55th Annual

FRIENDS OF THE PLEISTOCENE

1992 Reunion

THE LATE GLACIATION OF EASTERN NEW YORK STATE: Glacial Tongues and Bergy Bits

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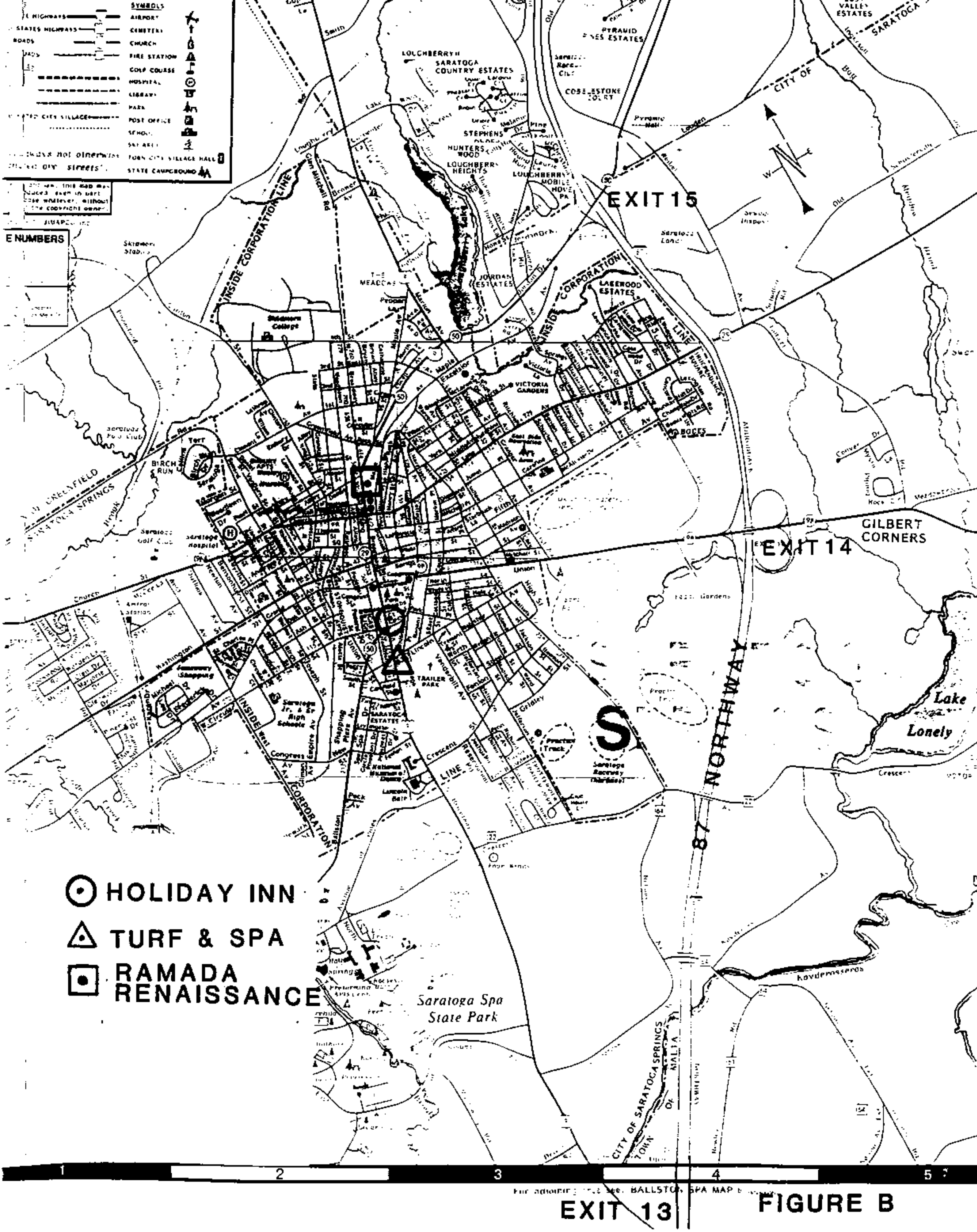
TABLE OF CONTENTS

LIST OF FIGURES



HOLIDAY INN
TURF & SPA
RAMADA RENAISSANCE
EXIT 15
EXIT 14
EXIT 13
GILBERT CORNERS
87 NORTHWAY
INSIDE CORPORATION LINE
CITY OF SARATOGA SPRINGS
TOWN OF MALTA
Saratoga Spa State Park
Lake Lonely
SARATOGA COUNTRY ESTATES
LOUGHBERRY HEIGHTS
JORDAN ESTATES
LAKEWOOD ESTATES
VICTORIA GARDENS
BIRCH RUN
PYRAMID PINES ESTATES
TRAILER PARK
E NUMBERS

FIGURE B

Trip log and stop descriptions

Trip begins at fire house in Scotia; intersection of Routes 5 and 147. See the topographic base map for trip stops.

Proceed NNW on Sacandaga Road (Route 147) 1.4 miles to sharp right turn before bridge. Road to this point crosses the Scotia Terrace at 280-290 feet ASL. About 70 feet of pebble and cobble gravel underlies the surface. This is one of the larger terrace remnants; most of the Village of Scotia and the Naval Depot are built on the terrace, which is bordered on the south by the Scotia Channel/Mohawk River floodplain and on the north by the Harding Channel.

Descend hill beneath the bridge and join Vley Road. The railroad and Vley Road follow the Harding Channel. This channel, with a base at 255 feet ASL, shared Mohawk discharge with the early Scotia Channel, after the formation of the Scotia Terrace, and eroded at least 35 feet of Scotia gravel. Proceed west on Vley Road 1.1 miles to Viele Road on the south. On the south side of the road are the production wells of the village of Scotia. Enter Viele and park at the turn-around at the end of the road.

SA-5 Walk to the edge of a pit excavated in the coarse gravel of the Scotia Terrace, and note the cobble lithologies, most of which are foreign to the eastern Mohawk basin. The water table in the terrace gravels fluctuates seasonally in response to draining of the Erie Canal during the non-navigation season. Canal level is lowered about 12 feet in winter, and some loss of head is experienced even at the edge of the aquifer. One can appreciate the water-yielding capability of the deposit, which has a transmissivity of 3 to 5 million gallons per day per foot of width.

Return to Vley Road and continue west for 0.5 mile. Intersection with Route 5.

SA-6 Glenville landfill to north through trees. Note the methane vent wells. We are standing at the sharply defined northern edge of the coarse gravel, at 260 feet ASL. The surface is the subtle western end of the Harding Channel. Figure 22 shows the relationships of landfill environs to the aquifer gravel. A stratigraphic review will explain the Mohawk glacial advance and the nature of the glaciolacustrine pre-Mohawk (Hell Hollow) and post-Mohawk (Lake Albany) deposits that form the substrates for the coarse gravels.

Turn east on Route 5 and proceed 0.4 mile to railroad overpass and park along Route 5.

SA-7 Walk along the edge of the Erie Canal for about 1000 feet to an exposure of east-dipping, graded beds (diurnal cycles ?) of Scotia Terrace cobble gravel. This clinoform unit is generally overlain by more level and continuously bedded gravel and suggests a prograding delta type of deposition of especially coarse grain

size.

Return to the vehicles and proceed 1.4 miles east on Route 5 (Mohawk Turnpike) to access road to the right that leads to the north end of Lock 8.

SA-8 The irregular surface leading to the lock is the bottom of the Scotia Channel. The channel is 2000 feet wide and 35 feet deep at this point. Discussion will focus on dynamics of channel flow, destination of Scotia Terrace gravel eroded by the channel water, and development of the big elbow of the Mohawk to the south.

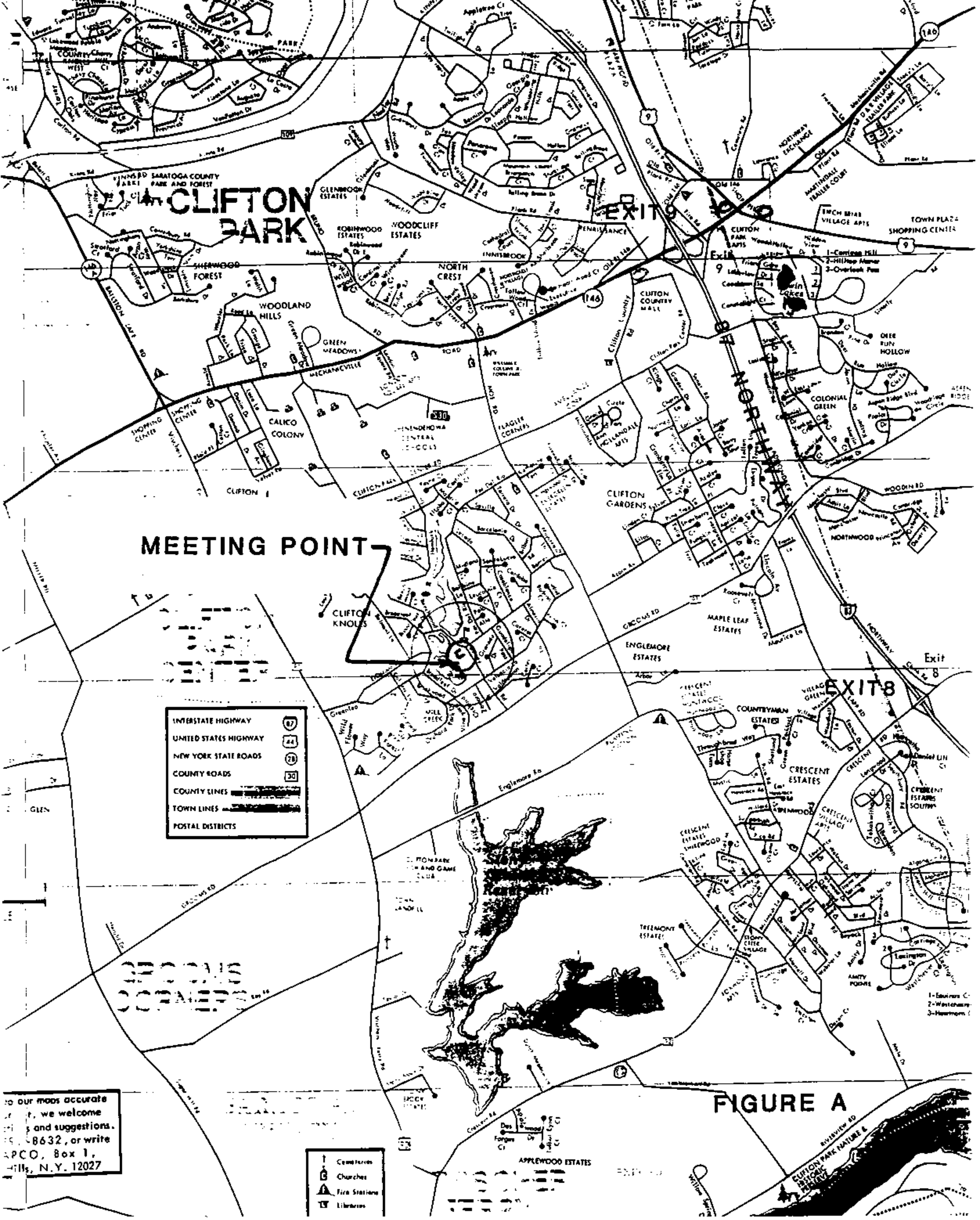
Return to Route 5 and proceed west for 4.2 miles to intersection with Route 103 at Lock 9. Turn SW across the Erie Canal for 0.5 mile on 103 to intersection with Route 5S. Turn east on Route 5S and proceed SW through Rotterdam Junction to Mabie Road. Turn SW on Mabie and park along the streets.

SA-9 Walk to the edge of a pit in the Rotterdam Junction Terrace and note the wells that produce for Schenectady Chemicals. The surface of the Rotterdam Junction Terrace at 250 feet ASL is the bottom of the Scotia Channel. About 40 feet of gravel was removed from what used to be Scotia Terrace here to form the present surface. The gravel is only 15 feet thick beneath the Schenectady Chemicals plant where a high point in the eroded glaciolacustrine subcrop approaches the land surface. To the northwest the Rotterdam Junction wells produce from about 35 feet of gravel that lies below the terrace. The aquifer has been mined by drag-lining, with no ill effects, and the pit is planned to extend to the northwest.

Walk south on the dirt road and cross the Rotterdam Channel to see the original Erie Canal, still holding water perched above the water table.

55TH ANNUAL REUNION
CLIFTON PARK, NEW YORK
MAY 15 TO 17, 1992
TENTATIVE SCHEDULE

Friday (May 15)	
6:00 to 11:00 pm	Registration & Welcoming Party Environmental Hydrogeology Corporation (EHC) Barney Road Clifton Park, New York
Saturday (May 16)	
8:00 am	Field Trip Leaves from Meeting Point (Figure A) EHC Parking Lot (South) Barney Road Clifton Park, New York
8:00 am to 12:00 pm	Field Trip Stops SA-1 to SA-4 Trip Leader: Bob Dineen
12:00 pm to 12:30 pm	Scenic Barrow Pit Picnic Box Lunch Provided
12:30 pm to 3:30 pm	Field Trip Stops SA-1 to SA-4 (Continued) Trip Leader: Bob Dineen
3:30 pm to 5:00 pm	Field Trip Stops SA-5 to SA-9 Trip Leader: Bob LaFleur
5:00 pm	Return to Meeting Place
	Banquet, Man O' War Room (Figure B) Ramada Renaissance Hotel Broadway (Street) Saratoga Springs, New York
7:30 pm to 8:00 pm	Cash Bar
8:00 pm to 11:00 pm	Dinner
Sunday (May 17)	
8:00 am	Field Trip Leaves from Meeting Point EHC Parking Lot (South) Barney Road Clifton Park, New York
8:00 am to 12:00 pm	Field Trip Stops SU-1 to SU-5 Trip Leader: Eric Hanson
12:00 pm to 1:00 pm	Lunch (on your own) Stop at Fast Food Restaurant *Shuttle back to meeting place for those who need to start early on the long drive home.
1:00 pm to 5:00 pm	Field Trip Stops SU-6 to SU-11 Trip Leader: Dave DeSimone
5:00 pm	Return to Meeting Place



CLIFTON PARK
SARATOGA COUNTY PARK AND FOREST
EXIT 9
EXIT 8
NORTHWAY
MEETING POINT
CLIFTON KNOLLS
CLIFTON GARDENS
ENGLEMORE ESTATES
MAPLE LEAF ESTATES
CRESCENT ESTATES
COUNTRYMAN ESTATES
CRESCENT ESTATES SOUTH
WOODLAND HILLS
SHERWOOD FOREST
NORTH CREST
ROBINWOOD ESTATES
WOODCLIFF ESTATES
GLENBROOK ESTATES
INNISBROOK
CALICO COLONY
GREEN MEADOWS
FLAGLER CORNERS
CLIFTON COUNTRY MALL
COLONIAL GREEN
DEER RUN HOLLOW
TOWN PLAZA SHOPPING CENTER
BIRCH BRIAR VILLAGE APTS
TREMONT ESTATES
AMITY POINTE
APPLEWOOD ESTATES
SHENENDEHOWA CENTRAL SCHOOLS
CLIFTON PARK NATURE & HISTORIC PRESERVE
MECHANICVILLE ROAD
GROOMS RD
Englemore Rd
INTERSTATE HIGHWAY
UNITED STATES HIGHWAY
NEW YORK STATE ROADS
COUNTY ROADS
COUNTY LINES
TOWN LINES
POSTAL DISTRICTS
Cemeteries
Churches
Fire Stations
Libraries
1-Carriage Hill
2-Hilltop Manor
3-Overlook Pass

FIGURE A

(Continued)

LIST OF TABLES

THE LATE GLACIATION OF EASTERN NEW YORK STATE
or Glacial Tongues and Bergy Bits

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GLACIAL LAKE ALBANY

INTRODUCTION

Glacial Lake Albany was one a series of proglacial lakes that developed in the Hudson-Champlain Lowlands as the Hudson-Champlain glacial lobe retreated. The receding ice margin formed the northern shore of the lakes. The lakes included Glacial Lake Hudson in the lower Hudson Valley and Glacial Lake Albany in the upper Hudson Valley. Glacial Lake Hudson extended from the Terminal Moraine, between Long Island and Staten Island, to Peekskill, NY. At its maximum extent, Lake Albany extended from Newburgh, NY to Glens Falls, NY. The water level of Lake Albany fell abruptly to the Lake Quaker Springs stage as the glacier retreated into the Champlain Valley. Lake Quaker Springs was an extension of Glacial Lake Vermont, a lake that occupied the Champlain Valley. Lower stages of Lake Vermont include Lakes Coveville and Fort Ann. These water levels can be traced into the Hudson Valley, where they become more "fluvial" and less "lacustrine." The series of proglacial lakes in the Hudson-Champlain Lowlands ended when the glacier retreated north into the St. Lawrence Lowlands.

Lakes Albany and Quaker Springs are recorded by sand and silt terraces, beaches, and deltas throughout the Hudson Lowlands (Woodworth, 1905). Generally, the elevations of the terraces and deltas have been tilted up to the north by glacial-isostatic rebound (Woodworth, 1905; LaFleur, 1965; Connally and Sirkin, 1986; DeSimone and LaFleur, 1986). Lake Albany stage deltas disappear north of Saratoga Lake and Lake Quaker Springs features can be traced north into the Champlain Valley (DeSimone and LaFleur, 1986). The terraces associated with Lakes Coveville and Fort Ann are not tilted to the same degree as the older lake features (Woodworth, 1905; LaFleur, 1965; Connally and Sirkin, 1986; DeSimone and LaFleur, 1986). The lower lake stages are recorded by fluvially-eroded terraces, spillways, catastrophic flood channels, small deltas, and infrequent beaches.

Lake Quaker Springs received large quantities of glacial meltwater from the retreating ice margin. Large tributary basins, particularly north and west of Albany, fed vast amounts of water into the later lake stages. Glacial Lake Iroquois, a large proglacial lake in the Ontario Lowlands, drained through the Mohawk Valley via the Rome-Little Falls outlet (Muller and Prest, 1985). The Iroquois waters entered Lake Albany at Scotia, NY, where they built cobbly terraces (LaFleur, 1979, 1983, 1992). Catastrophic floods from Lake Iroquois carved large, deep channels that were graded to the Quaker Springs through Coveville stages of Lake Albany (Stoller, 1920; LaFleur, 1965b; Hanson, 1977; LaFleur, 1983; Dineen and others, 1983; Dineen and others, 1988). Lake Iroquois and its successor, Lake Frontenac, eventually drained through the Covey Hill Spillway into Lake Vermont when the Ontario Lobe retreated north of the Adirondack Mountains (Chapman, 1937; LaFleur, 1965a; Muller and Prest, 1985).

As the glacier retreated into the Champlain Valley, Lake Vermont formed between the Hudson-Champlain divide and the ice front (Chapman, 1937). Waters escaping from Lake Vermont carved a series of channels near Fort Ann (DeSimone and LaFleur, 1985 and 1986). Lakes Coveville and Fort Ann are stages of Lake Vermont that were broad rivers in the Hudson Valley (LaFleur, 1965a).

The Lake Albany-Lake Vermont sequence of lakes ended when the ice front receded from the Covey Hill spillway into the St. Lawrence Lowlands (LaSalle, 1966). This event allowed the Champlain Sea to invade the St. Lawrence and Champlain Lowlands, and cut off the supply of glacial meltwater to the Mohawk, Champlain, and Hudson Valleys (Clark and Karrow, 1984; Muller and Prest, 1985). Lake Fort Ann was fed by waters that escaped from the Ontario basin along the north slopes of the Adirondacks and carved the Chateauguay and Covey Hill channels (Denny, 1974; Pair and others, 1990).

GEOGRAPHY

The mid-Hudson Valley region in eastern New York State is unique in that it is at the convergence of five physiographic provinces: the Mohawk and Hudson-Champlain Lowlands and the Adirondack, New England, and Appalachian Uplands (Fig. 1). The Appalachian Upland can be divided into the Helderberg and Catskill Plateaux and the New England Upland includes the Taconic Mountains and the Rensselaer Plateau. The Mohawk Lowland connects the Erie-Ontario and Hudson-Champlain Lowlands, while the Champlain Lowland is connected to the St. Lawrence Lowland. The St. Lawrence Lowland is connected to the Erie-Ontario Lowlands across the northwestern edge of the Adirondack Uplands.

The Hudson Lowlands are underlain by highly folded (east) to relatively undeformed (west) Lower Paleozoic shale and sandstone (Fisher and others, 1970). The eastern Mohawk Lowlands are underlain by half-grabens containing Lower Paleozoic shale, limestone, and sandstone, with infrequent but erosionally significant Precambrian metasediments in the Mohawk Gorge (Fisher, 1980). The New England Uplands, to the east of the lowlands, are underlain by metamorphosed Precambrian through Lower Paleozoic igneous and sedimentary rocks. The Catskill Mountains and Appalachian Plateau lie west of the lowlands. They are underlain by slightly deformed sediments of Middle to Upper Devonian age. The Adirondack Uplands are underlain by Precambrian metasedimentary and metaigneous rocks. The high ranges of the New England Uplands separate the Hudson and Connecticut Lowlands.

Several Cenozoic erosion surfaces are preserved near Albany (Table 1; Dineen, 1987). The higher surfaces define the skyline of the area from Newburgh to Schuylerville. The Catskill Erosion Surface is preserved in the High Peaks of the Catskill Mountains, and around the periphery of the High Peaks of the Adirondacks. The

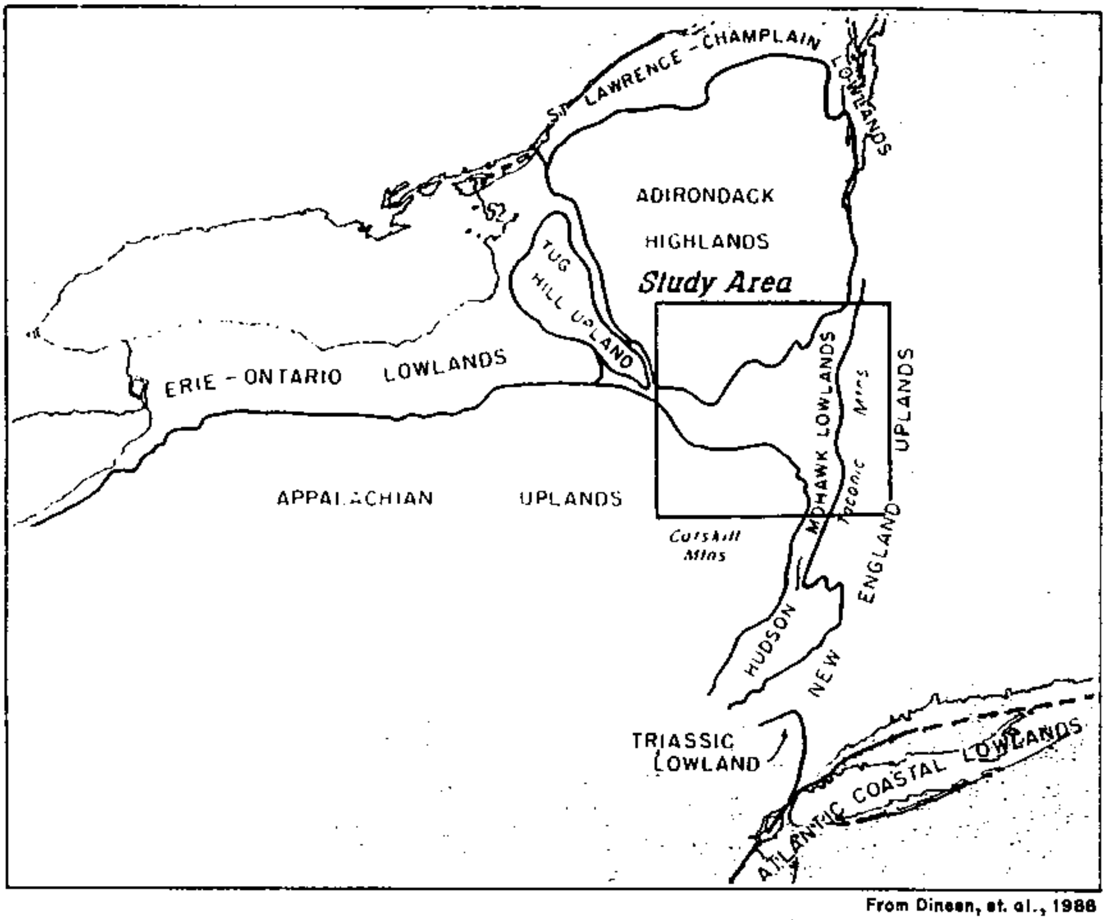

Figure 1.
PHYSIOGRAPHIC PROVINCES OF NEW YORK STATE

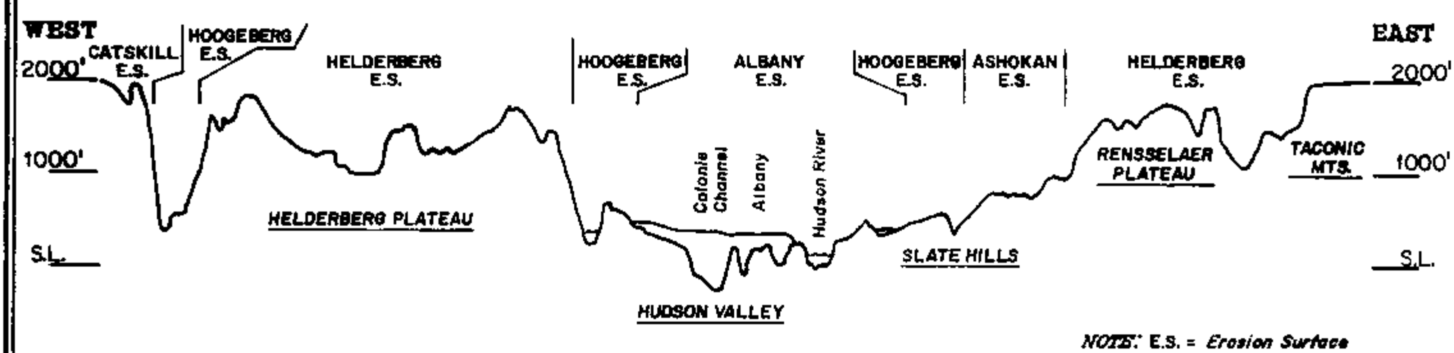

Figure 2.
EAST-WEST PROFILE OF THE PHYSIOGRAPHIC PROVINCES IN THE ALBANY AREA

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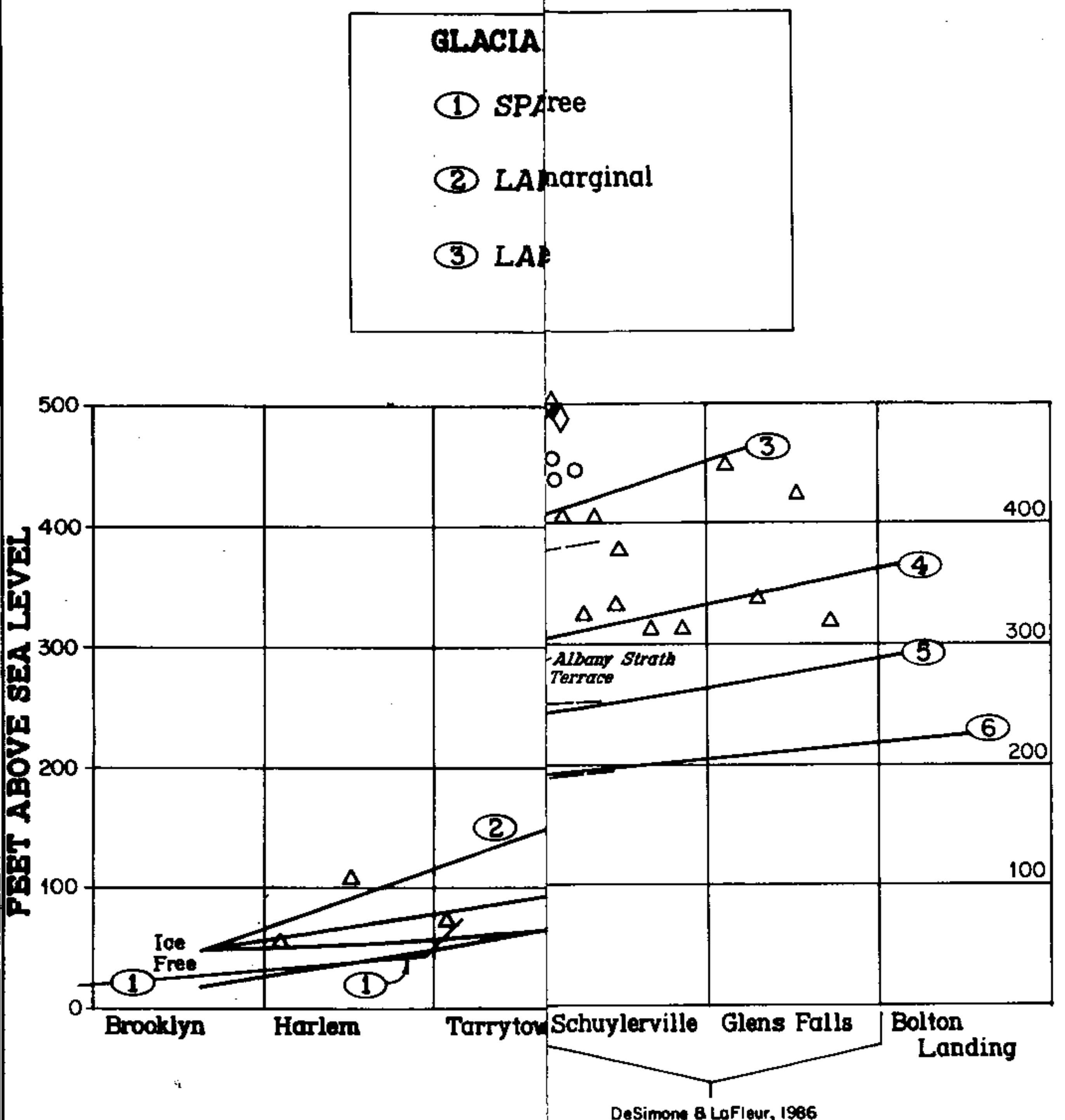
High Peaks of the Adirondacks are resistant intrusions of massive anorthosite that tower above the Catskill surface. The Helderberg surface rims the Catskills and the southern and eastern Adirondacks, and forms the crestline of the Taconics. The Monticello Erosion Surface caps the Rensselaer Plateau, and extends into the Mohawk Valley. The Ashokan and Hoogeberg surfaces form bedrock benches along the edges of the Hudson, Champlain, and Mohawk Valleys (Fig. 2). All of these erosion surfaces have been modified to a greater or lesser degree by glacial sculpting. Isolated patches of residual soils on high mountaintops in the Catskills (Rich, 1935), suggest that the highest surfaces were eroded less than lower surfaces.

Four southward-dipping strath terraces lie in the inner portions of the valleys (Dineen and others, 1983; Dineen, 1987). They include the glacial/interglacial Malta, Albany, and Guilderland Strath Terraces and the Holocene Castleton Strath Terrace. They are preserved along the interfluves of the pre- and interglacial drainage channels, and helped to control the width and depth of the later glacial lakes. The Castleton is drowned by the estuary of the Hudson River south of Troy and cannot be traced north of Cohoes (Fig. 3). It coincides with the depth of Holocene sediment, and was probably carved during Holocene time by the river as sea level progressively flooded the lower Hudson Gorge (Dineen, 1987). The Guilderland strath terrace coincides with the elevation of Lake Fort Ann and was modified by Lake Fort Ann waters (Fig. 3).

TABLE 1: EROSION SURFACES

Erosion Surface	Elevation in Feet Above Sea Level (from Newburgh to Schuylerville)	Gradient (ft/mile)
Catskill	4,000	
Helderberg	1,500 to 2,000	
Monticello	1,200 to 1,500	
Ashokan	700 to 1,100	
Hoogeberg	400 to 600	
Malta	250 to 390	1
Albany	210 to 260	1.25
Guilderland	105 to 195	4
Castleton	-90 to 20 (@ Kingston)(@ Cohoes)	1

The Hudson-Champlain Lowlands were drained by a trellis drainage network during the late Cenozoic (Fig. 4; Dineen and others, 1983; Dineen, 1987; Dineen and others, 1988). This network was rejuvenated in late preglacial or interglacial time, when deep, north-south trending parallel gorges and sharp breaks in the stream gradients were formed, apparently in response to lower sea level. The Colonie, Battenkill-Hudson, and Mohawk channels were the trunk streams, but only the Battenkill-Hudson channel can be traced south of Ravena (Dineen and Rogers, 1979; Dineen, 1987). Distinct breaks in gradient occur in these channels (Dineen and others, 1983), suggesting headward erosion of knickpoints, and hinting at ancient



E LEVELS, & STRATH TERRACES

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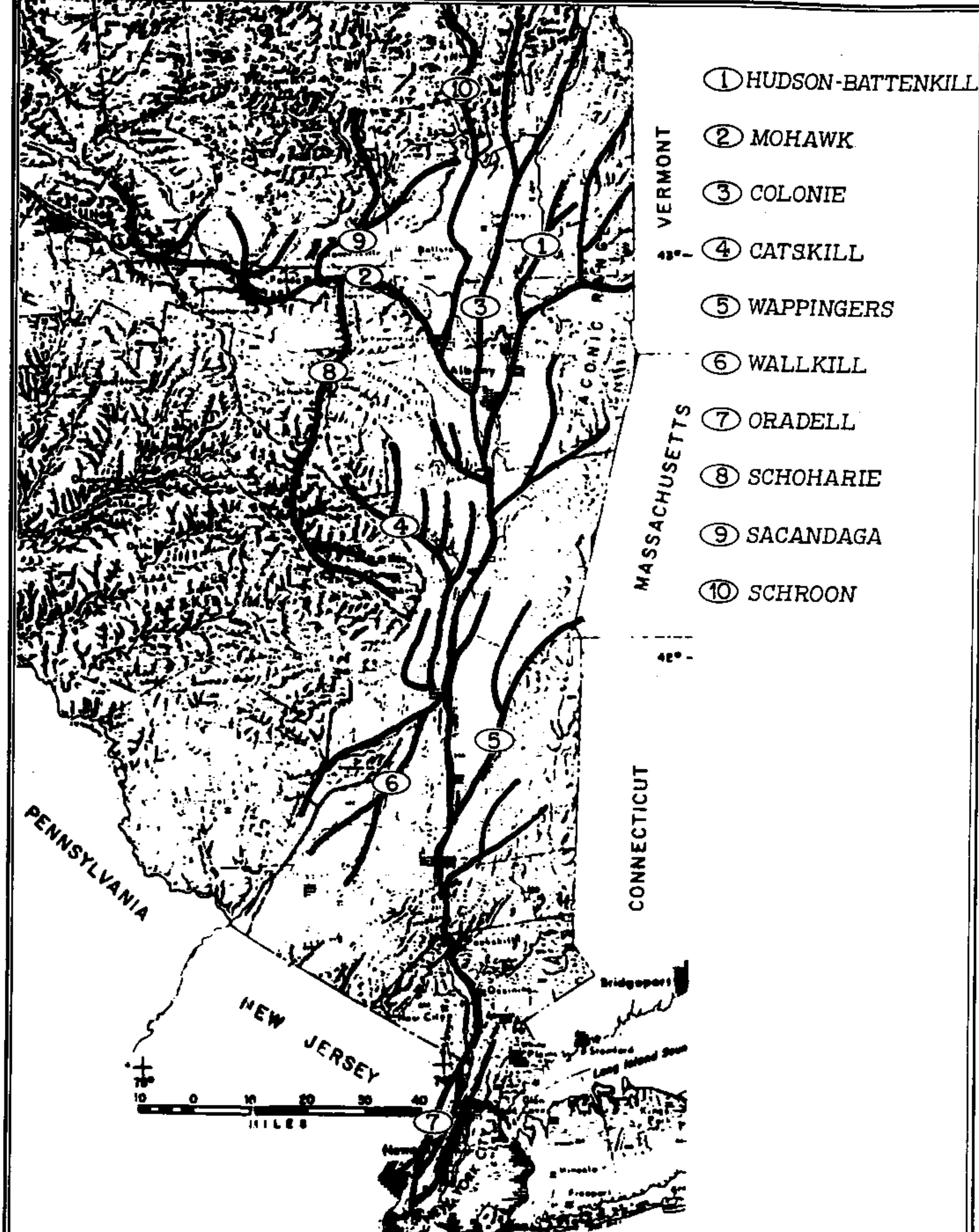

Figure 4.
PREGLACIAL DRAINAGE CHANNELS

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changes in base level.

The preglacial Mohawk River headed in the Deerfield Hills area near Utica (Fig. 4). Major tributaries formed on the lower portions of the tilted half-grabens. These tributaries include the West and East Canada Creeks and the Sacandaga Lake portion of the modern Sacandaga River.

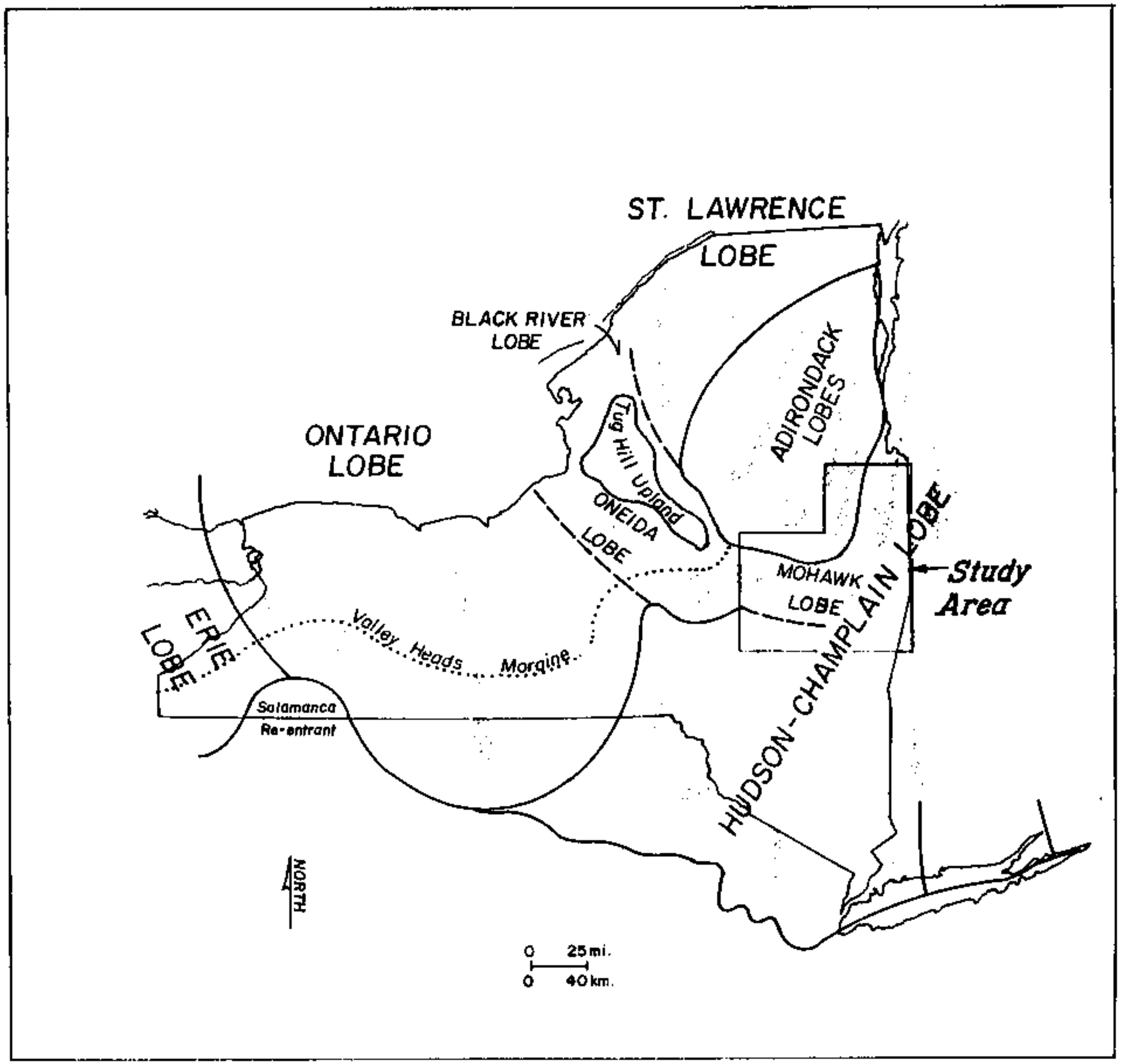
These large-scale geographic features controlled the flow of the Wisconsinan Glacier through eastern New York State (Fig. 5; Hughes and others, 1985; Dineen and Hanson, 1985; Dineen and others, 1988). The Adirondack Uplands split the flow of the eastern section of the Laurentide Ice Sheet into the St. Lawrence, Ontario, and Hudson-Champlain Lobes. Ice flow through the Adirondacks was weak, and the Ontario and St. Lawrence Lobes were able to push the Oneida and Black River sublobes into the western Mohawk Valley, while the Hudson-Champlain Lobe pushed the Mohawk sublobe into the eastern Mohawk Valley (Dineen and others, 1988; Ridge and others, 1990).

Large glacial lakes occupied the various lowlands as the Wisconsinan glacier retreated (Figs. 6 and 7). The lakes became younger from south to north and east to west. They communicated with each other as the retreating ice front exposed a succession of interconnecting spillways and cols (Figs. 1 and 6).

Glacial Lake Albany sediments lie in the Hudson-Champlain Lowlands of eastern New York (Figs. 6 and 7). These sediments form extensive clay terraces that border the Hudson River from Newburgh, NY, north to Glens Falls, NY (Merrill, 1890, 1891; Ries, 1890; Upham, 1903; Peet, 1904; Woodworth, 1905; Cadwell and Dineen, 1987; Cadwell, 1989). Similar clay terraces lie along the Hudson River south of the Hudson Highlands from Peekskill, NY, to Staten Island, NY (Merrill, 1890, Ries, 1890; Jones, 1899; Woodworth, 1905; Reeds, 1927). The river side terraces are underlain by scoured till (Upham, 1903; Peet, 1904; Woodworth, 1905; Cadwell, 1989), deltas (?) and kame terraces (?) (Cadwell, 1989), and lag gravels from Newburgh to Peekskill. The clay terraces grade into higher sand and gravel terraces away from the river.

SEDIMENTOLOGY

The earliest mappers in the Hudson Valley recognised the complex nature of the Lake Albany deposits. Eights (1852), Ries (1890), Jones (1899), Peet (1904), Woodworth (1905), Stoller (1911) described the wide-spread Hudson Valley clay deposits. Ries (1890), Upham (1903), Peet (1904), and Fairchild (1917, 1919) interpreted these clays as estuarine or marine deposits while Woodworth (1905), Stoller (1911, 1919, 1920, 1922), and Reeds (1927) cited abundant evidence that these were lacustrine deposits. This disagreement sparked a lively debate throughout the early part of this century! The argument was eventually settled in favor of



Modified from Dineen and others, 1988.

Figure 5.
GLACIAL LOBES IN NEW YORK STATE

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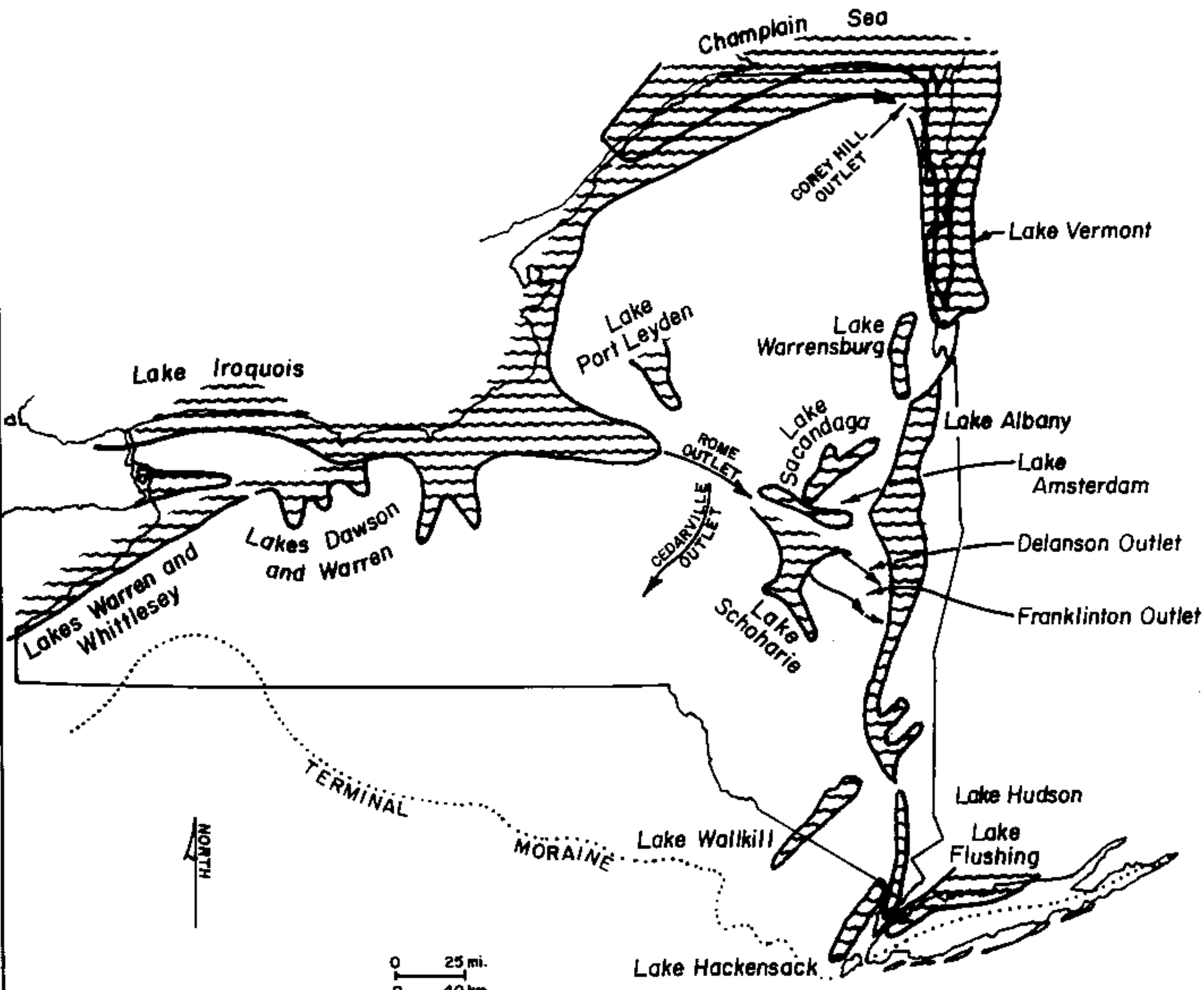

Figure 6.
GLACIAL LAKES IN NEW YORK STATE

Modified from Dineen and others, 1988.

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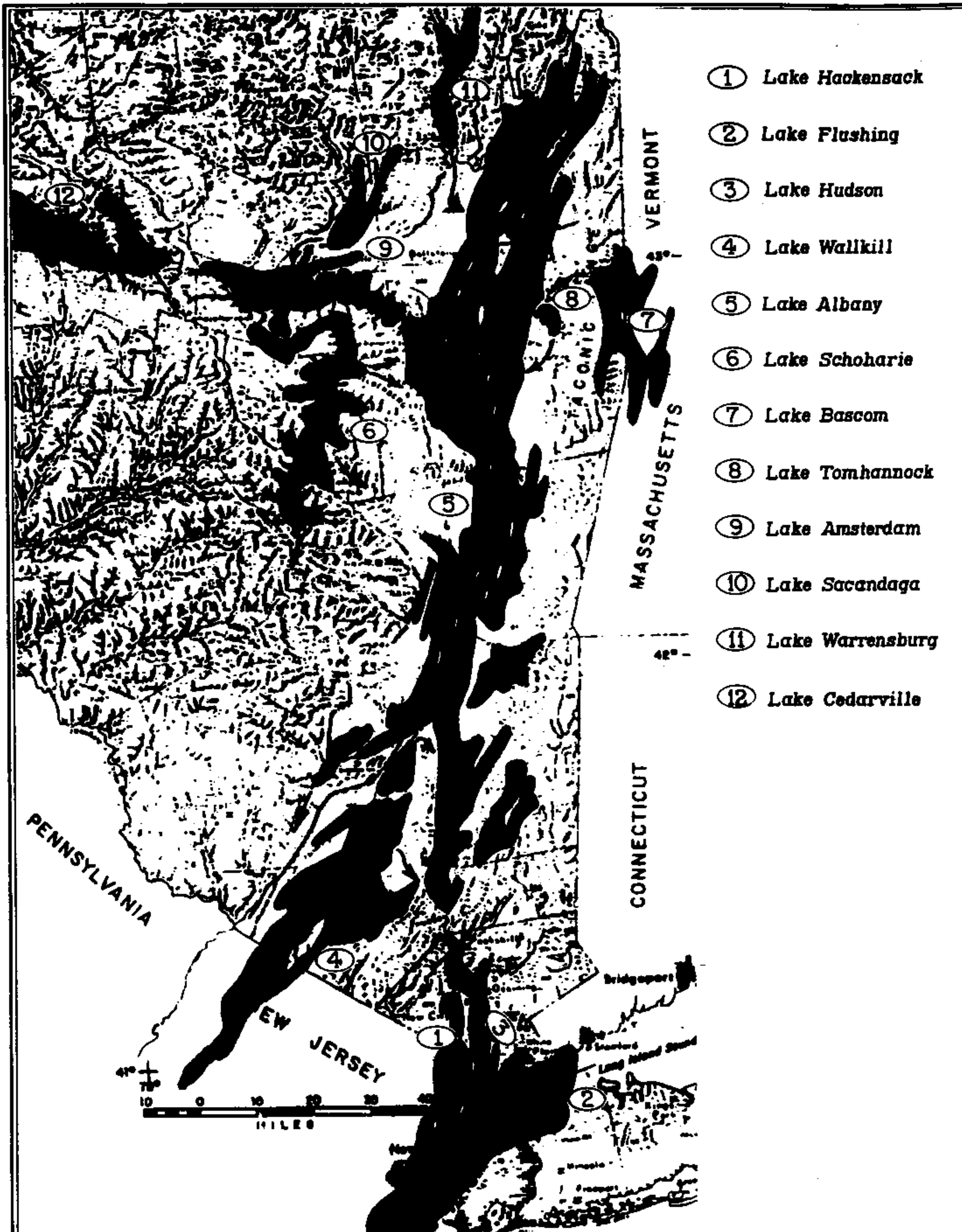


Figure 7.
GLACIAL LAKES OF THE HUDSON LOWLANDS

LMD-5/15/92

a lacustrine environment.

The sedimentary facies of Lake Albany can be deduced using the studies discussed above and more recent work by LaFleur (1965a, 1969a), Dineen and Rogers (1979), Dineen and others (1983), and Dineen and Duskin (1987), along with observations of exposures, drill cores, water well logs, and geophysical data. The sedimentary sequence begins with a basal ice-contact facies, which grades up to a deep-water facies, which then grades up into a shallow-water or nearshore facies. This sequence ends with fluvial and/or aeolian deposits (Fig. 8). The facies and geologic field trip stops that illustrate them are summarized below and in Appendix A.

<u>Basal Unconformity</u>: The base of the Lake Albany sequence generally is in abrupt, unconformable contact with the underlying bedrock or lodgement till. Many exposures, water wells, and test borings in the preglacial channels encountered large areas of till or bedrock under the lake clays, and the Basal Ice-contact Facies was missing.

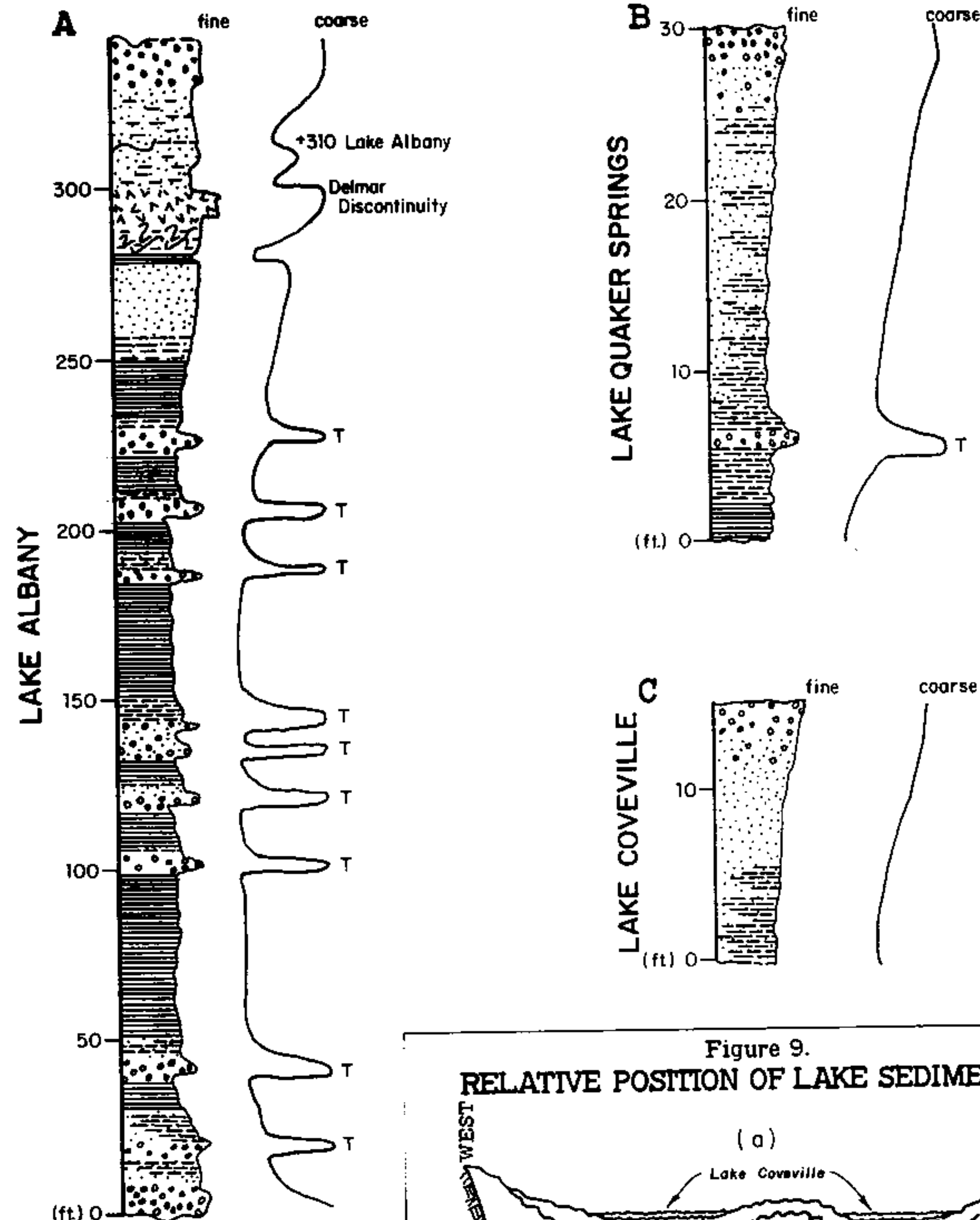
<u>Basal Ice-contact Facies:</u> The basal Lake Albany sediments are mostly faulted and deformed sand and gravel in kames, subaqueous fans and/or eskers that grade into valley-side kame deltas (LaFleur, 1969; Dineen and Rogers, 1979; Dineen and others, 1983). The basal, ice-contact facies tend to form ridges perpendicular to the ice fronts. The ridges trend northwest-southeast across the buried valleys.

<u>Deep-water Facies:</u> The middle portion of the Lake Albany sequence primarily is rhythmically bedded silt and clay. These beds grade up from ripple-laminated silt to rhythmites or "varves" of clay and silt or fine sand. The thickness of each couplet and the proportion of silt-to-clay in them decreases upward. This facies also includes turbidites, beds of contorted clay, striated boulders, and "rain-out" or dropstone till. The turbidites extend over large areas, based on geophysical interpretation of test borings (Dineen and others, 1983). The "rain-out" till and boulders are probably from icebergs (DeSimone, 1985; DeSimone and LaFleur, 1986) and floating ice shelves (Dineen and others, 1983). The locally contorted clay beds might have resulted from icebergs grounding on the lake bottom, by till impacting the bottom after release from icebergs, or by slumping of unstable bottom deposits. The bedrock topography controlled the locations of ice margins and their associated basal ice-contact deposits. The ice margin hesitated at sharp breaks in the preglacial valley gradients. Thick wedges of basal ice-contact deposits accumulated at those breaks (Dineen and others, 1983).

The bedrock topography also controlled the deposition of the clay. Each of the deep, narrow valleys (Fig. 4) acted as a sediment trap or basin, where thick wedges of silt and clay accumulated. The linear north-south valleys impeded east-west transport and

FACIES SEQUENCE OF THE HUDSON VALLEY GLACIAL LAKES

Figure 8.



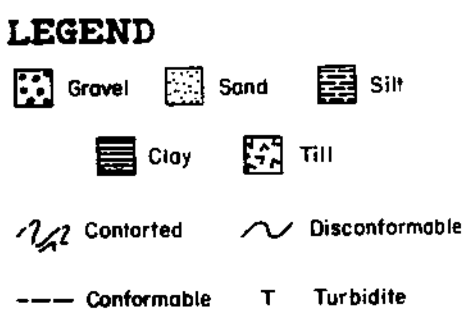
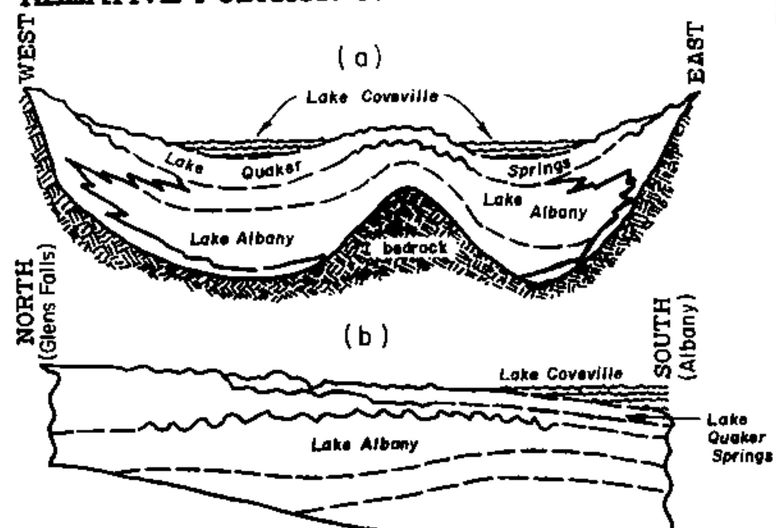


Figure 9.

RELATIVE POSITION OF LAKE SEDIMENTS



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facilitated north to south transport of sediment. In the Hudson Lowlands, the western valleys tend to reflect the contribution of the western tributaries, the central valleys derived most of their sediment from the ice-front, and the eastern valleys received sediment from the tributaries to the east. The interpretation of geophysical logs also suggests that the clay units are shingled systematically from south to north, and that the clays have compacted so that they sag towards the axes of the buried bedrock valleys (Fig. 9; Dineen and Rogers, 1979; Dineen and others, 1983).

The colors of the clay are clues to the origin of the deposits. The clay is rock flour ground from bedrock surfaces, thus distinct reddish clay beds occur where red beds are common, such as adjacent to the Esopus Creek and Cats Kill Valleys in the Hudson Valley and in the western Mohawk Valley near the Ontario Basin and the Hudson Valley at Colonie. Paleozoic redbeds are exposed in the southern Helderbergs and the Erie-Ontario Lowlands. The red Cambrian-Ordovician slates in the Batten Kill and Hoosic River Drainage Basins contributed red clay to the Battenkill-Hudson Gorge next to the Hoosic and Batten Kill deltas. The phyllites in the Hoosic and Batten Kill basins contributed green clays to the easternmost buried channel. White clay occurs near the mouth of the Kayderosseras Creek, a stream that drained the Adirondack Mountains. The blue or dark gray color of most of the clay deposits is from the dark Ordovician shales that lie along the Mohawk and Hudson Valleys. Oxidation and leaching causes the upper layers of the gray clay to become brown or yellow, and causes carbonate-sulfate concretions to form.

The deep water facies is thickest in the preglacial valleys, such as the Colonie and Hudson-Battenkill channels. The deepwater and shallow water facies pattern exhibit linear, north-south patterns as a result. The linear pattern are especially developed where the glacial lake levels coincided with the strath terraces. In these areas (such as the area from Newburgh to Kingston) the lakes had a strong estuarine appearance (Fig. 7). The interfluves between the channels formed elongate islands throughout the lake basin (see the area from Kingston to Glens Falls).

Shallow-water Facies: The higher wave and current energy in shallow water produced wide-spread blankets of yellow brown, fine to medium sand that are twenty to forty feet thick, and cover the deep-water clay throughout the Hudson Valley. The sand is usually planar to ripple laminated. The sand generally has a gradational contact with the underlying clay, except along the edges of the lake plain. It might contain a major contribution of wind-blown, lake deposited sand and silt, particularly in the Albany to Schuylerville area (Dineen and Rogers, 1979).

The shallow-water deposits began to accumulate as the lake basin filled with sediment and the lake levels fell (Fig. 8). Thus, they formed in response to the (lake) bottom coming up and the top

(water surface) coming down!

Near-shore Facies: The lake shore and adjacent environments produced several types of deposits and landforms. They include: deltas with topset, foreset, and bottomset beds; fluvial deposits, which include terrace deposits of the later lake stages, and grade into the deltaic topsets; off-shore bars and lagoons; and beaches, including wave-cut platforms. All these sediments were deposited in relatively high-energy environments.

Deltas: The deltas along the Hudson Valley can be classified as kame deltas, ice-marginal deltas, and ice-free deltas (Fig. 3). Glacial Lake Albany delta sizes tend to be greatest in the Albany area and decrease dramatically to the north (Hanson, 1977; Dineen and others, 1983; DeSimone, 1985; DeSimone and LaFleur, 1986). The deltas in the lower lake stages usually increase in size from south to north (Hanson, 1977; Dineen and others, 1983).

Kame deltas: Comprise outwash derived from the ice front. They have a collapsed ice-contact face on their proximal side. The Schodack, Pollock Road, Rensselaer, Waterford, Hampton, and Newtown Road kame deltas (Dineen and others, 1983) are good examples of this type of delta. They mark the edge of active ice.

Ice-marginal deltas derived their sediment as inwash from the highlands adjacent to the glacial lakes. The collapsed ice contact face is on the distal edge. The Red Hook, Catskill, upper Kinderhook, Voorheesville, upper Hoosic, and upper Batten Kill deltas are examples of ice-marginal deltas. Most were deposited next to stagnant ice.

Ice-free deltas: Lack ice-contact features, except for small kettle-holes. They are most abundant in the lower lake stages and include near-shore subaqueous fans and alluvial fan-deltas. Many of the deltas associated with the smaller tributaries have relatively "starved" topset sequences because many of the later lake stages were short-lived, only isolated blocks of melting glacial ice were present in the drainage basins of the deltaic streams, and stream sediment supply was dependant on local precipitation in basins that were being vegetated. The Kingston, lower Kinderhook, lower South Bethlehem, Schenectady, lower Hoosic, and lower Batten Kill deltas are examples of this type.

Beach Facies: Many rock ridges along the Hudson Valley are partially covered with gravelly silt. The gravel is usually quite angular and is derived from local rock. These ridges apparently were above wave-base in the lake, so winnowing of any till mantle or erosion of the rock face could take place. Lag deposits of local gravels accumulated around the ridge.

Beach Features also include planar-bedded gravelly sand, ripple-trough-laminated sand, boulder pavements, and wave-cut scarps

(LaFleur, 1965a; Dineen and Rogers, 1979; Dineen, 1982).

Off-Shore Bars: Drilling in the central Pine Bush of Albany County has outlined an off-shore bar and lagoon system (Dineen, 1982). The bar consist of planar to ripple laminated fine to medium grained sand in ribbons that are parallel to the paleo shorelines of the lakes. The sand ribbons enclose a lens of massive to laminated silt and silty clay. The clayey lens was a lagoon that was bordered on the shoreward side by a dunefield and on the lakeward side by an off-shore bar (Dineen, 1982).

Fluvial Facies: Fluvial deposits grade into the deltaic topsets. They include terraces that are associated with rivers that fed deltas, terrace deposits inset into pre-existing lake deposits during lower lake stages, and major lake outlet or catastrophic flood channel deposits. These deposits usually underlie terraces that are graded to deltas in adjacent lakes. Channels that are incised into older, higher lake deposits are associated with the lower lake stages and bear an unconformable relationship with the older lake deposits.

Floods from Lake Iroquois carved large, deep channels that were graded to the Quaker Springs through Coveville stages of Lake Albany (Stoller, 1911, 1919, 1920; LaFleur, 1965b, 1969a, 1979; Hanson, 1977). These magnificent channels are preserved in the upper Hudson Valley from Scotia to Schuylerville. Lake Iroquois and its successor, Lake Frontenac, eventually drained through the Covey Hill Spillway into Lake Vermont when the Ontario Lobe retreated north of the Adirondack Mountains (Chapman, 1937; LaFleur, 1965a; Muller and Prest, 1985). Waters escaping from Lake Vermont carved a series of channels near Fort Ann (DeSimone and LaFleur, 1985, 1986) that includes both outlet and catastrophic flood channels.

The floods also deposited large areas of slackwater clay and silt along the upper edges of the channels from Schenectady to Schuylerville (Dineen, Round Lake and Burnt Hills 7-1/2 minute quads, NYSGS Open File Maps; Hanson, 1977; Dineen and Hanson, 1983), and scoured longitudinal grooves in the Fort Ann channels (DeSimone and LaFleur, 1985, 1986). Overflow channels were cut across the interfluve between the Ballston and Colonie Channels by the floods. Fans of sand and gravel lie at the mouths of these overflow channels in the Clifton Park area (Dineen and others, 1983). The fans are often overlain by slackwater silt. The channel bottoms are locally covered with 5 to 50 feet of open-work cobbly gravel particularly at Scotia (LaFleur, 1983, 1992; LaFleur and Wall, this volume), at the outlet of Saratoga Lake (Hanson, 1977), and in the East Line area.

Lakes Coveville and Fort Ann were broad rivers in the Hudson Valley (LaFleur, 1965a; DeSimone and LaFleur, 1986). The Coveville stage was lacustrine in the Albany area, where numerous deltas, beaches,

spits, and deep-water Coveville deposits have been preserved (Dineen and Rogers, 1979; Dineen and Hanson, 1983).

Aeolian Facies: Dune Sand overlies many of the sand plains and deltas of the Hudson Valley and forms dunal ridges. The aeolian sand unconformable overlies the lacustrine deposits. The contact between aeolian sand and lacustrine sand is quite irregular, with 3 to 5 m (10 to 15 ft) of relief. Yardangs are present in the western (up-wind) portions of the dune fields. Both parabolic and linear dunes are present downwind. The cross-bedding and dune morphology suggest that the dune-forming wind came from the northwest (Dineen and Rogers, 1979; Dineen, 1982).

TERRACES, LAKE PLAINS, AND ISOSTATIC REBOUND

Merrill (1890), Ries (1890), Upham (1903), Peet (1904), Woodworth (1905), Fairchild (1917, 1919), and Stoller (1919, 1920, 1922) all used the elevations of the valley-side deltas, wave-cut cliffs, and beaches to define water planes (marine or lacustrine, depending on taste) that tilted up to the north at 2.5 ft/mile (Woodworth, 1905). Woodworth (1905) and Stoller (1919, 1920, 1922) also described multiple lake levels.

The clay and sand terraces and their associated deltas, beaches, and dunes constitute the primary evidence for several stages of Lake Albany. The water levels of the lake can be deduced from the delta elevations, but the elevations of the deltas and other lacustrine features show a significant degree of scatter, complicating the lake correlations (Dineen and others, 1988; Fig. 3). Secondary features include beaches, sand and clay terraces, kame terraces, tributary stream terraces, outlet and inlet channels, and catastrophic flood channels. This suite of features allows us to make a first approximation of the sequence of lake stages and of their relative ages and the magnitude of post-glacial rebound.

The multiplicity of deltas and beaches define many lake stages in the Hudson Valley (Fig. 3). The earliest stage is the highest, and the lakes become younger with decreasing elevation. The highest stage is defined by both ice-marginal and kame deltas, and is clearly a proglacial lake (Woodworth, 1905; Fairchild, 1917; Chadwick, 1928; LaFleur, 1965a and b; Connally and Sirkin, 1986). A lower, somewhat ice-free Lake Albany stage has been mapped in the Albany area (LaFleur, 1965a; Dineen and others, 1983, 1988). It was characterized by fewer kame deltas (Fig. 3). This level was considered the "Stable Lake Albany" stage by Connally and Sirkin (1986). Ice-marginal deltas persist until the Lake Coveville stage. Lake Coveville has not been split into substages. DeSimone and LaFleur (1986) were able to divide the Fort Ann Stage into three levels based on three overflow channels for Lake Vermont-Fort Ann stage in the vicinity of Glens Falls, NY. They confirmed the fluvial nature of the Fort Ann stage in the Hudson Valley. The

scatter of points representing deltas suggest that several intermediate "stable" substages probably exist for Lakes Quaker Springs and Coveville.

Lake Albany levels have been confidently traced from Glens Falls, NY, to Newburgh, NY (Woodworth, 1905; LaFleur, 1965a; Dineen and others, 1988). Lake terraces have been mapped south of the Hudson Highlands (Reeds, 1927), although terraces within the Hudson Highlands have been shown to be underlain by till (Peet, 1904; Woodworth, 1905) or terraced gravels (Cadwell, 1989). The lakes south of the Highlands were named Lakes Hackensack and Hudson by Reeds (1927). Woodworth (1905) thought that an ice margin at the north edge of the Hudson Highlands was the northern limit of Lake Hudson. Connally and Sirkin (1986) correlated Lake Hudson with their "Stable Lake Albany" stage. Dineen and others (1988) correlated Lakes Hackensack and Hudson with early Lake Albany, and considered them to be continuous from the Terminal Moraine to Glens Falls (See Table 2).

Lake Hackensack shorelines dip south at 2.25 ft/mi (0.43 m/km) and lie 40 ft (12 m) above the projection of the Lake Albany shoreline at Piermont, New York (Fig. 3; Reeds, 1927). The water plane is concave upward. Thus, Lakes Hudson and Hackensack are not the same lake.

Croton Point is a proglacial Lake Hudson delta at an elevation of 100 ft (30 m) ASL (Markl, 1971). Small deltas and erosional surfaces can be found along the Hudson gorge south of Sparkill Gap. The gradient of the water plane of Glacial Lake Hudson determined by these features is 1.19 ft/mile (0.23 m/km).

The "water planes" defined by the terraces, beaches, and deltas of the highest stage of Lake Albany tilt up to the north at 2.6 ft/mile to 2.7 ft/mile (0.49 to 0.51 m/km) (DeSimone and LaFleur, 1986). They include a series of "lake levels" defined by ice-marginal deltas that are concave upwards to the north. The concave-upwards ice-marginal lake levels form several sets. Three such levels can be traced between West Point and Catskill. They occur in the estuary-like section of the Lake Albany (Fig. 7). They are graded to successively lower lake levels, suggesting an eroding spillway as the Hudson Lobe retreated from Newburgh to Kinderhook. An additional set of concave-upward ice-marginal lake planes lies north of Schuylerville. The water planes of the lower stages are tilted to a lesser degree- those of Lakes Quaker Springs and Coveville are tilted at 1.5 ft/mile (0.28 m/km)(LaFleur, 1965a) to 1.6 ft/mile (0.3 m/km)(DeSimone and LaFleur, 1986).

The multiple Fort Ann stage tilts from 1.05 ft/mile to 1.0 ft/mile (0.2 to 0.19 m/km)(DeSimone and LaFleur, 1986; DeSimone, this volume). Thus, the water planes of the various lake stages converge to the south and meet at Newburgh (Fig. 3). The Fort Ann levels are coincident with the Guilderland strath terrace. Fort

TABLE 2: LAKE LEVELS

Lake Level	Elevations (in feet)	Gradient (ft/mile)
Hackensack	(Carteret, NJ to Newburgh)	
	70 to 250	2.25
Hudson	(Harlem to Newburgh)	
	30 to 120	1.19
Albany Sequence	(Newburgh to Schuylerville)	
Albany	130 to 420	2.65
Quaker Springs	130 to 310	1.5
Coveville	130 to 240	1.5
Fort Ann	130 to 190	1.5

Ann might have been controlled, in part, by the strath terrace.

The uppermost, proglacial Lake Albany stage was thought to be a product of a "peripheral bulge" that formed in front of the retreating ice sheet (Peet, 1904). The peripheral bulge was caused by a very rapid isostatic uplift of the earth's crust that formed a few tens of miles in front of the ice tongue. Fairchild (1917) considered the bulge to be a wave movement that was 100 miles wide and 100 feet high. The wave migrated north as the ice sheet retreated. Chadwick (1928) presented a theoretical argument for the presence and character of a peripheral bulge (see figs. 11 through 16 in Chadwick, 1928). Connally and Sirkin (1986) resurrected the peripheral bulge idea and described the upper Lake Albany stage as a very short-lived, very short lake that formed along or on the retreating ice margin. Unfortunately, no one has been able to predict unique geomorphic or stratigraphic features that would prove the existence of a bulge, although the concave-upwards curve of ice-marginal deltas on Lake Hudson-Hackensack and on the Lake Albany curve suggest that a short-lived, rapid period of rebound did occur.

Woodworth (1905), Fairchild (1917), LaFleur (1965a), and Connally and Sirkin (1986) noted "hinge lines" on the lake planes at the Hudson Highlands (at Newburgh) and at Glens Falls, where the terrace levels seem to undergo sharp changes in grade. The terraces are less steep or flatten out south of the Highlands and become significantly steeper north of Glens Falls. The locations of the hingelines, at either end of the Hudson Lowlands, suggest that either regional geology or differential ice thicknesses in the Hudson Gorge, Hudson Lowlands, and Champlain Lowlands caused significant "pulses" in isostatic rebound.

THE LAKE OUTLET PROBLEM

As noted above, Glacial Lake Albany was a northward-expanding proglacial lake that developed in the Hudson Lowlands as the Hudson-Champlain Glacial Lobe retreated. The lake extended to Glens Falls, New York, at its maximum extent (Chadwick, 1928; DeSimone and LaFleur, 1986). It received all the meltwater that spilled into the Great Lakes drainage basin from the retreating Wisconsinan Continental Ice Sheet (Stoller, 1922). The northern end of Lake Albany is fairly well mapped, whereas its southern end is not. The lake deposits have been traced as far south as Newburgh, New York (Woodworth, 1905; Cadwell, 1989). The dam or spillway that controlled the level of the lake has never been identified in the field.

Lake Albany was dammed by either a proglacial "peripheral bulge" (Peet, 1904; Fairchild, 1917; Chadwick, 1928; Connally and Sirkin, 1986) or by a recessional moraine between the Hudson Highlands and Long Island (Woodworth, 1905). The peripheral bulge would have developed as an isostatic response to glacial unloading of the

earth's crust and would have formed a "wave of uplift" in front of the retreating Hudson-Champlain Lobe (Peet, 1904). Thus, Lake Albany's early ice-contact stage might have been a localized, ice-marginal lake trapped between the peripheral bulge and the ice front (Chadwick, 1928; Connally and Sirkin, 1986).

Woodworth (1905) thought that Lake Albany was dammed by a recessional moraine in the Hudson Highlands and considered the till terraces in the Hudson Highlands gorge to be remnants of the recessional moraine. Reeds (1927) and Connally and Sirkin (1986) suggested that the dam was an extension of the Terminal Moraine between Staten Island and Long Island.

The difficulties in locating evidence for the dam or spillway of Lake Albany may be explained by the subsequent "relaxation" of the peripheral bulge (Peet, 1904; Connally and Sirkin, 1986) or by late glacial or post-glacial erosion of a recessional moraine dam (Woodworth, 1905; Reeds, 1927; Connally and Sirkin, 1986).

Dineen and others (1988) suggested that Glacial Lake Albany probably existed for 7,000 years and considered this to be a long time for either the Terminal Moraine or an earthen recessional moraine to dam a glacial lake that drained meltwater from the Great Lakes Basin. The dam would have been slowly weakened by the melting of buried ice blocks and by ground water and stream sapping and would have been further stressed by sudden rises in lake level caused by catastrophic bursts of water from the sudden release of proglacial lakes along the retreating flank of the ice sheet (DeSimone and LaFleur, 1986). Dineen and others (1988) suggested that the rocky sill of Sparkill Gap, a windgap at Piermont on the west shore of the Hudson River in Rockland County, NY, was the outlet for Lake Albany and that the Terminal Moraine was the dam.

The Glacial Lake Hudson water plane has a shallower gradient than the water plane of Lake Albany, although it appears to coincide with the water planes of the lower lake stages (Fig. 3). The projection of the Lake Hudson water plane lies 50 ft (15.2 m) above the floor of the notch in Sparkill Gap near Piermont NY (Averill and others, 1980; Dineen, 1986b; Dineen and others, 1988). The projection of Lake Albany's water plane lies between 30 feet ASL and sea level near Piermont, NY.

Fluvial sands can be traced south from the Sparkill Gap into the Hackensack Lowlands, where they overlie truncated Lake Hackensack deposits (figs. 15 through 18 in Lovegreen, 1974; fig. 15 in Averill and others, 1980). Water debouching through the spillway from Lake Hudson and early Lake Albany deposited the fluvial sequence in the Hackensack Lowlands (Dineen and others, 1988). The water would have been nearly sediment-free because the lakes would have trapped nearly all of the sediment available. Thus the fine alluvial sands in the Hackensack basin are probably locally-derived. Later outlets might have included Hell Gate and the East

River.

FOSSILS, RHYTHMITES, AND RADIOCARBON DATES

Fossils

The Lake Albany sediments are relatively fossil-poor. Ries (1890) mentioned finding a fragment of a diatom or desmid in the Hudson Valley clay. Peet (1904) described finds in clays of sponge spicules, fresh water diatoms, and insect tracks at Croton and leaves of Vaccinia oxycoccus at Albany. Hartnagel and Bishop (1922) noted several tantalizing fossil sites, including the Cohoes mastodon, found in a pothole 100 feet above the Mohawk river at Cohoes. They also mention a find at lock 6 of the Champlain Canal, near Fort Ann, where cones of Canadensis and spagnum moss were unearthed. Woodworth (1905) cites Asa Fitch's notes that an ash, beechnut, and butternut-rich bed was encountered beneath (fluvial?) sand in the Wood Creek Valley (between the Champlain and Hudson Lowlands) during the construction of the Champlain Canal. Fitch was very impressed by the fact that the trees were all tipped over, with their tops pointing south, opposite the present-day flow of Wood Creek. A Rangifer arcticus (caribou) antler was found in a gravel pit in the Schenectady Delta (Fisher and Ostrom, 1952).

Several fossils have been found recently. In 1977, Tom Engel (NYS Department of Environmental Conservation) found fossiliferous rhythmites in the excavation for the sewer treatment plant at Glenville Rest Home near Schenectady, NY. The fossils were mollusks that were in clay infilling the Ballston channel to an elevation of 240 ft (73.2 m). Eileen Jokinen described the fossils in Dineen and others (1988). She noted that the fossils suggested that Lake Albany contained vegetation and fish, at least in its later stages. Gail Ashley (personal communication, 1986) was able to identify Lebensporen (insect larva tracks) on silt laminae surfaces at the Powell-Minnock Claypit (stop 5 in Dineen and others, 1988).

In 1989, Laurie Williams of EHC sampled peaty sand in wells installed in dunes in a housing development east of Kendrick Hill, in the Town of Northumberland, Saratoga County. The sand was planar laminated, and disconformably overlaid rhythmically-bedded silt and clay. The peaty sand was overlain by at least 5 feet of planar laminated fine to medium sand. Cross-laminated fine sand blanketed the lower sands. The elevation of the ground surface was 317 feet, the silt and clay contact was at 294 feet. The stratigraphic relationships suggest that the sand was deposited in Lake Quaker Springs.

Rhythmites and other Varves

The Deep-water Facies is a series of silt-clay couplets 0.4 to 6 inches (1 to 15 cm) thick (Dineen and others, 1983). The couplet's base is usually on an eroded surface, and consists of planar to ripple-laminated silt or sandy silt that grades up to clay (Dineen,

1977). The percentage of silt systematically decreases upward (Dineen, 1977). The silt-rich couplets are thicker close to deltas or the ice front. Dineen (1977) measured a section with 600 rhythmites near Albany. The average rhythmite thickness was 0.5 to 2 inches (1.3 to 5 cm) in this sections. Extrapolating this average thickness, the section encountered in the deepest part of the Colonie Channel (Bore Hole 5 in Dineen and others, 1983) contained 1,450 to 5,800 couplets. Engineers, hydrologists, and the early mappers identified the couplets as "varves." LaFleur (1965a) did not find compelling evidence that they were annual accumulations of sediment, and suggested that the more-neutral term "rhythmites" be used.

<u>Radiocarbon Dates</u>
Table 3 summarizes 21 radiocarbon dates from the Hudson Valley. There is some scatter in the dates. The extrapolation from the age of the base-of-organic-rich sediment to the age of base-of-the-bog is fraught with peril, since sediment accumulation rates would have to be determined for each bog to calibrate the ages (Dineen, 1986a). The bogs were at different altitudes, had differing drainage basin areas and slopes, varied in their relationships to sources of aeolian sand and silt, colluvium, and stream sediment, and varied greatly in latitude. One date is apparently related directly to Lake Quaker Springs. It is from the well field in Northumberton (see above). Its bulk sample age was 11,050 yBP (GX-14348, #15 on Table 3) with an accelerator date (#10 in Table 3) on a single spruce needle of 11,770 yBP (Norton Miller, NYS Biological Survey, personal communication).

The systematically younger dates from south to north suggests that the Lake Albany stage lasted from 17,000+ to 11,800 yBP. The lake stages from Quaker Springs through Fort Ann lasted from 11,800 yBP to the inception of the Champlain Sea episode. Date Number 18 on Table 3 suggests that the Champlain Sea episode began prior to 10,700 yBP. Clark and Karrow (1984) favored a date of 12,100 yBP for Lake Iroquois, and 11,620 yBP, based on a pool of seven radiocarbon dates on marine shells in the St. Lawrence Valley, for the Champlain Sea. Muller and Prest (1985) favored a date between 11,800 and 10,630 yBP (± 11,200 yBP) for the Champlain Sea.

ICE MARGINS AND THE QUESTION OF READVANCES

Recent mapping has revealed a number of ice margin positions in the Hudson Lowlands (LaFleur, 1965a and b; Connally and Sirkin, 1967, 1970, 1971, 1986; Hanson, 1977; Dineen and others, 1983; DeSimone, 1985; DeSimone and LaFleur, 1985, 1986; Dineen, 1986a; Dineen and Duskin, 1987; Dineen and others, 1988; Dineen, 1992). These ice margins are marked by kame or till moraines, kame deltas, heads-of-outwash, and the chronologies of tributary lake basins.

While we enjoy some unanimity on the <u>locations</u> of the ice margins, we disagree on their <u>meaning</u>. Some workers have suggested that the

TABLE 3 -				RADIO-CARBON DATES Part 1
AGE IN C-14 YEARS	RANGE	LAB NUMBER	REFERENCE	LOCATION AND NOTES
1) 19,875	± 980	GX-8672	Dineen, 1986	Bear Swamp, base of bog, silt with trace organics, T-Pollen Zone
2) 16,650	± 660	GX-8488	Dineen, 1986	D'Hommeadeau Bog, Meadowdale Moraine, base of bog, silt with trace organics
3) 13,670	± 170	SI-4042	Connally and Sirkin, 1986	Eagle Hill Camp Bog, Red Hook III Moraine, pooled sample from 5 cores, A1 Spruce Pollen Zone
4) 13,260 13,210	± 125 ± 550	SI-6030 GX-9435	Connally, personal communication	Poughkeepsie IBM, dates organic material in folded clays under trough-cross-bedded gravel, from bogs along the shore of Lake Albany?
5) 13,150	± 200	I-4986	Connally and Sirkin, 1971	Pine Log Camp Bog, outwash train heading @ Hidden Valley Moraine, pool of 12 samples from base of bog
6) 12,850	± 250	L-1157a	Connally and Sirkin, 1970	New Hampton Bog No. 1, base of organics, retreat from Wallkill Moraine
7) 12,510	± 370	I-4137	Steadman and Funk, 1988	Orange County, NY, Dutchess Quarry Cave, caribou bone collagen, caribou in Wallkill Lowlands in Lake Albany time
8) 12,400	± 200	I-3199	Connally and Sirkin, 1971	Pine Log Camp Bog, outwash train of Hidden Valley Moraine, fragment of spruce log from base of bog

TABLE 3				RADIO-CARBON DATES - Part 2
AGE IN C-14 YEARS	RANGE	LAB NUMBER	REFERENCE	LOCATION AND NOTES
9) 12,150 12,870	± 210 ± 200	QC-297 QC-296	Averill and others, 1980	Oradell, NJ, peat overlying mastodon in erosional channel
10) 11,770	± 115	accel.	Dineen et al, in prep.	Schuylerville, spruce needle in moss, @ contact between varved clay and sand (Lake Albany-Quaker Springs transition)
11) 11,800	± 290	WAT-1190	Pair et al, 1988	Ogdensburg, marine shells in Champlain Sea clay
12) 11,590 12E) 14,100	± 255	QC-514	Dineen, 1986	Bear Swamp, base of organic silt, Spruce Pollen Zone; extrapolated age of 13,152 to 15,060 yr BP
13) 11,380	± 380	WAT-1301	Pair et al, 1988	Ogdensburg, marine shells in Champlain Sea clay
14) 11,280	± -	-	Steadman and Miller, personal communication	Elizabethtown, muskox collagen from glacial lake silt and clay
15) 11,050	± 450	GX-1434*	Dineen et al, in prep.	Schuylerville, bulk sample from laminated silty sand with moss, @ contact between varved clay and sand (Lake Albany-Quaker Springs transition)

TABLE 3				RADIO-CARBON DATES - Part 3
AGE IN C-14 YEARS	RANGE	LAB NUMBER	REFERENCE	LOCATION AND NOTES
16) 10,850 16E) 13,250	± 325	GX-8487 -	Dineen, 1986	D'Hommeadeau Bog, Meadowdale Moraine, base of organic silt, Spruce Pollen Zone; extrapolated age is 12,485 to 14,030 yr BP
17) 10,630	± 249	-	Muller and Prest, 1985	Salmon River, Malone, marine shells in Champlain Sea clay
18) 10,450	± 100	-	Steadman and Miller, Pers. communication	Norfolk, Beluga whale bone collagen (Champlain Sea)
19) 10,310 10,505	± 85 ± 100	SI-6016 SI-6016a	Connally, Pers. communication	Poughkeepsie IBM, peat folded into clay under sand and gravel
20) 10,280	± 270	J-5200	Weis, 1974	West 50th St., NYC, base of organic silt, unconformity on varved clay
21) 9,170 21E) 15,300	± 220	-	Ibe and Pardi, 1985	Hearts Content Bog, Catskill, base of organic silt; extrapolated date 13,170 to 17,375

margins were deposited by advancing ice (Connally and Sirkin, 1970, 1971, 1986; Dineen and others, 1983; Dineen, 1986a) while others have insisted that they are the product of systematically retreating ice (Duskin, 1985; DeSimone and LaFleur, 1985, 1986). The available surface and subsurface data is ambiguous (DeSimone and LaFleur, 1985, 1986; Duskin, 1985; Dineen and Duskin, 1987), but suggests that the ice margin oscillated. The evidence for large-scale readvances (on the order of tens of miles) should include widespread till sheets over folded or thrusted lacustrine or fluvial deposits, rejuvenation of outwash and tributary lake systems, and multiple tills. Local exposures of contorted clays or of till-over-stratified drift abound (Connally and Sirkin, 1967, 1970; Hanson, 1977; Dineen and Rogers, 1979; Dineen and others, 1983; Dineen, 1986; Dineen and Duskin, 1987), but continuous till sheets or other mappable features have not been found in the Hudson Lowlands north of Kingston. Dineen (1986) and Connally (1992) have suggested that the glacial deposits in the Wallkill Valley and on the dipslopes of uplands were deposited by surging ice that then stagnated in place.

Two areas do seem to have experienced readvances, however. Till occurs over compacted "varved" clay in the test borings for the Newburgh-Beacon Bridge and highly compacted gravel underlies uncompacted clay at the Mid-Hudson Poughkeepsie Bridge (Dineen and Duskin, 1987). Till-over-clay or -outwash has been encountered in water wells and test borings in the Casper Creek (Dineen and others, 1988) and Sprout Creek Valleys (Moore and LaFleur, 1982). Till-over-clay and -outwash has also been observed in the Wallkill Valley (Russell Waines, personal communication). These deposits might have been emplaced by a readvance from Poughkeepsie to the north flank of the Hudson Highlands (the Shenandoah Moraine of Connally and Sirkin, 1986) or could be pre-Woodfordian deposits that were overridden by the Woodfordian glacier.

The second area of till-over-something is in the Helderberg Plateau, where till overlies gravel or clay throughout the Cats Kill Valley (Dineen, 1986a). The till was probably deposited by the Middleburgh Readvance (LaFleur, 1969b) and are related to the Rosendale-Red Hook readvance (Connally and Sirkin, 1971, 1986).

EASTERN MOHAWK VALLEY

The eastern Mohawk Valley contains a record of the interactions between the Adirondack, Sacandaga, and Mohawk sublobes. The valley also records episodes of meltwater drainage from the western Mohawk Valley. The glacial deposits form thick, complex fills in a deeply entrenched preglacial drainage system (Dineen and Hanson, 1985).

The preglacial Sacandaga River drained the south-central Adirondacks. It flowed through the Sacandaga graben and entered the preglacial Mohawk River near Fonda, NY. The preglacial Luzerne River drained the eastern Adirondacks and entered the preglacial

Colonie Channel near Saratoga Springs. This stream network formed a rectangular drainage pattern along the bases of the tilted fault blocks. Differential weathering and erosion sculpted a series of east-facing escarpments along the strike-slopes of the faults. The preglacial Mohawk River cut across the rock structures and entered Colonie Channel south of Albany.

Chamberlain (1883) was one of the first geologists to realize that the Mohawk Valley was the key to correlating the glacial events between the Ontario and Hudson Lowlands. He and Brigham (1898) noted evidence for a west-flowing glacial lobe in the valley. Fairchild (1912, 1917) interpreted terraces and sandplains in the valley as part of a continuous glacial lake that "girded" the Adirondacks, a notion that was demolished by Stoller (1916), Miller (1923, 1925), and Brigham (1929, 1931). Stoller (1916) , Miller (1923, 1925), Chadwick (1928), and Brigham (1929) mapped evidence for proglacial lakes in the Sacandaga Valley and southern Adirondacks. Thirty years passed before LaFleur (1961, 1965, 1969) rekindled interest in the eastern Mohawk Valley.

Striae and drumlins were used to determine the directions of glacial movement (Yatsevitch, 1968; LaFleur, 1983; Dineen and Hanson, 1985). Several ice streams ice can be deduced from the movement indicators. The Hudson Lobe moved south, down the Hudson Lowlands. The Kayderosseras sublobe was a tributary to the Hudson Lobe that flowed down the upper Hudson-Schroon River lowlands. The Kayderosseras sublobe veered to the southwest along the face of the McGregor and Spruce Mountain ranges. The Sacandaga sublobe flowed southwest, down the Conklinville fork of the preglacial Sacandaga River. The Northville sublobe flowed south, out of the south-central Adirondacks. The Sacandaga and Northville sublobes were part of the Adirondack Lobe, and deposited silty sand-matrix diamicton with abundant boulders of metamorphic rocks. The Mohawk sublobe flowed west, up the Mohawk Valley, and deposited silty clay-matrix diamicton with many graywacke, shale, and dolostone boulders.

STRATIGRAPHY

The following stratigraphy (Table 4) is based on the work of Yatsevitch (1968) and LaFleur (1979, 1983, 1992). The Hell Hollow Till is a clay-rich diamicton with clast lithologies that were derived from the Mohawk Valley. It was deposited by westward flowing ice. The Hell Hollow (and associated proglacial lake beds and gravels) are weathered and disconformably overlain by the Little Falls gravels - a sequence of coarse cobble gravels with lithologies that were derived from the Ontario Basin. The Little Falls gravels are overlain by Mohawk Till- a fresh, silty clay-matrix diamicton with abundant eastern Mohawk stones. Locally, the till contains well-rounded cobbles that are reworked Little Falls gravel. The Mohawk Till is bedded towards its top, and grades into rhythmically bedded lake silt and clay of Lake Amsterdam. The Lake

Amsterdam beds are disconformably overlain by gravelly sand of the Fonda Sand Plain. This sequence is deeply entrenched, and overlain by the cobbly Scotia Gravel.

TABLE 4: MOHAWK VALLEY STRATIGRAPHY

Scotia Gravels
Fonda Sand Plain
Lake Amsterdam
Mohawk Till
Little Falls Gravel
Hell Hollow Till

The Mohawk Till is part of a complex of deposits that make up the Broadalbin Interlobate Moraine (Brigham, 1929; Yatsevitch, 1968; Dineen and Hanson, 1985). The Broadalbin Interlobate Moraine (BIM) is an extensive ridge that extends 15 miles (23 km) from Gloversville to Broadalbin and Perth. It was deposited between the Sacandaga and Mohawk sublobes. The BIM is 100 to 300 ft (30 to 100 m) thick and 0.25 to 4 miles (0.5 to 6 km) wide and is comprised of trough-bedded gravelly sand with many interbeds of diamicton. It contains abundant laminated sands to the west, and gravel beds to the east. Current-indicators indicate that it was deposited by meltwater flowing from east to west. Drumlinized till suggest that the BIM was deposited, in part, by the Yost readvance- a surge of the Mohawk sublobe that deposited the Mohawk Till (Yatsevitch, 1968). Glacial Lakes Amsterdam and Sacandaga occupied the lowland areas as the Yosts ice retreated (Dineen and Hanson, 1985). Extensive sandy deltas and sand and clay terraces remain from these lakes. Lake Sacandaga ended when the Luzerne sublobe retreated north of Conklinville (Dineen and Hanson, 1985). Lake Amsterdam ended when the Hudson Lobe retreated north of the West Hill-Fort Hunter outlet at Schenectady (LaFleur, 1983).

REGIONAL STRATIGRAPHY

The stratigraphy of eastern New York is summarized on Table 5. The 21 carbon-14 dates on Table 3 help to control the chrono-stratigraphy depicted on Table 5. Three readvances occurred within the dated interval, they were the Prattsville-Shenandoah, Middleburgh-Whitfield-Rosendale, and Yost-Delmar readvances. They took place during the Nissouri (?), Port Bruce, and Port Huron Stades. The Middleburgh and Yost readvances post-date a major recession of the glacier during the Erie Interstade, when the Mohawk Valley was temporally freed of ice and glacial sediments in the Hudson Gorge were deeply eroded.

Figures 10 through 19 depict ten phases in the history of Lake Albany. Many other phases can be considered using the ice margins determined by Dineen (1986), Connally and Sirkin (1973, 1986), and DeSimone and LaFleur (1986), and the lake stages depicted in Figure 2 and DeSimone and LaFleur (1986). Table 5 presents tentative correlations with the western Mohawk, eastern Mohawk, and Hudson

TABLE 5				Page 1
CARBON- 14 AGE/ STRAT	**WESTERN MOHAWK VALLEY**	**EASTERN MOHAWK VALLEY**	**HUDSON CHAMPLAIN VALLEYS**	**ADIRONDACK TACONICS**
10,000	Free Drainage	Free Drainage	Hudson Estuary (20) deep incision Lake Fort Ann III (19) Lake Fort Ann II	post- glacial uplift Champlain Sea (11; 13; 17; 18) Covey Hill Channels Chateaugay Channels
11,000	**(Ninemile Readvance) ROME TILL**	Scotia Gvl	Lake Fort Ann I Lake Coveville Lake	Elizabeth- town IM (14) Euba Mills IM L. Corinth N. Creek- Chester- town- Riverbank IM
12,000 TWO CREEKS INTER- STADE PORT HURON STADE	Hyper Iroquois Lake Amsterdam	Fonda Gravel Lake Sacandaga	Quaker Springs (9; 15) Moreau Pond IM Lake Albany II Niskayuna IM	Lake Warrens- burg Hidden Valley IM (5; 8) Randall Corners IM Conklin- ville IM
13,000 MACKINAW INTER- STADE	**(Little Falls Readvance) HOLLAND PATENT TILL**	Broadalbin ILM-Lake Amsterdam **(Yost Readvance) MOHAWK TILL**	Meadowdale IM (16E) Vly IM **(Delmar Readvance)** Bell Pond IM	Pine Bowl IM L. Bascom upper till @ Spiers Falls Lake Kinderhook

TABLE 5				Page 2
CARBON-14 AGE/ STRAT	**WESTERN MOHAWK VALLEY**	**EASTERN MOHAWK VALLEY**	**HUDSON VALLEY**	**ADIRONDACK TACONICS**
14,000	Lake Gravesville Lake Miller	800-ft Lake Schoharie 1100-ft Lake Schoharie Franklinton Ch.	Lake Albany Alcove-Red Hook III IM	Lake clays @ Spiers Falls
15,000 PORT BRUCE STADE	**(St. Johnsville Readvance) NORWAY TILL** **(Salisbury Readvance) HAWTHORNE TILL**	**(Pratts-ville Readvance) MOHAWK TILL (Tanners-ville Readvance)** rounded-cobble till @ Randalls Wallkill IM (6)	Whitfield-Red Hook I IM (3; 12E; 21E) **(Rosendale Readvance)** Lake Albany upper till @ Casper Creek Shenandoah Moraine Lakes New York and Hackensack	lower till @ Spiers Falls Pine Plains IM Lake Attlebury drumlin till @ Fishkill Creek
16,000 ERIE INTERSTADE NISSOURI STADE 17,000 +	Little Falls Gvl Shed Brook Discon-tinuity **WEST CANADA TILL WHITE CREEK TILL**	Little Falls Gvl Amsterdam Discon-tinuity **HELL HOLLOW TILL**	weath. zone @ Niskayuna unconform-ities in Wallkill Valley lower till @ Casper Creek	lower gravel @ Fishkill Creek

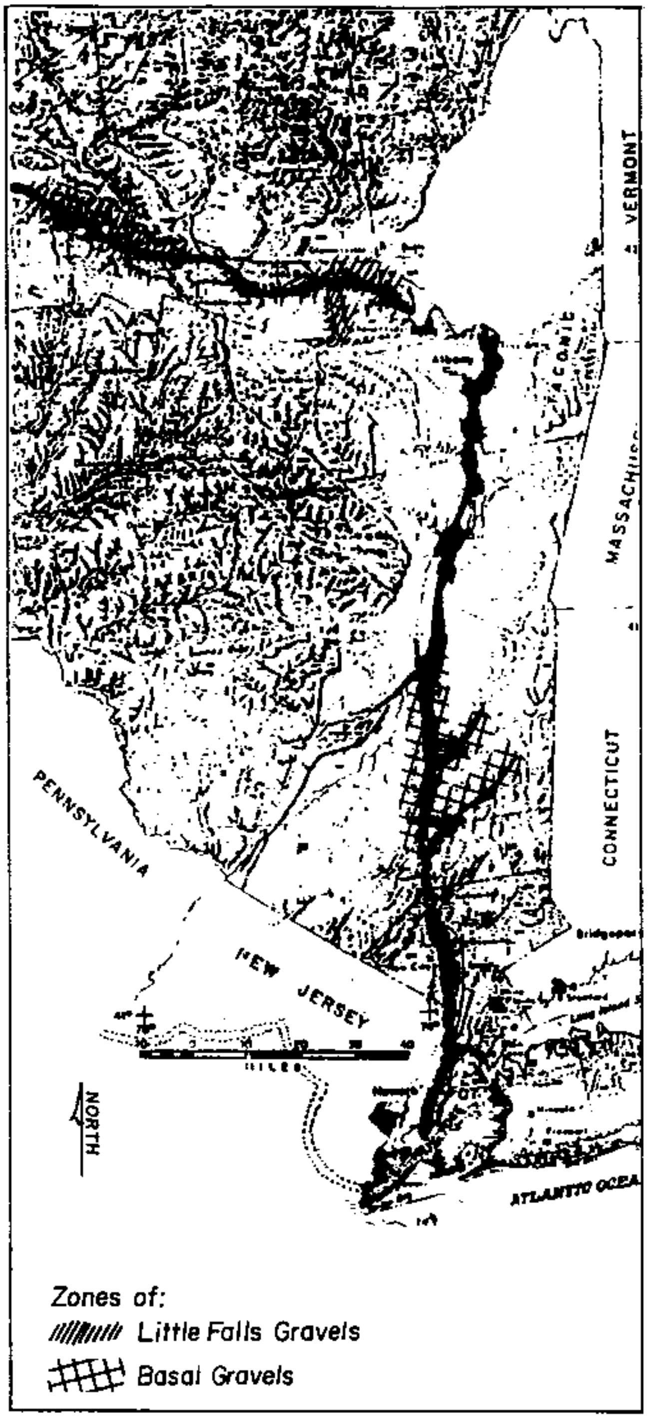

Figure 10.
Erie Interstade

LMD - 5/15/92

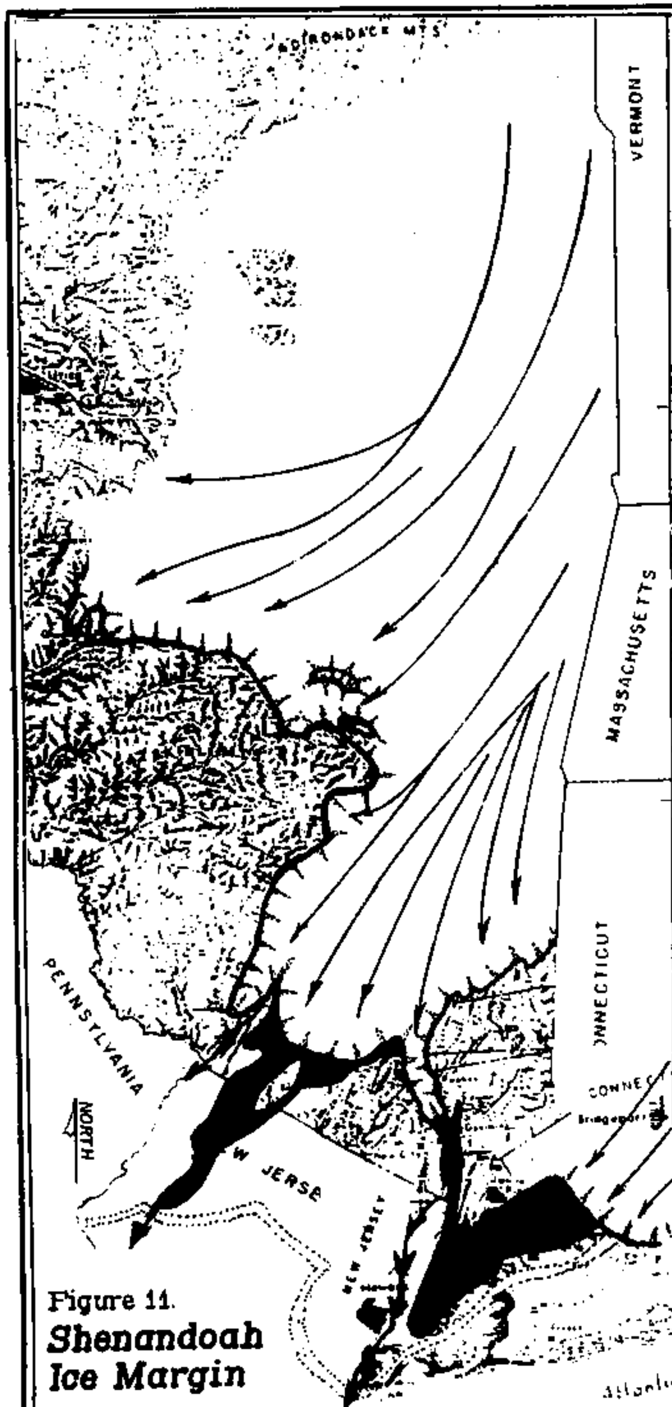

Figure 11.
Shenandoah Ice Margin

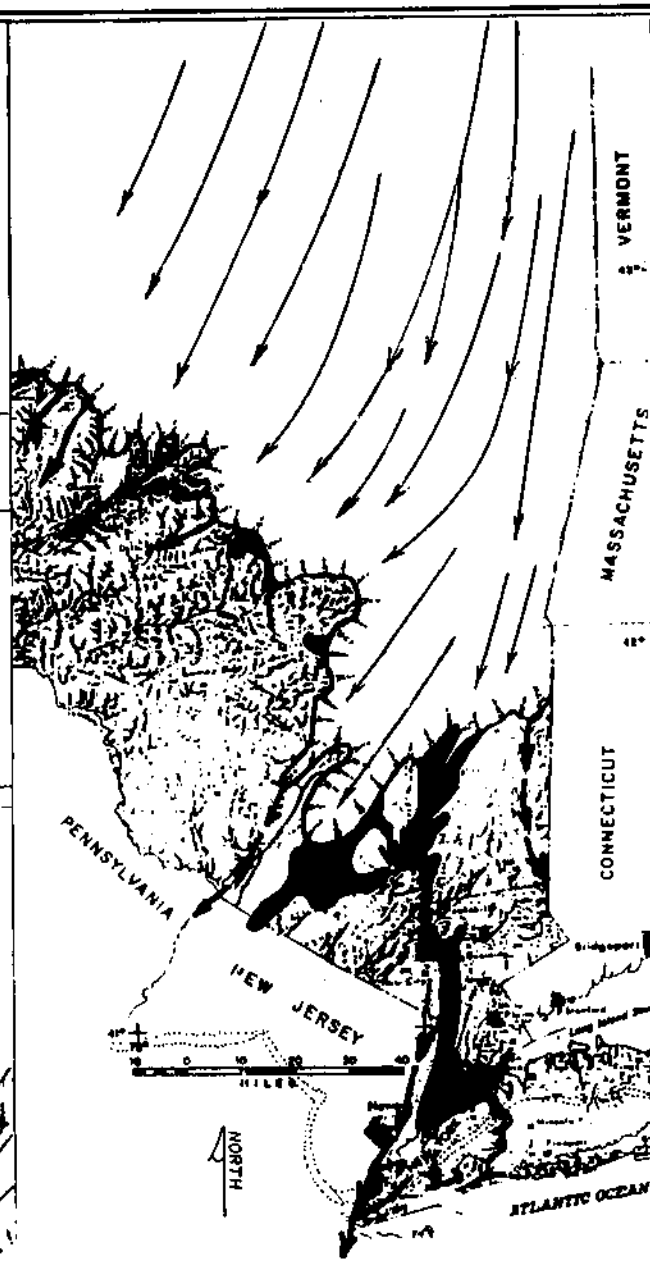

Figure 12.
Wagon Wheel Gap Ice Margin

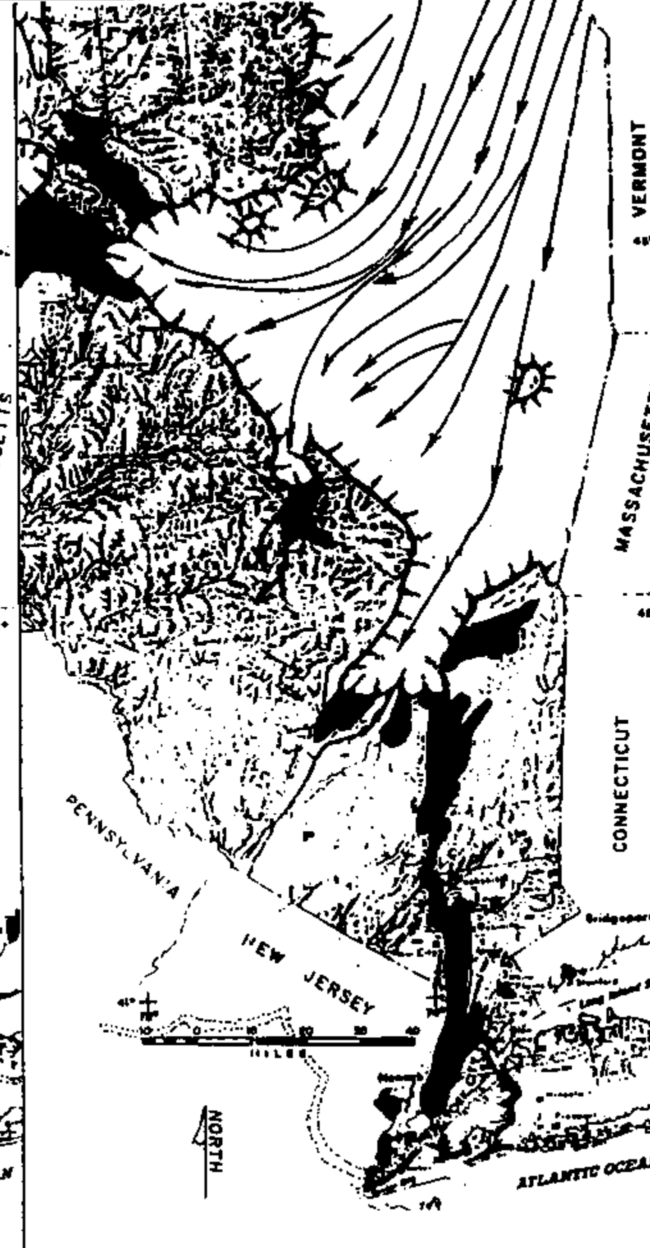

Figure 13.
Prattsville - Rosendale Readvance

LMD-5/15/92

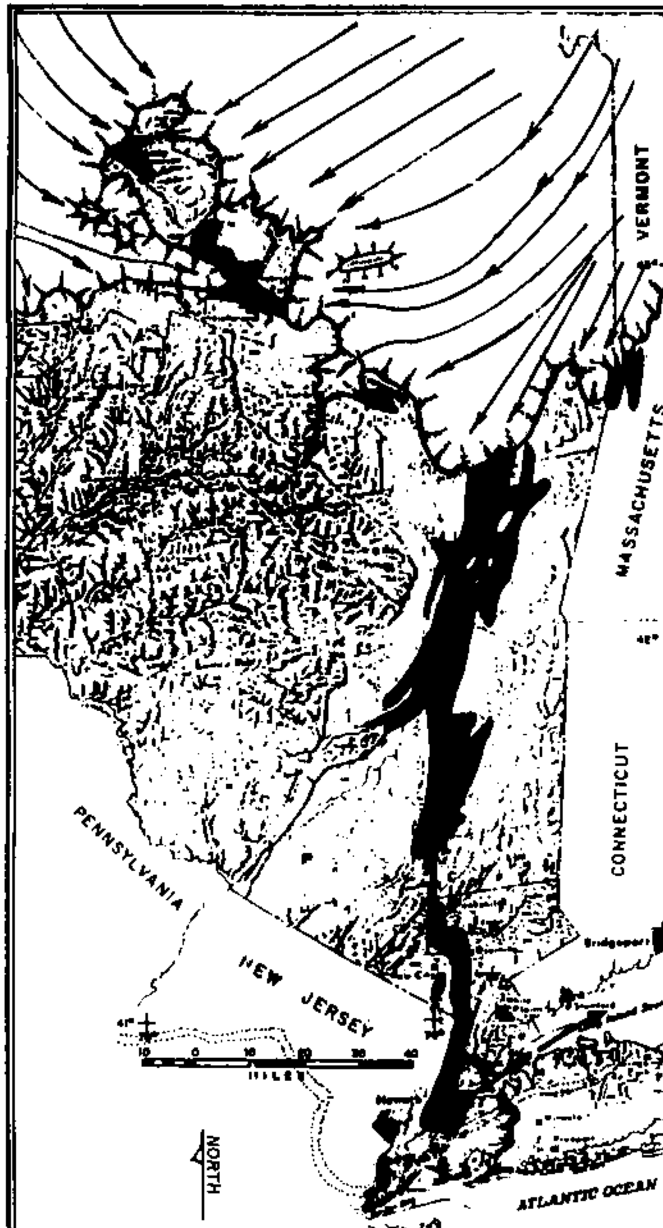

Figure 14.
Yost Readvance

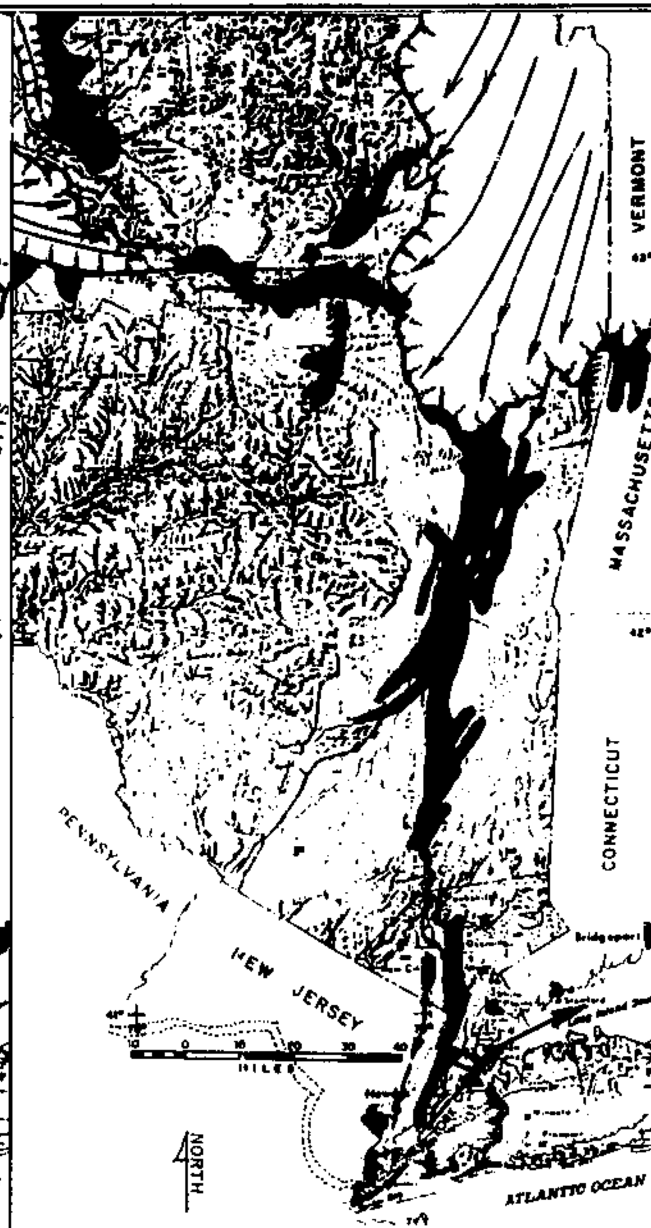

Figure 15.
Lakes Amsterdam and Albany

Figure 16.
Maximum Lake Albany

LMD–5/15/92

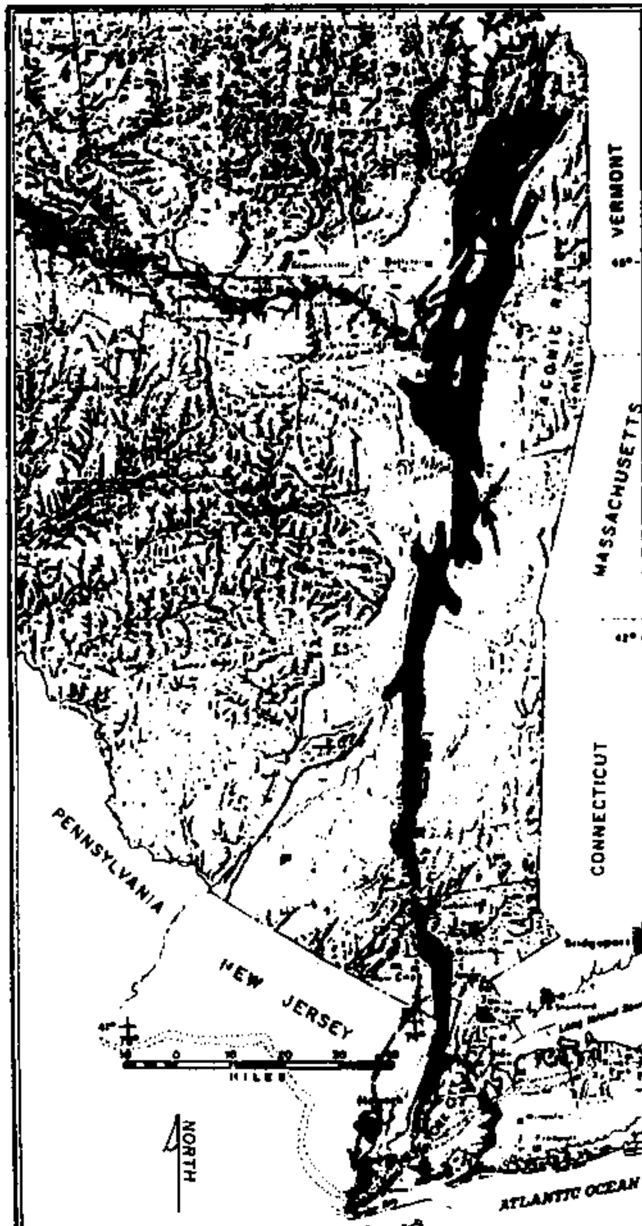

Figure 17.
Lake Quaker Springs

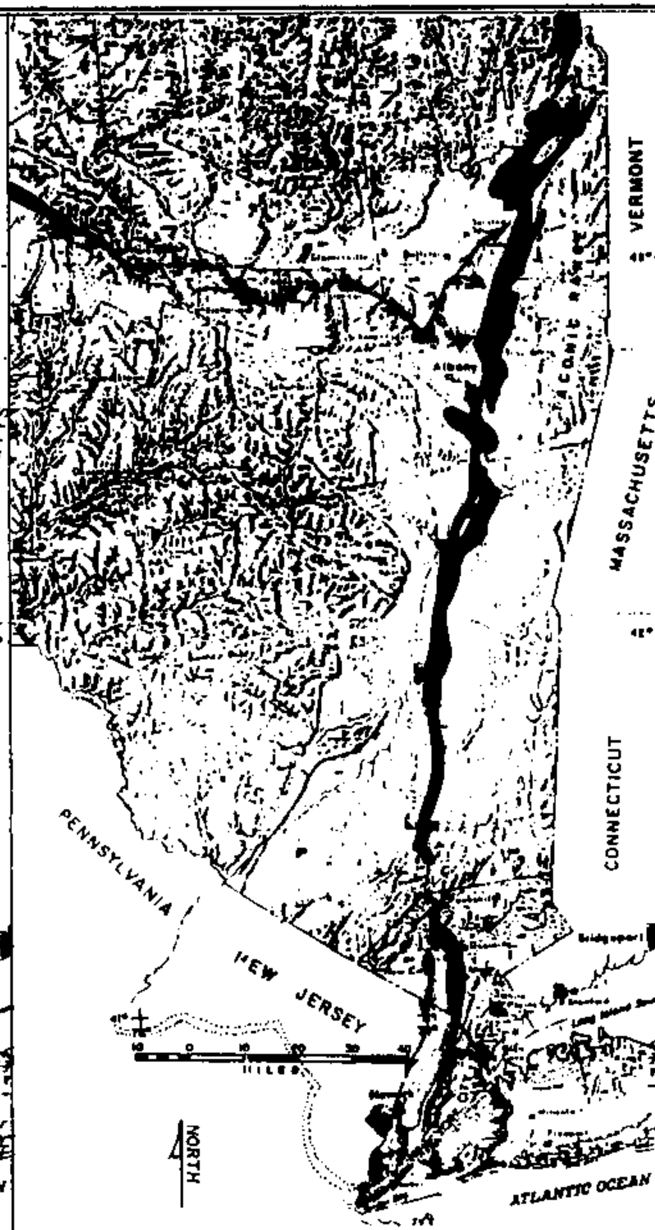

Figure 18.
Lake Coveville

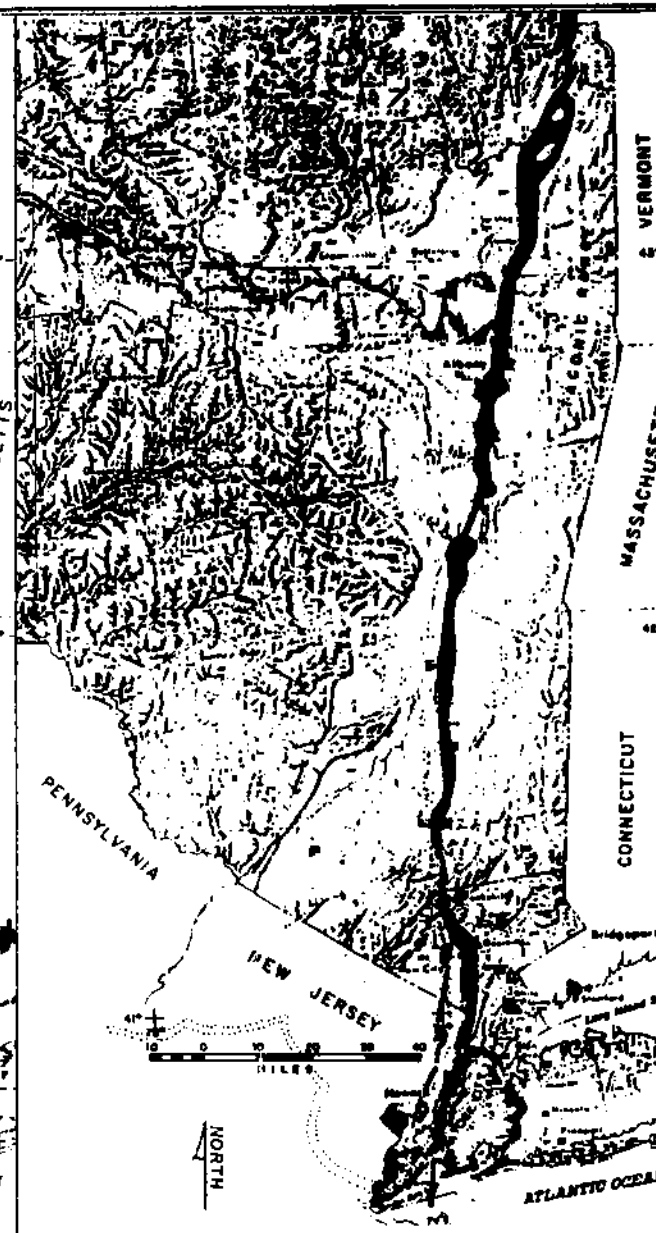

Figure 19.
Lake Fort Ann

LMD-5/15/92

Lowlands.

Pre-Erie Stade: The earliest glacial deposits on the Hudson Lowlands are best preserved in the tributary valleys. Multiple till exposures and borings are scattered throughout eastern New York. They include the West Canada Creek and White Creek Tills in the western Mohawk Valley (Ridge and others, 1990). The Hell Hollow Till occurs throughout the eastern Mohawk and Schoharie Valleys (LaFleur 1969, 1983). The older tills were also observed in the Rome to Utica area (Casey and Reynolds, 1989a, 1989b, and 1989c), in the southwestern foothills of the Catskills (T. Fluhr archives, NYSGS Open Files), and in the southern Adirondacks at Spiers Falls (Hanson and others, 1961; Connally and Sirkin, 1971), plus Hadley, Stoney Creek, and Broadalbin (Dineen and Hanson, 1985). A set of subglacial or ice-marginal meltwater channels were carved at Duanesburg during this retreat. These channels were subsequently filled with outwash, then buried with till during another readvance (Dineen, 1985, notes on the Duanesburg 7-1/2 minute compilation quadrangle in the NYSGS Open Files).

Ogdensburg-Culvers Gap-Sparkill Moraines, Glacial Lakes Wallkill, Hackensack, and Hudson-Flushing: This early phase is the retreat from the Terminal Moraine and the development of Lake Hackensack, with concurrent formation of the 600-foot (180 m) stage of Glacial Lake Wallkill, the Ogdensburg-Culvers Gap Moraine, Lakes Hudson and Flushing, and the Sparkill Moraine. The Connecticut Lobe blocked Long Island Sound as the Hudson Lobe retreated. This retreat began approximately 22,000 years ago (Sirkin, 1977, 1986).

As the ice front retreated, a readvance took place that involved the Hudson Lobe and its branch, the Mohawk sublobe. The Mohawk sublobe advanced to the west, depositing the Hell Hollow Till in the eastern Mohawk Valley (LaFleur, 1983) and the West Canada Diamicton in the western Mohawk Valley (Ridge, 1991).

FIGURE 10: Erie Interstade: The lower tills were deeply incised during the Erie Interstade. The Shed Brook discontinuity was formed in the western Mohawk Valley at this time (Ridge and others, 1990). The Shed Brook discontinuity was incised to 350 ft ASL at Rome (Casey and Reynolds, 1989a). It corresponds to a discontinuity on the Hell Hollow Till at Amsterdam and Scotia (LaFleur, 1983). The elevation of the discontinuity at Scotia is 190 ft (LaFleur, 1983).

Either the Shed Brook discontinuity, or an earlier erosion surface, has been found at 150 ft ASL at Utica (Casey and Reynolds, 1989c), and below sea level in the Hudson gorge, particularly in the Lower Hudson Sheet at the Kingston-Rhinecliff, Mid-Hudson Poughkeepsie, and Newburgh-Beacon bridges (NYSGS Open Files; Dineen, 1987; Dineen and Duskin, 1987). It might be present at Casper Creek (Dineen and others, 1988) and Wallkill Creek (Russell Waines, personal communication; Kopsick, 1974).

The Shed Brook discontinuity is overlain by the Little Falls gravel in the Mohawk Valley (LaFleur, 1983; Ridge and others, 1990). The elevation of the Little Falls gravel is 400 ft @ the confluence of the East Canada Creek and the Mohawk River (Ridge, 1991), 420 ft @ Little Falls (LaFleur, 1983), and 315 ft @ Tribes Hill (LaFleur, 1983).

Similar gravels are present in the lower Hudson Valley at Casper Creek (Dineen and others, 1988) and Fishkill Creek (Moore and LaFleur, 1982). The discontinuity in the Hudson gorge is overlain by coarse gravel at the Kingston-Rhinecliff, Mid-Hudson Poughkeepsie, and Newburgh-Beacon bridges (Dineen, 1987; Dineen and Duskin, 1987).

The elevations of the Shed Brook discontinuity and the Little Falls gravel require that the Hudson Lowlands be free of ice as far north as Schenectady, NY. The wide variation of the elevations of the Shed Brook discontinuity and the Little Falls Gravel is puzzling, however. More data is necessary to determine whether the surfaces are "true" representations of Middle Wisconsinan base levels and depositional surfaces, whether more than one erosion surfaces are involved, or whether the surface has been altered by folding and glacial/subglacial meltwater gouging. In any event, the Hudson Valley was (live) ice free to the latitude of Schenectady, NY during the Erie Interstade.

If the discontinuities in the lower Hudson Valley are correlative with the Shed Brook discontinuity, the Terminal Moraine had been breached during the Erie Interstade and subsequent Lake Albany was <u>not</u> dammed by the Terminal Moraine.

The period of time subsequent to the Erie Interstade was dominated by oscillatory glacial retreat.

FIGURE 11: Cuddebackville-New Hampton-Shenandoah Moraines, Glacial Lakes Wallkill and Hudson-Flushing: The Shenandoah Moraine was deposited against the Hudson Highlands during this readvance while the High Peaks of the Catskills rose as a nunatak above the resurgent Hudson Champlain Lobe. The Hudson Lobe deposited the Cuddlebackville in the Neversink valley and the New Hampton moraine in the Wallkill lowlands. Lakes Wallkill and Hudson-Flushing were trapped between the ice front and the Terminal Moraine. Lake Hackensack drained due to rapid isostatic uplift (perhaps a peripheral bulge), and Lake Hudson began to overflow through Sparkill Gap. The overflow eroded the emergent Hackensack clays. Many readvance features were deposited between Poughkeepsie and the Highlands, including drumlinized till over gravels at Fishkill and Casper Creeks (Moore and LaFleur, 1982; Dineen and others, 1988) and the Croton Delta (at Peekskill).

FIGURE 12: Wagon Wheel Gap-Phillipsport- Wallkill-Hyde Park Moraines, Glacial Lakes Wallkill and Albany (WAGON WHEEL GAP

MARGIN): The Hudson Lobe readvanced again and established a major ice margin position at the Wagon Wheel Gap-Phillipsport-Wallkill-Hyde Park. The moraines of this position are well developed, implying a relatively lengthy stillstand of the ice. The Tannersville(?) margin was established in the High Peaks of the Catskills, while the Salisbury-Mohawk readvance of the Mohawk sublobe deposited the Hawthorne Diamicton in the western Mohawk Valley. Lake Cedarville formed in the Mohawk Valley (Ridge, 1991). Lake Wallkill fell to its 400-foot (120 m) level and spilled eastward through the Moodna Creek Valley late in this phase, while Lake Hudson-Albany expanded into the mid-Hudson Valley.

FIGURE 13: Prattsville-Whitfield-Cottekill-Ulster Park-Red Hook Moraines, Glacial Lakes Grand Gorge-Schoharie, Warwarsing, Tillson, Albany, and Attlebury (PRATTSVILLE-ROSENDALE READVANCE): Lakes Schoharie, Warwarsing, Tillson, and Attlebury formed as the Middleburgh ice surged again. All the lakes (except the earliest stage of Lake Schoharie) spilled into Lake Albany. The ice retreated from this margin about 16,000 years ago (Tables 3 and 5). This event correlates with the St. Johnsbury Readvance of the Ontario Lobe into the western Mohawk Valley and the deposition of the Norway Diamicton (Muller and others, 1986; Ridge, 1990). Lake Cedarville filled the western Mohawk Valley while Lake Miller formed in the Black River Valley.

The receding Rosendale ice hesitated, leaving recessional moraines behind. Red Hook III on the east side of the Hudson gorge (Connally and Sirkin, 1986), and the Woodstock, Schoharie, and Alcove-Bell Pond recessional ice margins on the west side of the gorge (Dineen, 1986) formed and were abandoned. These retreatal ice margins can be traced using lake outlets, kame complexes, and discontinuous moraines (Dineen, 1986). The Red Hook III-Alcove ice margins can be dated using pooled samples from basal kettle-hole silt (Connally and Sirkin, 1986: C-14 date #3, Table 3) and extrapolated bog-bottom dates (Dineen, 1986: C-14 date #12E; Ibe and Pardi, 1985: C-14 date #21E).

FIGURE 14: Salisbury-Yost-Vly margins, Lake Albany (YOST-DELMAR READVANCE): The Hudson Lobe readvanced in the Mohawk Valley to the Yost Ice Margin, depositing the Yost Till (LaFleur, 1969; Dineen and others, 1988) and to the Vly margin in the Hudson Valley (Dineen and others, 1988). Lake Kinderhook formed in the upper Kinderhook Valley as the ice margin retreated from eastern Hudson Lowlands (DeSimone, 1989). The Holland Patent Till was deposited in the western Mohawk Valley by the Little Falls Readvance of the Ontario Lobe (Ridge and others 1990). Lake Sacandaga formed in the Sacandaga Basin and Lake Gravesville formed in the western Mohawk Valley (LaFleur, 1965; Ridge, 1991). The 1100 ft stage of Glacial Lake Schoharie drained into the Cats Kill Valley through the Franklinton notch. The Broadalbin interlobate moraine was deposited between the Mohawk sublobe of the Hudson Lobe and the Sacandaga and Schroon Lobes of the Adirondacks as the Yost ice

retreated (Dineen and others, 1985).

Lakes Schoharie, Amsterdam, Albany, Tomhannock and Bascom: The Hudson Lobe had retreated to the Spruce Mountain-Pownal ice margin by 15,000 years ago. Lake Schoharie fell through a series of water levels as the ice margin retreated in the Mohawk Valley (LaFleur, 1969). When the Mohawk Lobe retreated from the Mohawk Valley and exposed the Duanesburg-Bozen Kill outlets, Lake Schoharie fell from 800 to 660 ft (LaFleur, 1969) and Lake Amsterdam filled the lower Mohawk Valley. Lake Amsterdam eventually spilled through the Fort Hunter-West Hill Gap, while Lakes Tomhannock and Bascom expanded in the Taconic Mountains section of the New England Upland. Lake Gravesville expanded in the western Mohawk Valley (Ridge, 1991).

FIGURE 15: **Lakes Amsterdam, Albany, and the Meadowdale Moraine:** As the Hudson Lobe continued its retreat, proglacial Lake Amsterdam expanded into the middle to lower Mohawk Valley (LaFleur, 1969, 1983: DeSimone and LaFleur, 1985) and the Meadowdale recessional moraine was deposited in the Hudson Valley (DeSimone and LaFleur, 1986; Dineen, 1986, C-14 date 16E; Dineen and others, 1988). Proglacial lakes formed in the lower Black River Lowlands as the Ontario Lobe and Black River sublobe retreated from the western Mohawk Valley.

FIGURE 16: **Randalls Corners-Glen Lake-Carter-Manchester Moraines, Glacial Lakes Warrensburg and Albany:** As Lake Albany expanded to its maximum size, the glacier retreated into the Champlain Valley and Lake Warrensburg flooded the Upper Hudson River Valley. In the Hudson Lowlands, the ice deposited a series of recessional moraines at Niskayuna through Moreau Pond (Dineen, 1986; DeSimone and LaFleur, 1986; Dineen and others, 1988). The Randalls Corners and Conklinville moraines were deposited in the Adirondacks (Dineen and others, 1985). The Mohawk Valley carried relatively steady meltwater flow from western NY (LaFleur, 1983).

FIGURE 17: **Glacial Lake Vermont-Lake Quaker Springs:** The Hudson Valley became ice-free by 13,100 years ago. The water level of Lake Albany abruptly dropped to the Quaker Springs level (C-14 dates #9; Averill and others, 1980, #10 and #15; DeSimone and LaFleur, 1985, 1986; Dineen and others, 1988) and proglacial Lake Warrensburg formed in the upper Hudson and Schroon valleys (Dineen and others, 1985). Overflow from Glacial Lake Iroquois deposited the Fonda Gravels in the Mohawk Valley (LaFleur, 1983; Muller and Prest, 1985). The Hidden Valley recessional moraine (C-14 dates: #5 and #8 in Connally and Sirkin, 1971; DeSimone and LaFleur, 1986; Dineen and others, 1988) was deposited in the Adirondacks.

The Hudson Valley lake level fell to the Quaker Springs stage at the same time that catastrophic floods from the Appalachian Plateau, Ontario Basin, and the Adirondack Highlands began to flush over the Rome Outlet and through the Mohawk Valley. The floods carved channels between Schenectady and Saratoga Lake, and might

have caused the transition from the Albany to the Quaker Springs stages. The drop in lake level scoured near-shore areas and the lake bottom, and flushed peat balls and mats into the lake.

FIGURE 18: Glacial Lake Vermont-Coveville: Lake Iroquois began to fill the eastern Ontario Basin, and Lake Vermont expanded in the Champlain Valley, as the Hudson Valley lake fell to the Coveville level. The Coveville stage was fluvial-lacustrine throughout the Hudson Valley. Catastrophic floods still came through the Mohawk Valley from the Ontario Basin and the Adirondacks. These floods might have triggered the drop in lake levels from the Quaker Springs to the Coveville stage, and from the Coveville to the Fort Ann stage (LaFleur, 1983; DeSimone and LaFleur, 1986).

The margin of the glacier oscillated as a weakly rejuvenated Hudson-Champlain lobe deposited the Riverbank, North Creek, and Euba Mills moraines and heads of outwash in the upper Schroon Valley (Dineen, in prep.). Lake Coveville expanded into the Champlain Valley as the glacier retreated (DeSimone and LaFleur, 1986). Outflow from Lake Iroquois deposited the Scotia gravels (LaFleur, 1983; Muller and Prest, 1985).

FIGURE 19: Glacial Lake Vermont-Fort Ann: Lake Albany was overflowing through Hell Gate and the East River as Lake Coveville fell to Lake Fort Ann. Riverine conditions dominated the Hudson Valley from Fort Ann to Cohoes and from Newburgh to the Terminal Moraine. The Rome Outlet was abandoned as the retreating ice uncovered the Covey Hill Spillway and Lake Iroquois fell to Lake Frontenac. The ice front began to retreat from the northernmost buttresses of the Adirondacks, as ephemeral meltwater lakes formed along the ice margin (C-14 date: #14). The Chateauguay Channels and the Covey Hill outlet were carved into the northern slope of the Adirondacks (Denny, 1974; Clark and Street, 1984). The lake in the Champlain Basin fell through a series of levels as its outlet at Fort Ann was progressively eroded (DeSimone and LaFleur, 1985, 1986).

Lake Iroquois and post-Iroquois flood events might have caused the successively lower Fort Ann stages. When the ice margin retreated into Canada, the Champlain Sea flooded the St. Lawrence Lowlands (Denny, 1974; C-14 dates: #17, Muller and Prest, 1985, #18; #11, #13, Pair and others, 1988).

Downcutting was extremely rapid in the Hudson Valley during the Fort Ann stage (Woodworth, 1905; Newman and others, 1969; Hanson, 1977; DeSimone, 1977; DeSimone and LaFleur, 1986; Dineen and Duskin, 1987). The water plane of the Fort Ann stage of Lake Albany was graded to present sea level in the vicinity of the Narrows (Figs. 3 and 5). It appears to be erosional through most of the Hudson Valley. The Terminal Moraine dam failed, probably because of mass movements induced by jokulhlaups, ground water sapping along its outer margins, melting of enclosed ice blocks,

and erosion of the moraine's crest by flood overflow. The moraine dam was rapidly sluiced away and the Hudson's present course was established through the Narrows. Catastrophic floods severely eroded the lower Hudson Valley following the breaching of the Terminal Moraine. Over 200 ft (60 m) of lacustrine deposits were eroded from the lower Hudson gorge. The lake sequence in eastern New York came to an end and fluvial erosion began to dominate the Hudson Valley.

The Hudson River began to assume its present characteristics between 11,000 and 11,500 years ago, when the ice front pulled back enough to allow the Champlain Sea to invade the St. Lawrence Lowlands and the supply of meltwater to the Hudson Valley was cut off.

SUMMARY AND CONCLUSIONS

Lake Albany lasted from ~18,000 to 13,200 yBP and extended from the Terminal Moraine, between Staten and Long Islands, to Glens Falls, NY. The later Lake Albany stages ended 11,250 yBP. Lakes Hudson and Albany were continuous and out-lasted Lake Hackensack. Lake Hudson-Albany was dammed by the Terminal Moraine, and overflowed or spilled through Sparkill Gap into the Hackensack Lowlands. The overflow from the lake eroded the emergent Lake Hackensack clays, and deposited a ribbon of fluvial sand and gravel from Sparkill, NY, to Newark, NJ.

Several lake stages can be documented using these sand and clay plains, deltas, and beaches, including Lake Albany I and II (ice-contact and stable), Lake Quaker Springs, Lake Coveville, and Lake Fort Ann I, II, and III. The water planes of the various stages tilt down to the south because of glacial isostatic rebound. The tilts decrease from oldest (Lake Albany I) to youngest (Lake Fort Ann III). The water plane of Lake Albany tilts 2.6 ft/mile (0.49 m/km), while that of Fort Ann III tilts 1.1 ft/mile (0.21 m/km). Lakes Quaker Springs, Coveville, and Fort Ann I, II, and III project into the Champlain Valley, where they were stages of Lake Vermont. The Coveville and Fort Ann stages were riverine in the Hudson Valley.

The facies of the Lake Albany deposits change systematically from ice-contact sand and gravel at the base to rhythmically laminated silt and clay in the middle, distal beds, to silty sand in shallow water sediments, to gravelly sand in near-shore sediments. The sediment sequence was strongly influenced at the base by the retreating Hudson Glacial Lobe, and at the top by the shoaling effects of falling water levels. Lake Albany deposits are sparsely fossiliferous. Mollusks and trace fossils have been found in recently. The fossils that are present suggest a relatively stable water level for decades, and that vegetation and fish were present.

Moraine segments, kame deltas, and heads-of-outwash document many

recessional ice margins in Lake Albany. Possible glacial surges include the Prattsville-Shenandoah, Middleburgh-Rosendale, and Yost-Delmar. The Terminal Moraine became the dam for the lake. The Meadowdale margin coincided with the formation of Lake Sacandaga in the southern Adirondack Uplands, Lake Amsterdam in the Mohawk Valley, and Lake Bascom in the Taconic Mountains. The Carter margin was the last ice front in Lake Albany, coincided with the development of Lake Warrensburg in the eastern Adirondack Mountains, and marked the last mapped recessional position in the Hudson Valley.

The Quaker Springs and Coveville lake stages received several catastrophic floods from the Mohawk Valley. Later, catastrophic floods debouched from the Champlain Valley into the Hudson Lowlands during the Fort Ann stages. The catastrophic flood events seem to have coincided with the successive drops in lake levels. The Terminal Moraine dam probably collapsed during a catastrophic flood during the Fort Ann stage.

The Middleburgh Readvance is Woodfordian in age and correlates with the Valley Heads readvance in Central NY. Lakes Quaker Springs and Coveville were contemporaneous with Lake Iroquois, and Lake Fort Ann existed during Lake Frontenac time. The Hudson River assumed its present drainage area when the Hudson-Champlain Lobe retreated from the Covey Hill spillway and the Champlain Sea flooded the St. Lawrence and Champlain Lowlands.

None of the Hudson Lowlands lakes have been absolutely dated using <u>in situ</u> organic material or firm stratigraphic relationships. Dineen and others (1988) used "older than..." radio-carbon dates that infer the eastern New York lakes existed from 17,000 yr BP to 11,600 yr BP. Sirkin (1977) suggested that Lakes Hudson and Hackensack formed as the glacier retreated from the Terminal Moraine 21,000 years ago. The Hudson Valley lakes ceased to exist approximately 11,000 years ago, when the ice sheet retreated north of the Adirondack Mountains and the Champlain Sea subsequently flooded the St. Lawrence Lowlands (LaSalle, 1966; Clark and Karrow, 1984). Thus, the lakes existed for over 7,000 years.

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APPENDIX A: FACIES OF PROGLACIAL LAKE ALBANY AND FIELD STOPS THAT ILLUSTRATE THEM

Unconformable Contact with Striated Rock or Till:

Sharp contact between rhythmically laminated fine sand, silt, and clay.

Field Stops
- 8, 10, and 13 in Dineen and Rogers (1979)
- 12 in Dineen and Duskin (1987)

Basal Ice-contact Facies:

Faulted and deformed cross-bedded sand and gravel with interbedded flowtills in subaqueous fans and sinuous gravel ridges or eskers (Dineen and Hanson, 1983; Dineen and others, 1988). Locally, they grade into valley-side kame delta deposits (LaFleur, 1969a). The ice-contact deposits are subaqueous fans. The ice-contact deposits grade up and laterally into the Deep-water Facies.

Field Stops
- 8 in LaFleur (1961b)
- G-5 in Connally and Sirkin (1967)
- 9-5 in LaFleur (1969a)
- 3, 5, 7, and 8 in LaFleur (1979)
- 1, 5, 6, 7, and 16 in Dineen and Rogers (1979)
- 6 in DeSimone and LaFleur (1985)
- 4, 6, 9 and 12 in Dineen and Duskin (1987)
- 5, 6, and 7 in Dineen and others (1988)

Deep-water Facies:

Rhythmically bedded silt and clay that grade up from ripple-laminated silt to rhythmites of clay, silt, and fine sand. The thickness of each couplet and the proportion of silt-to-clay in them decreases upward. Contains beds of ripple-laminated sand or sand and gravel, and lenses or beds of contorted clay, striated boulders, and till.

Field Stops
- 6 and 8 in LaFleur (1961b)
- 12 and 13 in LaFleur (1965b)
- 9-2 in LaFleur (1969a)
- 3, 4, 5, 7, and 8 in LaFleur (1979)
- 1, 2, 12, 13, 14, 15, and 16 in Dineen and Rogers (1979)
- 6 and 7 in DeSimone and LaFleur (1985)
- 4, 6, 11, and 12 in Dineen and Duskin (1987)
- 7 in Dineen (1987)
- 6, 7, and 9 in Dineen and others (1988)

Shallow-water Facies:

Silty gravel adjacent to ridges of till or gravel.

Yellow brown, planar- to ripple-laminated, fine to medium sand, generally has a gradational contact with the underlying clay. In the Kingston area, the sand is trough-bedded with many cut-and-fill structures (Dineen and Duskin, 1987).

Field Stops
8 in LaFleur (1979)
2, 3, 13, and 15 in Dineen and Rogers (1979)
3 and 6 in Dineen and Duskin (1987)

<u>Near-shore Facies:</u>

Fan-shaped accumulations of trough-cross-bedded sand and gravel topsets, planar-cross-bedded gravelly sand foresets, and ripple-laminated to planar-bedded sand, silt toesets and bottomsets.

Field Stops
5 and 6 in LaFleur (1961b)
13 in LaFleur (1965b)
5 in LaFleur (1969a)
9-4, 9-5, and 9-7 in LaFleur (1979)
1, 2, 5, 9, 10, 11, 12, 13, and 16 in Dineen and Rogers (1979)
1 and 4 in DeSimone and LaFleur (1985)
3 in Dineen and Duskin (1987)
7 in Dineen and others (1988)

<u>Fluvial Facies:</u>
Fluvial deposits are trough-cross-bedded sand and gravel to poorly-sorted gravelly sand that are poorly stratified with horizontal bedding. They bear an unconformable relationship with the older lake deposits.

The sediments in the floors and terraces of the outlet or catastrophic flood channels tend to be thin but coarse with boulder lags. Slackwater deposits occur along the channel sides. The channel floor is usually grooved parallel to the long axis of the channel, tear-drop shaped hills often occur on the floors of the channels.

Field Stops (Fluvial Facies)
5 and 11 in LaFleur (1961b)
4, 8, and 10 in LaFleur (1979)
2, 3, and 4 in LaFleur (1983)
15 in Dineen and Rogers (1979)
7 and 11 in Dineen and Duskin (1987)

Field Stops (Channels)
3, 6, 17 and 18 in LaFleur (1965b)
6 in LaFleur (1979)

9 in DeSimone and LaFleur (1985)
8 and 10 in Dineen and others (1988)

Beach Facies: Beaches include planar-bedded, imbricated gravelly sand, ripple-trough-laminated sand, boulder pavements, and wave-cut scarps.

Field Stops
9 in LaFleur (1961b)
12 and 14 in LaFleur (1965b)
9-4, 9-6, and 9-8 in LaFleur (1969a)
1 in LaFleur (1979)
3 in Dineen and Rogers (1979)

Aeolian Facies: Aeolian sand is yellow brown, planar-cross-laminated medium to fine sand that unconformable overlies the lacustrine deposits. Occurs in parabolic and linear ridges.

Field Stops
3 and 4 in Dineen and Rogers (1979)

APPENDIX B
FIELD TRIP STOPS

The field trip stops will be in the eastern Mohawk Valley (Saturday) and Hudson Valley (Sunday). The southernmost stop will examine the Lake Albany sequence, the westernmost stops will examine an interlobate moraine, and the northernmost stops will examine catastrophic flood channels. The field stops are located on Figure 9.

STOP SA-1: YOST PIT (Randall Quad): This exposure is a small borrow pit that lies behind a farmhouse. The pit is excavated into the base of the 420 ft Fonda Terrace (LaFleur, 1983). The pit exposes over 20 m of glacial sediment that documents the Yost Readvance.

19.00 m	10YR 5/4 Bleached cross-laminated fine sand, upper 2 m are contorted. 10 cm cobbles are common (Lake Amsterdam- ice proximal beds)
14.40 m	10YR 5/4 Imbricated cobbly gravel, imbricated down to east (sediment transport to the west), cobbles are 10-20 cm, subangular to subrounded (subaqueous outwash)
13.20 m	10YR 6/4 Compact silt-matrix-supported very fine sand diamicton, subhorizontal partings, trace cobbles (basal till)
11.20 m	10YR 6/4 Sand and gravel, planar cross bedded 15-20 N90W, trace cobbles and boulders
6.00 m	Boulder pavement- scattered facetted and striated boulders on truncation surface
6.00 m	10YR 3/1 Compact, laminated, matrix-supported diamicton, alternating beds of silty fine sand and very fine sand, with facetted cobbles, upper 1.5 m is oxidized
3.45 m	10YR 3/1 Compact, matrix-supported, massive diamicton, very fine sandy silt matrix with trace gravel, sheared at top
2.90 m	10YR 6/1 Very compact, matrix-supported diamicton, slightly gravelly fine sandy silt matrix, many faint lenses of cross-laminated silt, many partings, gravel axes are horizontal, weathered in upper 20 cm
1.20 m	10YR 8/2 Cross bedded pebbly sand and gravel, cross beds dip 20 N60W, upper 20 cm is sheared, upper 40 cm is leached
base	

The clasts in the diamictons are predominantly from the Mohawk Valley.

STOP SA-2: HERBA PIT (Broadalbin Quad): This pit is on the axis of the Broadalbin Interlobate Moraine. The clasts are predominantly from the Sacandaga Graben and Mohawk Valley. Over 9 m of sediment have been described here.

9.70 m	7.5YR 6/8 matrix-supported diamicton, sand and silt matrix, some cobbles
9.00 m	10YR 6/3 trough cross-bedded cobbly gravel
8.60 m	10YR 7/6 lenticular, trough cross-bedded cobbly sand and gravel
6.00 m	10YR 5/4 compact, matrix-supported massive to laminated diamicton, with a sandy matrix, some laminae of fine sand
4.80 m	10YR 7/2 interbedded massive to planar bedded sand and massive sand diamict, greasy rotten shale clasts are common, might be a soil zone, shearing in lower 20 cm
3.90 m	Truncation Surface
3.90 m	10YR 7/2 cross bedded cobbles and sand, cross beds dip 10 S80E

STOP SA-3: REX EXCAVATING PIT (Gloversville Quad): This instructive pit exposes the core of the Broadalbin Interlobate Moraine. It contains several till and lacustrine sequences that suggest either an oscillating ice margin or multiple readvances. A possible soil zone lies between the tills (is there a soils scientist in the house?). A similar exposure lies 1.1 miles to the east. The clasts were derived from the Adirondack Mountains and the Mohawk Valley. A typical section from this pit includes:

11.30 m	10YR 4/2 matrix supported diamicton, with a silt matrix, faintly laminated with sand stringers at base, fissile, sheared or interlaminated with
9.70 m	10YR 6/4 irregularly laminated fine sand and silt, rotten shale stones are present
8.00 m	10YR 4/6 laminated, matrix-supported diamicton, silt matrix
7.70 m	10YR 6/3 laminated fine sand, cobbly, locally cemented
6.60 m	10YR 5/3 planar cross bedded gravelly sand, trace boulders, cross beds dip 10-15 N60W
2.00 m	Silt matrix supported diamicton, striated boulders

STOP SA-4: TWIN CITIES SAND AND GRAVEL PIT (Gloversville Quad): This pit is over 300 m long and 14 m high. It is at the distal portion of the Interlobate Moraine and contains distinct deltaic features. Both gravity and thrust faults are common in the east (proximal) side of the pit.

14.00 m	(west only) cross-laminated fine sand (aeolian)
12.00 m	Laminated (west) to massive sandy matrix-supported diamict, fills low spots in lower unit, soil zone at top.
10.00 m	Planar cross bedded fine sand with lenses of trough laminated sand, faulted to east, contains contorted lenses of diamict to east.
base	

STOP SU 1- POWELL-MINNOCK CLAYPIT (Ravena Quad): The Powell-Minock claypit is on the Ravena 7-1/2 minute quadrangle in Coeymans, Albany County, NY. The pit supplied the Powell-Minnock brickworks from the turn of the century until the nineteen fifties. Powell-Minnock presently make their bricks from the Esopus Shale.

The claypit is in a clay plug that fills the mouth of the buried Colonie Channel. The Colonie meets the preglacial Battenkill-Hudson under the tidal flats of the present Hudson River. The floor of the Colonie Channel beneath the pit lies at an elevation of 110 ft (33.3 m) below sea level (Dineen and others, 1983). A composite stratigraphic section has been measured at the pit, a few grain-size analyses have been run (Gail Ashley at Rutgers), and measurements of paleomagnetism have been performed (William Brennen, SUNY at Geneseo). Gail Ashley noted the presence of insect larvae tracks in the silt rhythmites at this site. The lithic section is as follows:

19.4 m	10YR 4/2 clay and 10YR 8/2 fine silt rhythmites, 1 to 10 cm (0.5 to 4 in) thick, 5 cm (2 in) average. Rhythmites are planar, with some convoluted and flame structures. Flames are overturned to the west. The sequence fines upward and is 50% clay. "Clay dogs" or doughnut-shaped concretions are present. (WEATHERED DEEP-WATER FACIES)
16.4 m	10YR N/5 rhythmic clay with less than 5% silt. Rhythmites are 0.5 to 10 cm (0.2 to 4 in) thick, thinning upward. One ripple-laminated, fine sand bed occurs under the top rhythmite. The size of the clay is 8 phi.Insect larvae tracks were found in this layer. (DEEP-WATER FACIES)
14.9 m	10YR 2/2 mottled clay and fine-sandy silt in 1 to 10 cm (0.5 to 4 in) thick rhythmites. The rhythmites thicken and become coarser upwards. The silty members are climbing ripple-laminated, the ripples climb to the N90E and are capped with fine sand. (DISTAL TURBIDITES)
12.3 m	10YR 5/2 contorted, fine to medium sand, with some 1 to 2 cm (0.5 to 1 in) laminae of clay. The sand contains abundant flame structures that are overturned to the west. Blebs of clay are abundant at the base of this bed.(TURBIDITE)
11.4 m	10YR N/4 massive to poorly layered clay. Laminae are overturned to the S80W, the top of the bed has ridges that trend N57E to N10E. (TURBIDITE)
9.7 m	10YR N/4 rhythmic clay and silt, with dropstones near the top of the unit. The dropstones are as large as 2cm by 2cm by 8cm (1 by 1 by 3 in) and are rounded. Rhythmites are 4 to 15 cm (2 to 6 in) thick, and are 50% clay.(DEEP- WATER FACIES)
9.1 m	10YR N/4 massive to contorted clay with 5 to 10% silt and 50 cm (20 in) thick fine sand

	beds. (TURBIDITE)
7.5 m	2.5Y 7/2 silt and 10YR N/4 clay and silt rhythmites. Silt is ripple-laminated to cross-bedded. Some 2 to 5 cm (1 to 2 in) rounded pebbles occur in the clay. One 100cm by 100cm by 200cm (40 by 40 by 80 in) striated graywacke boulder or dropstone was present. The bedding is deformed under the large dropstone. The rhythmites are 8 to 35 cm (6 to 12 in) thick, decreasing in thickness and silt content upwards, they dip towards the southwest (dips range from S20E to N70W). Some vertical N15E joints are present. (ICE-CONTACT FACIES)
4.1 m	10YR 5/2 very fine to medium sand, with 25N60W climbing ripples at the base, N60W ripple-trough-laminae in the center, and 20S80W climbing ripples at the top. This unit fines upward. (ICE-CONTACT FACIES)
2.8 m	5Y 5/3 to 5Y 6/3 fining upward, ripple-laminated sand, silt, and clay. (ICE-CONTACT FACIES)
2.6 m	10YR 5/2 cross-bedded cobbly sand and gravel. Beds are 10 to 15 cm (4 to 6 in) thick and fine upwards. Some imbricated cobble beds, with a silty sand matrix. Cobbles are rounded and range in size from 2 to 12 cm (1 to 5 in). The cross-beds dip from S40W to S10W. The sequence is faulted from the base to 4.1 m (13.5 ft). These normal faults trend S50W and dip 35S. (ICE-CONTACT FACIES)

Test borings in the claypit suggest that the ice-contact sand and gravel extends to a depth of 170 ft (52 m), that it overlies 5 to 10 ft (1.5 to 3 m) of till, and that its top is quite uneven.

STOP SU-2: MASICK (ROTTERDAM) PIT (Rotterdam Junction Quad): The Masick sand and gravel pit lies in the Rotterdam Junction 7-1/2 minute, quadrangle in Schenectady County, at an elevation of 340 ft (104 m). The pit is in the southwest corner of the Schenectady Delta, at the point where the Normans Kill entered Lake Albany.

The pit exposes 10 to 20 ft (3 to 6 m) of 10YR 5/4 planar cross-bedded, cobbly sand and gravel at its base, the beds dip towards the southeast. They are overlain by 5 ft (1.5 m) of 10YR 5/4 trough-cross-bedded sand and gravel, the trough axes trend east-west. The tops of the sand and gravel units are sheared, and overlain by 1.5 to 10 ft (0.5 to 3 m) of 10YR 4/4 compact, fissile, matrix-supported, bouldery, sandy, clay diamicton. The diamicton is overlain by 1.5 to 13 ft (0.5 to 4 m) of 10YR 4/3 climbing ripple-laminated (base), to planar and ripple-trough- laminated (center), to climbing ripple-laminated (top) coarse to fine sand, fining upward. The ripples climb S0E to S50E. The sand beds are overlain by 13 ft (4 m) of trough-bedded, gravelly sand. The troughs are 8 to 12 in (20 to 30cm) deep and 5 to 10 ft (1.5 to 3

m) wide, and trend southeast. The top of the sequence is 6 to 10 ft (2 to 3 m) of 10YR 4/3 planar to trough-laminated fine to medium sand.

The basal sands and gravels are a kame delta associated with a recessional ice margin at Schenectady. The ice front readvanced across the kame delta, and deposited the gray, compact till. The glacial Normans Kill deposited the upper gravelly sands over the abandoned, till-covered kame after the ice retreated. Lake Albany did not fall to the Lake Albany II stage until after the 340-foot (109 m) Normans Kill and Schenectady delta was deposited.

STOP SU-3: WUNDERLICH (Pollock Road) PIT (Niskayuna Quad): The Wunderlich pit was Stop 1 in Dineen and Rogers (1979). It is in the Pollock Road kame delta, in northeastern Albany County, on the Niskayuna 7-1/2 minute quadrangle. Meltwater from the Niskayuna ice margin deposited a network of eskers from Clifton Park to Pollock Road (Dineen and others, 1983). These eskers grade into the northern edge of the Pollock Road kame delta. The delta was deposited in Lake Albany at the Niskayuna ice margin. It was abandoned, and the ice front had retreated out of the area, by the time the lake level fell to the Albany II stage.

The pit exposes 20 ft (6 m) of 10YR 6/3 normal faulted, trough-cross-bedded sand and gravel. The beds dip to the southwest, the faults are down to the north. These ice-contact deposits are overlain by 65 ft (20 m) of 10YR 5/4 gravity and thrust-faulted, planar cross-bedded gravelly sand, with contorted beds near the top of the unit. The cross-beds dip to the south. The top of the section is a 3 to 20 ft (1 to 6 m) 10YR 6/4 matrix-supported, bouldery, silty sand diamicton. The central beds grade into rhythmically laminated silt and clay to the south. The upper till does not extend to the south.

The sequence in the Wunderlich pit is very similar to that in the Masick Pit. Their deposits are probably contemporaneous and record the same minor readvance.

The Schenectady delta was deposited as the Hudson Lobe retreated from the Meadowdale to the Niskayuna margins (Hanson, 1977). Lake Albany clay filled the preglacial Alplaus-Ballston channel as the delta grew. The delta continued to grow, and the animals at Glenville continued to thrive, as the glacier retreated to the Hidden Valley-Glen Lake-Carter-Manchester ice margin (Fig. 7c). The lake level fell to the Albany II stage and the delta was abandoned. The emergent delta deflected the Mohawk River flow into the Alplaus-Ballston channel.

STOP SU-4: EAST LINE CHANNEL (Round Lake Quad): Catastrophic floods from the Ontario Basin started to carve out the clay in the Alplaus-Ballston channel, and to scour a series of overflow channels across the interfluve between the preglacial Alplaus-

Ballston and Colonie channels (Hanson, 1977; Dineen and others, 1983). Lobes of sand and gravel were deposited in Lake Albany II near Shenendahowa and Clifton Park (Dineen and others, 1983). Slackwater deposits from later floods backfilled the interfluve overflow channels. Most of the flood waters passed through the East Line, Drummond, and Ballston Creek channels into Lakes Quaker Springs and Coveville. A bedrock sill developed at East Line between the Drummond and Ballston Creek channels. The overflow channel gravels are a source of groundwater in southern Saratoga County.

The Drummond channel carried the floods across the present site of Saratoga Lake and formed the Schuylerville plunge pool. At the same time, flood overflows deepened the Ballston Creek channel. The Ballson Creek channel eventually "pirated" flood flow into and through the Round Lake basin. By the Fort Ann stage, the overflows from the catastrophic floods had reduced the floor of the Rexford channel (south, near Schenectady) enough to divert the the Mohawk River into its present course.

A controversy has been raging (kind of) between Dineen and Hanson concerning the origin of Ballston, Round, and Saratoga Lakes. Hanson cites abundant evidence for the existence of recessional ice margins at Niskayuna, Round Lake, and Saratoga Lake (Hanson, 1977). He suggests that thick blocks of stagnant ice were buried under the fluvial deposits at these margins. The lakes are in kettle holes left by the melted ice blocks.

Dineen points out that Round and Saratoga Lakes overlie the intersections of the Drummond and Ballston Creek channels with the preglacial Colonie channel (Dineen and others, 1983). He suggests that the lakes occupy large scour holes carved into the relatively soft glacial deposits that fill the Colonie channel. He also suggests that Ballston Lake also lies in a deep scour hole.

STOP SU-5: LUTHER FOREST GRAVEL PIT (Round Lake and Mechanicville Quads): The many private water companies within the Towns of Clifton Park and Malta (southern Saratoga County) derive their water supplies from a series of esker, outwash fans, and esker delta complexes. These complexes include an esker with a delta or fan at its terminal end. The Saratoga eskers form a semi-continuous line of eskers that is approximately 15 miles long. The complex lies along the eastern valley wall of the preglacial Colonie Channel (Dineen and others, 1983). There is a poorly developed esker complex along the western wall of the channel as well. The majority of the esker deposits are buried under Lake Albany sediments. The locations of the eskers have been documented through the use of hundreds of well borings drilled during numerous water exploration projects in southern Saratoga County.

The esker segments are from 2 to 5 miles long, with an average length of 3.8 miles. Their average cross-sectional width is 100 to

200 feet. The esker segments are offset along the paleo-ice margins. The offsets vary from 1,500 to 10,000 feet. The length of the eskers can be used to determine the width of the stagnation zone (esker length = stagnation zone width). The stagnation zone width averaged 3 to 5 miles.

Using the stagnation zone model and plotting the locations of active ice margins allows estimates of readvance distances to be made. The Wunderlich Pit (Stop SA-3) is at the southern terminous of the Niskayuna Ice Margin (Hanson, 1977; Dineen and others, 1983; Dineen and others, 1988). The readvance till that overlies the esker delta was deposited by a readvance from the Grooms Corners Ice Margin, a distance of 4 miles. This was the Delmar Readvance of Dineen and Rogers (1979).

The fans and deltas were deposited in the open waters of Lake Albany (elevation 340 ft at Niskayuna). Many of the deposits do not contain topset beds, indicating that the did not build up to the surface of the lake before the glacier retreated to the next margin. They are subaqueous outwash fans. The Luther Forest Delta has topset beds.

The series of esker segments, subaqueous outwash fans, and esker deltas are being utilized by both private water companies and by the aggregate extraction industry. In recent years, it has become common practice to install water supply wells in the gravel pits once the pits are reclaimed.

HUDSON LOWLAND LAKE LEVELS

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Late glacial and post-glacial isostatic crustal rebound in response to ice unloading tilted formerly horizontal glacial lake shorelines in the Hudson Lowland. Water level indicators permit reconstruction of these former lakes and identification of the most stable water levels. (Figure 3 and 20). The topset-foreset contact, either observed in exposures or interpreted from delta morphology, is the most useful indicator for Lakes Albany and Quaker Springs. Kame terraces graded to Lake Albany, river terraces along major and minor tributaries, and erosional terraces developed in lake plain clay along outlet channels were also useful water level indicators. Well developed beach profiles, while not plotted on the figure, extend for several kilometers along the eastern side of the Hudson Valley from North Greenbush to Middle Falls. These beaches formed along the shore of Lakes Albany and Quaker Springs.

Tilted water planes drawn through these data identified six stable lowland water levels: Lake Albany, Lake Quaker Springs, Lake Coveville, Fort Ann I, Fort Ann II and Fort Ann III (DeSimone 1985, DeSimone and LaFleur 1985, 1986). The Albany, Quaker Springs and Coveville water planes are clearly curved and asymptotic southward. The Albany water plane terminates in the Hudson Lowland (DeSimone 1983) but the Quaker Springs and Coveville water planes extend into the Champlain Lowland and link with their counterpart phases of Glacial Lake Vermont (Chapman 1937, LaFleur 1965).

Lakes Albany and Quaker Springs were <u>bona fide</u> lakes in their character within the Hudson Lowland as illustrated by numerous deltas, beach segments and thick rhythmite deposits. However, by the time water levels dropped to the Coveville phase, a more fluvial character can be inferred. The Coveville phase in the Hudson Lowland represents a transition from predominantly lacustrine to predominantly fluvial and includes features attributable to both natures. The sandy Hudson Falls Delta and prominent terrace gravels along both the Batten Kill and the Hoosic River were coeval with Lake Coveville. Elsewhere, Coveville features consist of terraces cut into the lake plain clay which are often capped with a meter or so of well sorted sand. This sand may be heaped into dunes which must have formed in post-Coveville times.

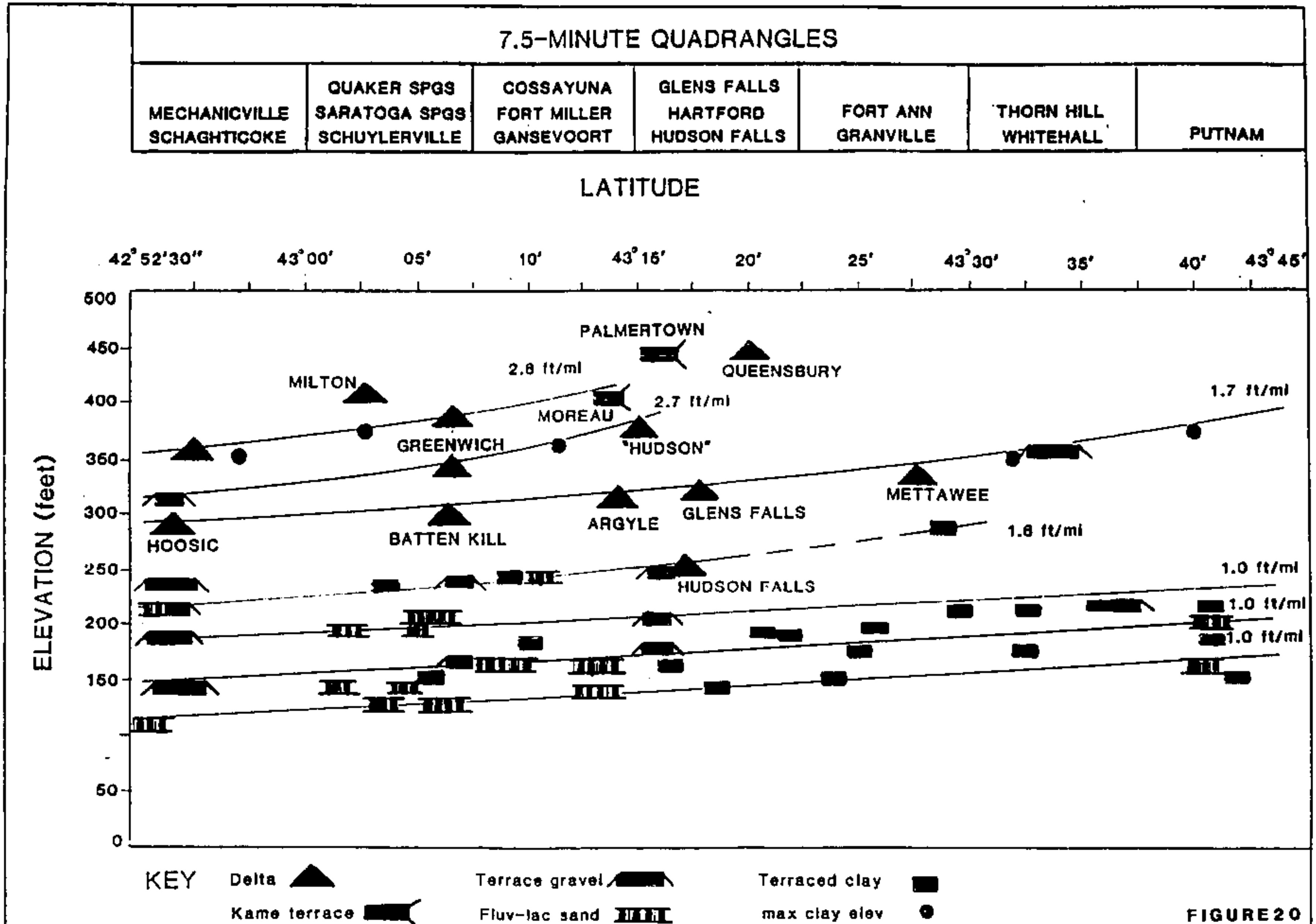

FIGURE 20

LAKE LEVELS NORTHREN HUDSON VALLEY

Fort Ann waters in the Hudson Lowland were predominantly fluvial in character as demonstrated by the flat, low gradient profiles and abundant erosional terraces with or without a cap of sand and gravel. One of the most extensive Fort Ann sand and gravel terraces is being excavated along River Road in the Town of Schaghticoke approximately 4.5 miles north of Stop SU-6. Look for it as we drive north as optional Stop SU-11.

LAKE LEVEL TRANSITIONS

A lake level transition or interval of falling water level could have been gradual, either through a continuous or step-wise lowering, or catastrophic in nature. Discharge, catastrophic or otherwise, from one basin to another would have impacted the water level in the receiving basin, probably lowering water levels by erosion of a sill or outlet channel. Readvances into a lake basin and subsequent meltwater influx during retreat would have similarly impacted water levels. To date, readvances and probable correlative water level transitions have not been satisfactorily linked.

The cause of the transition from Lake Albany to Lake Quaker Springs has long been a puzzle. Could there have been an ice readvance into the northern Hudson Lowland which precipitated the decline in water level? One of us has recently proposed that the demise of Lake Albany was the result of a readvance of the Hudson Lobe correlative to the Stanwix Readvance of the Oneida and Black River sublobes in the western Mohawk Valley (Table 6, Table 7 DeSimone and Lovejoy 1992). This Stanwix equivalent deposited the extensive Palmertown Kame terrace and Moreau Pond Kame moraine along the grounded western limb of the Hudson Lobe. Stagnation at the readvance limit resulted in the massive Glen Lake stagnation zone deposits most recently mapped in detail by Connally (1973). A re-examination of these sediments may shed light on this new interpretation.

Stop SU-6 provides us with indirect evidence of this readvance. The typically thick rhythmite sequence long attributed to deposition in Lakes Albany and Quaker Springs is interrupted here by a persistent layer of underflow and interflow sands approximately 1 meter thick. Numerous lenses and pods of a pebbly, clayey diamicton occur primarily along one discrete horizon within the sands. Soft sediment deformation accompanied deposition of these "till balls" as they settled through the water column and plunked into the soft lake bottom ooze. These dropstone tills and the interruption of the normal deepwater rhythmite sequence

CORRELATION OF POST-ERIE LITHOSTRATIGRAPHIC AND MORPHOSTRATIGRAPHIC UNITS AND EVENTS IN THE MOHAWK AND HUDSON LOWLANDS

DE SIMONE 30 JAN 92

WESTERN MOHAWK	EASTERN MOHAWK	HUDSON (EVENTS)	HUDSON	ICE MARGIN #
LAKES DAWSON, AVON & "HYPERIROQUOIS" DRAINAGE TO GLACIOMOHAWK RIVER	MOHAWK COBBLE GRAVELS & MOHAWK OUTLET CHANNELS	QUAKER SPRINGS; GLENS FALLS ▵; BATTENKILL ▵	WELCH HOLLOW KM & KT	12
			SOUTH GRANVILLE M	11
			ARGYLE KD, TAMARACK SWAMP KM	10
			VALLEY CROSSROADS SF	9
			HARPER ROAD SF	8
			BLACK VALLEY M, PATTEN MILLS KT	7
ROME D		HUDSON ▵	CARTER SWAMP KM, LAKE SUNNYSIDE KM	6
STANWIX READVANCE			PALMERTOWN KT & MOREAU POND KM, CLARK POND KM, GLEN LAKE STAGNATION ZONE	5
	LAKE AMSTERDAM; 440-400 FT CRANESVILLE OUTLET	LAKE ALBANY; SCHENECTADY ▵	WILTON KM, BATTENVILLE M,	4
		HOOSIC ▵	ARCHDALE KM, HIDDEN VALLEY M, WOODLAND LAKE KM (CORINTH ICE TONGUE)	3
	WEST HILL OUTLET		WILLOW GLEN KD, COUNTRY KNOLLS KD,	2
	500 FT; FONDA SAND PLAIN	POESTENKILL ▵	NORTH HOOSICK M, HALFMOON KD, GROOMS KD, WATERFORD SF	1
POLAND FORMATION: HOLLAND PATENT D, BARNVELD READVANCE	800 FT; DELANSON OUTLET		HOOSICK FALLS KM, SPEIGLETOWN-MELROSE KT, POLLOCK ROAD KD	A
	LAKE SCHOHARIE		LOUDONVILLE KM, GUILDERLAND KT, RENSSELAER SF, NORTH GREENBUSH KT SYCAWAY KT	B
NORWAY D, HINCKLEY READVANCE	INTERLOBATE M, YOSTS READVANCE	SCHODACK ▵	MEADOWDALE M, WYNANTSKILL KM, VOORHEESVILLE KD, WEMPLE & HAMPTON SF,	C
	1180 FT		EAST GREENBUSH-SCHODACK KT, NEW SALEM KM WEST SAND LAKE-BURDEN LAKE KM	D
HAWTHORNE D, SALISBURY READVANCE	MOHAWK D		McKOWNVILLE M, PINE SWAMP-NASSAU LAKE KM, SAND LAKE-GLASS LAKE KM	E

Table 6.

FEATURES OF PROPOSED POST-ERIE CORRELATION SCHEME

STRENGTHS

1. Satisfies two till stratigraphy of lower Mohawk (Hell Hollow site) and northern Hudson (West Milton site) separated by free drainage of Erie Interstadial.

2. Readvance margins distinguished by high volumes of ice marginal sediment.

3. Readvances have an impact on levels of local and regional lakes.

    A. Lake Schoharie is a multi-phase lake.

    B. Salisbury limit in Hudson Valley separates older 180 ft Lake Albany deposits from main 330 ft Lake Albany deposits.

    C. Stanwix readvance in Hudson Valley corresponds to lowering of Lake Albany to Lake Quaker Springs.

    D. Oneida and Black River ice readvances trigger lake level changes in lower Mohawk valley.

4. Reasonable agreement with existing radiocarbon dates and with varve years for intervals between readvance pulses.

WEAKNESSES

1. Not all readvance margins have been identified as of readvance origin...TESTABLE.

2. Morphostratigraphic tracing of ice margins is imprecise.

3. Inherent and perhaps unrecognized biases of individual workers.

4. Simplistic assumption that readvances and events are contemporaneous for Oneida, Black River, Adirondack, Mohawk & Hudson ice.

Table 7.

is consistent with a readvance of the Hudson Lobe into the northern part of Lake Albany. While the western limb of the Hudson Lobe was grounded west and north of Glens Falls, the eastern limb at the terminus floated in Lake Albany. The readvance and subsequent rapid retreat of the floating margin by calving could have been responsible for the drop of Lake Albany to Lake Quaker Springs. Where exposures have been studied, evidence for calving above and below this horizon is not nearly so prominent.

Lake Albany never recovered. Woodworth (1905, p.177) first observed the "falling off in altitude of the deltas successively northward from that of the Batten Kill" and noted their failure to coincide with an Albany water plane. He concluded, "...these lower deltas were not made in the waters of Lake Albany." The Glens Falls Delta (320 ft.), Batten Kill Delta (310 ft.), Argyle Kame Delta (320 ft.), Mettawee Delta (330 ft.) and Fair Haven Delta (360 ft.) were all deposited into stable Lake Quaker Springs at or near its maximum level (DeSimone 1985). Optional Stop SU-9 provides an opportunity to see a complete stratigraphic section through a "thin" portion of the Batten Kill Delta.

Mohawk Outlet Channels:

There is considerable evidence that the Quaker Springs-Coveville and Coveville-Fort Ann I transitions were initiated by high discharge through the glacial Mohawk Valley and eastern outlet channels which resulted from water level fluctuations in the Ontario Lowland (Stoller 1922, LaFleur 1975, 1979, 1983, Hanson 1977). Extensive erosion in the lower Fish Creek valley from Grangerville to Coveville indicates the Grangerville Channel carried high discharge during the Coveville-Fort Ann I transition (DeSimone 1977, p35). The channel topography is strongly fluted in the southeastward flow direction; lake clay remnants are preserved in the shadows of eroded drumlins; and till and bedrock were scoured from 240-200 ft. elevations within the channel. Optional Stop SU-10 takes us to this distal portion of the Grangerville Channel.

Fort Ann Outlet Channels: The Fort Ann channels include the primary Fort Edward channel (Woodworth 1905, p198), the Durkeetown channel (Chadwick 1928, p914), and the Winchell channel (DeSimone 1985). These Champlain Lowland outlet channels were broadly defined during the Coveville phase and extensively excavated through clay, till, and bedrock during the Fort Ann phase. The Hudson-Champlain canal and the Hudson River below Fort Edward occupy the Fort Edward channel. Dead Creek, the distal portion of the Moses Kill, and the headwaters of Wood Creek occupy the broad Durkeetown channel. Winchell Creek and the distal portion of Big

(Mill) Creek occupy the Winchell channel. All of these modern streams are underfit and follow entrenched courses with narrow floodplains on the channel bottoms. Eroded clay, till, and bedrock comprise the channel escarpments with predominantly eroded clay in the channel bottoms.

Clark and Karrow (1984) and Pair, Karrow and Clark (1987) correlated falling water levels in the Ontario and St. Lawrence Lowlands (lower Iroquois and post-Iroquois) with drainage around Covey Hill into Lake Fort Ann in the northern Champlain Lowland. This inflow to Lake Fort Ann increased outflow from the southern end of the Champlain Lowland along three Fort Ann channels (Fort Edward, Durkeetown, Winchell). Accompanying erosion deepened these outlet channels until the water level stabilized at a lower elevation.

Stops SU-7 and SU-10 provide us with an opportunity to see features associated with two of the three Fort Ann Outlet Channels. At Stop SU-7, we can overlook the well-defined Winchell channel. Stop SU-8 will give many of you an opportunity to see the classic potholes described by Chapman (1937) which were scoured by high discharge along the primary Fort Edward channel.

These water level transitions link the deglacial histories of adjacent basins and thus represent a valuable correlation tool. Figure 21 presents a modified correlation table for water levels in the Ontario, St. Lawrence, Champlain and Hudson Lowlands. The original version of this table was offered by DeSimone as an alternative to Dineen's scheme (Dineen, DeSimone and Hanson 1988). The obvious gap in these data lies with the undifferentiated nature of Fort Ann waters in the Champlain Lowland. Can multiple Fort Ann levels be recognized there? It sounds like a good dissertation project to us!

DE SIMONE 27 APR 92

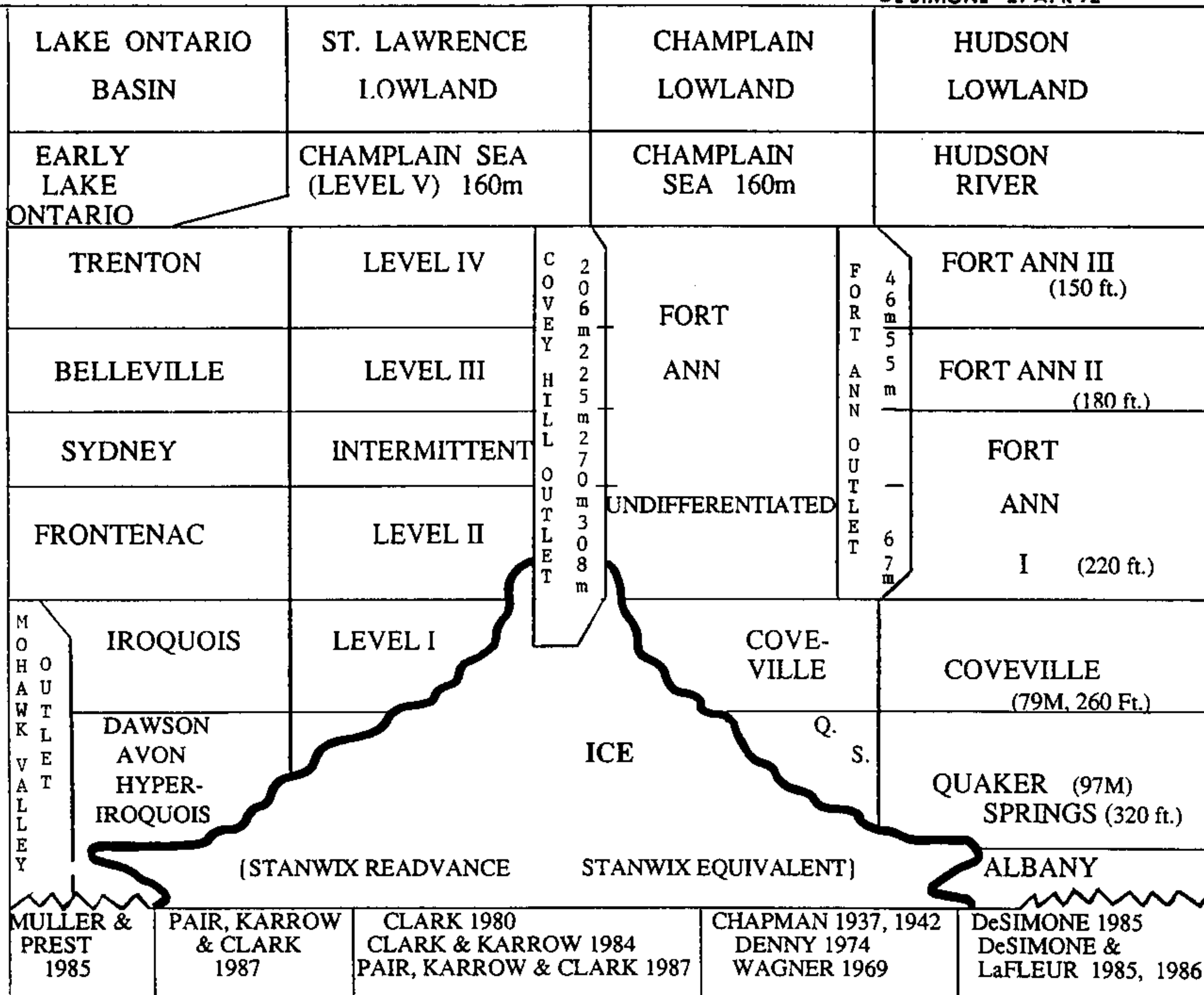

FIGURE 21
CORRELATION CHART

STOP SU-6: BERGY BITS AND THE FALL OF LAKE ALBANY

Location: Troy North 1:24,000 quadrangle; from junction NY40 and NY142, proceed 3.2 mi. north on NY40, turn west onto RC122 (Calhoun Road) for 1.8 mi., turn south onto Old River Road and proceed 0.5 mi. to parking area adjacent to Champlain Canal Lock 1.

Exposures of Lakes Albany and Quaker Springs rythmites lie along the riverside road for a mile. The road has been closed for several years because of large landslides that occurred after heavy seasonal rains hit during the spring thaw. The deluge coincided with the inability of the Town of Schagticoke to pay for repairs. Observations in the Albany area have shown that the clays are potentially unstable if the surface slopes exceed 12 degrees, especially after heavy rains or if the slopes have been undercut or loaded. This section of the road must be kept open for local residents, so after each slide, the road is cleared and the slope undercut again. The continuous horizontal laminae of clay, silt, and fine sand are occasionally interrupted lenses or pods of pebbly clay diamicton that range from a few centimeters to half a meter in diameter. The clays are typically contorted or disturbed under the diamicton lenses with flame and dish structure evident. The enclosing sediment for the single, prominent diamicton horizon consists of bedded sands, approximately 1 meter thick, sediment much coarser than the deepwater rhythmites above and below in the section. The compact, pebbly, unsorted nature of the diamicton pods and the deformation of the underlying sediments indicate that these pods and lenses settled through the water column and impacted the soft, oozy lake bottom. Hence, these diamictons are dropstone tills, presumably derived from icebergs drifting in the prevailing currents of Lakes Albany and Quaker Springs. The stratigraphic position of these exposures and their northern location suggest that these deposits date from the late Lake Albany and early Quaker Springs stages. The numerous dropstone till lenses and pods along one discrete horizon in the rhythmites of the Deepwater Facies represents an interval of accelerated ice margin retreat by calving.

STOP SU-7: FORT ANN CHANNELS - WINCHELL CHANNEL OVERLOOK

Location: Hartford 1:24,000 quadrangle; from junction NY40 and NY149, proceed 2.7 mi. west on NY149, from junction of Towpath Road and NY149, in Smiths Basin, proceed east on NY149 for approximately 3.2 mi.

The eastbound approach most dramatically illustrates

this eroded channel which carried some of the outflow from Lake Fort Ann in the Champlain Valley. Mill (Big) Creek enters the channel from the south through a bedrock gorge and meanders across the channel bottom toward Smith's Basin. At Smith's Basin, the Winchell Channel rejoins the primary Fort Edward Channel. Clay, till and bedrock are locally exposed along the flanks of the channel while lake clay with an organically enriched soil floors the channel. This pattern is generally true for the Durkeetown and Fort Edward Channels except that there is considerably more bedrock exposed to the north on the floor of the Fort Edward Channel.

As Washington County is largely rural and economically depressed, several attempts have been made to site private and/or regional landfill facilities in the County. One of us was involved with the preliminary site investigation for one of these landfill proposals. The former farm acreage included land on the flank and across the bottom of the Durkeetown Channel. The expansive clay floors seemingly offer an attractive site for landfilling. However, these channels are clearly pre-Woodfordian (or pre-glacial) bedrock valleys. It is possible that during ice retreat, locally extensive subaqueous fan gravel and sand may lie beneath the lacustrine clays and silts at the surface. Thus, there is the potential for valuable confined aquifers in these channels. In addition, the depth to bedrock and surface drainage characteristics on these channel floors will be highly variable. For the most part, the channel floors are wet, poorly drained areas occupied by grossly underfit streams. Wetland vegetation may be present. These factors all mitigate against selecting channel bottom areas as landfill sites. The thicker clay plains at elevations of 240-300 ft. are better suited to landfills, if we must have them at all.

STOP SU-8: FORT ANN CHANNELS - CHAPMAN'S POTHOLES

Location: Fort Ann 1:24,000 quadrangle; approximately 2 mi. north of the village of Fort Anne along US4.

The whitewater must have been something here during Fort Ann time. The potholes are concentrated at an area where the Fort Edward channel bifurcated around a bedrock knoll. Chapman (1937) interpreted these potholes as the product of major stream discharge from Lake Fort Ann in the Champlain Valley into the Hudson Valley. Three distinct Fort Ann levels are documented by a series of terraces, some in the outlet channels, that have been mapped in the Fort Ann area (DeSimoe, 1985; DeSimone and LaFleur, 1986). The water planes (DeSimone and LaFleur, 1986) and the presence

of the potholes suggest that the Fort Ann stage was fluvial in the Hudson Valley.

The Fort Ann water levels can be correlated with the water levels in the St. Lawrence and Ontario basins (DeSimone and LaFleur, 1985, 1986). Pulses of high discharge and rapid water level changes in the St. Lawrence and Ontario Lowlands (Clark and Karrow, 1984) coincide with periods of enhanced erosion in the Fort Ann outlet channels (DeSimone and LaFleur, 1985, 1986). The secondary Durkeetown and Winchell channels (DeSimone and LaFleur, 1986) might have been used as overflow routes for part of the enhanced discharge. The stable level of Fort Ann fell after each pulse of increased discharge.

OPTIONAL STOP SU-9: BATTENKILL DELTA STRATIGRAPHY

Location: Schuylerville 1:24,000 quadrangle; from junction US4 and NY29, proceed east on NY29 for 4.8 mi.; from junction NY40 and NY29, proceed west on NY29 a few tenths of a mile, turn north onto Windy Hill Road and continue 0.8 mi. to Tracy Bros. excavation on your right.

Depending upon exposure status and the groundwater level in the deeper parts of the pit, the stratigraphy here reveals a basal till which is silt-rich and gray-colored with rounded and subrounded clasts of local shale and carbonate lithologies. Rhythmically bedded lake sediments overlie the till. Sand and gravel of the Batten kill Delta overlies the rhythmites.

Well logs from the delta reveal a changing pattern to deposition. When ice retreat first uncovered the pre-glacial Battenkill channel south of this location, subaqueous fan deposition buried this channel with sand and gravel. The locus of deposition migrated westward and northward as ice retreat continued. This thick wedge of the Gilbert-type delta extends westward and is traversed by NY29. Sand and gravel thins to the north and south of NY29.

In the Battenkill channel, deposition of lake clays followed subaqueous fan gravel sedimentation and a thin slice of delta sand and gravel caps the sequence. Thus, two separate glacial aquifers exist here. A surface unconfined or water table aquifer exists across the delta while a confined aquifer (sealed by lake clay) exists in the old Battenkill channel. This resource awaits use by industry or for residential purposes.

OPTIONAL STOP SU-10: GRANGERVILLE FLUTING AND THE COVEVILLE

PLUNGE BASIN

Location: Schuylerville 1:24,000 quadrangle, from junction of US4 and NY32 in Village of Schuylerville proceed west on NY32 for approximately 1.9 mi. to junction with Degarno Road. From this location you can loop southward to Coveville and note the exposed bedrock floor of the Grangerville Channel above the plunge basin. You can also loop north to Grangeville and note the flute topography with pressure clays in the water flow shadows of the eroded drumlins along Degarno Road and the Burgarne crossroads.

This channel carried outflow from the Mohawk Valley during the Coveville-Fort Ann I transition. Fish Creek, the outlet of Saratoga Lake, flows along the channel floor for most of its length but turns northeastward and has carved a post-glacial bedrock gorge en route to its junction with the Hudson River in Schuylerville. A small drumlin field, which formed beneath the Hudson Lobe as ice flowed from the north-northeast, obstructed outflow along the Grangeville channel. Discharge scoured the drumlin flanks and eroded away most of the glaciolacustrine sediment which may once have nearly buried these drumlins. The result, which shows beautifully on the surficial map of the area, is a topography strongly fluted in the southeastward flow direction of Mohawk discharge. The numerous shadows or tails preserve isolated remnants of glaciolacustrine clays and silts. This topography strongly contrasts the original and well-preserved drumlin summits.

The distal end of the Grangerville channel prominently illustrates the erosive effects of Mohawk Valley discharge. The local shale bedrock has been stripped bare and scoured. At the junction of the Grangerville channel with the Hudson, a dry waterfall and plunge pool is preserved above the western bank of the Hudson River. Unfortunately, the land is privately owned and not accessible. Your best view is from the river. Hence, this is an "optional" stop.

OPTIONAL STOP SU-11: FORT ANN SANDS

Location: Schagticoke 1:24,000 quadrangle; proceed about 4.5 mi. north along River Road from our "Bergy Bits" stop. Permission to enter the pit must be obtained from Andy Clemente @ (518) 273-5800.

The horizontally stratified sand and pebble gravel here is much thicker than usual for Fort Ann sand deposits.

Undoubtedly, the Hoosic River to the north was responsible for the added sediment influx to the Fort Ann "river". Interestingly, the Hoosic River seems to have fairly extensive deposits graded to one or more of the Fort Ann levels, but the Battenkill's deposits of similar age are restricted to small terraces within the lower gorge of the Battenkill west of Middle Falls. Similarly, while the Hudson River deposited a large delta into Lake Coveville, there are no comparable Fort Ann deposits attributable to the Hudson. These observations suggest that the erosive power of the rushing Fort Ann waters dwindled dramatically as you proceed southward from the Champlain Valley along the primary Fort Edward channel. It would be instructive to estimate discharge and flow velocity ranges for each of the lake level transitions.

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Distribution and Emplacement of Coarse Gravels in the Eastern Mohawk Valley

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The City of Schenectady and several neighboring towns and villages derive their water supplies from a complex of gravel deposits that adjoin the Mohawk River/Erie Canal. About 20 million gallons per day are withdrawn from wells, most of which induce infiltration from river/canal water. Interest in protecting the Mohawk Aquifer is high, and several well fields have recently been hydrogeologically investigated and tested to determine radius of influence and groundwater gradients. Landfill closure investigations provide new stratigraphic details that relate aquifer sediments to older glacial/deglacial units.

In a 1965 USGS study of the water resources of the eastern Mohawk Valley, John Winslow and others carefully mapped the extent of a high-yield gravel aquifer that is bordered by sandy deposits of lesser productivity. A high-energy fluvial environment was assigned to the coarse gravel, and a more moderate floodplain environment for the mixed sand and gravel found in pits cut into terrace surfaces. At that time, there was little understanding of the complex hydraulic history of the glacial and deglacial Mohawk drainage - particularly the idea that episodic, high discharges (hlaups) could be responsible for transport and deposition of the thick coarse gravels of exceptional transmissivity (up to 5 million gallons per day per foot of width).

A review of the role of episodic high discharges in Mohawk Valley stratigraphy is found in LaFleur (1983). The source of the coarse bright gravel that comprises the most productive part of the Mohawk aquifer is the Little Falls Gravel (LFG), cited above (1983), and formally named by Ridge, Franzi, and Muller (1991). Reworked masses of LFG are found at St. Johnsville, Randall, Fort Hunter, Rotterdam Junction, Wyatts, and Scotia, where abundant outcrops show generally uniform and texturally simple bedding forms.

The disjointed distribution of the gravel masses along the Valley axis was though (LaFleur, 1983) to be the result of at least three hlaups during which the gravel would have been tracked only so long as the floods lasted, in a framework of falling valley base levels.

These high discharges were believed adequate to cause Hudson Lowland lake levels to decline. Alternatively, a more or less continuous high-level discharge occurred, accompanied by declining lake levels that fell for reasons other than flood surcharge. Once the source beds were sufficiently eroded to allow easier access for discharge, then a depositional record of the ongoing flood might diminish or disappear.

Between Schenectady and Saratoga Springs, a system of channels carried Mohawk water north and east to Glacial Lakes Quaker Springs, Coveville, and Fort Ann. Occasional very high discharges are suspected agents of erosion of the walls of the main Ballston and Drummond Channels, where bank overtopping, particularly at East Line, rather than piracy by headward erosion, seems responsible for initiation of the Anthony Kill Channel that drained to Lake Coveville at Mechanicville.

Scotia Gravel

The coarse gravel mass that underlies the 290 foot terrace, prominent in the Village of Scotia, is called the Scotia Gravel. It exceeds 100 feet in thickness beneath the terrace summit and was originally deposited through nearly the entire width of the Mohawk Valley. The main part of the Scotia Gravel is found east of the Hoffmans Fault, where the Mohawk Valley floor widens from 2000 feet to about 7000 feet. Several miles separate most of the western segments, which are much less extensive. In each case where a gravel mass is found, there is coincidence with a valley widening. This is not surprising, but what is odd is the apparent total absence of gravel in the narrow valley reaches.

Preliminary calculations of discharge for the Scotia Gravel transport give at least 300,000 cfs, for a flow depth of 20 feet and a 7000 foot width. Water velocity was about 3.2 fps.

Scotia Channel

Following deposition of the Scotia Gravel, a lessened discharge cut the Scotia Channel to a depth of 35 feet and a width of 2000 feet. In the narrow reaches of the Mohawk Valley the difference in width between the channel and the valley walls that carried the Scotia Gravel discharge is small. On the wider valley floor, the 290 foot scotia Gravel Terrace was incised sharply and the gravel mass was segmented into terraces and subsidiary channels. Examples are the Rotterdam, Glenville, and Hoffmans Channels. These channels are all related to the Scotia Channel, but are about half the size.

At Schenectady, the valley floor includes a large elbow south of the river. The outside valley wall is the eroded back side of the Lake Albany Schenectady Delta. A sweeping path of Scotia Channel migration is seen here. It is not certain how much scour and fill the channel migration accomplished, but it is interesting to note the Schenectady and Rotterdam well fields, located in that strath, produce from gravel and sand that appears less well sorted compared with other aquifer producing zones. A similar stratigraphy is seen at the Rotterdam Junction well field where up to 50 feet of silty and sandy gravel overlies the coarse (Scotia) gravel production zone. If these upper, more poorly sorted units are Scotia Channel fill rather than older Scotia Gravel, then an enormous volume of coarse gravel has been carried downstream through the various channels that dissect the Albany lake plain.

Discharge for the Scotia and related channels is calculated at about 220,000 cfs. The St Lawrence at Ogdenburg is rated at 295,200 cfs and the Niagara at Buffalo at 203,000 cfs. The Mohawk at Little Falls averages 2799 cuf, and at Cohoes (which includes the Schoharie) averages 5747 cfs. Significant hlaups from the western Mohawk basing may have contributed to the overall Mohawk flood history, but it appears that "Great Lakes" discharge involving Iroguis and other large drainages are responsible for the movement of the coarse gravels and their subsequent erosion.

Flow nets drawn in parts of the aquifer suggest that groundwater underflow down-valley is vigorous and only partly controlled by the river level, which in summer lies pooled between locks. Where the canal occupies an outside meander bend, flooding scour may keep the bottom clean and allow surface water to recharge the aquifer, whereas in the long straight reaches between bends the canal should be more silty bottomed and less connected to the aquifer. Winslow and others (1965) suggested the turbulence just downstream from Lock 8 is helpful in keeping the bottom clean enough to allow easy infiltration to the Rotterdam/Schenectady aquifer during pumping. There may be alternating gaining and losing reaches of the canal during the navigation season, for reasons of channel pattern irrespective of local effects of lock action. In any case, about 30 million gallons of groundwater per day flow through Rotterdam Junction segment of aquifer in summer, unimpended by pumping which is a small fraction of that total.

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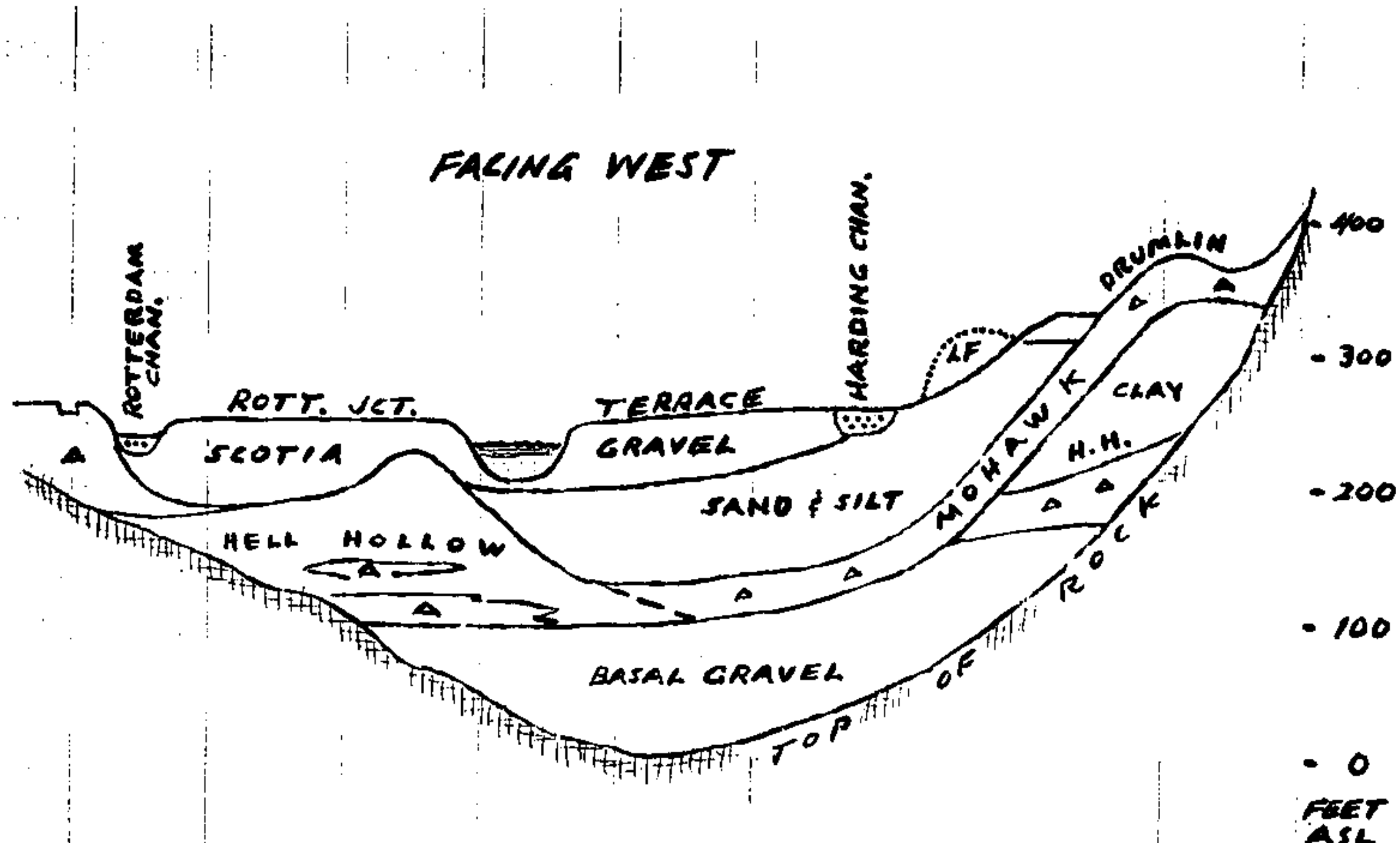
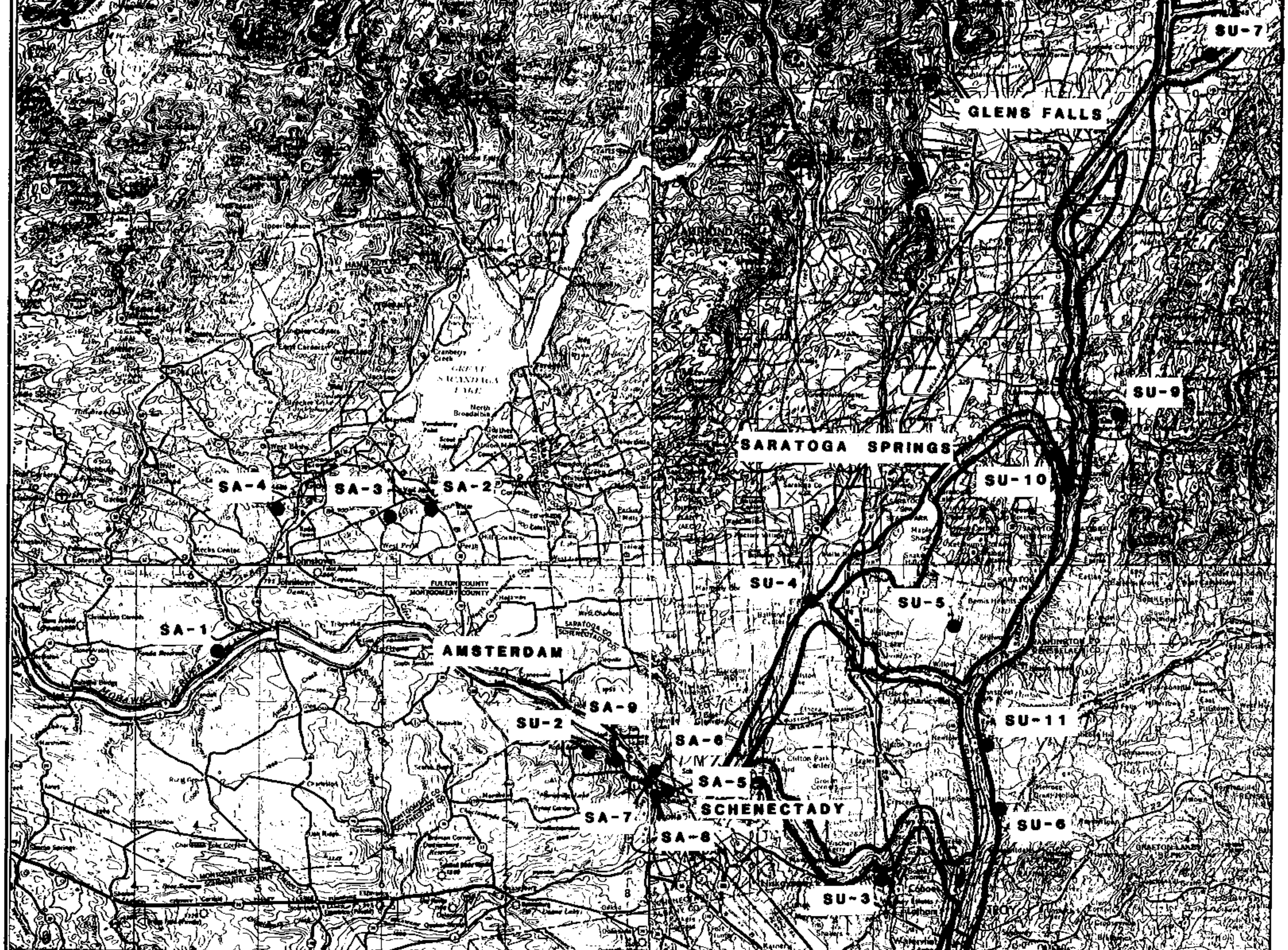


FIGURE 22

MOHAWK CHANNEL CROSS-SECTION



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