fig. 3). Some upland striae near and east of the 73° W meridian trend roughly north-south and suggest that the influence of the Connecticut Valley ice lobation extended as far west as that line.

The long axes of drumlins (Fig. 3) show a pattern similar to that of the striae, though they show less dispersion in azimuth. Many of the drumlins here, as in eastern Massachusetts (Hansen, 1956; Stone and Peper, 1982), contain cores of a till that probably dates from one or more pre-late-Wisconsinan glaciations. However, the present shapes of some of the drumlins and the azimuths of their long axes appear to have been determined during the retreat of the last ice sheet. Most of these drumlins trend southeast, but some on the west side of the Connecticut Valley and in the lower Housatonic River valley trend southerly to southwesterly and several in the uplands show deflection of ice flow by high ground.

During the later phases of deglaciation, the movement of the ice was even more strongly influenced by the local topography. Many striae and drumlins within the Housatonic River lowland show southerly deflection of ice flow. Locally, striae record multiple directions of ice movement. For example, striae exposed in the floor of a former pit just east of Lenox (locality 2, Fig. 3) trend in various directions but nearly all are included in two groups. One group ranges in azimuth from 120° to 130°, averaging about 125°; the other ranges from 165° to 185° with an average of about 175°. Presumably, the more easterly striae were cut earlier by ice that moved southeast across the top of Lenox Mountain, whereas the more southerly ones were cut later by ice that moved south within the valley and that flowed around the north end of Lenox Mountain. Similar deflections of ice flow are recorded by striae west of South Egremont (locality 3, Fig. 3) and in the northeast corner of the study area (locality 4, Fig. 3).

In addition to the "Richmond boulder train," Taylor (1910) described a second fan of amphibolite boulders, "the Great Barrington Train," that was also derived from The Knob (Fig. 3). Unlike the boulders in the Richmond fan, those in Taylor's "Great Barrington Train" are small, well rounded, and deeply weathered. They also have a different distribution – they extend to the south within valleys rather than to the southeast across uplands; they are confined to the west side of the Taconic range as far south as State Line; they are distributed to and beyond Great Barrington, to a distance nearly twice as far from The Knob as the Richmond boulders are recognized. Taylor suggested that these Great Barrington boulders were derived from a talus deposit that had accumulated and been weathered in preglacial or interglacial time. The talus deposit lay on the southeast slope of The Knob where presumably it was protected in the lee of The Knob while the thick ice moved southeast across it. Only later did it become subject to erosion, after the ice had thinned enough so that it was deflected to the south by the Taconic Range. However, this interpretation has been questioned by Kelley and Newman (1975); citing new detailed bedrock mapping by N. M. Ratcliff, they concluded that the boulders in the Great Barrington Train may not be derived solely from the outcrops at The Knob.

West of about 73° W longitude, the striae, boulder-train, and drumlin data and the ice-margin alignments required to dam in the glacial lakes all consistently indicate that, throughout the deglaciation, ice related to the eastern flank of the Hudson-Champlain lobe prevailed. The data further suggest that the ice of that eastern flank remained active during glacier thinning and retreat and that its margin developed a highly irregular pattern of sublobes that extended south and east in the larger valleys of Berkshire County. For the purpose of this report, ice that deglaciated all of western Massachusetts west of 73° W longitude is included in the eastern flank of the Hudson-Champlain lobe.

Till Deposits

Sandy till of late Wisconsinan age forms a discontinuous mantle over the area that is absent on many steep slopes and hill crests. Hummocky till, locally mantled by boulders, masks bedrock topography in places and is inferred to range up to 15 m in thickness. No till ridges, boulder lines, or other such features known to us in western Massachusetts are comparable to those interpreted as moraines or moraine segments in coastal southern New England (Flint and Gebert, 1976; Goldsmith, 1982; Stone and Peper, 1982). A transverse, reportedly till-mantled linear ridge (Fig. 6a) 2 km north of Queechy Lake, New York has been interpreted as a moraine (Kelley and Newman, 1975) but alternatively the ridge may be chiefly an ice-marginal lacustrine deposit laid down on the valley bottom in Glacial Lake Kinderhook. Taylor's unpublished maps compiled from detailed field studies completed after 1903 include significantly fewer areas identified as terminal moraines than shown on his 1903 map. Many of the areas that he still accepted as moraines are coarse-grained, ice-marginal heads of meltwater deposits, though some are hummocky till.

In a few places erosional channels cut in till indicate positions of glacier-ice margins on hillslopes. Taylor (1903, p. 338) described such a "border drainage" channel that is well preserved (Fig. 4). A shallow valley, now

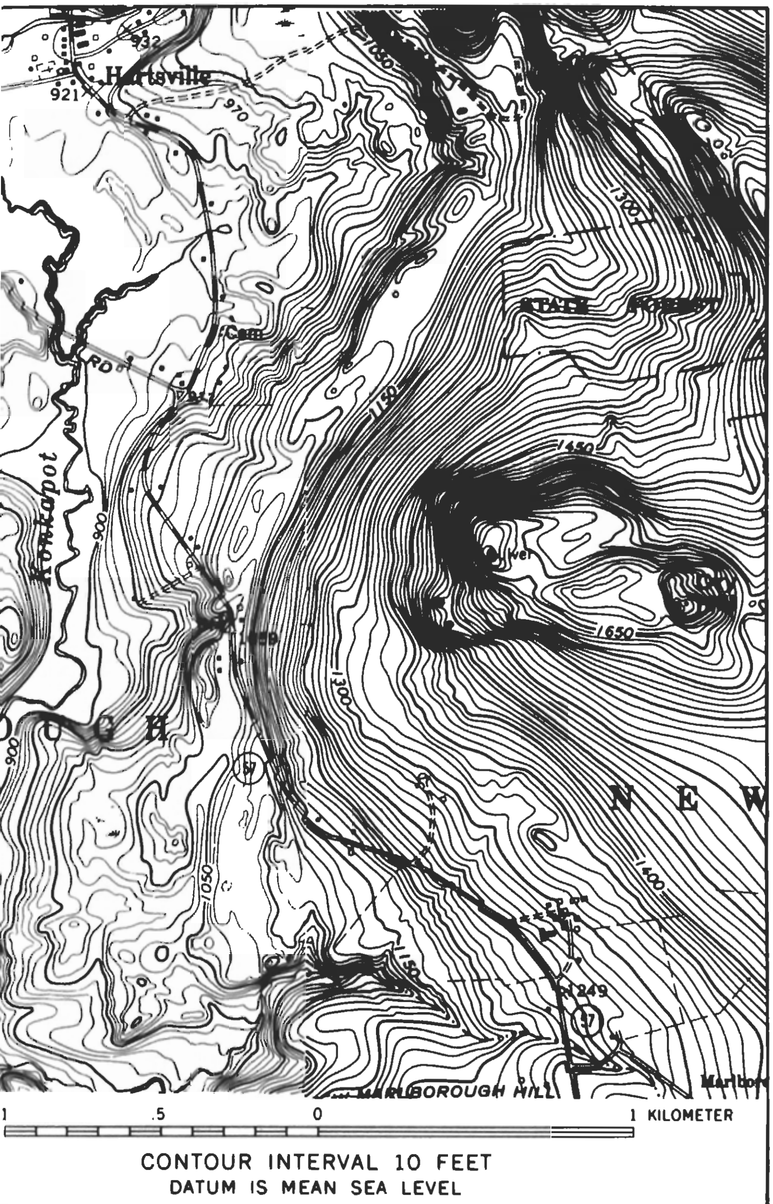

Figure 4

dry, slants diagonally down a mountainside about 1.3 km southeast of Hartsville and extends more than 2 km southwest. It is held in by a ridge of till, mantled by stratified drift, imposed on a slope that faces northwest. As Taylor noted (1903, p. 331), the development of such a river bed "would be altogether impossible without the immediate presence of (an) ice front to serve as one of the retaining banks." All such channels known to us in the study area slope in southerly directions (see Taylor, 1903, fig. 8), which indicates a consistent control of the aspect of erosional meltwater flow away from the retreating ice margin.

Meltwater Deposits

Because large, laterally continuous recessional moraines are not present in western Massachusetts, we base our deglaciation stratigraphy and the pattern of ice-margin retreat on the relationships of meltwater deposits, chiefly deposits laid down in glacial lakes. Similar deposits in central eastern New York, including extensive glacial lake deposits in the Kinderhook Creek and Hoosic River basins, are compatible with the system of map units that comprise the deglaciation stratigraphy of the region (Fig. 5). Additional map units in New York, such as deltas, kames, kame moraines and other local deposits that D. H. Cadwell (unpub. data) has identified as moraines, also may be included in the deglaciation stratigraphy.

Deposits of twelve major glacial lakes in western Massachusetts, southwestern Vermont, and central eastern New York comprise the basic stratigraphic elements that are related to the recession of the Hudson-Champlain lobe in the region (Fig. 5). The deposits typically comprise a variety of morphologic features (Fig. 6) that form discrete, mappable sedimentary units called morphosequences (Koteff and Pessl, 1981; "sequences" of Jahns, 1941). Morphosequences, chiefly ice-marginal and non-ice-marginal deltas (ice-contact and non-ice-contact morphosequences of Koteff and Pessl, 1981, Fig. 5), are correlated with, and combined within the map units of major glacial lakes of the region.

Ice-marginal ridge and channel. The 349 m (1,145 ft.) hill 1.3 km east-southeast of Hartsville forms the northeast end of a ridge that extends southwest, unbroken for 1.5 km, to the road just northwest of BM 1059. South of the break at the road, the ridge extends south for nearly another kilometer. The ridge consists in part of stratified drift (sand to bouldery gravel) deposited by meltwater that came off adjacent ice but is probably largely till. It slants diagonally down the northwest-facing slope of Dry Hill. The channel southeast of the ridge was the course of an ice-marginal stream; it is now dry except ephemerally after heavy rains, though a small stream coming from an artesian well up the hill to the east crosses it near the benchmark. A photograph taken across the channel to the ridge was published as Fig. 4 of Taylor, 1903. Scale, 1:24,000; contour interval 10 ft.

The map units shown in Figure 5 include deposits of all stages of each lake. They include glaciofluvial beds in tributary valleys that are graded to deltas, delta topset, foreset, and bottomset beds, and lake-bottom sediments. The latter are chiefly silt and clay but include ice-marginal lacustrine sand and gravel. The glaciofluvial and deltaic facies include discontinuous deltaic kame terraces in major valleys as well as ice-marginal deltas. They are distinguished from the finer-grained, lake-bottom facies in Fig. 5. The delta deposits consist of as much as 5 m of fluvial topset gravel and sand overlying foreset beds of sand, pebbly sand, and silt. Sets of foreset beds commonly are 3 to 10 m thick, and generally dip 15° to 25°. They include beds with planar and ripple-drift bedforms and commonly fill channels that are 10 to 50 m wide (Fig. 7a). Glaciolacustrine deltas in the region, although sparsely distributed, are similar morphologically and sedimentologically to deltas of other glacial lakes (Stone and Force, 1982). Commonly, sand and gravel foreset beds in deltas in the Housatonic-Hoosic marble lowland are cemented by calcium carbonate (Fig. 7b).

Coarse-grained ice-marginal deltas are typical of upland glacial lakes and are common in parts of larger, lowland lakes. They contain topset beds of cobble-to-boulder gravel overlying foreset beds of pebble-to-cobble gravel and sand with planar bedding. Outsize, commonly angular boulders are often present; these are considerably larger than the more rounded clasts that comprise the bedding. They generally record the presence of glacier ice in the immediate vicinity at the time of deposition, although some boulders may have been rafted in by icebergs. Collapse deformation is common, including warped bedding, folds, and high-angle faults (Fig. 8). Ice-channel deposits in ridges and ice-marginal deposits in discontinuous flat-topped terraces comprise ice-contact parts of individual deltaic and fluvial deposits (Fig. 6). Delta bottomset beds of silt and fine sand grade laterally into lake-bottom silt and clay deposits that are locally as much as 75 m thick in the lower Housatonic River valley (Norvitch and Lamb, 1966). Lake-bottom sediments are exposed or are known from subsurface data in most of the glacial lake basins; these include the clay deposits of Glacial Lake Bascom in the Hoosic River basin (Taylor, 1916).

The positions of ice-marginal deltas and correlated lake spillways, and the superposed and inferred intertonguing relationships of deltaic deposits over extensive lake-bottom sediments, indicate the extent and shape of the controlling glacial lakes. The distribution of deposits and the altitudes of correlated deltas and lake spillways of the principal glacial lake units of the study are are summarized in Table 1. The variation in altitudes of deposits and spillways of Lakes Housatonic, Bascom, Kinderhook, and Shaftsbury indicate postglacial isostatic tilt of the region of about 0.75-0.85 m/km (4-4.5 feet/mile) upward to the north.

Lake Kinderhook

The method of combining morphologic and sedimentologic features in a single map unit is well illustrated by the numerous deposits laid down during four stages of Glacial Lake Kinderhook (Table 1). Lake Kinderhook came into existence when the level of the ice in the Kinderhook Creek drainage dropped below the level of the head of the valley northeast of Hancock, Massachusetts. However, deposits in the upper part of the Kinderhook valley in Massachusetts are at altitudes of 360 and 344 m (1,180 and 1,130 feet), and no stable spillways appropriate to these altitudes have been identified; these deposits are probably ice-marginal terraces not controlled by stable levels of Lake Kinderhook.

The four lake stages recognized in this report (Table 1; Fig. 9) occupied the Kinderhook valley from above Hancock, Massachusetts, to Brainard, New York. They are named for the locations of their respective spillways. From oldest to youngest, they are the Queechy Lake stage, the Canaan Road stage, the New Britain stage, and the Ashley Hill stage. The position of accordant ice-marginal deltas in the northern part of the Queechy Lake-stage basin and the deltas deposited into the lower lake stages by tributary streams and by outwash in the main valley indicate that four different water planes existed within the basin at different times, each controlled by an associated spillway.

Coarse-grained deltas of the Queechy Lake stage are present in the northern part of the basin, 4 to 6 km north of Stephentown (localities 1 and 2, Fig. 9), at the sides of valleys or at the mouths of tributary valleys. They have flat fluvial surfaces at altitudes of 1,090 and 1,110 feet (332-338 m); the thickness of the deltaic sediments ranges between 20 to 35 m. Erosion by meltwater streams and minor erosion by meteoric streams has truncated the sides and fronts of many of these deltas. Elsewhere, the uncollapsed topography of the frontal slopes of digitate deltas is preserved, as in the large delta 1 km north of Stephentown (locality 3).

Some coarse-grained deposits of the Queechy Lake stage were laid down beneath the margin of the ice in deep lake water; for example, the collapsed ice-marginal gravel and sand exposed in laterally continuous beds that dip southerly at less than 10° in the Lebanon Springs landfill are assigned to the Queechy Lake stage (locality 4, Fig. 9). Sources of meltwater and of sediments for other deltas appear to have been at the heads

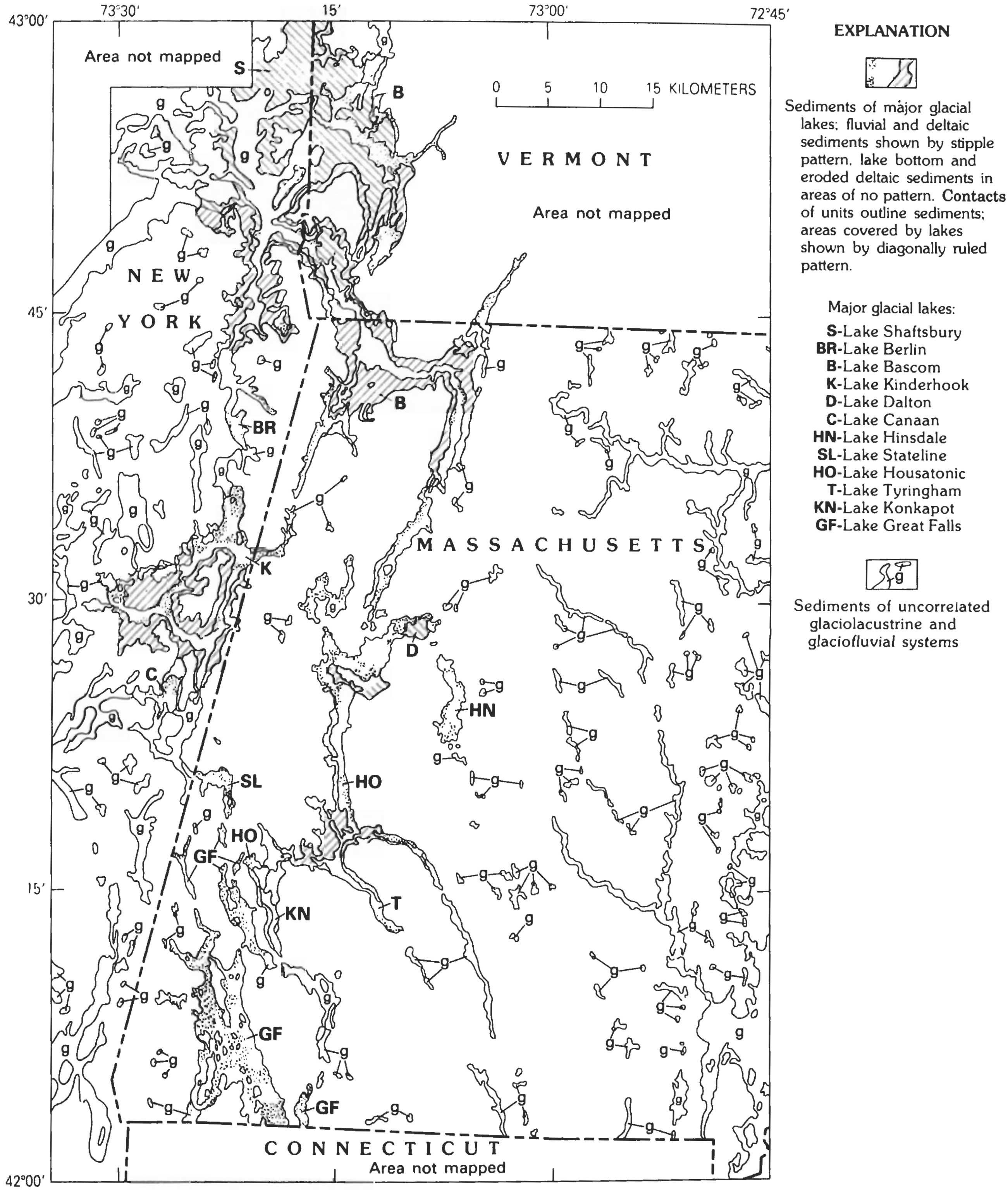

Figure 5 **Preliminary map of late Wisconsinan stratified deposits of western Massachusetts and adjacent parts of New York and Vermont. Data in Massachusetts from published quadrangle reports and unpublished data of C.R. Warren, F.B. Taylor, B.D. Stone, J.D. Peper, and J.H. Hartshorn. Data in New York from Cadwell (1981) and unpublished data of D.H. Cadwell and B.D. Stone. Data in Vermont from Stewart and MacClintock (1970) and modified from Shilts (*in* Stewart and MacClintock, 1969).**

of tributary valleys that drained to deltas in the main valley north of Stephentown. A large pit 2 km west of Stephentown (locality 5) exposes noncollapsed delta foresets that dip south to southwest. The topset-foreset contact in this delta is at about 329 m (1,080 feet) altitude.

Several deltas of the Canaan Road stage are present southwest of Stephentown. These occur at the mouths of tributary valleys (localities 6 and 7, Fig. 9) and stand at altitudes of 317-323 m (1,040 to 1,060 feet).

Two flat surfaces, at altitudes of 287-290 m (940-950 feet) 1 km north of Stephentown (locality 9), are correlated with the New Britain stage of Glacial Lake Kinderhook. Both surfaces are underlain by glaciofluvial sand and gravel, and the morphology of these deposits suggests that the glaciofluvial sediments overlie finer-grained beds, probably foresets. Their positions within the valley and their relationships to higher deltas show that these two deltas were built by meltwater streams that eroded older deposits and deposited reworked detritus as inset deltas. The nearly flat surface at altitudes of 290-297 m (950 to 975 feet) located at the head of Wyomanock Creek southeast of Stephentown proabably resulted from lateral planation and fluvial deposition when meltwater streams in upper Kinderhook Creek basin were graded to the New Britain stage of Glacial Lake Kinderhook.

Deltas of the Ashley Hill stage (locality 10, Fig. 9) lack evidence of ice-marginal deposition. They are in the western part of the drainage basin, at altitudes of 212-218 m (695 to 715 feet). The large delta at the confluence of Kinderhook and Wyomanock Creeks appears to have been scarped in postglacial time and originally may have filled the valley; the delta probably correlates with small meltwater terraces that are cut into older deposits farther up the Kinderhook Creek valley. A terrace at 212 m (695 feet) 2 km northeast of Brainard (locality 11) probably was cut by lateral planation by meltwater in the Kinderhook Creek valley just after Lake Kinderhook was finally emptied.

ICE MARGIN RETREATAL POSITIONS, MAP UNITS, AND MODE OF DEGLACIATION

Selected ice margin positions related to the 12 principal stratigraphic units, the distribution of deposits, the extent of lake-water planes, and identified lake spillways are shown in figure 10a. The ice-margin positions are inferred from: 1) ice-contact morphology and sedimentary facies of individual ice-marginal deposits (morphosequences); 2) positions of dams required to impound drainage basins behind glacial-lake spillways; and 3) ice at the heads of tributary valleys required as sources of meltwater and of glacially derived sediment. These retreatal ice-margin positions are directly related to the map units and show the local trend, shape, and pattern of the ragged margin of stagnant ice related to the Hudson-Champlain lobe in the region.

The pattern of these retreatal ice-margin positions and their altitudinal relationships indicate that the general direction of ice-margin retreat was toward the northwest, though local trends commonly were related to the pattern of major valleys and the cols between mountain massifs. The deposits of the upper stages of Lake Housatonic and deposits of Lakes Hinsdale, Dalton, and Konkapot step down to lower and younger deposits of Lake Housatonic in the Housatonic lowland. Deposits of Lakes Bascom, Berlin, Shaftsbury and Kinderhook show a similar stepwise lowering of these respective lake levels to the northwest. These relations strongly demonstrate that the direction of the regional retreat of the ice margin was from southeast to northwest. The local stratigraphic relationships between ice-marginal deposits (Fig. 6) and the succession of glacial lakes in all of the basins in the area indicate that the deglaciation by the eastern flank of the Hudson-Champlain lobe occurred principally by stagnation-zone retreat. The width and extent of a coherent and grounded marginal zone of stagnant ice probably varied with valley size and morphometry. For example, Cadwell (1981) emphasized some evidence for the extent of residual stagnant ice in the Kinderhook Creek basin. The inferred ice-margin positions (Fig. 10a) suggest valley sublobes as long as 15-20 km in parts of the Housatonic lowland.

REGIONAL CHRONOSTRATIGRAPHY AND DEGLACIATION HISTORY

The trend and alignment of local ice-margin retreatal positions and the succession of proglacial lakes with the requisite meltwater drainage paths and ice-margin dams, indicate the relative chronostratigraphic regional relationships among map units. These relationships, including correlated retreatal ice-margin positions referred to numbered deglaciation phases in this report, are summarized in the correlation diagram (Fig. 11). Extensions of ice-margin lines that are controlled by stratigraphic evidence are illustrated in Figure 10b. The detailed pattern of these preliminary regional retreatal positions is derived from a projection of glacier surface slope of about 20 m/km around major topographic elements. These regional ice-margin positions are best controlled in the lower Housatonic lowland and in the Kinderhook and Hoosic basins where the patterns indicate that glacial sublobes, fed by active ice moving through cols between upland areas, were present.

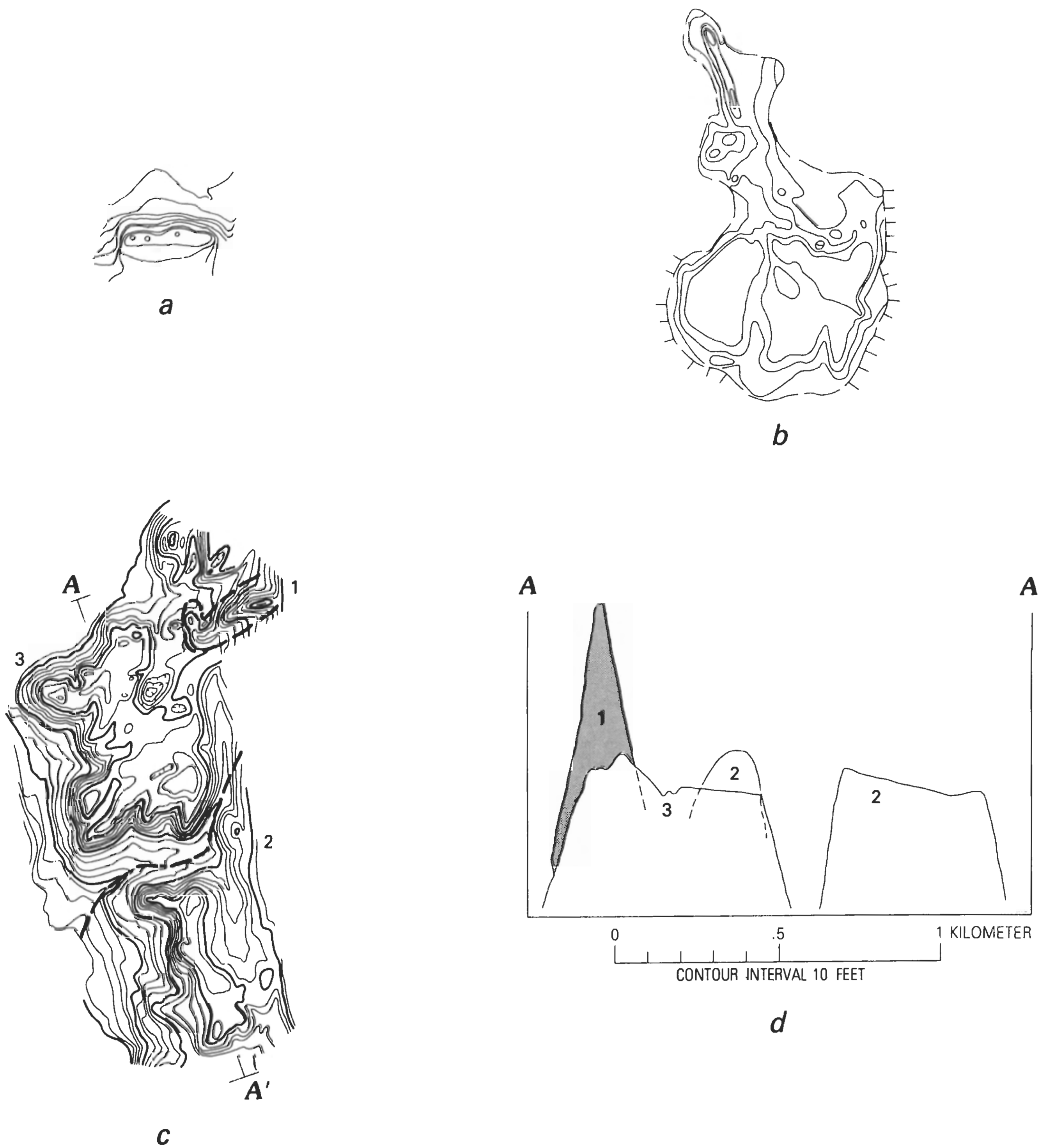

Figure 6 **Morphology and profiles of ice-marginal morphosequences (scale 1:24,000; contour interval 10 ft). a) Probable ice-marginal lacustrine deposit of Queechy Lake stage of Lake Kinderhook (identified as small moraine by Kelly and Newman, 1975, p. 278); b) ice-marginal delta, glacial Lake State Line; c) successive deposits: (1) ice-channel deposit, (2) older ice-marginal delta, (3) younger ice-marginal delta in Konkapot River basin at Mill River village (map unit Qcd2 of Holmes and Newman, 1971); d) projected topographic profiles of deposits shown in c.**

TABLE 1. PRINCIPAL GLACIAL LAKES AND ASSOCIATED SPILLWAYS AND DEPOSITS

Map Symbol on Fig. 5	Lake name (reference)	Controlling spillways (altitude, in feet)	Deposits in or graded to the lake or its spillway (altitude, in feet). Deposits have been raised relative to the controlling spillway by postglacial tilt of 0.75-0.85 m/km (4-4.5 ft./mi.) up to the north
HO	Lake Housatonic (Taylor, 1903; modified in this report)	1. Uncertain; probably stratified drift NW of Shaw Pond (near 1,400)	Sand, some gravel, in a kame terrace (to 1,435; lake level near 1,415)
		2. Saddle to West Branch Farmington River SE of Greenwater Pond (1,385)	Sand and gravel in collapsed kame terraces above Greenwater Brook (to 1,375)
		3. Uncertain; probably ice against north slope of Beartown Mountain (about 1,275)	Sand and gravel in delta east of Lee (to 1,275)
		4. Uncertain; probably ice against north slope of Beartown Mountain (about 1,180 to 1,060)	Sand and gravel in kame terraces east to northeast of Lee (to 1,180; prominent terrace at about 1,065)
		5. Ice Glen (1,035)	Sand and gravel in Lenox Dale Delta (to 1,065; lake level near 1,035)
		6. Brookside Col (965)	Gravel and sand in Glendale Delta (to 985) Lake sand and silt in Stockbridge basin (to 835) Lake sand in southeastern part of Pittsfield (to 995) Sand and gravel in kame terrace, Pittsfield (to 1,045)
T	Lake Tyringham (new name)	1. Saddle to Clam River, West Otis (1,365)	Sand and gravel in kames, eskers, etc. (to 1,380)
		2. Uncertain; probably ice against north slope of Beartown Mountain (1,050?)	Lake sand with a few pebbles (to 1,025)

Table I (continued)

Map Symbol	Lake name (reference)	Controlling spillways (altitude, in feet)	Deposits (altitude, in feet)
GF	Lake Great Falls (new name, glacial Lake Falls Village of Holmes and Newman, 1971; glacial Lake Sheffield of Newton, and others, 1975)	Great Falls, Connecticut (635)	Lake clays (to 675) Sand, some gravel, in kame terraces (to 740) Sand and gravel in valley trains (outwash) (to 800 on Williams River; to 995 on Alford-Seekonk Brook; to over 950 on Green River; to 1,060 on Konkapot River)
HN	Lake Hinsdale (new name)	Saddle to West Branch, Westfield River (near 1,495)	Gravel and sand in Bullards Delta (to 1,560; lake level near 1,500) Sand, some gravel, in kames and kame terraces (to 1,505) Lake sand, some silt (below 1,450) Sand, some gravel in Hinsdale Delta (to 1,535)
KN	Lake Konkapot (new name)	Konkapot Col (965)	Sand, some gravel, in kame terraces and eskers (to 950) Sand, some gravel, in Open Heart Delta (east of Brookside Col) (to 1,045; lake level near 985)
D	Lake Dalton (new name)	1. Ice against Day Mountain (to 1,375) 2. Rock spur west of Day Mountain (1,155)	Sand and gravel in kames and kame terraces (to 1,375) Gravel and sand in Dalton Delta (to 1,185; lake level 1,160: Holmes, 1968)
B	Lake Bascom (Dale, 1906)	Saddle to Housatonic River in NE corner of Pittsfield (1,005)	Sand, some gravel, in deltas and kame terraces (to 1,070) Sand, silt, and clay in lake beds Sand and gravel in deltas in Vermont Shilts (<u>in</u> Stewart and MacClintock, 1969) (lake level reportedly about 1,100)
SL	Lake State Line (new name)	Uncertain; possibly saddle 1.5 km south of Maple Hill (925), possibly drift or ice east of Maple Hill	Sand and gravel in kame terraces (to 965) Sand and gravel in delta deposited from Furnace Brook (to 945; lake level near 935) Sand, some pebbles, in delta from Baldwin Brook (to 940)

TABLE 1 (continued)

Map Symbol	Lake name (reference)	Controlling spillways (altitude, in feet)	Deposits (altitude, in feet)
K	Lake Kinderhook (new name)	1. Outlet of Queechy Lake (1021) or outlet of Beebe Pond (1,025)	Sand and gravel in deltas N, E and W of Stephentown (to 1,095, lake level near 1,075, 2 km W of Stephentown; to 1,110 4 to 6 km N of Stephentown)
		2. Saddle crossed by Canaan Road 3 km S of New Lebabon Center (1,005)	Sand and gravel in deltas, including several at mouths of valleys from NW (e.g. to 1,055 3.5 km WSW of Stephentown)
		3. Saddle 1 km SW of New Britain (905)	Sand and gravel in kame terraces, in deltas (to 965; good examples 2.5 km SW of Stephentown, 940, and 1 km N of Stephentown, 965), and in a stream-cut surface 2 km SE of Stephentown (to 975)
		4. Probably the saddle 1 km SE of Ashley Hill (705)	Sand and gravel in deltas, mostly or entirely of material reworked from above deposits (to 715 2.5 km ENE of West Lebanon)
C	Lake Canaan (new name)	Saddle 2 km S of Canaan Center (985)	Small sand and gravel deposits (lake mentioned by Taylor, 1903, p. 329)
BR	Lake Berlin (new name)	Saddle at Cherryplain, N.Y. (1,115)	Sand and gravel in deltas (to 1,120)
S	Lake Shaftsbury (Shilts and Behling, 1967; Shilts, *in* Stewart and MacClintock, 1969)	Saddle at Potter Hill, in SW part of town of Hoosick, N.Y. (895)	Deltaic deposits in Vermont (to 1,100)

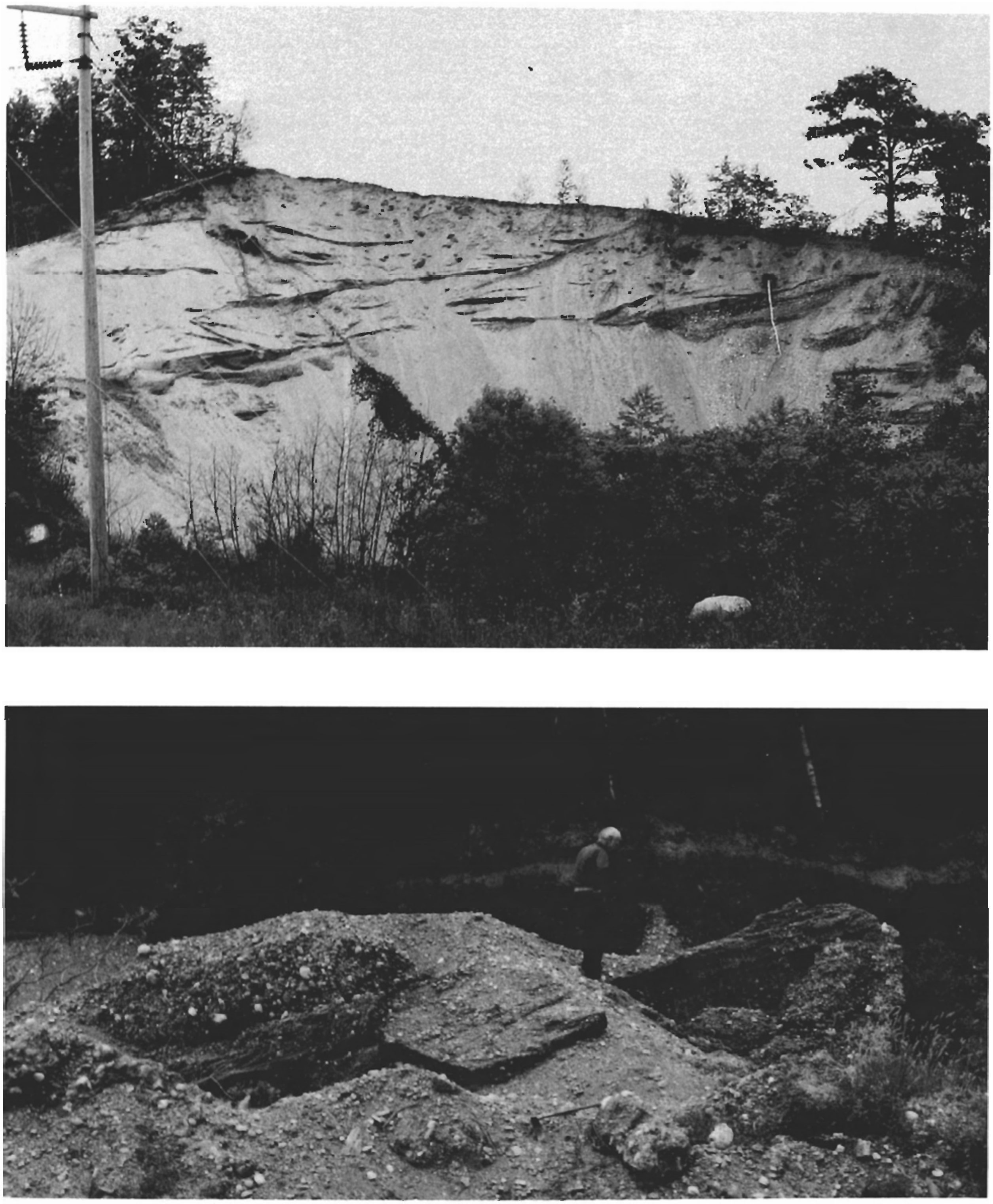

Figure 7 **Delta foreset beds deposited in Lake Housatonic. a) Foreset beds fill a channel and dip away from pit face; view is southeast, east of Lee, Massachusetts; b) Gravel foresets of the Glendale delta cemented by calcium carbonate deposited by groundwater; the beds stand in undisturbed position, the uncemented parts of the gravel have been removed.**

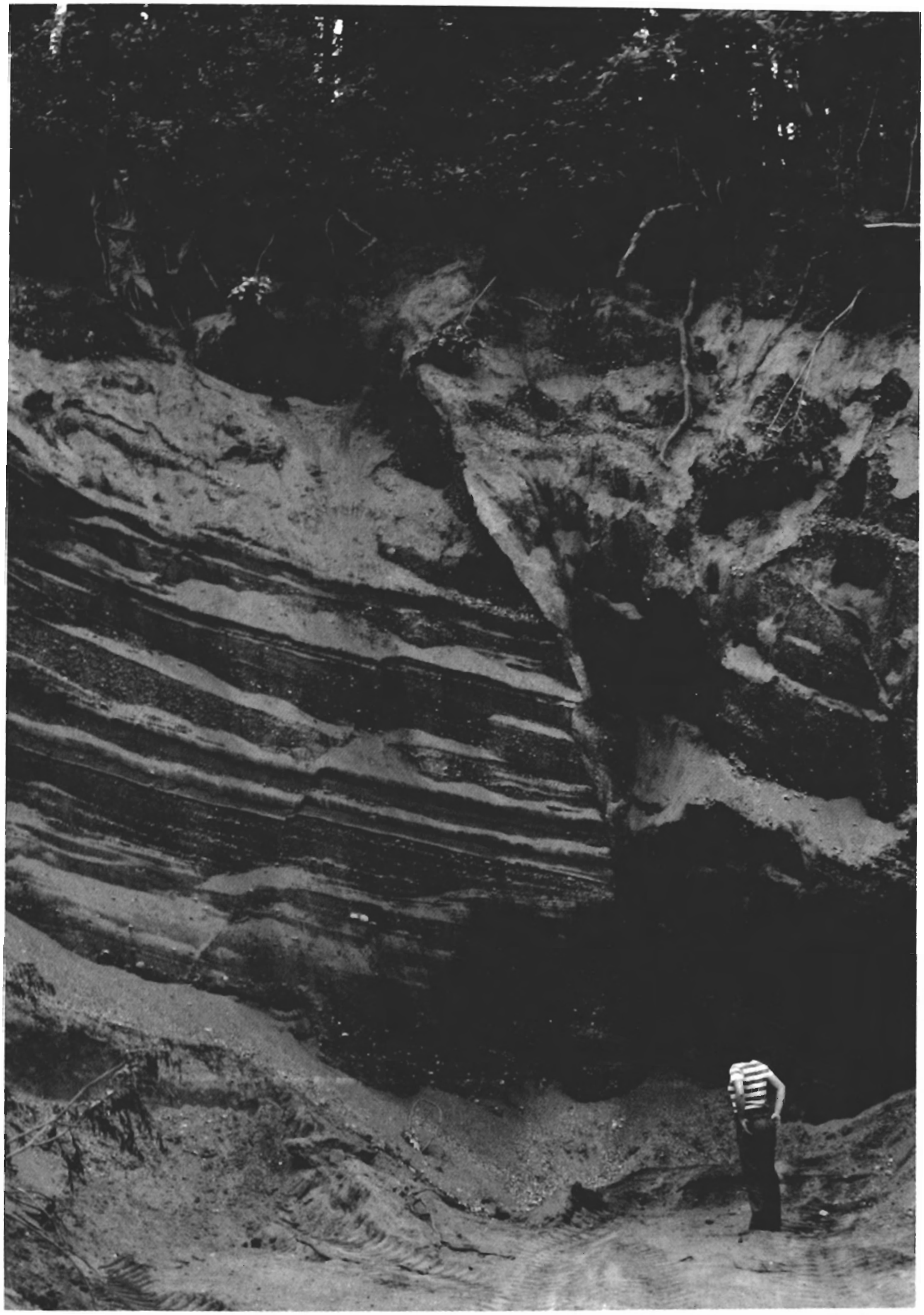

Figure 8 Coarse delta foreset beds deformed by normal faults related to melting of buried ice and subsequent collapse of overlying or adjacent sediments. Glendale delta of Lake Housatonic.

Historical Summary of Deglaciation

Figures 10b and 11 synthesize our deglaciation stratigraphy and summarize our conclusions about the mode of deglaciation of the region. The following is a narrative historical summary, relating local and regional stratigraphic relationships to the deglaciation processes.

As the surface altitude of the Hudson-Champlain lobe in western Massachusetts progressively lowered during the early phases of deglaciation, nunataks emerged above the ice surface in the southern part of the area, probably east of South Sandisfield (phase 1 of Warren and Harwood, 1978). The relative timing of emergence of other, higher nunataks, such as Mount Greylock and others in northern New England (Borns and Calkin, 1977; Goldthwait, 1968; Goldthwait and Mickelson, 1982; Gerath, 1978), is not known. As deglaciation continued to phase 6 (Fig. 11), initial deposits of Lakes Housatonic and Tyringham were deposited in upland basins. Both of these lakes persisted through phase 9, as the ice margin retreated from drainage divides at the heads of Greenwater Brook, at North Otis, at Hop Brook, and at West Otis. The oldest deposits of glacial Lake Great Falls which came into existence when the ice retreated from the rock ridge at Great Falls, Connecticut, possibly predate phase 6.

Glacial Lake Hinsdale came into existence in phase 7 when the ice surface dropped below the level of the divide between the Housatonic River and the West Branch Westfield River, south of Hinsdale. In phase 8 the ice margin 8 km south of Dalton stood against a saddle at 560 m (1,835 feet) altitude and sent a meltwater stream down the valley of Cady Brook. This meltwater deposited a delta in Lake Hinsdale which is herein called the Bullards Delta. The topset beds of the Bullards Delta are as much as 6 m thick and include much cobble gravel; the largest stones present are rounded boulders about 50 cm (20 inches) in diameter. No outsize boulders are present. The position, morphology, and textural composition of the delta indicate that it was not constructed at the ice margin, but rather by meltwater from Cady Brook. Thus the ice margin at the head of Cady Brook stood above 560 m (1835 feet) at a time when the ice in the Lake Hinsdale basin, below 457 m (1,500 feet), had retreated at least 4 km to a position north of the Bullards delta.

As the surface of the ice in the Stockbridge Basin lowered during phases 8 and 9, the meltwater that had been impounded in Lakes Housatonic and Tyringham was released through a lower outlet to the south around the north and west sides of Beartown Mountain. Ice-marginal deltaic deposits of Lake Housatonic (Fig. 7a) at 366-389 m (1,200 to 1,275 feet) east of Lee record one stage of the lake level, probably in phase 9. A transverse sand ridge across the width of the Tyringham valley at Tyringham village probably marks an ice-margin position of the same age, though the deposit was not built up to the surface of Lake Tyringham. With further ice retreat at about the end of phase 9, the water level of Lake Tyringham dropped to the level of Lake Housatonic and an arm of that lake extended into the Tyringham valley.

By phase 10, the level of glacial Lake Housatonic stood at about 323 m (1,060 feet) east of Lee. The ice margin impounding this lake extended south on the west side of Beartown Mountain as far as the Konkapot Col, where it stood above 305 m (1,000 feet) and sent meltwater loaded with debris down the Konkapot River valley. The phase-10 age of this ice-front position is inferred from the discontinuous meltwater terraces in the Konkapot River valley that appear to be graded to the outwash fan at Ashley Falls, referred to phase 10 by Warren and Harwood (1978). In the upper Housatonic valley, the ice impounding glacial Lake Hinsdale had retreated by phase 10 to the site of Hinsdale, where it deposited a sandy ice-marginal delta so massive that it dammed the preglacial valley.

During retreatal phase 11, ice in the Konkapot Creek valley north of the Konkapot Col retreated some 6 km without depositing much stratified drift. Then, a considerable volume of sand and gravel was laid down in the ice-marginal Open Heart Delta, named for the Open Heart Camp which owns it. The Open Heart Delta was deposited in Lake Konkapot, which was held in by ice that stood to the north. The lake level stood at about 300 m (985 feet) altitude, controlled by the Konkapot Col. The ice in the Housatonic valley had dropped to a level too low to cross Three Mile Hill on the west side of Lake Konkapot, but ice west of the Brookside Col, between Monument Mountain and Three Mile Hill, stood against that col at a level high enough to prevent Lake Konkapot from spilling there.

Also during phase 11, the ice in the Stockbridge basin retreated from the highlands north of Lee, dividing into two sublobes. Ice from the north, flowing down the east side of Lenox Mountain, retreated to the position of the Lenox Dale delta. The sublobe from the west, flowing around the south end of the West Stockbridge Mountain, retreated to a position against the hill west of Ice Glen. The level of Lake Housatonic at the Lenox Dale delta was about 315 m (1,035 feet), higher than the level projected northward from the Konkapot Col, that Taylor (1903, p. 351) identified as the controlling spillway of Lake Housatonic when the Lenox Dale delta was deposited. Rather, the lake level was controlled by a spillway at the meltwater channel Ice Glen. In the upper Housa-

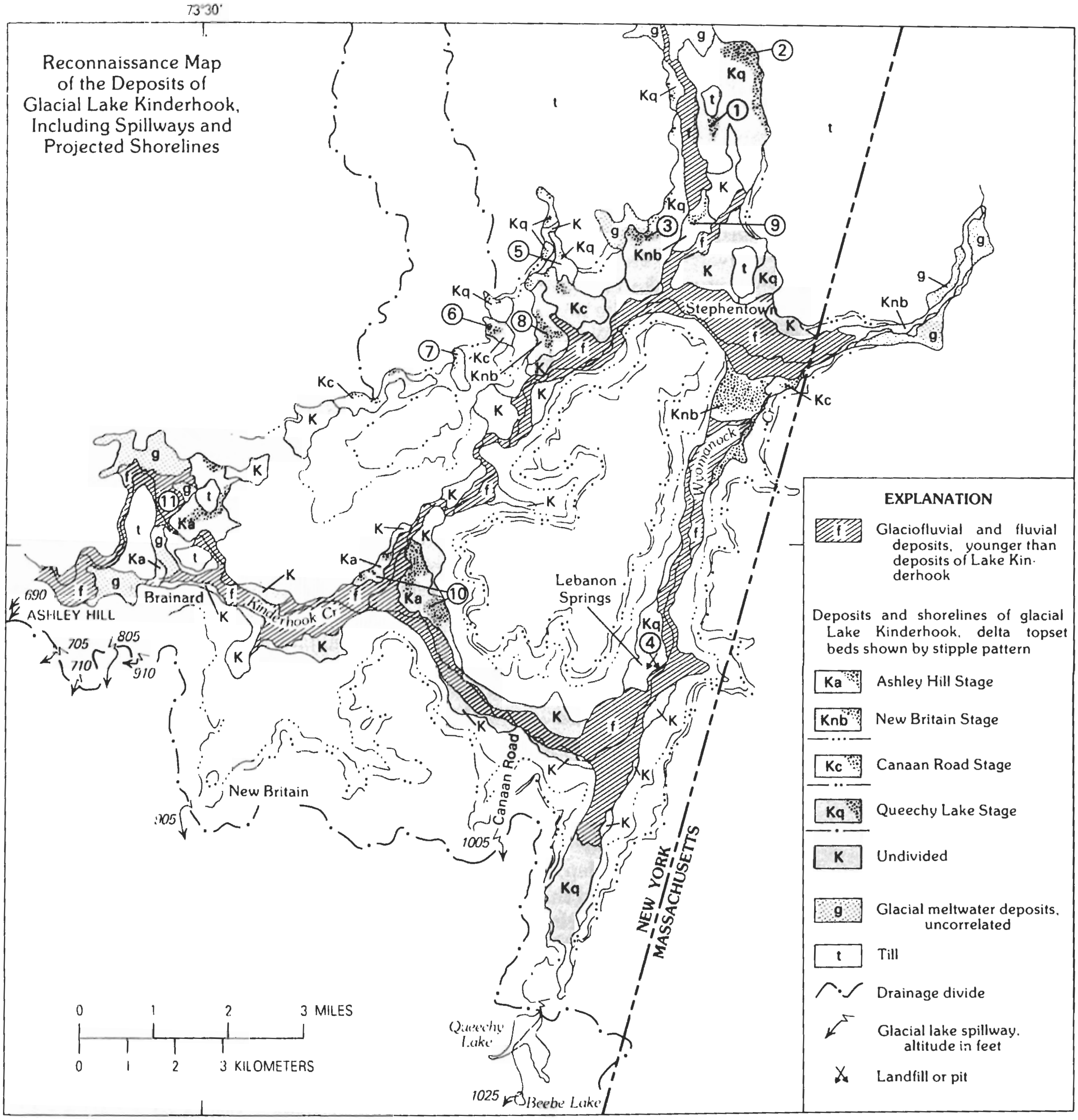

Figure 9 Reconnaissance map of the deposits of glacial Lake Kinderhook, showing spillways and projected shorelines. Data from unpublished studies of B.D. Stone and D.H. Cadwell. Numbers refer to localities cited by number in the text.

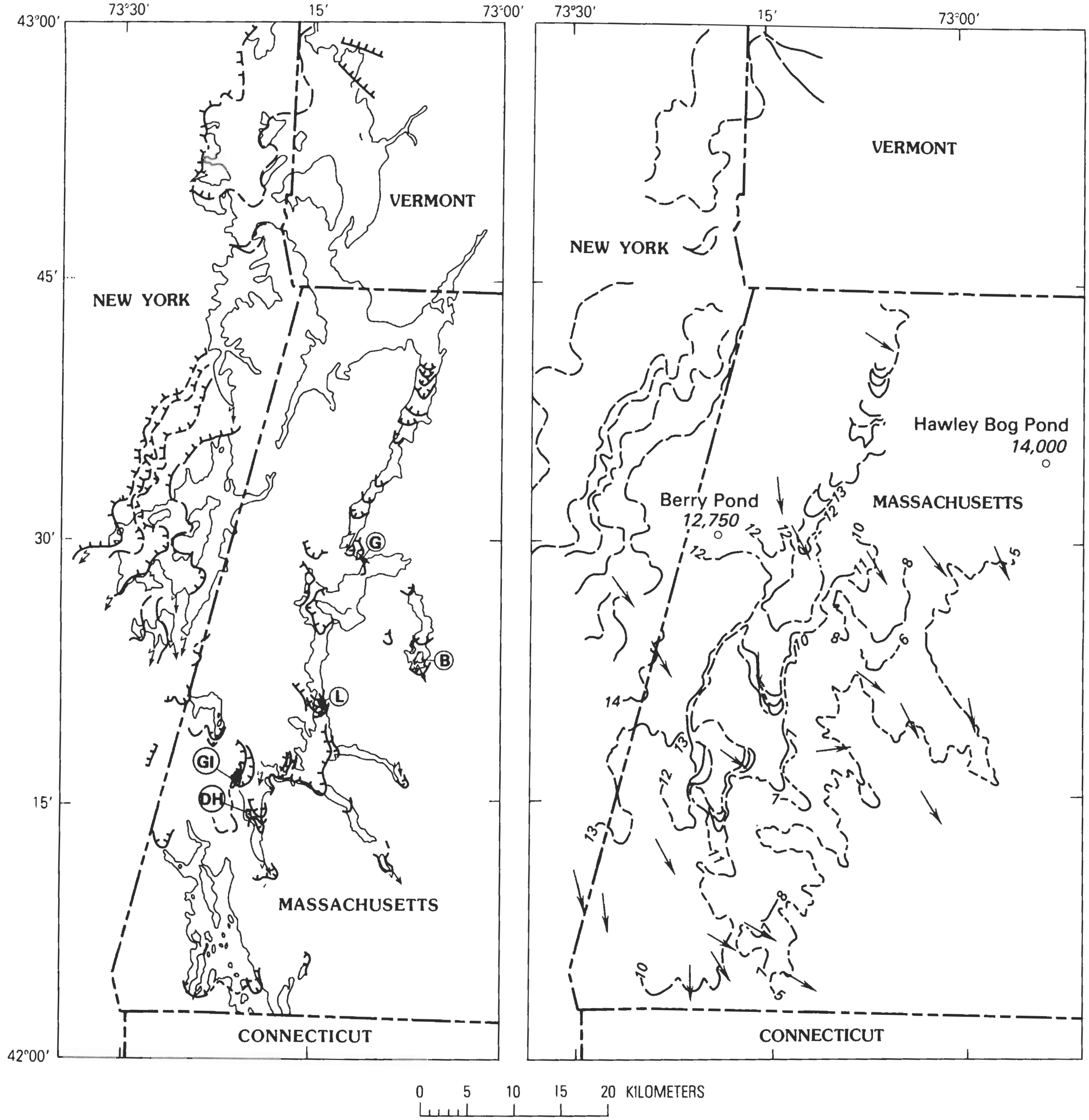

Figure 10 Maps showing ice-margin retreatal positions in western Massachusetts and adjacent parts of New York and Vermont. a) Ice-margin positions and extent of correlated glacial lake deposits or glacial lakewater planes, and lake spillways. b) Preliminary regional ice-margin positions, selected striations inferred to be late-glacial in age, and radiocarbon ages of postglacial deposits. B – Bullards Delta. G – The Gulf. Gl – Glendale Delta. OH – Open Heart Delta. I – Ice Glen.

tonic valley, the ice retreated in phase 11 from the Hinsdale delta, causing a new, lower lake, glacial Lake Dalton, to form between it and the Hinsdale delta. The runoff drainage from the Lake Hinsdale basin, cutting down in the sandy barrier of that delta, was superposed across a rock spur on the side of the preglacial valley, creating a rapids.

In phase 12, the ice margin withdrew from the positions against the Open Heart Delta and the hill northwest of Ice Glen. The water level of Lake Housatonic therefore dropped to the level of Lake Konkapot. At about the same time the level of the ice in the Housatonic valley west of the Brookside Col also was lowered, so that the enlarged Lake Housatonic discharged through the Brookside Col at an altitude of 294 m (965 feet). Meltwater from the ice in the Stockbridge Basin built the Glendale Delta (Taylor, 1903, p. 351; see Figs. 7b and 8) into Lake Housatonic in three successive ice-marginal deposits during phase 12.

Taylor (1903, p. 251), finding the Lenox Dale and Glendale Deltas in the same lake basin at Stockbridge, correlated them and made this correlation a key point in his mapping of recessional ice fronts. However, new topographic control, exposures in the deltas, and new estimates of the amount and pattern of postglacial uplift preclude this correlation. The Lenox Dale delta is correlated with a Housatonic Lake stage controlled by the Ice Glen outlet, and the Glendale delta is correlated with the Brookside Col stage.

North of Lenox Dale, lake sands in the southeast corner of the city of Pittsfield were deposited in Lake Housatonic while it was at the Brookside-Col level during phase 12. Downtown Pittsfield is built on a fluvial terrace laid down during phase 12 between two valley-ice sublobes that stood above 320 m (1,050 feet) on the east and west. The sublobe on the west flowed from the west across a saddle in the Taconic Range, occupying the basin of Onota Lake; the eastern sublobe flowed from the north, up the valley of the Hoosic River. A third sublobe, also from the north but coming from the west side of Mount Greylock, occupied the basin of Lake Pontoosuc.

At the same time, the east flank of the Hoosic valley ice sublobe stood above 366 m (1,200 feet) just north of the northeast corner of Pittsfield. There meltwater flowing south along the edge of the ice poured through a saddle known as The Gulf (Fig. 10a), and built into glacial Lake Dalton a gravelly delta on which downtown Dalton now stands.

During phase 13, glacial Lakes Housatonic and Dalton were drained. Glaciofluvial morphosequences graded to the subsequent fluvial terrace level in the Housatonic valley head at four places: 1) southeast of Maple Hill on the Williams River, 2) west of Maple Hill at the head of Alford-Seekonk Brook, 3) at the head of Scribner Brook, a tributary of Alford Brook, and 4) above North Egremont on the Green River. The ice front retreated to a position near the head of the Hoosic River drainage and kame terraces as high as 361 m (1,185 feet) were deposited east of what is now Cheshire Reservoir.

Glacial Lake Bascom did not come into existence until phase 14, when the ice retreated from the divide at altitude 306 m (1,005 feet) at the head of the Hoosic River in Pittsfield. Lake Bascom lengthened and deepened as the ice margin retreated down the Hoosic River valley into Vermont. Shilts' (Shilts and Behling, 1967) Lake Bascom shoreline at an altitude of 335 m (1100 feet) was based on altitudes of ice-marginal deltas of the lake near Shaftsbury Center, Vermont. Glacial Lake State Line also is correlated with phase 14. This lake occupied a basin that is no longer closed as it has free drainage to the south down the Williams River to the Housatonic. The nature of the dam that impounded it is uncertain; it may have been either drift or a block of residual, dead ice in the Williams River valley southeast or east of Maple Hill.

Deposits of the four stages of Lake Kinderhook and their associated ice-margin positions illustrate the north-northeast trend of the edge of the Hudson-Champlain lobe that is parallel to the strike of bedrock-controlled topography of the Taconic Range and of the Rensselaer Plateau. Ice-marginal deltas of glacial Lake Berlin in the Little Hoosic River valley suggest that a valley-ice sublobe, fed by active ice in the lower Hoosic River valley, persisted during ice-margin retreat in the upper reaches of the valley. Preliminary ice-margin positions at the north end of the Little Hoosic valley, northwest of the northwest corner of Massachusetts, are shown in Figure 10b. These suggest correlation of deposits of Lake Kinderhook and initial deposits of Lake Berlin with ice-marginal deltas of Lake Bascom in the Williamstown, Massachusetts, basin. Deposits of Glacial Lake Canaan are about the same age as deposits of the Canaan Road stage of Lake Kinderhook.

When the ice margin retreated to a position northwest of the lower Little Hoosic valley, the level of Lake Berlin dropped to the level of Lake Bascom. Lake Bascom continued to expand northward to its maximum extent near Shaftsbury Center, Vermont. Deglaciation of the col at Potter Hill, New York, caused the impounded water level in the Hoosic-Wallomsac drainage to lower some 60 m to the level of Glacial Lake Shaftsbury (Shilts and Behling, 1967). The ice-margin position required by the position of ice-marginal deltas and the Potter Hill Spillway of Lake Shaftsbury shows the continuity of the

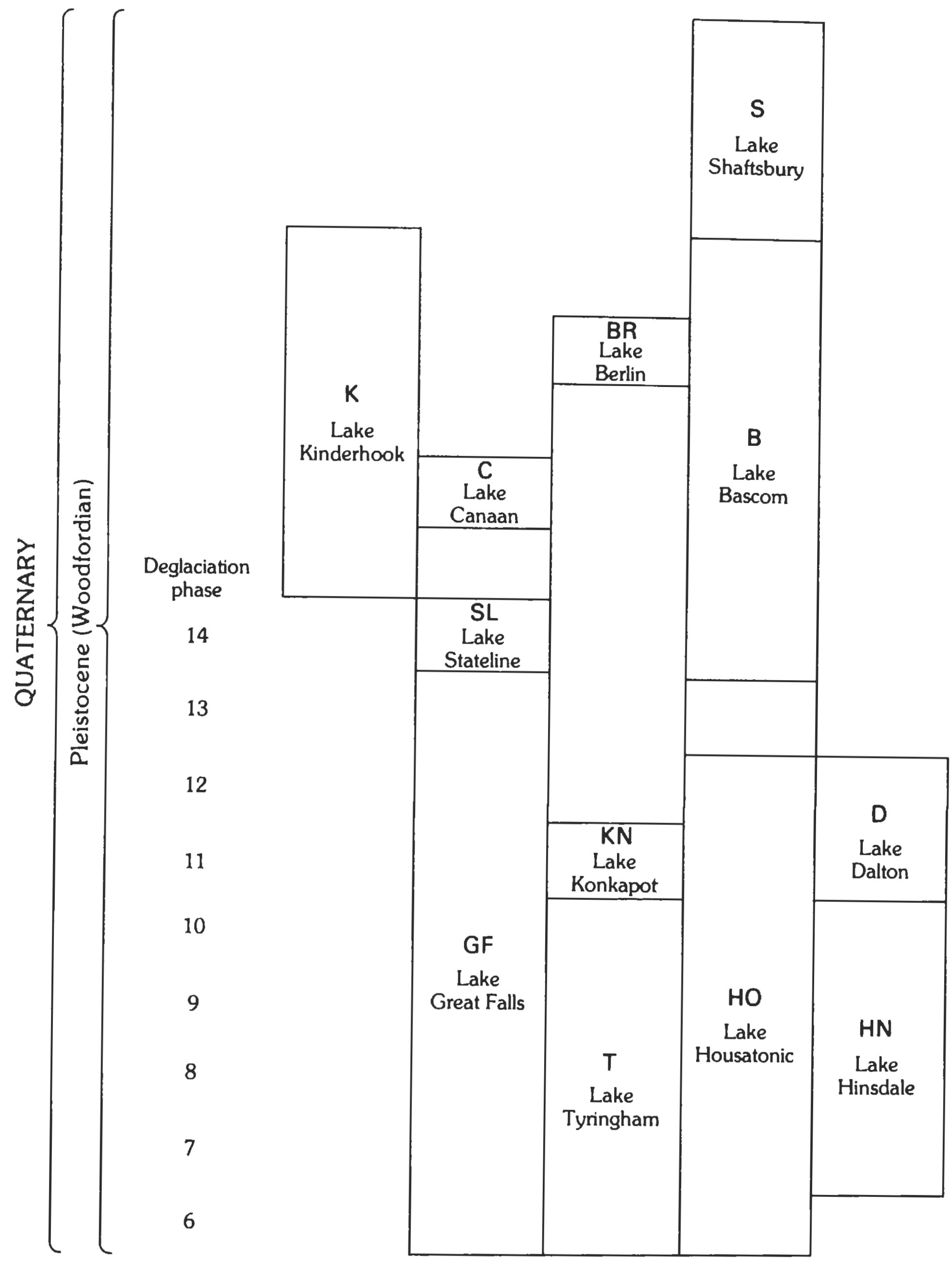

Figure 11 Preliminary correlation of principal meltwater deposits of lettered map units of Fig. 5, also in Table 1, of western Massachusetts, central eastern New York, and southwestern Vermont, shown in Fig. 5. Deposits include deltaic, lacustrine, and fluvial sediments deposited in or graded to the glacial lakes. Deglaciation phase numbers are those defined in, or extended from, those of Warren and Harwood (1978). Glacial Lake Great Falls probably came into existence in or about phase 6, but the deposits of the lake in Massachusetts are in phases 10-13.

north-northeast aspect of the eastern flank of the Hudson-Champlain lobe, parallel to other ice-margin positions deduced by Cadwell (1981, and unpublished data) and suggested by ice-marginal deposits in the Troy, New York area (LaFleur, 1965).

Radiometric Ages of Deglaciation

Radiocarbon ages of peat and organic carbon in limnic sediments that overlie glacial sediments in the area include: 1) 14,000 ± 130 yrs BP (WIS-1122) in limnic sediments containing spruce pollen in Hawley Bog Pond (Fig. 10b) in the uplands near Hawley, Massachusetts, (W.A. Peterson, University of Massachusetts; Bender and others, 1981); 2) 12,680 ± 480 yrs BP (OWU-481) at the base of the spruce-pollen zone from Berry Pond (Fig. 10b), west of Pittsfield (Whitehead and others, 1973); 3) 12,680 ± 200 yrs BP (Y-2247a) in sediments of Queechy Lake (Stuiver, 1975); and 4) 12,750 ± 230 yrs BP (RL-245) on postglacial material in the Ellsworth, Connecticut, area (Kelley, 1975). These minimum-age estimates of deglaciation seem compatible with generally older dates of post-glacial sediments and extrapolated dates based on sedimentation rates north of the late Wisconsinan-age terminal moraine in New Jersey (Cotter and others, 1982; this volume) and New York (Connally and Sirkin, this volume). Within such a regional chronologic framework, the deglaciation stratigraphy of western Massachusetts and adjacent New York and Vermont may be as old as 15-16,000 yrs BP.

CONCLUSIONS

Based on new field work and compilation of detailed map data of western Massachusetts, we conclude the following:

1) The general ice-movement direction in western Massachusetts and adjacent parts of New York and Vermont during the early phases of late Wisconsinan deglaciation was southeasterly. During later phases of deglaciation, major lowland areas deflected local valley-ice sublobes southerly and easterly.

2) Because laterally continuous moraines, multiple drift sheets of late Wisconsinan age, or readvance stratigraphies are not known in western Massachusetts, and because till moraines and ice-marginal (kame) moraines are not numerous in adjacent New York and Vermont, regional, climatically driven readvances of the eastern flank of the Hudson-Champlain lobe probably did not occur during deglaciation of the region.

3) Stratified deposits in this region are chiefly lacustrine, including deltaic, associated fluvial, and lake bottom sediments in all major lowlands and in upland tributary basins that drain northwest. Ice-marginal deposits are chiefly deltaic (ice-contact lacustrine morphosequences), which show successive ice-margin retreatal positions within each glacial lake basin and which indicate deglaciation of all basins by a process of stagnation zone retreat.

4) Positions and altitudes of ice-marginal deltas, extensive lake bottom deposits, and correlated glacial lake spillways of 12 principal glacial lakes indicate intrabasin and interbasin correlations of ice-margin retreatal positions. These correlated regional ice-margin positions illustrate the consistent northnortheast trend and the shape of the ice margin related to the eastern flank of the Hudson-Champlain lobe during deglaciation of the region. Active valley-ice sublobes, controlled by the positions of cols through the uplands and the geometry of lowlands, continued to supply meltwater and sediment to the stagnant ice along the ice sheet's margin during deglaciation.

5) The radiometric ages of postglacial organic sediments of the region indicate that deglaciation of western Massachusetts occurred before 14,000 years ago. They are compatible with suggested ages of deglaciation of the middle reaches of the Hudson lowland as early as 15-16,000 years ago.

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