ACTIVE AND STAGNANT ICE RETREAT: DEGLACIATION OF CENTRAL NEW YORK

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INTRODUCTION

The depositional landforms of glacial origin on the Appalachian Plateau of central New York are atypical of classic continental environments. They consist of an unconventional landform assemblage formed by retreating ice-tongues at the margin of the Laurentide Ice Sheet. Landforms representative of backwasting occupy most through-valleys,while widespread stagnation and downwasting occurred in non-through valleys. A modified ice-tongue model includes valley tongues 20 km long and emphasizes the significance of inwash as a primary source of sand and gravel, and subglacial sediment flow within tunnels as the major source and transport mechanism for silt.

Previous work

This region received little published attention between Fairchild's 1925 description of the glacial landscape (kame and kettle topography, pitted plains, terraces, proglacial lakes and hanging deltas) and Coates' definition of the till shadow concept in 1966. Woodfordian facies of "bright" and "drab" drift were the subject of study in the Binghamton/Elmira area, (Denny, 1956; Moss and Ritter, 1962; Coates, 1963), but consideration of depositional environments along the main Susquehanna Valley remained unreported until described by Fleisher (1977a) and Melia (1975). Fleisher (1984, 1985) discussed the distribution of landforms as ice-marginal indicators and proposed an upper Susquehanna Lake chain, including glacial lakes Cooperstown, Davenport, Middlefield, Milford, Oaksville and Otego. Krall (1979) mapped and named the Cassville-Cooperstown Moraine (5 km south of the Otsego Lake at Index), which he interpreted as a evidence for readvance (circa 14,000 years BP) based on topographic correlation with moraines in the Hudson Valley. Sales, et al., (1977) provides a general summary of Otsego Lake geology, including bedrock, glacial and bottom samples. The first report of systematic sampling of lake sediments was by Yuretich (1979, 1981), who documents shallow stratigraphy, mineralogy and geochemistry in 25 short cores(30-60cm) from representative locations throughout the lake basin.

MacNish and Randall (1982) used well and boring log data to establish Quaternary aquifer properties and to provide generalized stratigraphic associations between surface data and landforms. They suggested various geomorphic and hydrologic settings to depict deglacial conditions related to rate and mode of retreat (active vs. stagnant).

The first suggestion of significant stagnant ice sedimentation was by Fleisher (1986a), who recognized anomalously large floodplain areas confined within high outwash terraces and valley train. These "mega-kettles", known as dead-ice sinks, developed in early post-glacial time by retarded downwasting of detached and buried remnant ice masses.

Regional Setting

The region is characterized by a coarsely dissected plateau of Middle Devonian strata that dip gently to the south-southwest at less than 10 degrees. Major north-south oriented valleys consist of broad, U-shaped troughs separated by wide, low-relief divides. Local topographic relief ranges between 250-300 m, but valley floor well data (Randall, 1972) indicate bedrock relief is considerably greater, averaging 300 m but reaching 400 m in places.

The Susquehanna River flows south, then southwest through a glacially modified, deeply incised, meandering valley (Figure 1). This valley and its strongly asymmetric tributary pattern of south-flowing streams are considered vestiges of a pre-glacial drainage system of great antiquity (Fleisher, 1977b, Sarwar and Friedman, 1990). Larger north-south oriented valleys are invariably asymmetric, with steeper slopes facing west and more gently inclined slopes to the east, from which the majority of tributary runoff originates. This topographic configuration is completely unrelated to bedrock strike and dip. While the cause of the valley asymmetry is not well understood, it appears to have played a significant role in late-glacial deposition of inwash-sourced sediment.

GLACIAL DEPOSITION

Upland diamict on divide areas is generally thin, ranging from a discontinuous veneer to 5 m, except in till shadows in the lee of divide ridges and hills, where test borings indicate thicknesses reach 40-60 m (Resource Engineering, 1986). Valley floor deposits consist of sand and gravel interstratified with sand, extensive silt and some clay in thicknesses that commonly reach 70-120 m. By correlation with deposits in adjacent drainage basins, the chronology is interpreted to be Woodfordian and range from 16,000 to14,500 years BP (Cadwell, l972b). Fleisher (1987) interpreted stratigraphy and deglacial landforms to be consistent with single-stade development. Consistent with this are generalized diagrams of MacNish and Randall (1982) that suggest single-stade settings based on regional well data and general topographic expression.

Listed below are landform assemblages for through and non-through valleys. These are interpreted to indicate active ice retreat (backwasting) in through valleys and stagnation (downwasting of detached ice masses, as well as ice-tongue collapse) in non-through valleys.

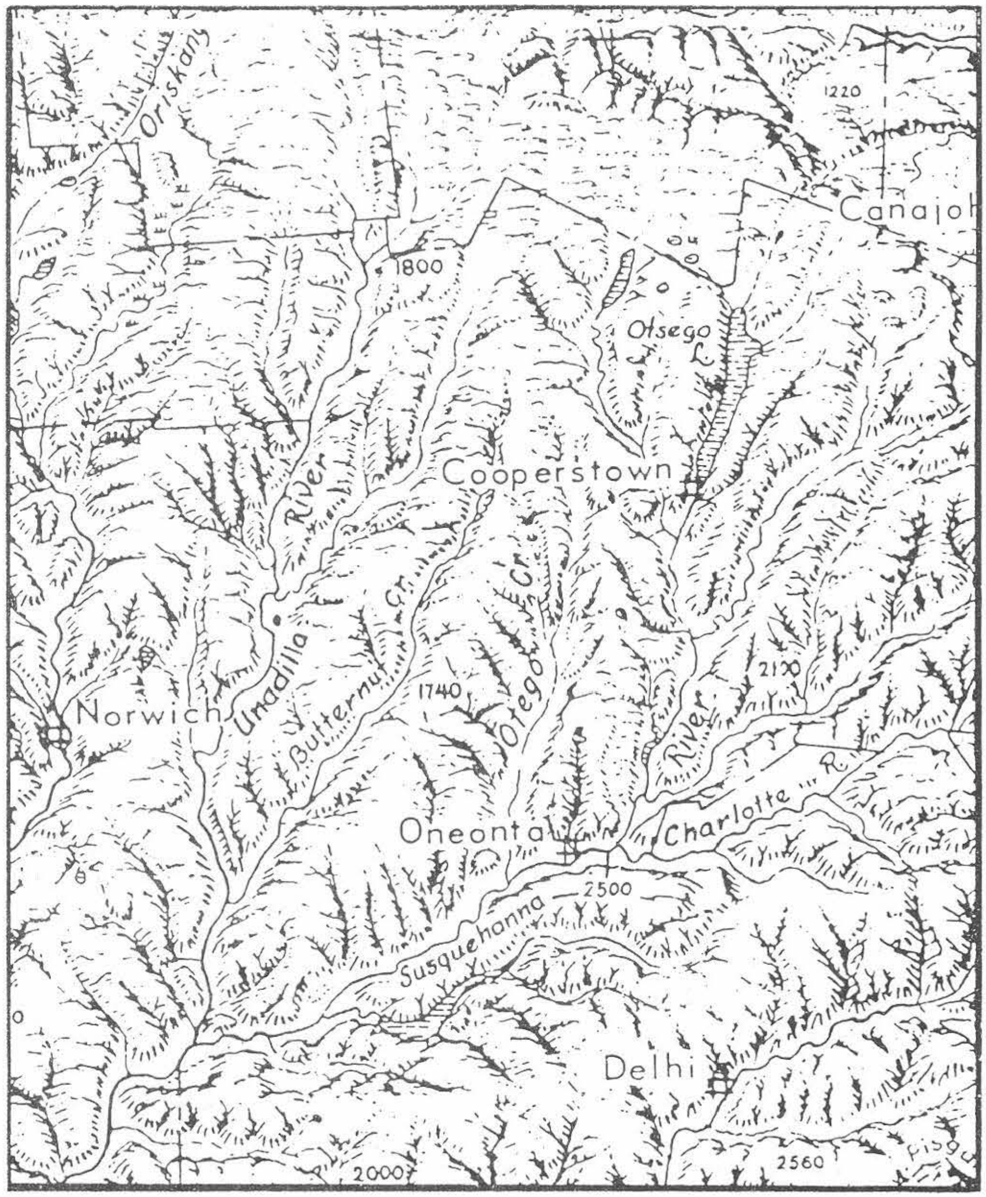

Figure 1 Index Map

THROUGH VALLEY ASSEMBLAGE	NON-THROUGH VALLEY ASSEMBLAGE
(produced by backwasting)	(produced by downwasting)
kame moraines/valley train	kames and kame fields
high gravel terraces (deltaic)	discontinuous gravel plain remnants
lacustrine plains	dead-ice sinks
kame fields	eskers
dead-ice sinks	

Kame moraines general occur in association with valleys trains that grade upvalley to the base of the moraine. A conspicuous kame and kettle topography, with local relief of 10-15 m, identifies moraines. Most moraines still occupy the full valley width, although breached by modern drainage and rise 25-30 m above the floodplain. They are poor morphostratigraphic units owing to the loss of topographic expression on valley walls and are essentially absent from upland areas. The moraine/valley train landform association is interpreted to be indicative of an active ice-marginal position (actively flowing ice) because it is assumed that a high-discharge, hydrological connection with meltwater from the main ice sheet was required to aggrade thick sand and gravel in the form of a massive valley train. Local, clusters of small scale ice blocks were occasionally incorporated within some gravels as indicated by the pitted nature of valley trains. The cause of ice block concentration in some areas and not others remains to be determined, but may be related to clusters of grounded ice-islands and bergs in emptied proglacial lakes across which inwash accumulated. This may account for why 20 m of pitted outwash lies over 110 m of silt in the vicinity of Milford Center on the Susquehanna near Oneonta.

Paired and non-paired "kame terraces" are common and typically contain conspicuous deltaic foreset and topset bed. They stand 25-30 m above the floodplain and are graded remnant patches of other planar gravel deposits and hanging deltas. These features are commonly referred to as kame terraces, but actually appear to have originated by meltwater streams discharged from lateral englacial tunnels where ice-tongues were in contact with proglacial lakes. In this sense, they are deltas formed by continuous aggradation during slow, active ice retreat.

Lacustrine plains occur commonly throughout the region and occupy large segments of many through and non-through valleys. They are underlain by deposits of silt and "quicksand" (driller's term for saturated coarse silt and fine sand that liquifies when vibrated) reported in driller's logs to exceed thicknesses of 100m (Randall, 1972). However, the sparse occurrence of well developed hanging deltas and paucity of other strandline features seem inconsistent with the extent of lakes suggested by thick silts and extensive lacustrine plains. Perhaps this indicates very rapid rates of lacustrine sedimentation in short-lived lakes.

Kame fields are found more commonly in non-through valleys, although not exclusively. They are of limited lateral extent (seldom exceeding 1-2 km across), with local relief similar to that of moraines, yet found independent of valley train. Invariably, kame fields are found at the confluence of upland tributaries and main valleys. As a landform of ice-cored origin, downwasting of remnant ice masses appears to be the primary mode of deposition.

Another landform of ice-cored origin is the dead-ice sink. It's presence is apparent only when viewed in the context of adjacent landforms. Conceptually, it is an exceptionally large kettle that occupies the full valley floor width. Although formed in much the same way as a kettle, sinks are many times larger. They appear as anomalously broad floodplain areas (3-4 km in diameter) bounded upvalley and down by terraces, valley train, kame fields or moraines. Implicit to their occurrence is the burial of large stagnant ice mass remnants of collapsed ice-tongues or detached from the terminus of retreating active ice tongues. Downwasting ultimately creates a valley-floor depression that serves as a sediment sink during late-glacial time, or post-glacial time if melting is sufficiently retarded.

Very few eskers are know within the upper Susquehanna. However, those that have been mapped all are short (less than a km) and of low relief (10-14m). They typically occur as part of the non-through valley landform assemblage and are assumed to indicate stream deposition beneath dead-ice.

VALLEY ICE-TONGUE MODEL

The valley ice-tongue model, proposed by Moss and Ritter (1962), was used by Cadwell (1972a) to correlate the valley facies of ice-marginal positions with isolated stratified drift on adjacent slopes in the Chenango drainage basin. Fleisher (1977a) applied ice-tongue depositional environments to account for the common occurrence of deltaic sand and gravel kame terraces and explain patterns of ice-marginal landform distribution within valleys of the Susquehanna drainage basin. MacNish and Randall (1982) proposed rate of ice-tongue retreat as a controlling factor in the development of landforms and stratigraphy. Although discussion is limited to descriptive generalities, their schematic diagrams clearly relate rate of retreat to types of deposits. Slow retreat favored the formation of deltaic kame terraces in much the same way described by Fleisher (1977a). Hesitation during retreat permitted aggradation to occur across full valley width to form a recessional moraine, although no specific moraine forming process (i.e. sediment source or transport mechanism) is identified. Conversely, rapid retreat precluded lateral accretion by outwash streams and glaciofluvial sand and gravel was interbedded with lacustrine sediments. A completely different set of landforms, including eskers, and heterogeneous stratigraphic units of limited lateral continuity are attributed to stagnant ice conditions.

Shortcomings to the earlier ice-tongue model

Although the earlier valley ice-tongue model has been useful, it fails to account for several curious aspects of regionally extensive deposits and does not portray sufficient details pertaining to depositional environments. For example, the current model does not address the following observations:

1. landforms in through valleys differ significantly from those in non-through valleys
2. kame-moraines
   a. lose topographic expression on valley walls and across divides
   b. consist of crudely sorted and stratified sand and gravel, not till
   c. are graded to thick valley train through valleys
   d. are difficult to correlate from one valley to the next
3. thick silts occupy valleys that lack well developed strandline features, such as hanging deltas and beach deposits

Furthermore, it is difficult to apply this model to deglacial conditions that would account for:

1. an adequate source of coarse and fine sediments
2. sediment transport mechanisms to account for the contrast between stratified valley deposits and till-like upland drift
3. a single-stade environment that would result in 20 m of pitted sand and gravel over 80 m of silt (lacustrine ?)

While the ice-lobe model is helpful in understanding the distribution of glacial landforms within valleys and provides a conceptual model to guide regional correlation Fleisher (1986b), it does not apply well to upland areas where end and recessional moraines are virtually lacking. Therefore, the valley ice-lobe model must be modified to better accommodate field evidence.

Modified ice-tongue model

As a starting point in the modification of the ice-tongue model, it was assumed that "the glacier profile is related to the hydraulic and strength properties of potentially deformable bed material" (Boulton and Jones, 1979). In central New York, the subglacial zone of the Laurentide ice sheet would have been charged with saturated, fine sediment of low permeability derived from the erosion of lower Paleozoic shales and siltstones as the ice moved southward onto the Appalachian Plateau and into the Susquehanna drainage basin. As indicated by Boulton and Jones, when "bed transmissibility is low, water pressure builds up, the bed begins to deform, and a lower equilibrium profile will develop". Proglacial, ice-contact lakes into which subglacial meltwaters discharged during deglaciation would detain free downvalley flow, thereby contributing to the development of positive pore pressure build up within saturated bed material. This in turn reduced basal shear strength, facilitated deformation and resulted in lower ice-tongue gradients.

Two methods of gradient calculation were considered. The Mathews (1974) method is summarized by the general expression,

$$h = A\,x^{0.5}$$

where h = ice thickness, A = a coefficient that varies with different values of basal shear stress, and x = distance from glacier terminus.

The Ridky and Bindschadler (1990) method is based on Nye (1952) and Hughes (1981) and involves a much more sophisticated approach, which was used to develop ice thickness profiles and flowlines for Late Wisconsinan ice in the Finger Lakes region of central New York. Their calculations, based on basal shear stress values of 0.5, 1.0, and 1.5 bars, compare favorably with profile data generated by using the general Mathews equation (Table 1).

Table 1 shows how these gradients may be used to estimate the length of ice-tongues that rise gradually upvalley before joining the ice sheet margin on adjacent divides. Because local bedrock relief (from beneath valley fill to outcrops on divides) averages 300 m, the length of an ice-tongue would be approximated when ice thickness equals bedrock relief, as shown in Table 1. General agreement between ice-tongue height values that span 300 m, as determined by both methods of calculation (at comparable basal shear stress), suggests limited extrapolation to shear stress values less than 0.5 bars is valid. This technique suggests ice-tongues may have reached lengths in excess of 20 km.

TABLE 1. ICE-TONGUE SURFACE GRADIENTS

	Mathews*					Ridky & Bindschadler**		
	Basal shear stress (bars)							
	0.10	0.30	0.50	1.00	1.50	0.50	1.00	1.50
Distance from terminus (km)	Height of ice surface (m)							
2	57	100	134	182	<u>223</u>	142	204	<u>251</u>
4	81	141	189	<u>257</u>	<u>314</u>	191	<u>278</u>	<u>345</u>
6	100	172	223	<u>314</u>	385	221	<u>327</u>	409
8	115	199	268	363	445	294	417	511
10	129	223	<u>287</u>	406	498	<u>295</u>	432	538
12	141	244	<u>314</u>	445	545	<u>397</u>	547	663
15	157	<u>273</u>	352	498				
20	182	<u>314</u>	423	575				
30	223	385	518					
40	257	445						
50	<u>287</u>							
60	<u>314</u>							

Underlined sets of data indicate height values that bracket 300 m, which is required to exceed local bedrock relief of divides above valley floor. Each set corresponds to a predicted range of ice-tongue lengths.

* Mathews method

$h = A x^{0.5}$, where h = ice thickness, A = a coefficient that varies with different values of basal shear stress, and x = distance from glacier terminus.

** Ridky and Bindschadler Method
(see original reference for derivation of formula)

These calculations assume specific quantitative conditions that may be more precise than accurate for the purpose of developing a conceptual working model. Therefore, it should be emphasized that the most important aspect of the modified model is the extensive ice-free uplands on which inwash processes could function, not the finite interpretation of ice-tongue length (Figure 2).

An additional significant modification of the ice-tongue model involves the dynamics of longer ice-tongue movement leading to stagnation on the scale of an entire ice-tongue (20 km long). Ice-tongue stagnation has been proposed as the result of the progressive development of a negative ice budget within through valleys caused by restricted internal flow within thinning ice on headward divides (Fleisher, 1986a). Figure 3 (schematic longitudinal profiles) illustrates how ice-tongue starvation developed in non-through valleys, whereas active ice movement within through valleys (open-to the-north) was sustained by continued internal flow.

The occurrence of dead-ice sinks in through valleys implies still another cause for stagnation that involved the detachment of large ice masses (a few km in diameter) from the ends of actively flowing ice-tongues during backwasting, as indicated by Fleisher (1986a). A possible detachment mechanism, proposed by Mulholland (1982) suggests *"tongues of ice reached a critical level of reduced compressive strength after which they could no longer transmit directed basal shear stress. When this critical thickness was achieved, a line of failure would develop some distance up glacier, separating thick, strong, active ice from thin, weak, stagnant ice"* (see Figure 4).

The terminus of the Bering Glacier, Alaska, contains such structures and provides a modern analog for this detachment mechanism. Here, low angle shears rise from within the glacier to separate tabular plates of ice a few meters thick (Figure 5). Similar remnant ice plates are buried beneath several meters of outwash gravel and lacustrine sediment 2 km from the glacier terminus. Here,ice foliation and multiple shear planes are oriented semi-parallel to those observed in the Bering (Figure 6). This suggests detachment occurred during retreat from a 1966 surge terminus (Fleisher, 1991).

SEDIMENT SOURCES

Evenson and Clinch (l987) document the significance of meltwater in moving sediment from upvalley sources to the glacier terminus. Their study identifies specific mechanisms responsible for down-glacier transport of inwash derived from ice-free tributary valleys, ice-dammed lakes, slope processes and re-worked older deposits, as well as inwash from tributaries beyond the existing ice limit.

The modified ice-tongue model proposed here combined with the inwash concept advanced by Evenson and Clinch lead to mechanisms of sediment transport that account for the origin of all sediment associated with Laurentide deglaciation. Central to the modified model are ice-tongues 20 km long within valleys between ice-

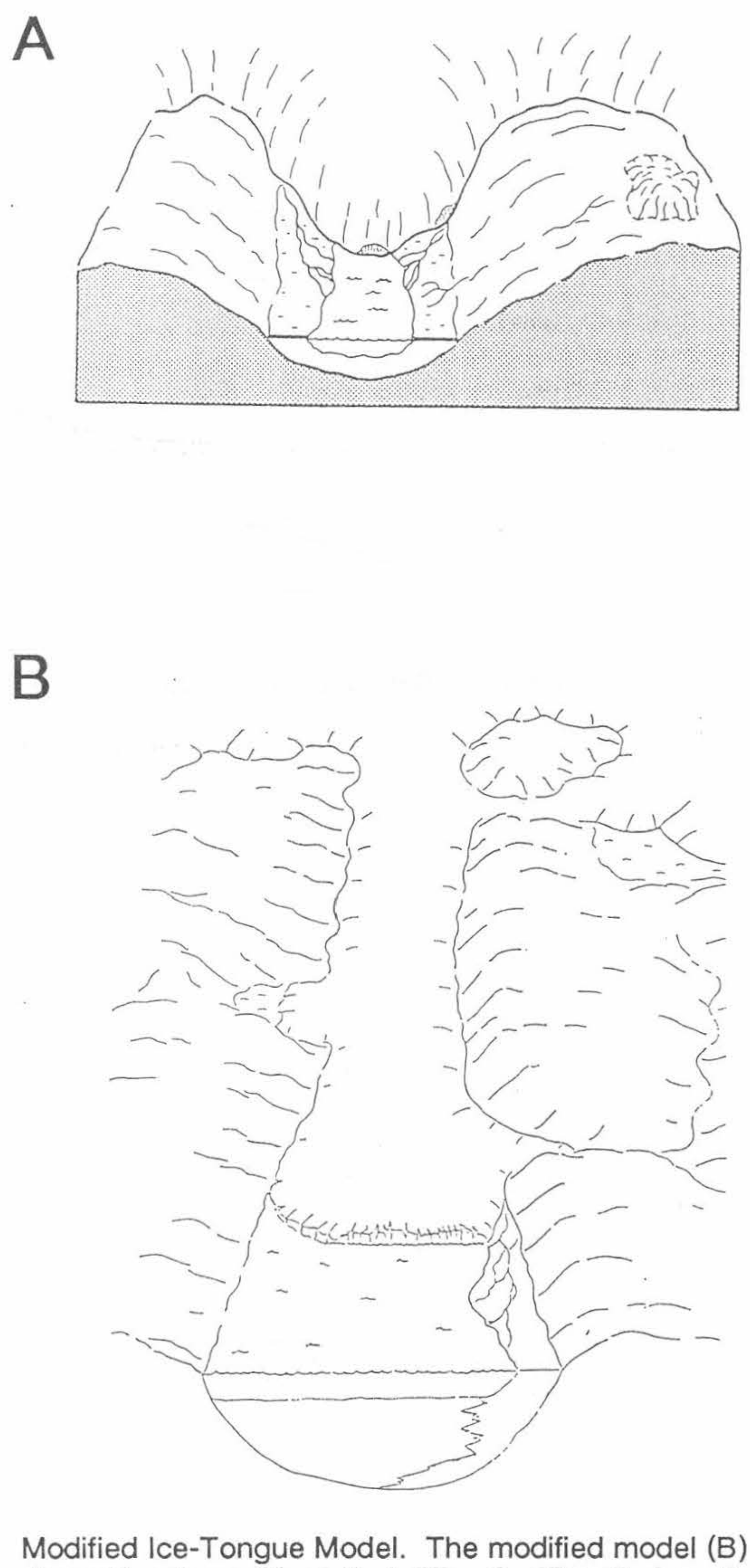

Figure 2 Modified Ice-Tongue Model. The modified model (B) depicts depositional conditions that differ significantly from those of the earlier model (A) and includes several possible sediment sources and debris transport mechanisms.

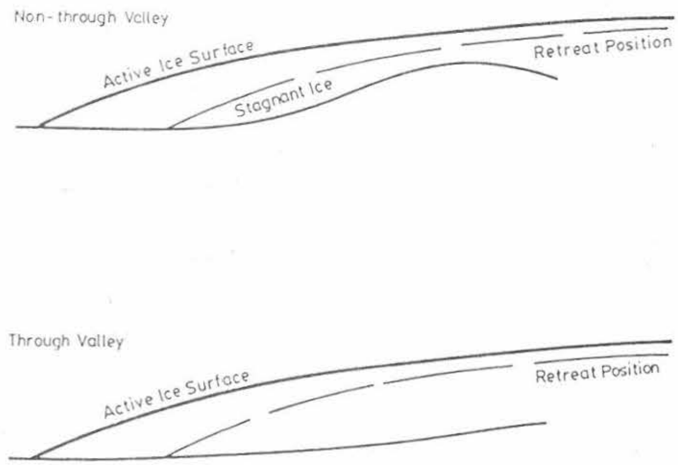

Figure 3 Schematic longitudinal profiles

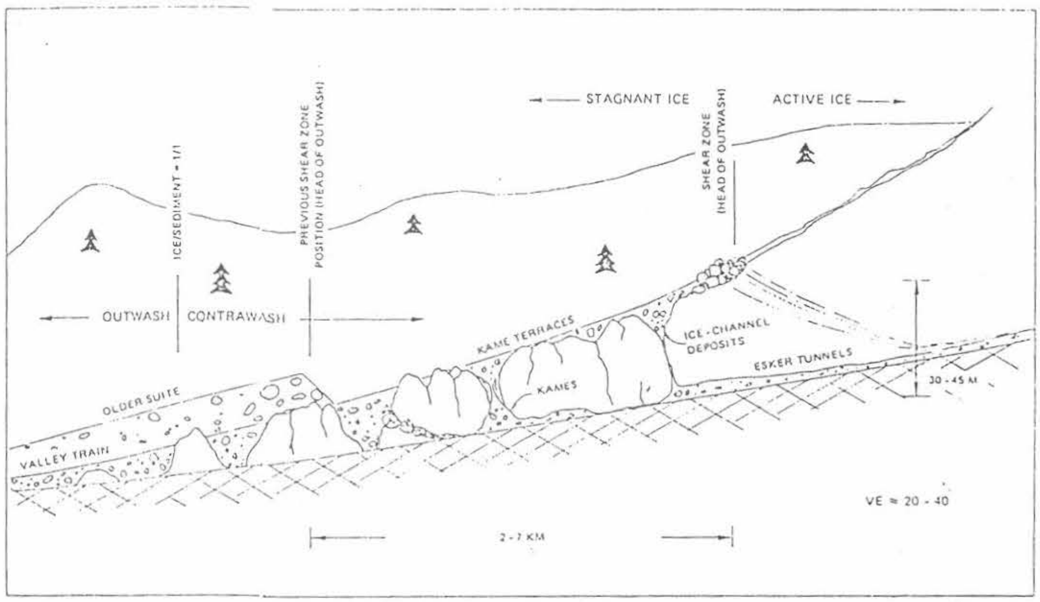

Figure 4 Detachment of stagnant ice masses (from Mulholland, 1982)

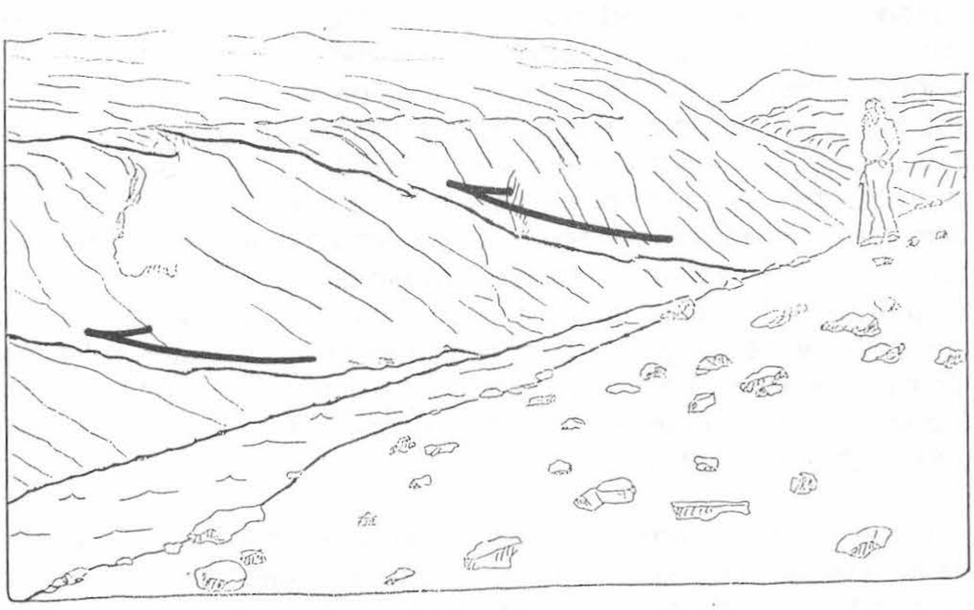

Figure 5 View up-glacier of rising shear planes on vertical canal wall at the Bering Glacier terminus.

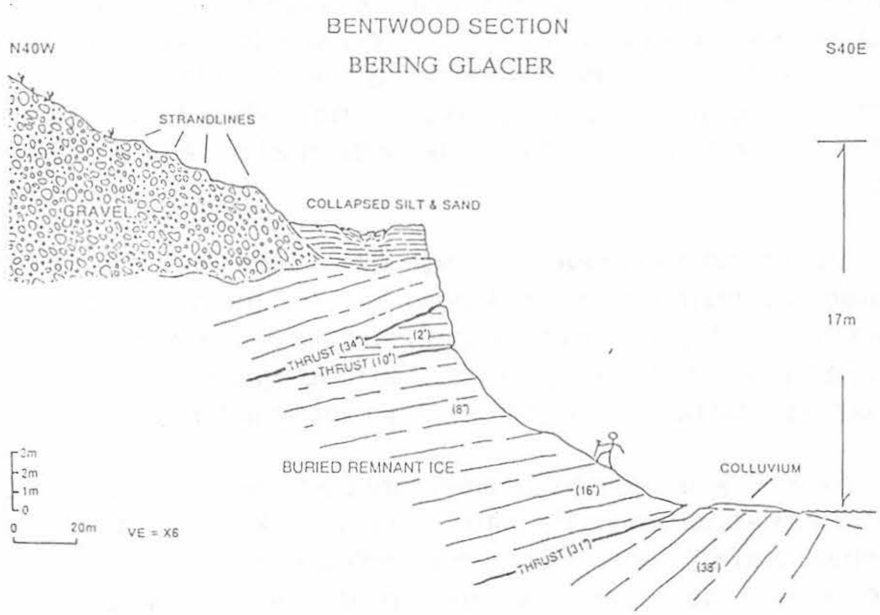

Figure 6 Shear planes separate four tabular ice plates (Bentwood Section). Foliation in each plate dips in the upglacier direction and is truncated by shears. Overlying sediment includes lacustrine silt and sand disturbed by subsidence as remnant ice below melts (from Fleisher, 1991).

free uplands and deglaciated tributaries. Consequently, bouldery till originally deposited in tributary valleys was re-worked and transported to the surface of adjacent ice-tongues, where re-sedimentation sorted fine from coarse material. Downglacier transportation by supraglacial streams fed accumulating deltas in ice-contact proglacial lakes. Coupled with uniform, slow retreat this process formed continuous lateral deltas that aggrade upvalley as space is made available during retreat. These join with kame terraces deposited by associated streams along the glacier/valley wall interface. Fluvial transport of tributary inwash directly into a proglacial lakes formed deltas at grade with deltaic terraces. Active ice-tongue flow concentrated supraglacial inwash at stable ice margins to form moraines and associated valley trains, whereas less active flow allowed debris to remain concentrated in areas near tributary confluences. Where protected from supraglacial streams, such deposits formed kame fields through downwasting.

While tributary inwash seems to be an adequate source of coarse sediment, a comparison of sediment volume to upland area suggests the inwash process could not have yielded sufficient fines to account for the volume of fine sand and silt within the valley fill. Uplands are mantled by silt-rich lodgement till that is generally thin, except in till shadows where it reaches thicknesses in excess of 40-60 m. To test the possibility that upland drift may have served as a significant source of fines, the volume of fines in the valleys was compared to the area of adjacent uplands.

Well data from the upper Susquehanna and all major tributaries indicate the valleys contain approximately 58 cubic km of fine sediment fill. Distributing this volume as a uniform blanket over the entire upland area (4455 sq. km) would add l6 m to the existing till mantle. This amount of upland erosion in rills, gullies, mass wasting and sheet wash would have produced conspicuous evidence of dissection, which does not exist. Furthermore, winnowing fines from that much till would produce widespread lag gravel, which is also lacking. While these processes may account for some fines, the bulk of proglacial lake silts must have been derived from other sources.

Most sand and gravel is interstratified within fine sand and silt, which establishes contemporaneous deposition. Yet, the source of fines does not appear related to inwash sources of the sand and gravel. Assuming minimal silt from supraglacial and englacial sources, and inadequate amounts from uplands, the only possible remaining source is the subglacial sediment load of valley ice-tongues.

Extensive fine clastic sedimentary bedrock was exposed to Laurentide ice on the northern Appalachian Plateau. Here, a thick lower Paleozoic section contains 60-70% shale and siltstone. Subglacial bedload derived from these rocks would consist of fine-grained, deformable sediment incapable of draining a constant influx of meltwater. These conditions favor development of subglacial channels and conduits through which excess water would drain (Boulton and Hindmarsh, l987). As water discharge increases, so does the piezometric gradient, which in turn raises water pressure values within the sediment to equal ice overburden pressure. As Boulton and Hindmarsh suggest, this leads to sediment liquification in the terminal zone and the "flow of liquified sediment into the proglacial environment."

Similar conditions are proposed to have existed beneath Laurentide ice-tongues on the Appalachian Plateau. The saturated subglacial sediment, driven by englacial water, served the dual function of providing; l) low basal shear stress leading to low gradient, long ice-tongues and 2) large discharge of silt-rich sediment directly into ice-contact, proglacial lakes. Therefore, it is proposed this process provided a major source of fine sediment by a transport mechanism of subglacial sediment flow. Because non-through valleys were incapable of maintaining steady-state ice flow conditions, ice-tongues collapsed, downwasted and left non-time-transgressive inwash in an assemblage of dead-ice landforms (Hughes, l987; Kaszycki, l987, Mullins and Hinchey, 1989).

REFERENCES CITED

Boulton, G. S. and Hindmarsh, R. C. A., l987, Sediment deformation beneath glaciers: rheology and geological consequences, Journal of Geophysical Research, Vol. 92, No. B9, pages 9059-9082.

Boulton, G. S. and Jones, A. S., l979, Stability of temperate ice caps and ice sheets resting on beds of deformable sediment, Journal of Glaciology, Vol. 24, No. 90.

Cadwell, D. H., l972a, Late Wisconsin chronology of the Chenango River valley and vicinity, New York. Doctoral dissertation, SUNY at Binghamton, l02 p.

Cadwell, D. H., l972b, Late Wisconsin deglacial chronology in the northern Chenango River Valley: New York State Geological Association Guidebook, 44th Annual Meeting, p. D1-D15.

Cadwell, D. H., and Dineen, R. J., 1987, Surficial Geologic Map of New York, Mohawk-Hudson Sheet.

Coates, D.R., 1963, Geomorphology of the Binghamton area: in Geology of South-Central New York: D. R. Coates, ed., New York State Geol. Assoc. 35th Ann. Meeting, p. 97-116.

Denny, C. S., 1956, Surficial geology and geomorphology of Potter County, Pennsylvania: U. S. Geol. Survey Prof. Paper 288, 72 p.

Evenson, Edward B. and Clinch, J. Michael, 1987, Debris transport mechanisms at active alpine glacier margins: Alaskan case studies, Geological Survey of Finland Special Paper 3, p. 111-136.

Fairchild, H. L., 1925, The Susquehanna River in New York and evolution of western New York drainage: N. Y. S. Mus. Bull. 256, 99 p.

Fleisher, P. Jay, 1977a, Glacial Geomorphology of the Upper Susquehanna Drainage: Section A-5, p. 1-22, in Wilson, P. C. (ed.), Guidebook to Field Excursions, New York State Geological Association, 49th Annual Meeting, State University College at Oneonta, Oneonta, New York.

_______, 1977b, Deglacial Chronology of the Oneonta, New York Area: p. 41-50, in Cole, J. R. and Godfrey, L. R., (ed.), Proceedings of the Yager Conference at Hartwick College; Hartwick College, Oneonta, New York.

_______, 1984, Topographic Control of Ice-marginal Deposition and Landform Development, Upper Susquehanna Drainage Basin, in Rickard, L. V. (ed.), The State Education Department, The University of the State of New York, Empire State Geogram, vol. 20, no. 1, p. 15.

_______, 1985, A Procedure for Projecting And Correlating Ice-Margin Positions: Journal of Geological Education, v. 33, no. 4, p. 237-245.

_______, 1986a, Dead-ice Sinks and Moats: Environments of stagnant ice deposition: Geology, v.14, no. 1, p. 39-42.

_______, 1986b, Late Wisconsinan Stratigraphy, Upper Susquehanna Drainage Basin, N. Y.: in Cadwell, D. H., Dineen, R. J. (eds.), The Wisconsinan Stage of the First Geological District of Eastern New York: New York State Museum Bulletin #455, p. 121-142.

_______, 1987, Quaternary stratigraphy and landform evidence for stadial interpretation, central New York State, Geological Society of America, Abstracts with Programs, Vol. 19, No. 1, p. 14.

_______, 1990, Glacial geology of Charlotte Creek Valley and Pine Lake Archaeology Site: unpublished report, 19 p.

Fleisher, P. Jay, Mullins, H. T., Yuretich, R. H., 1990, Seismic stratigraphy of glacial Lake Cooperstown, Geological Society of America, Abstracts with Programs, Vol. 22, No. 2, p. 16.

Fleisher, P. Jay, 1991, A Revised Valley Ice-Tongue Model for the Appalachian Plateau, Central New York: Geological Society of America, Abstracts with Programs, v. 23, no. 1, p. 30.

Franzi, D. A., E. H. Muller, P. J. Fleisher, and D. H. Cadwell, 1990, Ice-Marginal Glacial Environments of the Bering Glacier Piedmont Lobe, A Possible Analog for the Late Pleistocene of New York: Geological Society of America, Abstracts with Programs, v. 22, no. 2, p. 17.

Gonsalves, Michael, Joseph Nossal, P. Jay Fleisher, and D. H. Cadwell, l99l, Ice-Contact Lake Sedimentation, Bering Glacier, AK: A Model for Late Glacial Deposition in Central New York: Geological Society of America, Abstracts with Programs, v. 23, no. 1, p. 36.

Gustavson, Thomas C. and Boothroyd, Jon C., 1987, A depositional model for outwash, sediment sources, and hydrologic characteristics, Malaspina Glacier, Alaska: A modern analog of the southeastern margin of the Laurentide Ice Sheet, Geological Society of America Bulletin, v. 99, p. 187-200, August 1987.

Halter, Eric F., Lowell, Thomas V., and Clakin, Parker E., 1984, Glacial erratic dispersal from two plutons, Northern Maine: Geological Society of America, Abstracts with Programs, V. 16, no. 1, p. 21.

Hughes T. J., 1981, Numerical reconstruction of paleo-ice sheets in Denton, G. H. and Hughes, T. J., eds. The last great ice sheets: New York, John Wiley and Sons, Inc., p. 221-261.

Kaszycki, C. A., 1987, A model for glacial and proglacial sedimentation in the shield terrane of southern Ontario, Geological Survey of Canada Contribution 35486, p. 2373-2391.

Koteff, Carl, 1974, The Morphologic Sequence Concept and Deglaciation of Southern New England in Coates, D. R., ed., Glacial Geology, SUNY-Binghamton, p. 121-144.

Krall, D. B., l977, Late Wisconsinan ice recession in east-central New York, Geological Society of America Bulletin, v. 88, p. l697-l7l0.

MacClintock, Paul and Apfel, E. T., 1944, Correlation of the drifts of the Salamanca re-entrant, New York: Geol. Soc. America Bull., v. 55, p. 1143-1164.

MacNish, R. D. and Randall, A. D., 1982, Stratified Drift Aquifers in the Susquehanna River Basin, New York: New York State Department of Environmental Conservation Bulletin 75, 68 p.

Mathews, W. H., 1974, Surface Profiles of the Laurentide ice sheet in its marginal areas, Journal of Geology, Vol. 13, No. 67, p. 37-43.

Melia, M. B., 1975, Late Wisconsin Deglaciation and Postglacial Vegetation Change in the Upper Susquehanna River Drainage of East-Central New York: Master's Thesis, State University College, Oneonta, NY, 139 p.

Merritt, R. S., and Muller, E. H., 1959, Depth of leaching in relation to carbonate content of till in central New York: Am. Jour. Sci., v. 257, p. 465-480.

Morrow, William H., Jr., 1989, Hydrogeology of the otego Creek Valley, Otsego County, New York, SUNY-Oneonta Master's Thesis.

Moss, J. H., and Ritter, D. R., 1962, New evidence regarding the Binghamton substage in the region between the Finger Lakes and the Catskills, New York: Am. Jour. Sci. v. 260, p. 81-106.

Mulholland, J. W., 1982, Glacial stagnation-zone retreat in New England: Bedrock control: Geology, v. 10, p. 567-571, November 1982.

Mullins, H. T. and Hinchey, E. J., l988, personal communications

_______, l989, Erosion and infill of New York Finger Lakes; Implications for Laurentide ice sheet deglaciation: Geology, v. l7, p. 622-625.

Nye, J. F., l952, A comparison between the theoretical and the measured long profile of the Unteraar Glacier, Journal of Glaciology, v. 2, p. l03-l07.

Powers, M. C., 1951, Journal of Sedimentary Petrology, v. 23, p. 118, in Compton, R. R., 1962, Manual of Field Geology, John Wiley and Sons, Inc., p. 215.

Randall, A. D., l972, Records of wells and test borings in the Susquehanna River Basin, New York, New York State Department of Environmental Conservation, Bulletin 69.

Randall, A. D., 1973, A Contribution to the Late Pleistocene Stratigraphy fo the Susquehanna River Valley of New York, U. S. Geological Survey, Albany, New York.

Resource Engineering, l986, Engineering report: Subsurface investigation for groundwater sources, prepared for the Village of Cooperstown #8526.

Ridky, R. W. and Bindschadler R. A., 1990, Reconstruction and dynamics of the Late Wisconsin "Ontario" ice dome in the Finger Lakes region, New York: Geological Society of America Bulletin, v. 102, p. 1055-1064.

Sales, J. K., Harman, W. N., Fleisher, P.J., Breuninger, R. and Melia, M. B., 1977, Preliminary geological investigation of Otsego Lake, in Wilson, P. C. (editor), Guidebook to Field Excursions: New York State Geological Association, SUNY-Oneonta, p.(A-6) 1-26.

Sarwar, G. and Friedman, G.M., 1990, Former presence of post-Devonian strata covering the Adirondacks; Evidence from fluid-inclusions: Geological Society of America, Abstracts with Programs, Vol. 22, No. 2, p. 67.

Yuretich, R. F., 1979, The bottom sediments of Otsego Lake: 12th Annual Report, Biological Field Station, Cooperstown, N.Y., SUNY-Oneonta, p. 40-50.

__________, 1981, Sedimentary and geochemical evolution of Otsego Lake: 14th Annual Report, Biological Field Station, Cooperstown, N.Y., SUNY-Oneonta, p. 91-108.

CASE STUDIES

The following case studies discuss examples of deglacial processes and resulting landforms at specific locations within the upper Susquehanna drainage basin.

Case #1: A Case for remnant ice, inwash and sediment source, Glacial Lake Cooperstown, Cooperstown, New York

Case #2: Remnant ice as base level controls for stratified drift aggradation: Otego Creek, New York

Case #3: Inwash sediment sources, Valley of Charlotte Creek, central New York

Case #4: Implications of pebble count data; confluence of Unadilla River and Tallette Creek, Columbus Quarters, New York

Case #1: A Case for remnant ice, inwash and sediment source, Glacial Lake Cooperstown, Cooperstown, New York

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INTRODUCTION

Otsego Lake is the headwaters for the Susquehanna River near the northern extent of the eastern Appalachian Plateau (Figure 1). The historic village of Cooperstown is situated on its southern shore and the hamlet of Springfield Center is a few kilometers from the northern shore. Otsego Lake lies within one of several through valleys in the region. To the west is Canadarago Lake, to the east Cherry Valley, all oriented N10E as part of a regional stream pattern, along which ice flow was facilitated. Local relief varies from 700' in the vicinity of Cooperstown to 900' northward. Bedrock relief is observed to be greater at the southern end of the lake where valley fill is thickest.

Oaks Creek drains Canadarago Lake and joins the Susquehanna River 4 km south of Cooperstown near the hamlets of Index, Hyde Park and Phoenix Mills. Red Creek enters the Susquehanna Valley 1.5 km south of Cooperstown from a non-through valley on the divide east of Otsego Lake. The Susquehanna Valley at Otsego Lake is conspicuously asymmetric, as are most through valleys in this region. Much steeper west-facing slopes are unrelated to the gentle southern dip of Devonian strata. Short streams drain small, first-order catchment on the eastern slopes, whereas streams off western slopes flow from second order drainage basins and cover a significantly larger area than those from the east.

GLACIAL LANDFORMS

The conspicuous kame and kettle topography at Index (5 km south of Cooperstown) is the Cassville-Cooperstown moraine, named and interpreted by Krall (1977) as the terminal deposit of a readvance here and at Cassville, 50 km to the northwest. The moraine grades downvalley into a short, deeply incised valley train. Both consist of sand and gravel, which is only slightly less well sorted and stratified in the moraine. Local relief on the moraine ranges from 7-20 m, with kettle frequency diminishing toward the valley train. Krall mapped the northwest trend of the moraine along Oaks Creek valley to Fly Creek and Oaksville, where it served as a dam for Glacial Lake Oaksville, the precursor to Canadarago Lake (Fleisher,1977). To the east, the moraine is correlated with kames in the valley of Red Creek.

The terrain upvalley from the moraine, between Index and Cooperstown, is anomalous. It consists of a unique assemblage of landforms unlike those typical of either active or stagnant ice retreat, as defined by MacNish and Randall (1982). Although the narrow floodplain here is gentle enough to support tight river meanders, the adjacent valley floor is 7-10 m higher and much too irregular to qualify as a

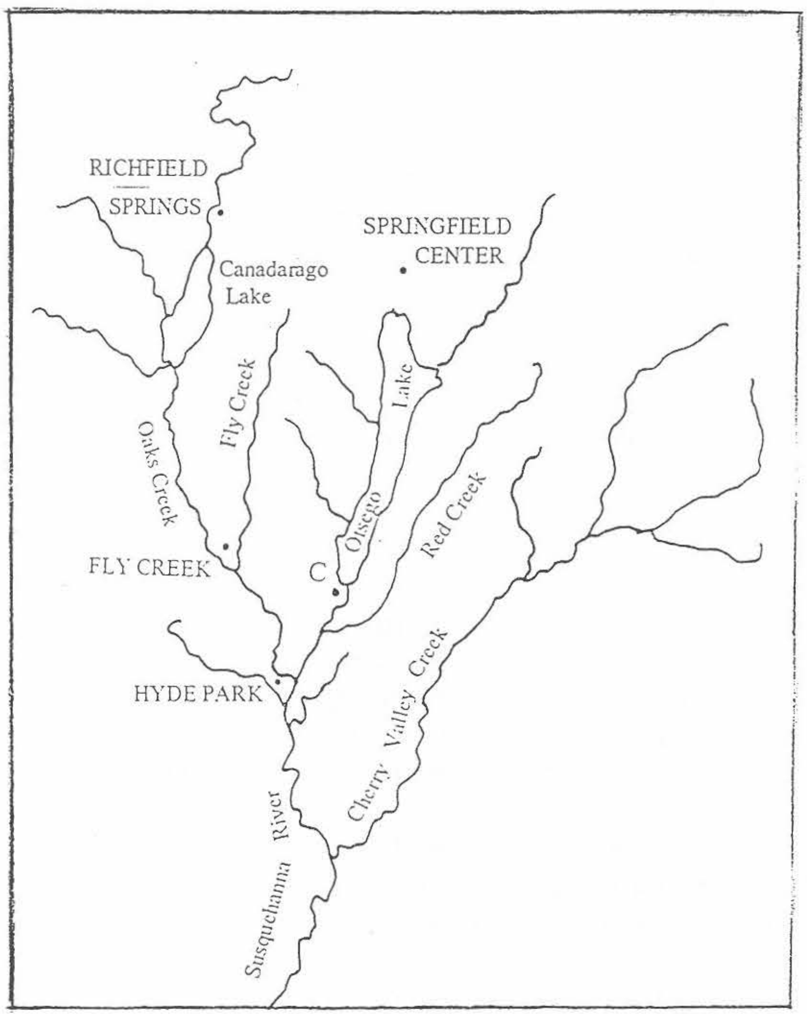

Figure 1 Index map

lacustrine plain, although composed of silt at depth. It rolls gently, suggestive of subdued terrain, but is not consistent with landforms immediately upvalley or down.

This *misfit terrain* gradually yields upvalley to gravel terrace remnants at the mouth of Red Creek and others at the mouth of a few small first-order streams on the valley walls immediately south of Cooperstown. The village of Cooperstown is situated upon the dam for Otsego Lake (possibly a moraine), referred to here as the *"Doubleday Ice Margin"*.

ON-LAND GLACIAL STRATIGRAPHY

A clear record of valley fill stratigraphy is represented by published well and boring data (Randall,1972), technical reports (Resource Engineering,1986, unpublished) and through personal communication with local water well drillers. Subsurface data shown in Figure 2 represents a semi-continuous stratigraphic record of deposits formed during retreat from the Cassville-Cooperstown margin between Index and Cooperstown.

While silt behind a moraine might suggest lake sedimentation, the lack of a lacustrine plain and the absence of strandline features (i.e. hanging deltas at tributary mouths) indicate otherwise.

FORMATION OF MARGINAL-ICE CLEAT AND DEAD-ICE SINK

Projection of a low ice gradient (consistent with a basal shear stress of 0.3-0.5 bars) northward from the Cassville-Cooperstown margin depicts an ice-tongue surface rising to the main ice margin along north-facing upland slopes north of Otsego Lake. This means the entire lake basin held an ice-tongue approximately 20 km long, while adjacent slopes were ice-free and subject to inwash processes.

Retreat from the Cassville-Cooperstown margin produced landforms and stratigraphy atypical of normal active ice retreat. In order to account for this, the concept of a *"marginal-ice cleat"* is introduced.

The cleat concept is based on field observations at the terminii of many active glaciers in Alaska, where active ice appears to rise from within the glacier on deep seated shears. Such is the case for structures along the retreating terminus of the Bering Glacier, central coastal Alaska. Here, discretely separate, semi-horizontal plates of ice, bound above and below by shear planes, form a cleat of passive ice on which active ice has risen. Several km from the retreating ice front, remnant ice plates have been detached and buried beneath foreland drift (Figure 3) during the last 20-30 years (Fleisher, 1991).

It is suggested here that similar rising structures developed during retreat from the Cassville-Cooperstown margin (Figure 4a) in a series of detachment planes (shears), as proposed by Mulholland (1982) for New England deglaciation.

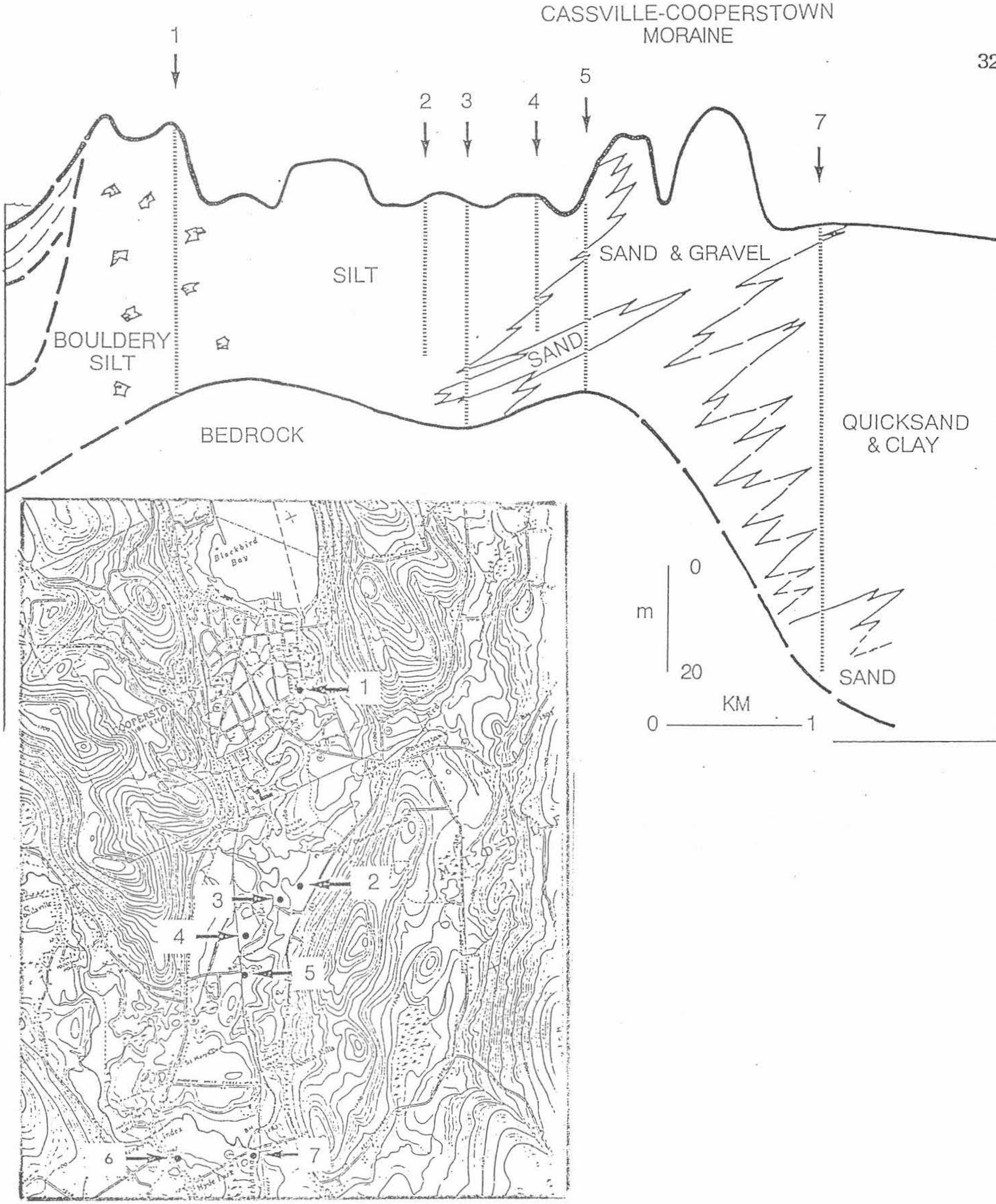

Figure 2 Well data and cross section. On-land subsurface data are derived from several locations between Cooperstown and Index, as shown on the index map. The correlation of stratigraphic units and their occurrence beneath associated landforms is illustrated. (from Fleisher, et al, 1990)

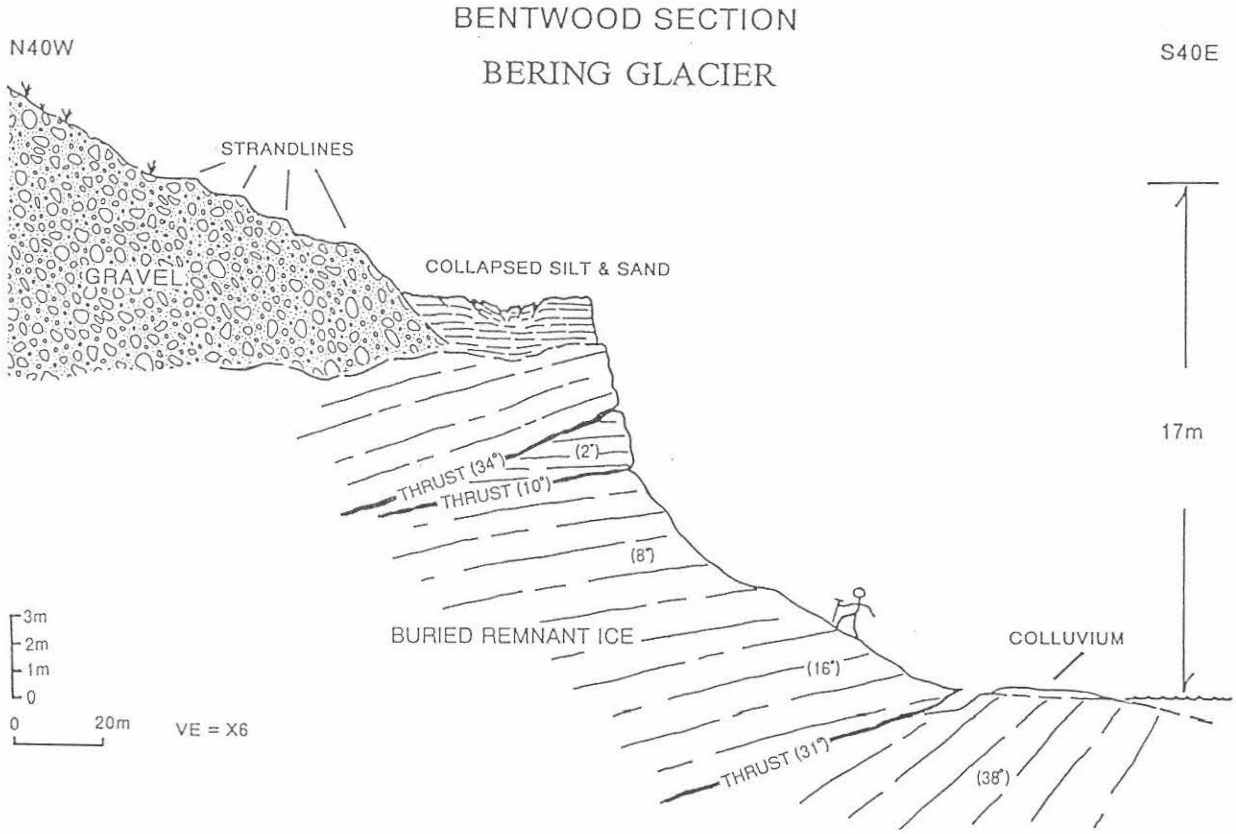

Figure 3 Detached remnant ice beneath foreland drift several km from retreating ice front. (from Fleisher, 1991)

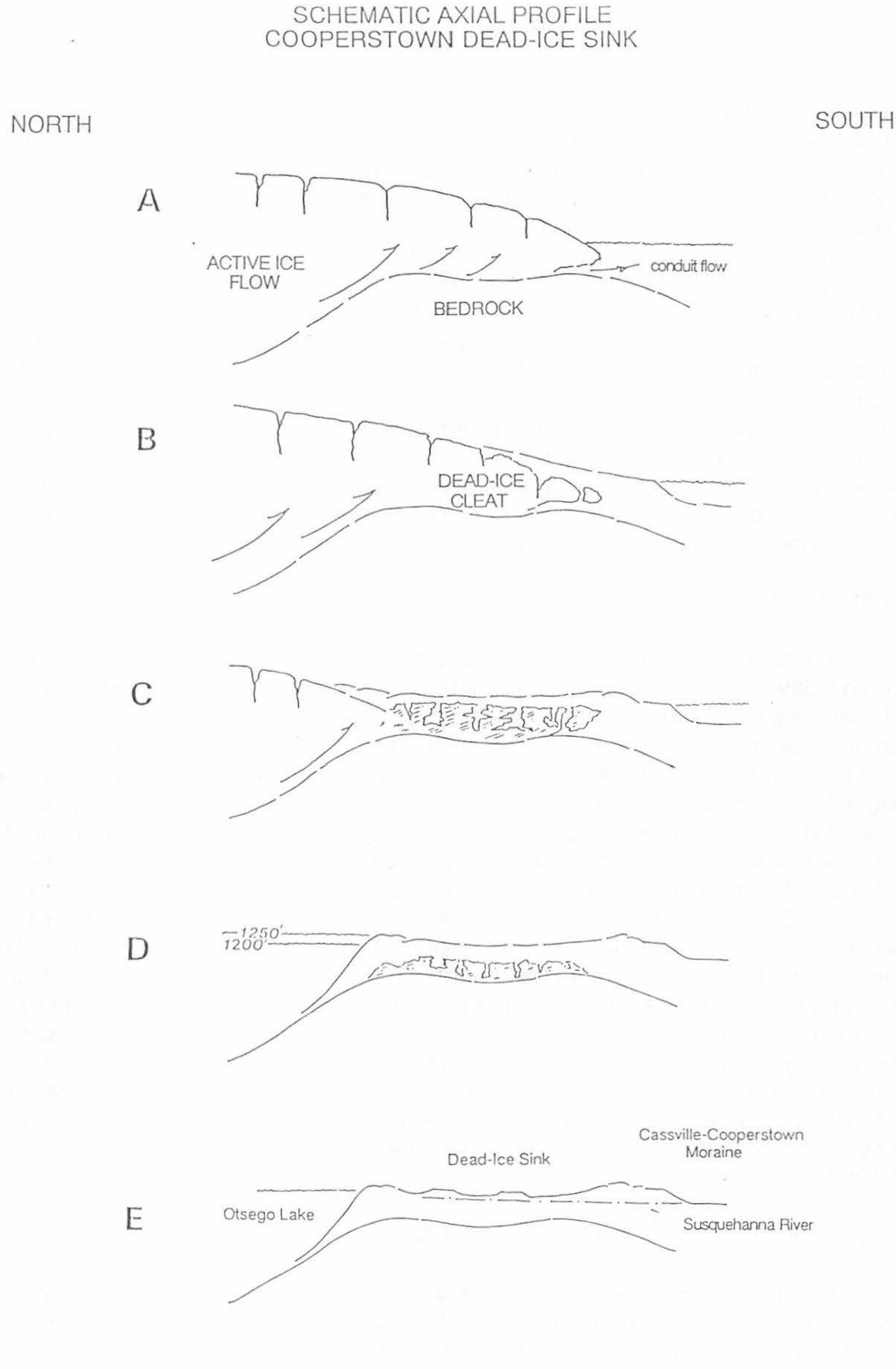

Figure 4 Dead-ice sink development. Schematic diagram of rising structures in the terminal zone of the retreating ice-tongue leads to detachment of marginal-ice cleat and subsequent development of dead-ice sink. (from Fleisher, et al, 1990)

Detachment of a marginal ice-cleat (a stack of imbricate ice plates) 65-70 m thick, 1 km (3,000') wide and 3 km (9,000') long filled the valley, as active ice rose against its upvalley side at the Doubleday margin (Figure 4B). Inwash, primarily from Red Creek and Willow Creek, spread an insulating cover across the cleat causing retarded melting and slow, gradual subsidence. Coarse sediment from these tributary sources was supplemented by silt-ladened overflow from incipient Glacial Lake Cooperstown. Local ephemeral ponds and lakes collected silt over buried ice, analogous to ice-contact lakes reported by Cadwell, et al.,1990 and Gonsalves, et al., 1991. Continuously changing water levels adjusted to new outlets in downvalley ice-cored deposits. High turbidity favored rapid silt accumulation in a constantly adjusting lake system, which precluded the development of distinct strandline features. Semi-continuous subsidence over a buried ice mass of this dimension qualifies this environment as a dead-ice sink (Figure 4C) (Fleisher, 1986). However, ongoing sedimentation filled the sink as it developed, creating a unique association of landforms and stratigraphy (Figure 4D). The final result is a valley containing 65-70 m of silt capped by inwash gravel, yet lacking lacustrine landforms (Figure 4E).

GLACIAL LAKE COOPERSTOWN

Fleisher (1977) documented incised hanging deltas and near-surface lacustrine silt and clay north of Otsego Lake as evidence for a higher lake stage referred to as Glacial Lake Cooperstown. Hanging deltas are found at the mouths of all west-slope streams, as at Brookwood Point, Threemile Point, Fivemile Point, Sixmile Point, and from the Thurston Hill and Allen Lake areas. The Cassville-Cooperstown moraine was originally assumed to have formed the dam for Lake Cooperstown at an elevation of 1250'. While this elevation has since been confirmed by strandline mapping north of the lake (Cadwell, 1988), the lack of lacustrine landforms between the moraine at Index and the modern lake shore to the north suggests the dam was actually farther upvalley. Elevations on the Doubleday Margin (at the village of Cooperstown) range between 1220-1240', whereas hanging deltas stand at a consistent 1250 feet. How can the dam be lower than the lake? Although the Doubleday Margin is not high enough now, it is suggested that at the time of original impoundment it would have been partially ice-cored and, therefore, higher. As buried ice slowly melted, surface elevation gradually diminished and the dam was eventually breached. A conspicuous alluvial fan at the mouth of Willow Creek on the western end of Doubleday Margin indicates inwash kept pace with subsidence and forced the spillway eastward to its present location where incision ultimately lowered lake level to the modern elevation of 1200 feet.

SEISMIC INVESTIGATION OF OTSEGO LAKE

Complementing on-land evidence and further documenting the depositional history of Glacial Lake Cooperstown are the high-resolution seismic reflection profiles obtained by Mullins and Hinchey (1988) along twenty-one east-west oriented transverse profiles one north-south oriented longitudinal (axial) profile (Figure 5).

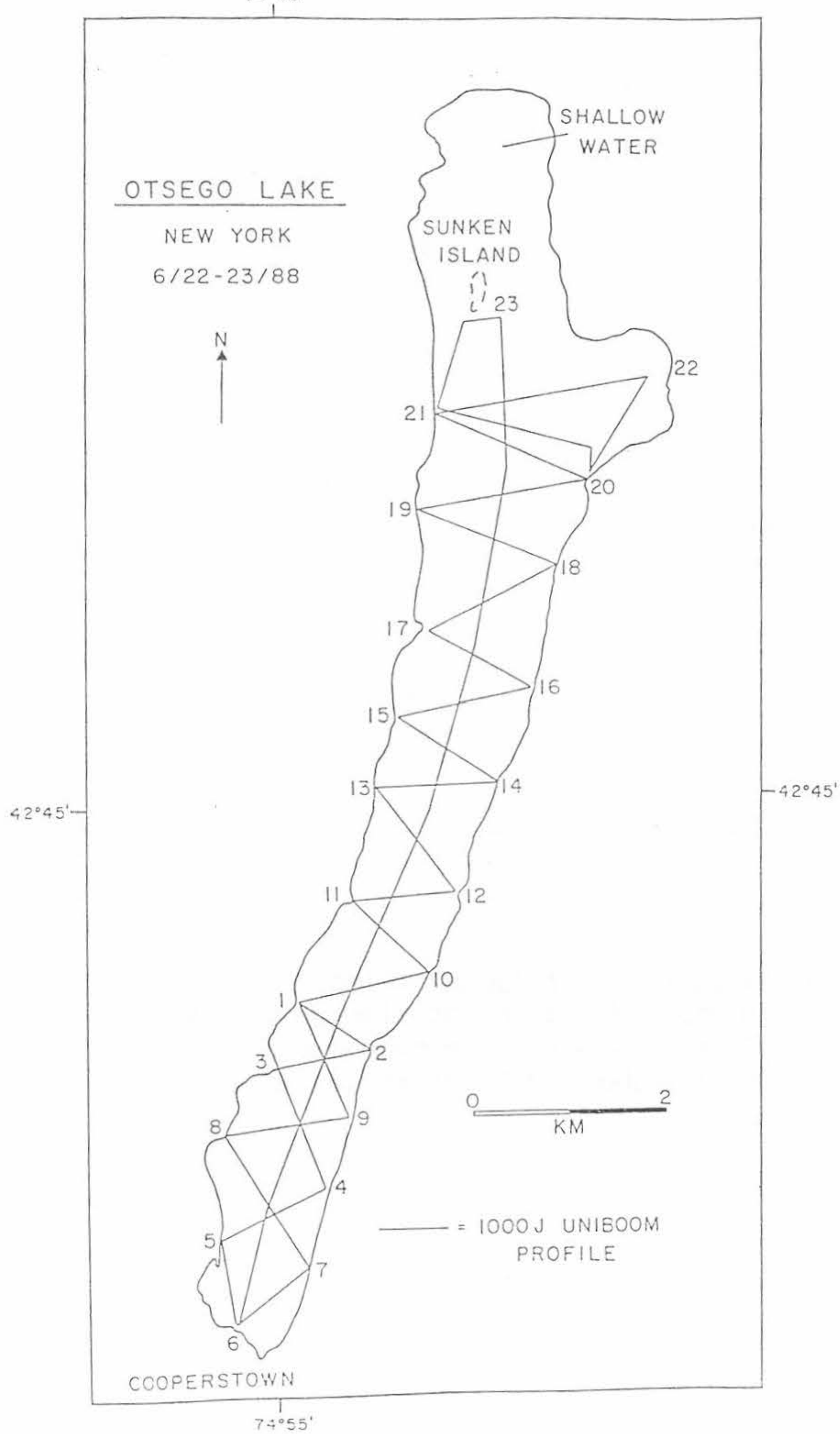

Figure 5 Index map of seismic traverses. Trackline map illustrating location of uniboom seismic reflection profiles. (from Fleisher, et al, 1990)

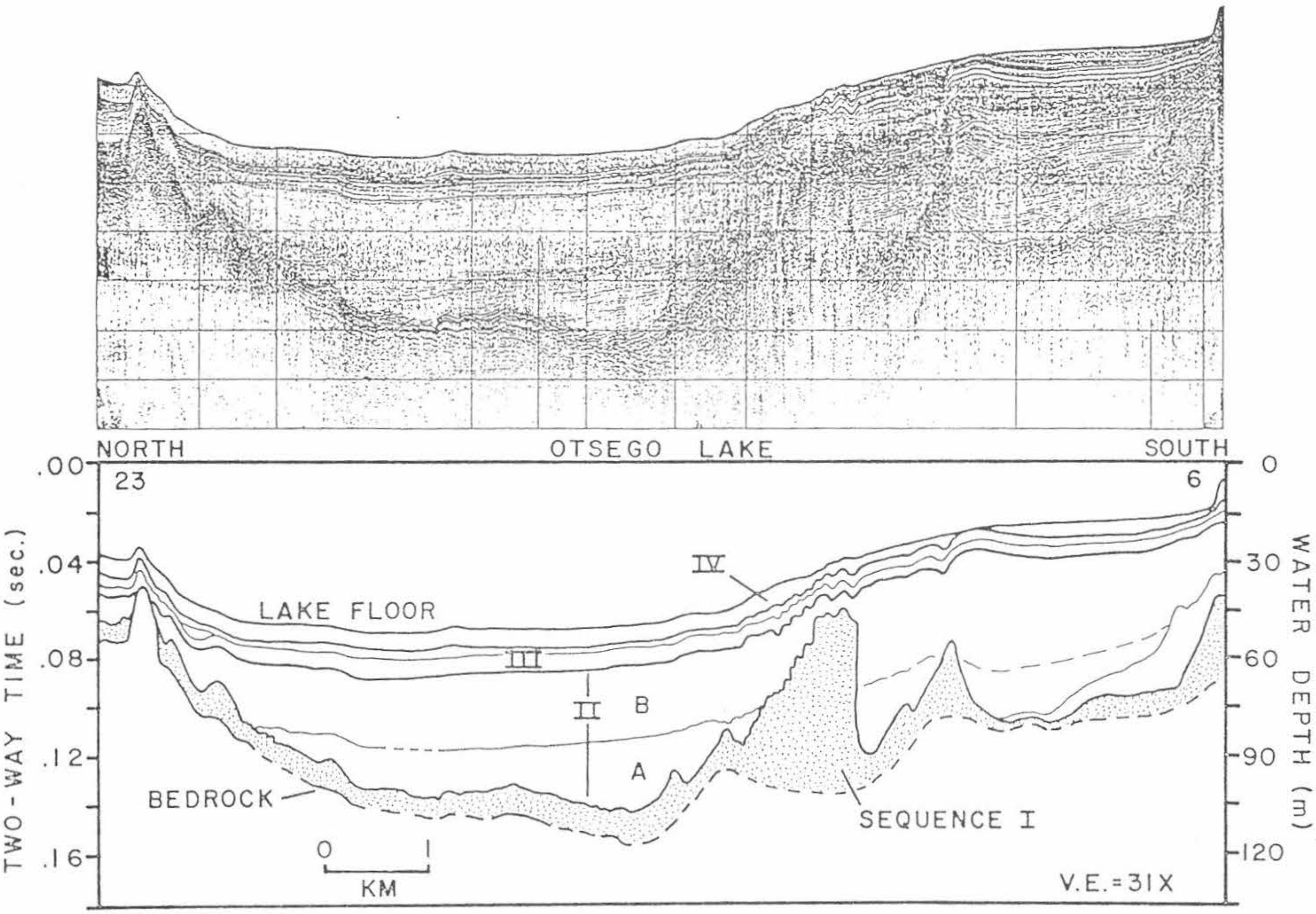

Figure 6 Axial profile. Photographic (top) and line drawing interpretation (bottom) of axial seismic reflection profiles from Otsego Lake. Roman numerials label depositional sequences. Vertical exaggeration = X31. (from Fleisher, et al, 1990)

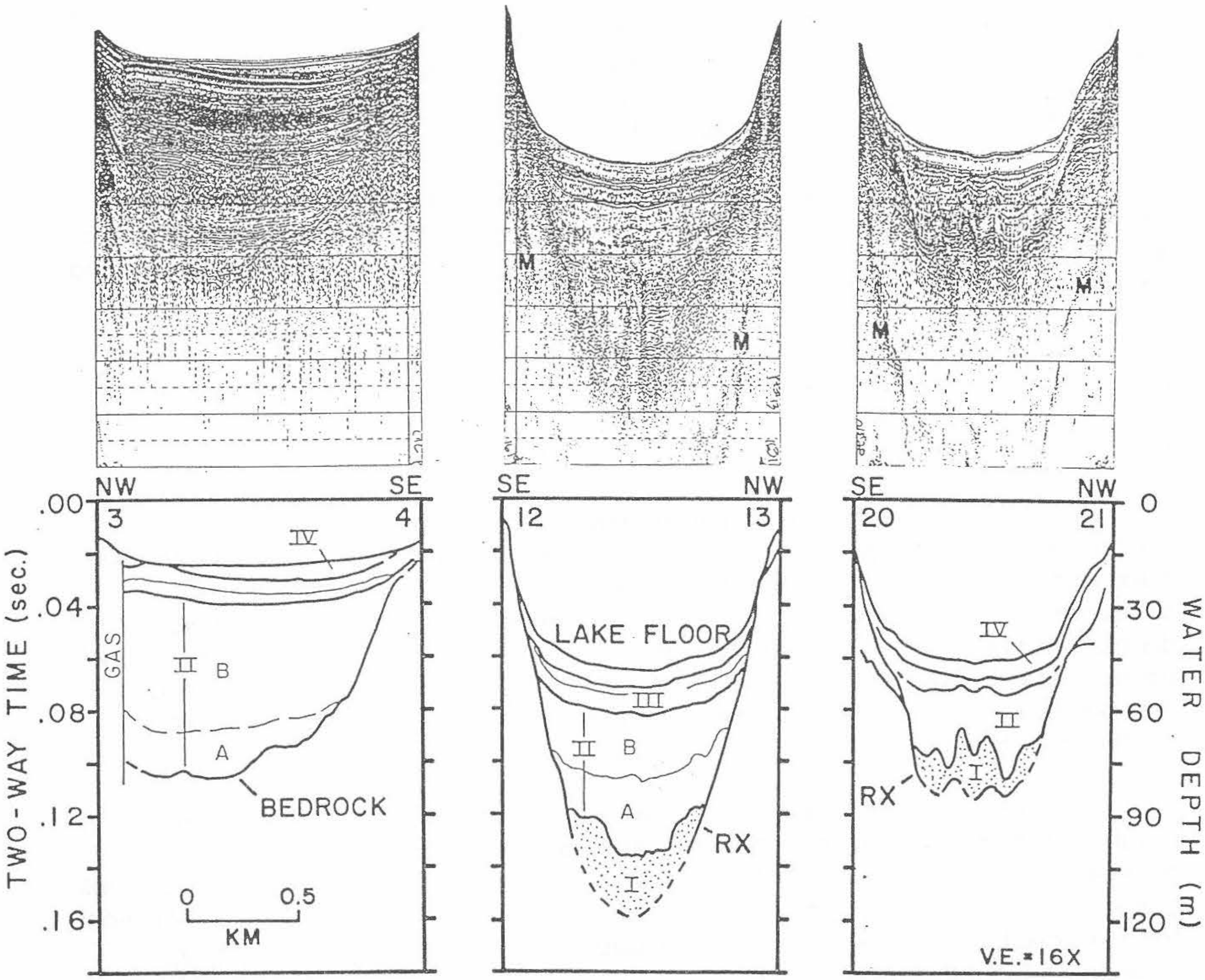

Figure 7 Transverse profiles. (from Fleisher, et al, 1990)

Figure 6 is the north-south axial line and Figure 7 contains transverse examples from the southern, central and northern portions of the lake. From these data measurements of water depth, sediment thickness and depth to bedrock were made. Figure 6 includes interpretations of depositional sequences within the sediment column (Fleisher, Mullins and Yuretich, 1990).

A maximum of 88 m of Quaternary sediment rests upon the bedrock basin of Otsego Lake. The thickest sediments are found in the southern-third of the lake basin and thin by more than 50% to the north. This information correlates with water-well log data that shown 55 m of bouldery silt beneath the Doubleday Margin.

Bedrock lies as much as l32 m below the modern lake level in the central portion of the basin. Bedrock rises at both ends of the lake to define a closed basin.

OTSEGO LAKE STRATIGRAPHY

Based on a combination of reflector terminations and vertical variations in seismic facies, Four depositional sequences have been identified within the sediment column of Otsego Lake (Fleisher, Mullins and Yuretich, 1990). The oldest unit, sequence I, has a seismic character interpreted to be coarse-grained, poorly sorted diamict. It is relatively thin (20 m) throughout much of the basin but forms three distinct ridges (possibly eskers or suballuvial fans) up to 60+ m thick in the southern third of the lake basin. Sequence I rises above lake level beneath Cooperstown to form the modern dam for Otsego Lake.

Sequence II is the thickest and appears to dip slightly northward in Figure 5. It is subdivided into two sub-sequences, IIA and IIB. Sequence IIA is characterized by high-frequency continuous reflections, whereas sequence IIB displays a seismic facies change that grades from continuous reflections in the south to a transparent (reflection-free) seismic facies northward. The continuous, high-frequency reflections suggest wide-spread, but temporally variable deposits, such as rhythmites, whereas the transparent seismic facies implies massive, non-stratified sediments of uniform texture that may have accumulated very rapidly.

Sequence III is a relatively thin unit that maintains a more or less uniform thickness along the north-south axial profile. Sequences II and III account for as much as 68 m of the total 88 m (77%) of sediment fill within Otsego Lake.

Sequence IV, the youngest, is not entirely continuous and generally appears as a layer less than l0 m thick. It is thickest along the flat lake floor, where it is about 6 m thick. This sequence is interpreted to be a massive, unstratified deposits, and represents modern (post-glacial) lacustrine deposition.

ANALYSIS OF DIP DIRECTION

Apparent dips measured on crossing and axial profiles were resolved at profile intersections for true dip direction through the use of the standard Schmidt net technique. The true dip is less than 2.2 degrees (most less than 1.0 degree), and vary only slightly from place-to-place. Generally, magnitude of dip appears to diminish up-section and northward.

A conspicuous eastward dip is shown on several transverse profiles (Figure 8). Analysis of dip direction in Sequence IIA and B indicates azimuths lie between N 79 E and N 88 E in profiles 1-2 and 2-3 in the vicinity of Brookwood Point and Leatherstocking Falls. Similarly, the dip direction in Sequence IIB along profiles 10-11, 11-12 and 12-13 lies between N 74 E and S 75 E northward of Threemile Point for more than a km . Consistent eastward dips continue in IIB south of Brookwood Point in profiles 4-5 and 8-9, where azimuths range from N 84 E to S 68 E, all of which is summarized in Figure 9. Furthermore, bedding-plane troughs marking the deepest part of the depositional basin shift eastward in progressively younger layers. These data indicate a consistent sediment source from the west, which is taken as evidence for tributary inflow. Several unnamed streams occupy drainage basin on the western valley wall and appear to have supplied a source of sediment to the southern-third of the lake basin.

A less conspicuous eastward dip direction is present in the central portion of the basin between Threemile and Fivemile Points along profiles 12-13, 13-14 and 14-15. Here, stratification is well represented in Sequence A, while IIB lacks bedding. This condition continues northward, where general symmetry is seen in profiles 15-16 and 16-17 within 1 km north and south of Fivemile Point. Here, the axial profile reflects the true dip of northward onlap in Sequence A (Figure 6) indicating continued sediment movement and infill from the south. Bedding symmetry within crossing lines in the vicinity of Fivemile Point suggests infill from Shadow Brook via Hyde Bay equaled that from streams flowing from western slopes.

Asymmetry is seen again in the northern part of the lake with increasing significance from Fivemile Point northward along profiles 17-18, 18-19, 19-20 and 20-22. However, dips here favor inclination to the west, which indicates sediment was primarily derived from Shadow Brook to the northeast.

INTERPRETATION OF SEISMIC RECORD

Seismic character and pattern of occurrence suggest sequence I is a late glacial diamict implaced at the base of a retreating ice-cliff. Deposition is thought to have involved sediment flow through subglacial conduits and channels that delivered suspended silt to the ice-contact, proglacial lake environment of Glacial Lake Cooperstown. Fountain discharge driven by englacial hydrostatic head (Gustavson and Boothroyd, 1987) created local thickening above bedrock pinning points in the southern part of the basin.

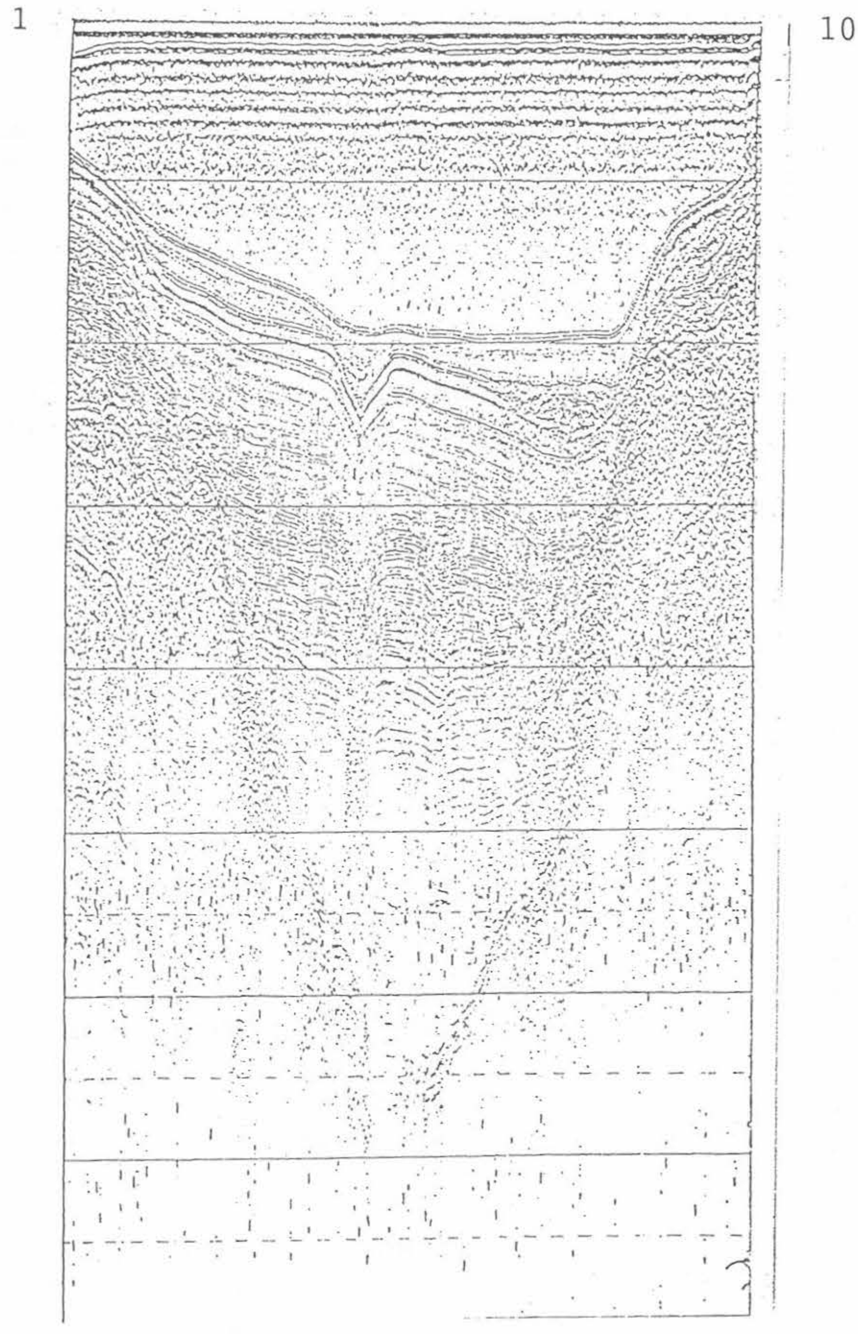

Figure 8 Asymmetric bedding. Transverse profile between stations 1 and 10 shows conspicuous eastward dip direction. Vertical exaggeration = X16. (from Fleisher, et al, 1990)

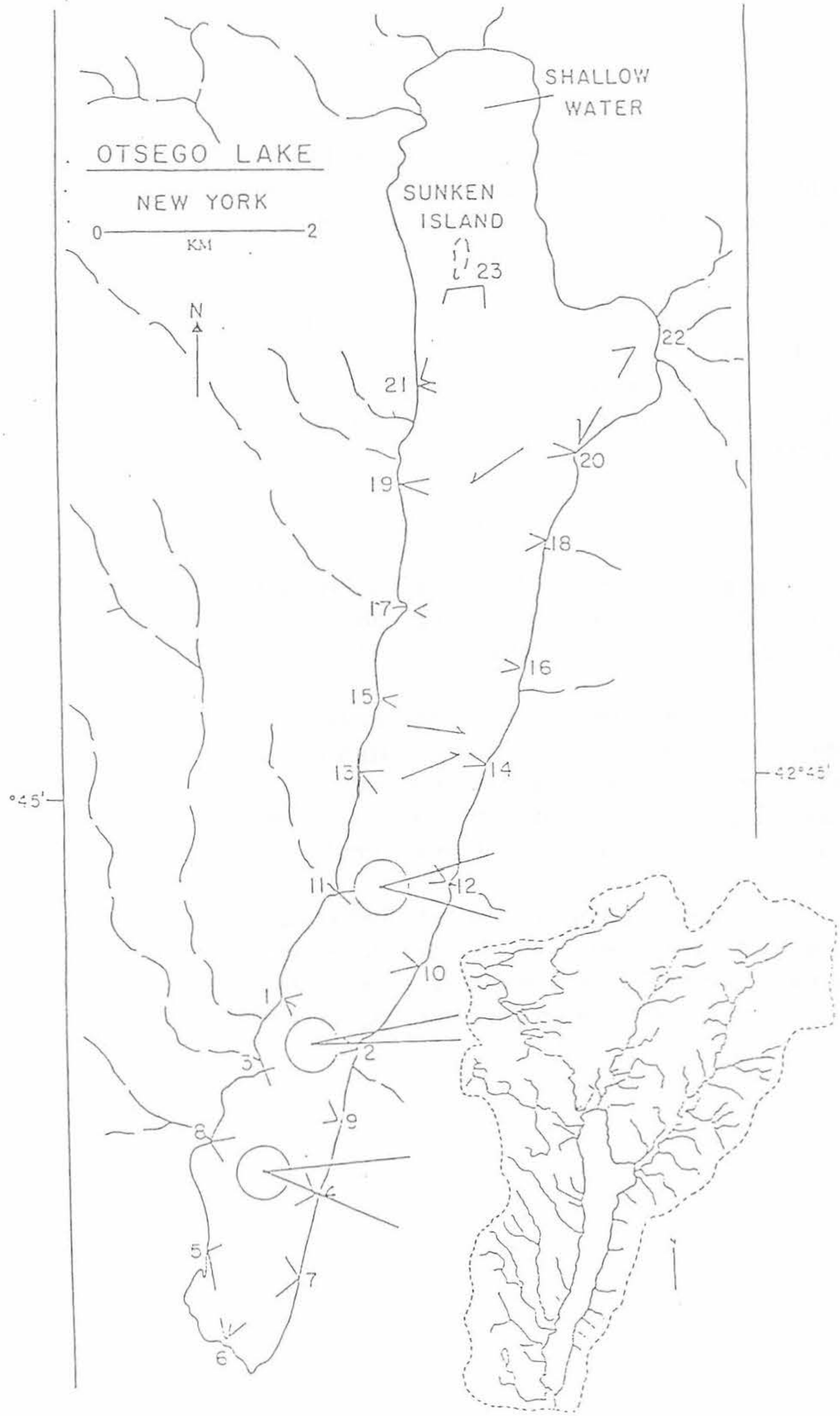

Figure 9 Dip direction for sequence II and tributary inflow sources. Small arrows indicate true dip direction calculated from apparent dips shown on seismic profiles. Data from two or more crossing profiles are summarized in "pie segments". Inset map shows asymmetry of drainge basin. (from Fleisher, et al, 1990)

Sequence II is attributed to sediment transport mechanisms that involved turbidity flows, suballuvial flows and debris flows, possibly related to flashy tributary streams, associated with extremely rapid silt sedimentation in a highly turbid ice-contact lake.

Thickening of A and B in the southern portion of the basin appears associated with tributary inflow that initially introduced sediment from the west, and was followed by northward bottom flow around "highs" in Sequence I. Disturbed bedding in sequence II is thought to be related to subsidence, collapse and compaction of glacial and post-glacial units.

CONCLUSIONS

Glacial Lake Cooperstown was dammed at 1250' by the Doubleday ice-margin at Cooperstown, not behind the Cassville-Cooperstown moraine 5 kilometers downvalley, as originally postulated by Fleisher (1977).

Depositional landforms and stratigraphy between the Cassville-Cooperstown margin at Index and the Doubleday margin at Cooperstown are inconsistent with standard interpretations of active ice retreat (fast or slow). Instead, the anomalous features suggest stagnation and inwash. This report suggests that retreat included the formation of a marginal-ice cleat and ultimate development of a dead-ice sink.

The eastward dipping beds of Glacial Lake Cooperstown indicate fluvial sources to the west. However, the primary source of silt in sequence II is thought to have been subglacial sediment flow in basal conduits that discharged directly into the lacustrine environment from the retreating ice-tongue terminus.

Case #2: Remnant ice as base level controls for stratified drift aggradation: Otego Creek, New York

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INTRODUCTION

The Otego Creek valley is a non-through valley with a glacial landform assemblage typical of downwasted ice, which includes kame fields, discontinuous gravel plain remnants, dead-ice sinks and eskers (Fleisher, 1991). Morrow (1989) and Fleisher (1986a, 1986b) described the processes responsible for these landforms.

Morrow's mapping units include ablation moraines (equivalent to kame field), outwash and kame deltas (similar to gravel plain remnants), local lake plains, dead-ice sinks, alluvial fans, an esker, and modern flood plain as shown in Figure 1.

Morrow's map is used to identify and locate upper-level planar landforms. The purpose of this study is to display and describe their longitudinal gradients and postulate the base level to which they are graded. The possibility that remnant ice masses served as temporary base levels during ice-tongue collapse is considered.

METHOD

Once planar landforms were identified on 7.5 minute quadrangles, longitudinal projected profiles were constructed within the stream valley between the hamlets of Laurens and Hartwick, a distance of 9.7 miles (Figure 2). The profile represents the highest planar landforms along a strip consisting of contiguous 1000-foot wide Block (pixels) aligned along the general valley thalweg. The maximum elevations of planar landforms, with a minimum dimension of 1000 feet within each block, were plotted at the centers of each block and used to construct the profile. Otego Creek was also plotted.

DISCUSSION OF OBSERVATIONS

Three general elevation groups were observed: 1140-1160' between Laurens and just south of Blood Mills Road, 1200-1260' from Blood Mills Road to about 1500' north of Laurens town line, and 1300-1340' from Jones Crossing to Hartwick (alluvial fans and patchy kame moraines were not included) (see Figure 1).

The profile lends itself two possible interpretations. The first involves a single profile (Figure 2A) on which all high planar surfaces fall. This is graded to an elevation of 1140 +/- 10', which corresponds to Glacial Lake Otego within the Susquehanna Valley at the mouth of the Otego Creek (Melia, 1975; Fleisher, 1977). The second (Figure 2B) suggests the projection represents three different profile segments at 1300-1340', 1200-1260', and 1150-1160', each graded to different local base levels. The

FIGURE 1. QUATERNARY LANDFORMS, MT. VISION QUADRANGLE (FROM MORROW, 1989)

Surficial Mapping Units

Abm — Ablation Moraine: Extensive, hummocky terrain, not associated with massive outwash. Consists of silt, sand and gravel in both well sorted and poorly sorted intervals. Potential for large scale aquifer development is variable depending on permeability and compaction.

Ow — Outwash: Well sorted, stratified sand and gravel, deposited by proglacial fluvial action. Contains a variable amount of silt, but is generally very permeable and has high potential for aquifer development.

kd — Kame Delta: Stratified sand and gravel that has been deposited into a proglacial or ice contact lake and has internal deltaic structures. Very permeable and has high potential for aquifer development.

lp — Lake plain: Primarily deposits of sand, silt and/or clay deposited in a lacustrine environment. Poor potential for aquifer development.

DIS — Dead-ice sink: Variable texture (size and sorting), identified as an anomalously broad flood plain that is bounded by high outwash and/or ablation moraines. These confining units often show collapse structures adjacent to the sink.

t/r — Till/bedrock: Till (e.g. clay, silt-clay, boulder clay), resting on bedrock, thickness variable.

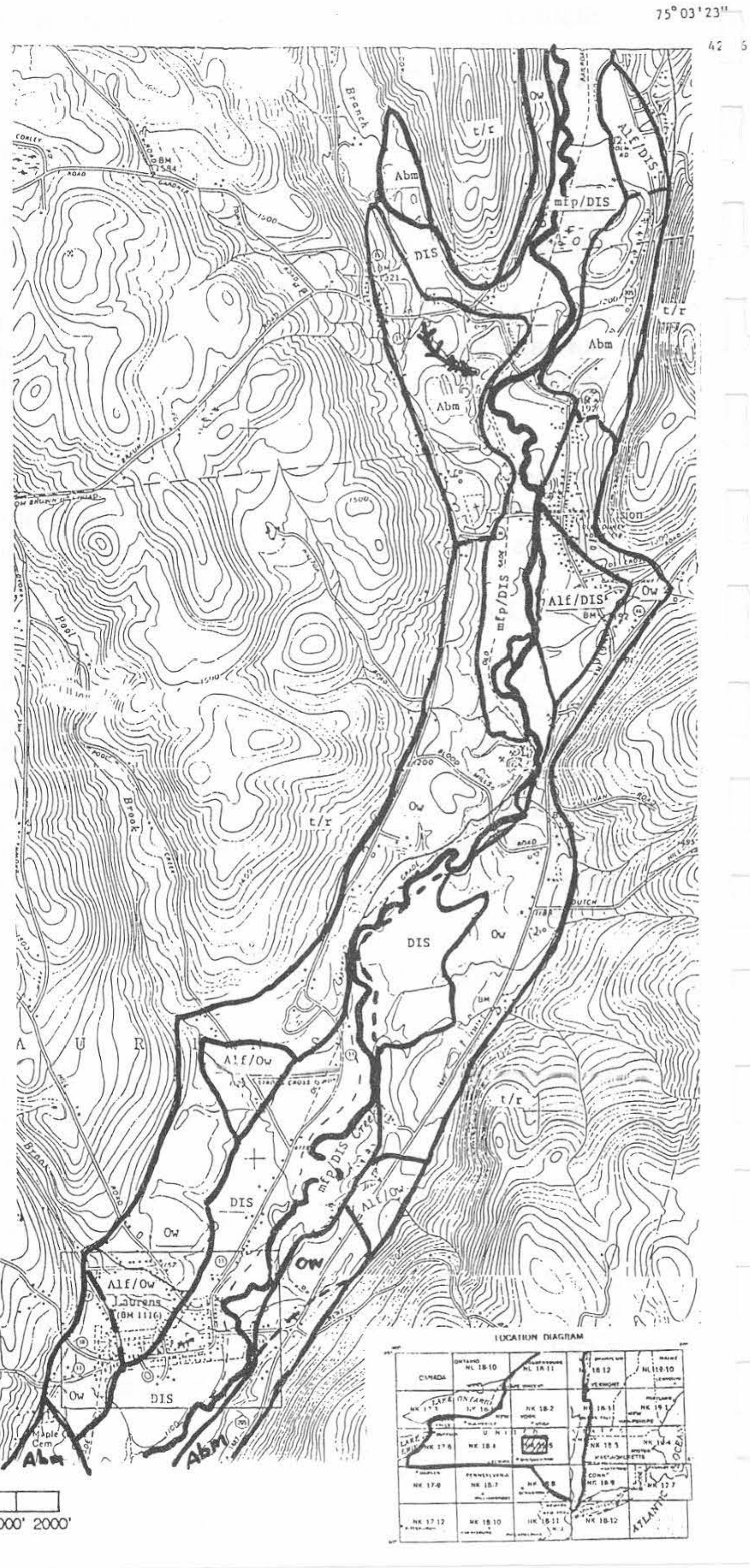
mfp — Modern flood plain: Post-glacial flood plain deposits of silt, low permeability, thickness 1-5 meters.

Alf — Alluvial fan: Fan shaped accumulations of poorly stratified silt, sand and boulders, at the foot of steep slopes, generally permeable.

NOTE: Where two units are separated by a slash, the first unit overlies the second. Example: afp/DIS, modern flood plain deposits overlie a dead-ice sink.

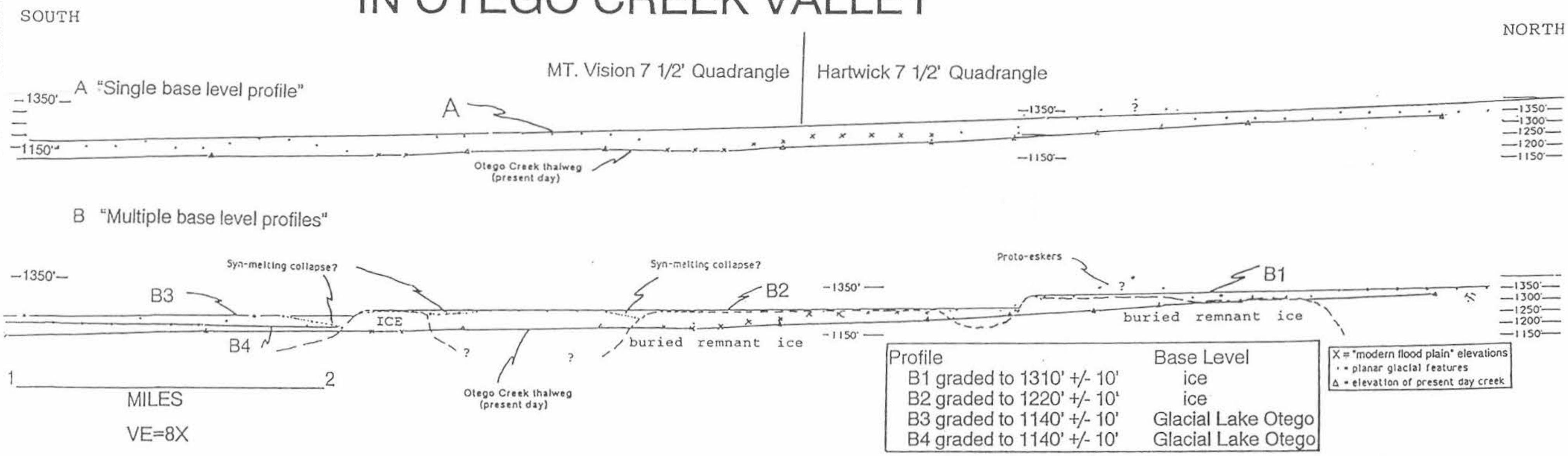
Map Symbols

Esker



0' 1000' 2000'

FIGURE 2. PROJECTED PROFILES: PLANAR FEATURES IN OTEGO CREEK VALLEY



surfaces represented by these profiles do not resemble a "shingled sequences" configuration, as presented by Mullholland (1982) in New England or morphosequences similar to those suggested by Koteff (1974). An alternative of local application is that several remnant ice masses occupied the valley at various times during ice-tongue collapse and obstructed meltwater movement during the deposition of planar sand and gravel features.

Air photo observations support the occurrence of one or more small-scale sinuous gravel ridges (proto-eskers) between Jones Crossing and Hartwick. This supports the notion of englacial fluvial transport within stagnant and downwasting ice.

CONCLUSIONS

The best fit for the data suggests that detached remnant ice masses impeded meltwater flow, which in turn formed ponds and lakes into which deltaic terraces were deposited. Problems with this model stem from the need to develop a time-frame which would account for both the sequence of deposition and source of meltwater necessary to move the sediment. It is also suggested that englacial and subglacial tunnels and conduits provided avenues for sediment transport.

Case #3 Inwash sediment sources, Valley of Charlotte Creek, central New York

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INTRODUCTION

Remnant-ice landforms and the moraine at West Davenport are thought to have developed in association with ice buried beneath inwash. Here, it appears as though tributary inwash from Kortright Creek provided significant supraglacial aggradation to bury large segments of the valley ice-tongue and produce extensive ice-cored deposits that ultimately downwasted in place.

GLACIAL DEPOSITS OF CHARLOTTE CREEK VALLEY

Glacial landforms and stratigraphy of the Charlotte Creek valley (Figure 1) are consistent with a regional pattern seen elsewhere throughout central New York. Deglaciation was characterized by retreating ice-tongues in valleys dammed by moraines, ice-cored inwash and thickly aggraded valley train (Fleisher P.J., et al., 1990). Consequently, many ice-contact lakes formed and their associated deposits are well developed and widely distributed. Such is the case for Charlotte Creek valley.

Four general types of depositional landforms shown in Figure 2 are:

1. Moraine and Pitted Valley Train
2. Dead-ice Sink Complex
3. Pitted Hanging Deltas
4. Lacustrine Plain

Moraine and Pitted Valley Train (Outwash/Inwash)

The moraine at West Davenport spans the full valley width at a maximum elevation of 1280' and is breached by Charlotte Creek at a floodplain elevation of 1180' through the moraine. Local relief on the kame and kettle topography of the moraine is about 60'. Contiguous with the moraine to the west is a dissected, valley train at a general elevation of 1220'. It is pitted by shallow kettles and spotted by occasional kames. The floodplain of Charlotte Creek is 60' lower, widening westward to its confluence with the Susquehanna. A north-south profile across the valley and through the moraine illustrates the association of landforms with stratigraphy at depth (shown in Figure 3), and shows no indication of a break in the depositional record. Therefore, these deposits, as well as those of the entire valley, are thought to be the product of a single deglacial event.

Dead-ice Sink Complex

The term "dead-ice sink" (Fleisher,1986) describes a landform similar to a kettle, but of much larger scale. As with a kettle, a block of ice is buried within valley-floor

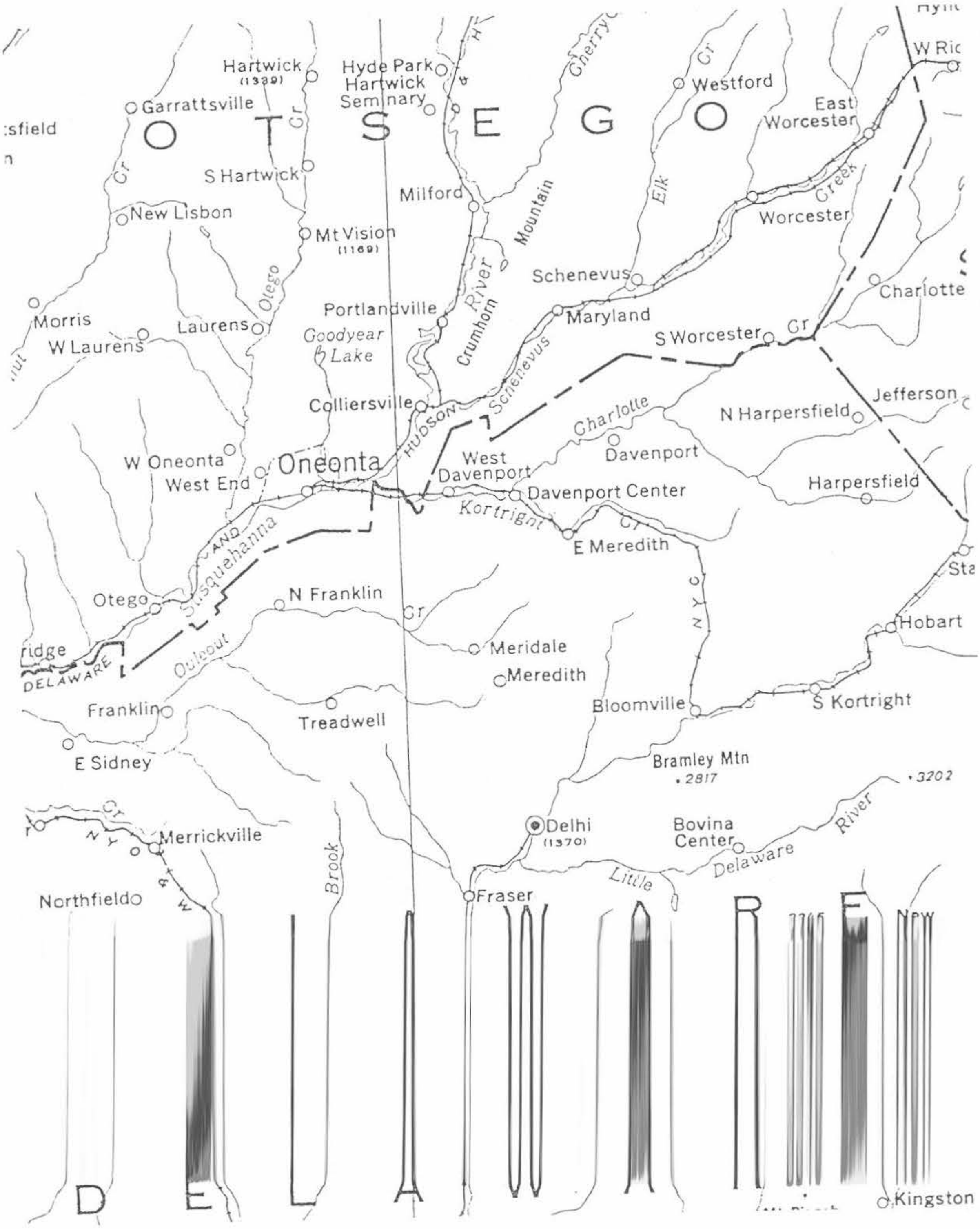

FIGURE 1 Index Map. The West Davenport/Davenport Center area is located near the mouth of Charlotte Creek Valley, 4 miles upstream from the Susquehanna confluence.

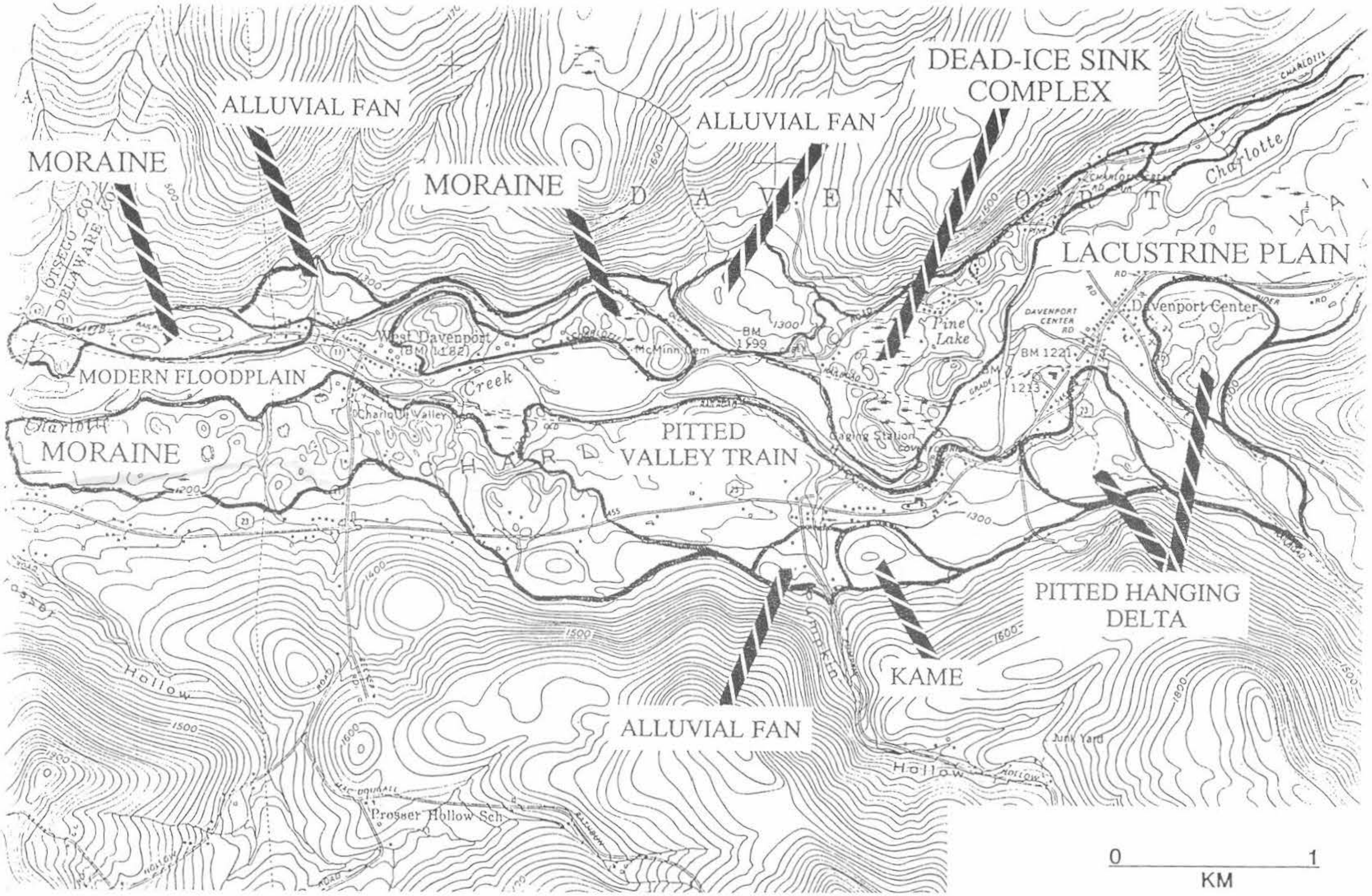

FIGURE 2 Map of Glacial Landforms. Landforms in the West Davenport/ Davenport Center area between the moraine at West Davenport and lacustrine plain of Glacial Lake Davenport.

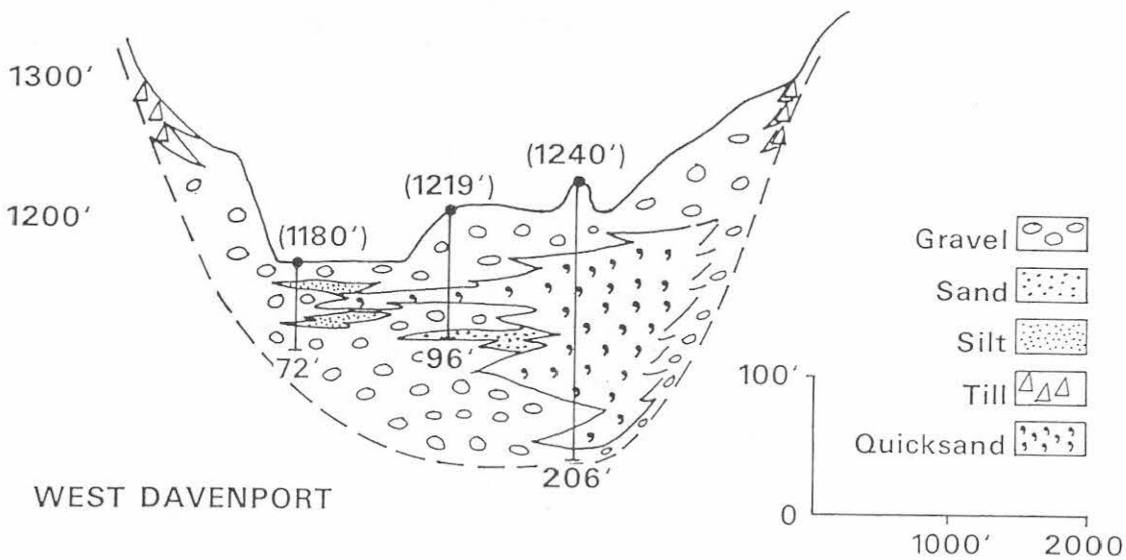

FIGURE 3 Cross section of Quaternary deposits within moraine at West Davenport. Upvalley view of interstratified gravel, sand and "quicksand" (coarse silt). Well data from Randall, 1972.

material (usually stratified drift) and ultimately melts to create an exceptionally large closed depression. However, the distinction between a kettle and a sink lies in the size of the resulting depression. Dead-ice sinks commonly occupy a significant portion of the entire valley floor and are recognized by anomalously broad floodplains within otherwise continuous valley trains and/or paired outwash terraces. Deglacial conditions leading to sink formation involve the detachment of a large ice mass from the retreating ice-tongue, with subsequent burial beneath a valley train of outwash and inwash deposits. Retarded melting due to insulation by overlying debris results in the slow development of a valley-floor depression in which highly diverse deposits accumulate. Where more than one large ice mass is involved, a complex of juxtaposed sinks and kames form. A dead-ice sink complex at Pine Lake lies between the West Davenport moraine and Davenport Center lake plain. Here, local relief exceeds 100' and pitted valley train (outwash/inwash) occurs along the south side of the valley adjacent to the sink complex at elevations of 1280' to 1300'. The floodplain forms the floor of the sink at 1200' (BM on abandoned railroad bridge at 1213').

Associated with the moraine/dead-ice sink complex are several landforms at elevations between 1300-1380'. These include an incised hanging delta/fan complex at 1300' on north side of the valley between West Davenport and Pine Lake, an alluvial fan remnant at 1360' at the mouth of Pumpkin Hollow on the south side of the valley and a dissected alluvial fan at 1260-1280' on the north side of the valley at West Davenport. These are interpreted to be indicative of significant inwash accumulation.

Pitted Hanging Delta

Possibly the most significant landform within the entire valley is the pitted hanging delta that stands at 1280' at the mouth of Kortright Creek in Davenport Center. Comparable elevations on the moraine downvalley suggest the moraine served as the dam for Glacial Lake Davenport (Fleisher, 1977), into which the delta grew. More importantly, the surface of the delta contains several distinct kettles that are in excess of 40' deep. This indicates the delta aggraded onto and across grounded ice in Glacial Lake Davenport immediately adjacent to the dead-ice sink complex and that remnant-ice and ice-cored drift accumulated within the valley while upland slopes were ice-free. These ice-contact, alluvial deposits are interpreted as inwash from upland tributaries.

Another pitted hanging delta can be seen at Butts Corner, where Middle Creek enters Charlotte Creek valley. Borrow pit excavations reveal 20-30' of alluvium from Middle Creek covers deltaic gravels containing a topset/foreset contact near 1320', thereby establishing a downvalley spillway at 1300-1320'. Smaller hanging deltas at similar elevations occur elsewhere within the valley. East of Butts Corner a variety of landforms appear to have served as local dams for ponding upvalley as suggested by hanging deltas that rise in steps through Fergusonville, Simpsonville and South Worcester. Remnant-ice masses may have also provided temporary base levels for scattered strandline deposits.

Lacustrine Plain

A semi-continuous lacustrine plain rises from a valley floor elevation of 1210' at Davenport Center to 1240' at Davenport 2 miles to the east. In the main Charlotte Creek valley, the lacustrine plain is interrupted by various forms of outwash, inwash and previously ice-cored, landforms, which appear as subdued, gently rolling terrain 20-40' above the lacustrine plain and a comparable elevation below a few valley train remnants. The upvalley floodplain climbs to 1280' at Fergusonville, 1320' at Simpsonville and 1400' in South Worcester. These abrupt rises support the notion of a lake and pond chain within the valley. In some places the lacustrine plain is covered by modern floodplain sediments five feet thick, as seen in archaeology dig site excavations a few hundred meters south of Pine Lake (Fleisher, 1990, unpublished report).

CHRONOLOGY OF GLACIAL EVENTS

Summarized below are the events interpreted to have occurred in the vicinity of Pine Lake during deglaciation, beginning with late glacial and continuing into post-glacial time.

Glacial Time

Deglaciation of Charlotte Creek valley pre-dated full retreat of the Susquehanna valley ice-tongue. Separated and detached ice masses remained within the valley long after upland slopes and tributary valleys were ice-free. The Susquehanna ice-tongue and associated ice-cored deposits dammed the mouth of Charlotte Creek valley and impounded an early phase of Glacial Lake Davenport at 1300-1320'. Graded to this elevation are hanging deltas near West Davenport, Butts Corner and Fergusonville. Dissected remnants of a delta-like feature at 1300' extend across the valley at the mouth of Dona Brook and may have at one time divided Lake Davenport into eastern and western segments.

Inwash from Kortright Creek buried remnant ice masses at Davenport Center and effectively covered the downvalley ice-tongue to the west. This contributed to the debris mass that ultimately formed the moraine at West Davenport .

Late-Glacial Time

As the Susquehanna valley ice-tongue retreated from its position behind the Oneonta Moraine, an ice-cored dam remained at the mouth of Charlotte Creek buried beneath inwash primarily from Kortright Creek (Figure 4). In adjustment to local base level changes within the Susquehanna, the spillway for Lake Davenport was lowered to 1280', where it remained while several deltas formed, including the ice-cored delta at Davenport Center.

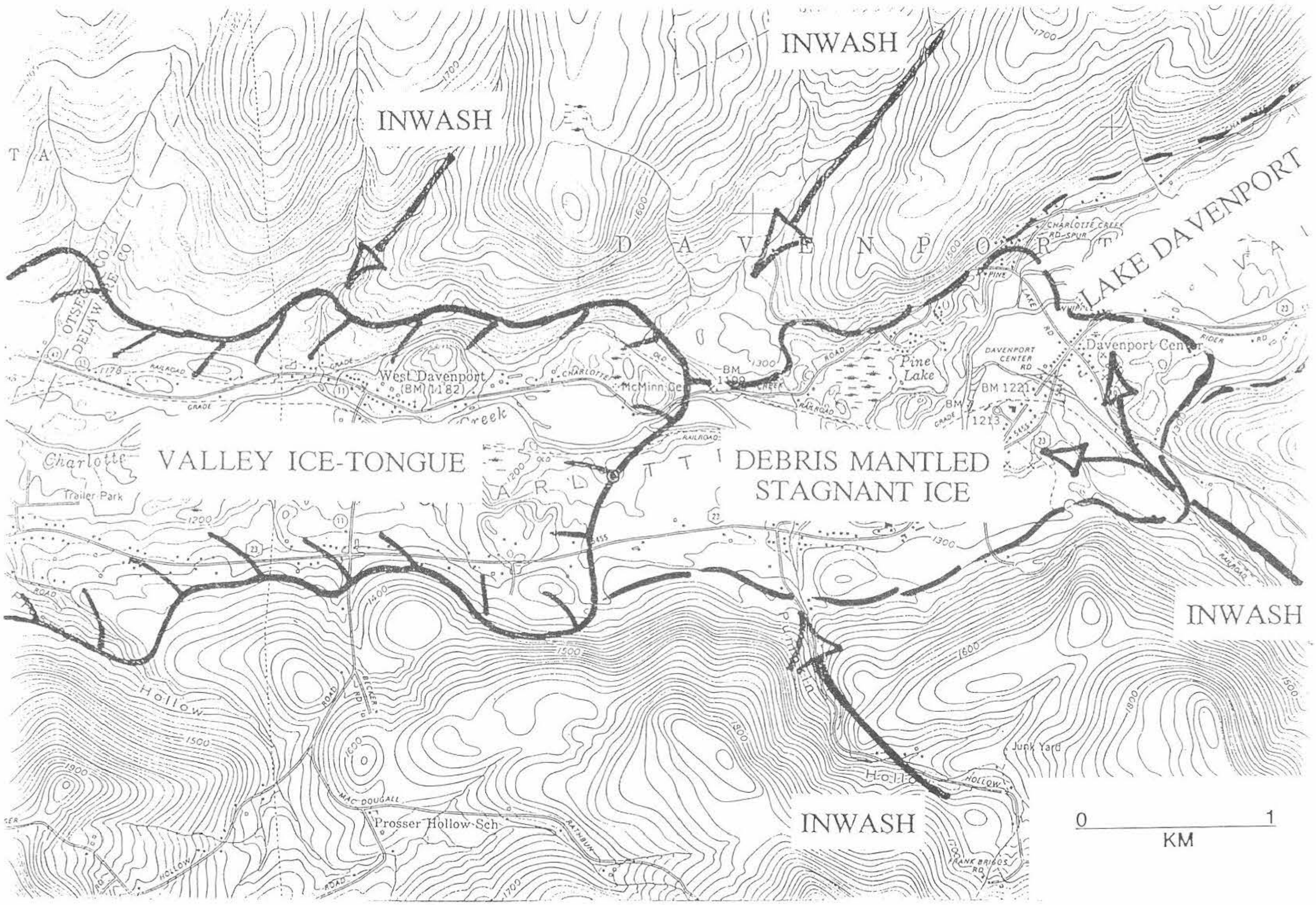

FIGURE 4 Late Glacial Retreat Configuration.

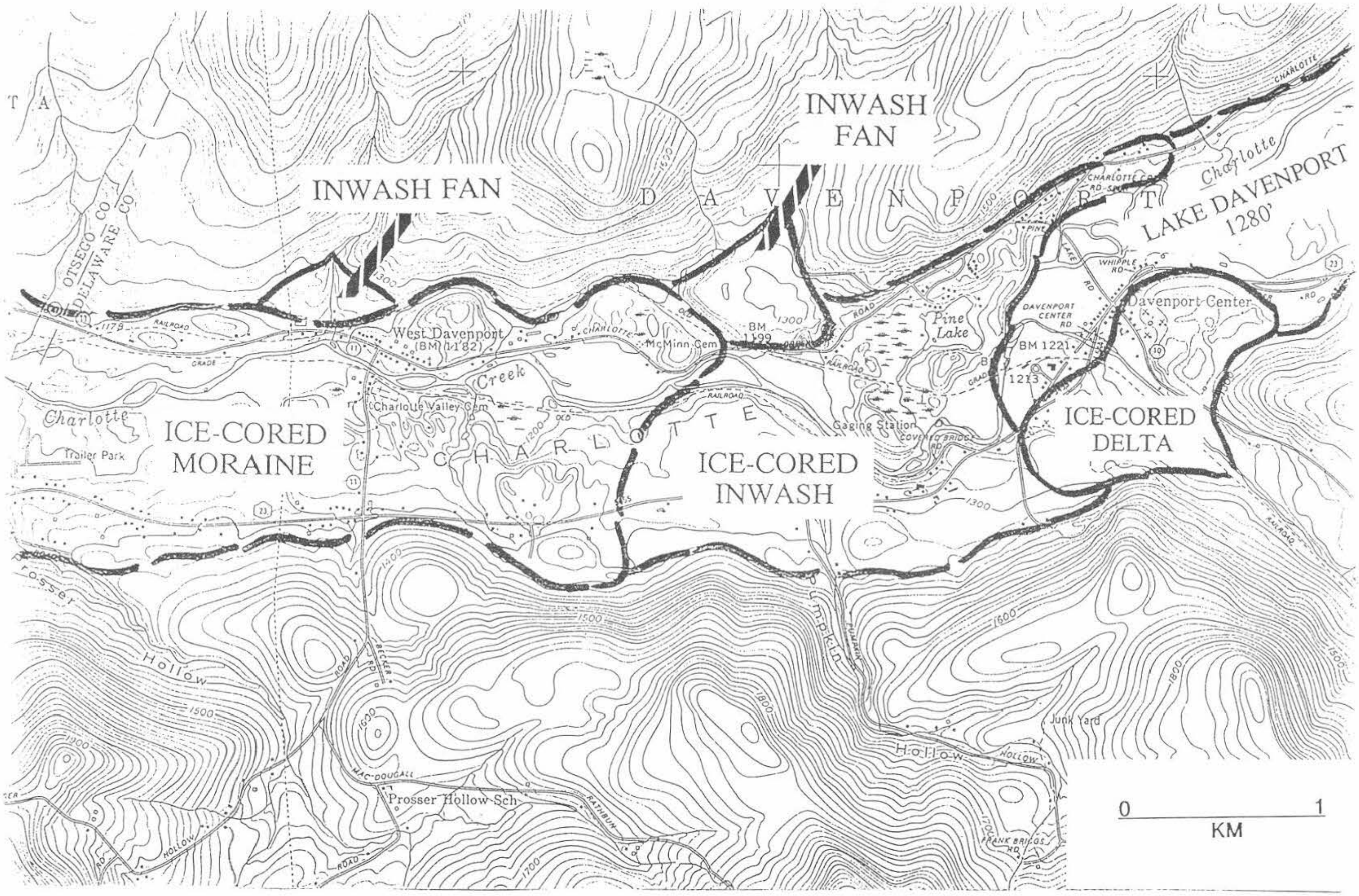

FIGURE 5 Early Post-Glacial Deposits.

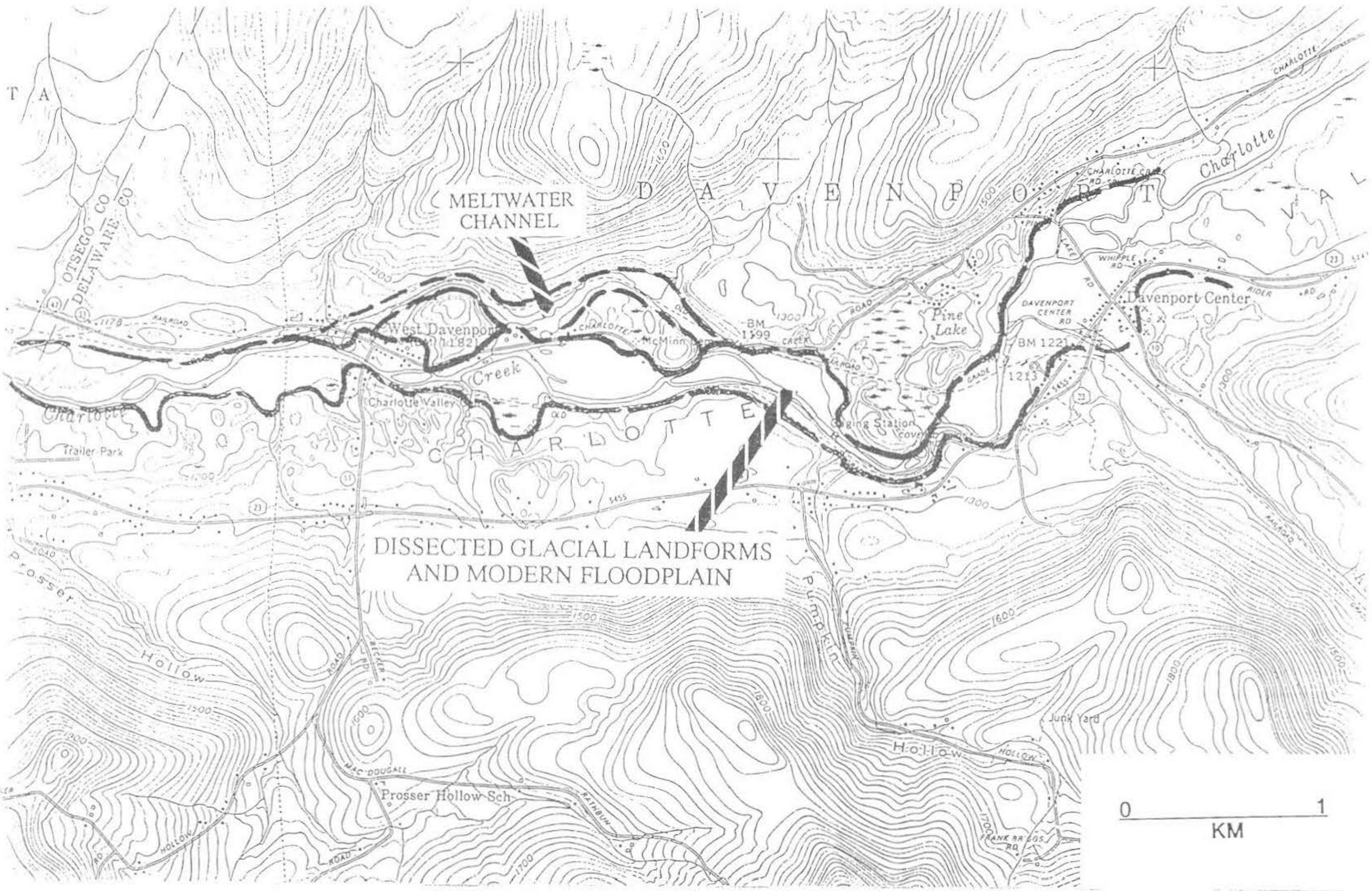

FIGURE 6 Holocene Dissection

Downwasting of all ice-cored material progressed, while the moraine and sink began to develop. As the spillway elevation was lowered, Lake Davenport emptied, as delta/fan complexes were incised and the lacustrine plain emerged.

Early Post-Glacial Time

Retarded melting of partially buried ice masses eventually produced a variety of collapse features in association with the newly-formed lacustrine plain. The terrain at Pine Lake began to form as downwasting lead to the development of pitted valley train, kame and kettle topography and a dead-ice sink complex (Figure 5). These landforms developed during this period of time as downwasting slowly progressed and ice-cored valley deposits were lowered to river level, as Holocene sedimentation kept pace with subsidence.

Holocene Time

As Charlotte Creek established its course on the newly-formed valley floor, floodplain aggradation began (Figure 6). Lateral planation trimmed the edges of kames and terraces, as overbank discharge spread blankets of silt on glacial gravel and lacustrine silt, sand and clay. Through the course of aggradation and meandering, the creek re-occupied some locations several times and floodplain sediment accumulated. Silt layers and channel sands were observed in trench excavations at the Pine Lake archaeology dig of 1990.

APPENDIX A

PEBBLE COUNTS IN THE CHARLOTTE CREEK DRAINAGE BASIN

Steve Carley and P. J. Fleisher, Earth Sciences Department, SUNY-Oneonta

INTRODUCTION

In a paper dealing with the geology of the Charlotte Creek Valley, Fleisher (1980) interpreted many of the landforms as inwash features formed in association with remnant ice. Ice-free uplands provided a source for alluvium washed onto remnant ice blocks between Davenport Center and West Davenport (Figure 1). Subsequent downwasting formed a moraine and dead-ice sink complex. A major source of inwash is thought to have been Kortright Creek. This study tests the usefulness of pebble count data to determine sediment provenance in an area where exotic pebbles comprise less than 5% of the total sample. If this method of analysis is valid for samples such as these, then the source of the gravels associated with the dead ice complex and the moraine may be substantiated. Halter, et al., 1984, suggest that sediment transported by Laurentide ice in Maine was deposited within 10 kilometers of its source. If this applies to the Appalachian Plateau as well, a high percentage of exotics beyond 10 km would suggest transport mechanisms other than glacial were significant.

The possibility of selective lithologic attrition by weathering or transport mechanisms is important in this area because susceptible siltstone and shale are common. This is an important factor when evaluating the meaning of pebble count data.

PEBBLE COUNT SAMPLE SITES (Figure 2)

Kortright Creek

The modern stream flows over glacial deposits from the valley head to the village of East Meredith, where it encounters a bedrock knick point. All samples taken in this valley (Sample 1-9) were obtained upstream from the bedrock knick point from holes dug in the drift.

Pitted Hanging Delta

Kettles in the surface of this delta and prominent foreset beds indicate that it prograded from the Kortright Creek Valley, across grounded ice, and into glacial Lake Davenport (Fleisher, 1990). The upper elevation of the dissected delta lies at 1280 feet, 40 feet above the modern creek. Sample 10 was taken from an active borrow pit within the core of the delta adjacent to country Route 10 along Kortright Creek. Sample 11 was taken from foreset beds in an active borrow pit on the distal northern side of the delta a few hundred yards from sample 10.

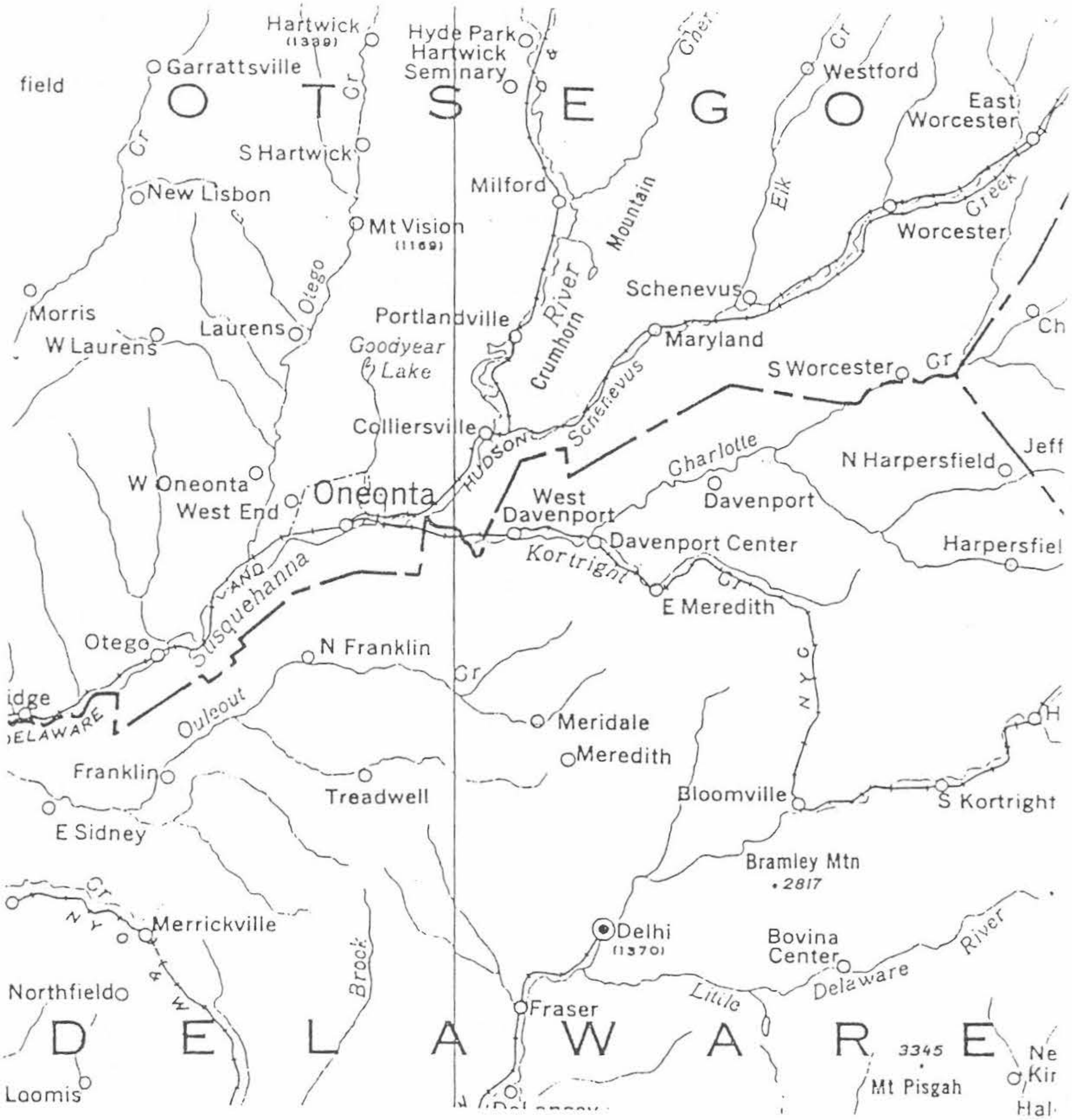

Figure 1 Index Map

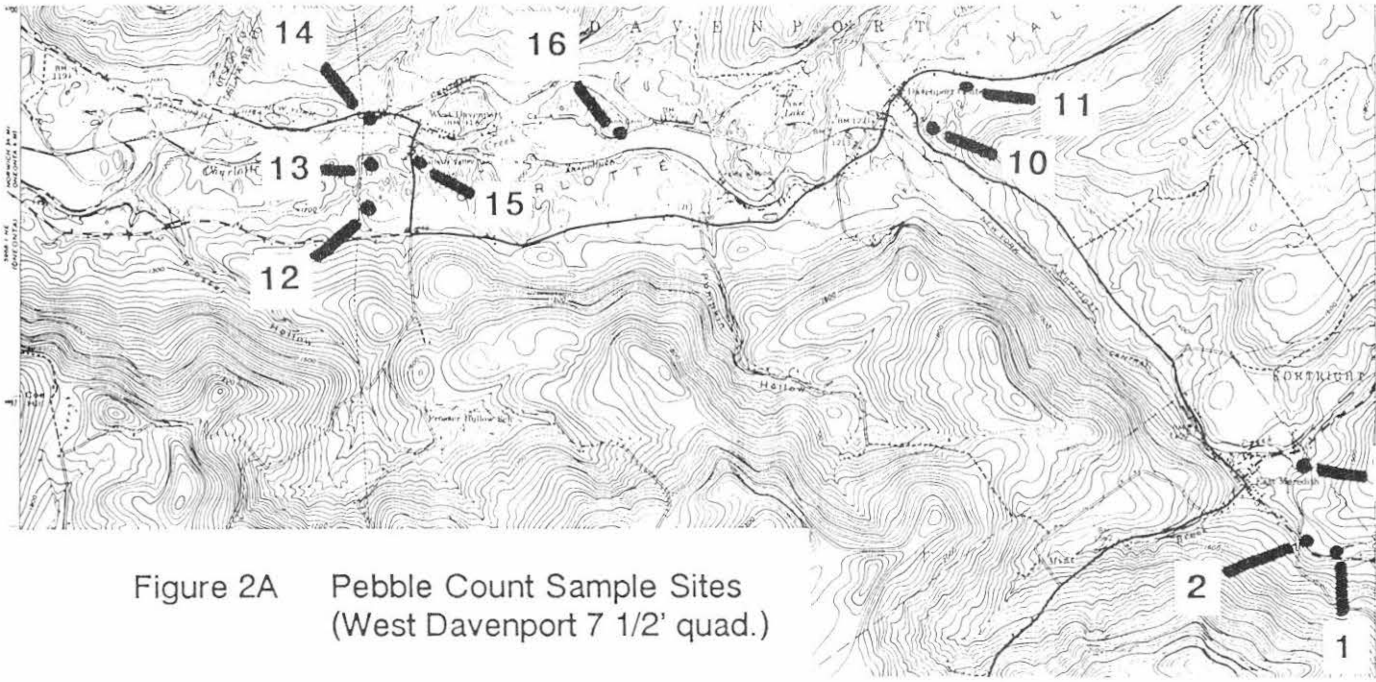

Figure 2A Pebble Count Sample Sites (West Davenport 7 1/2' quad.)

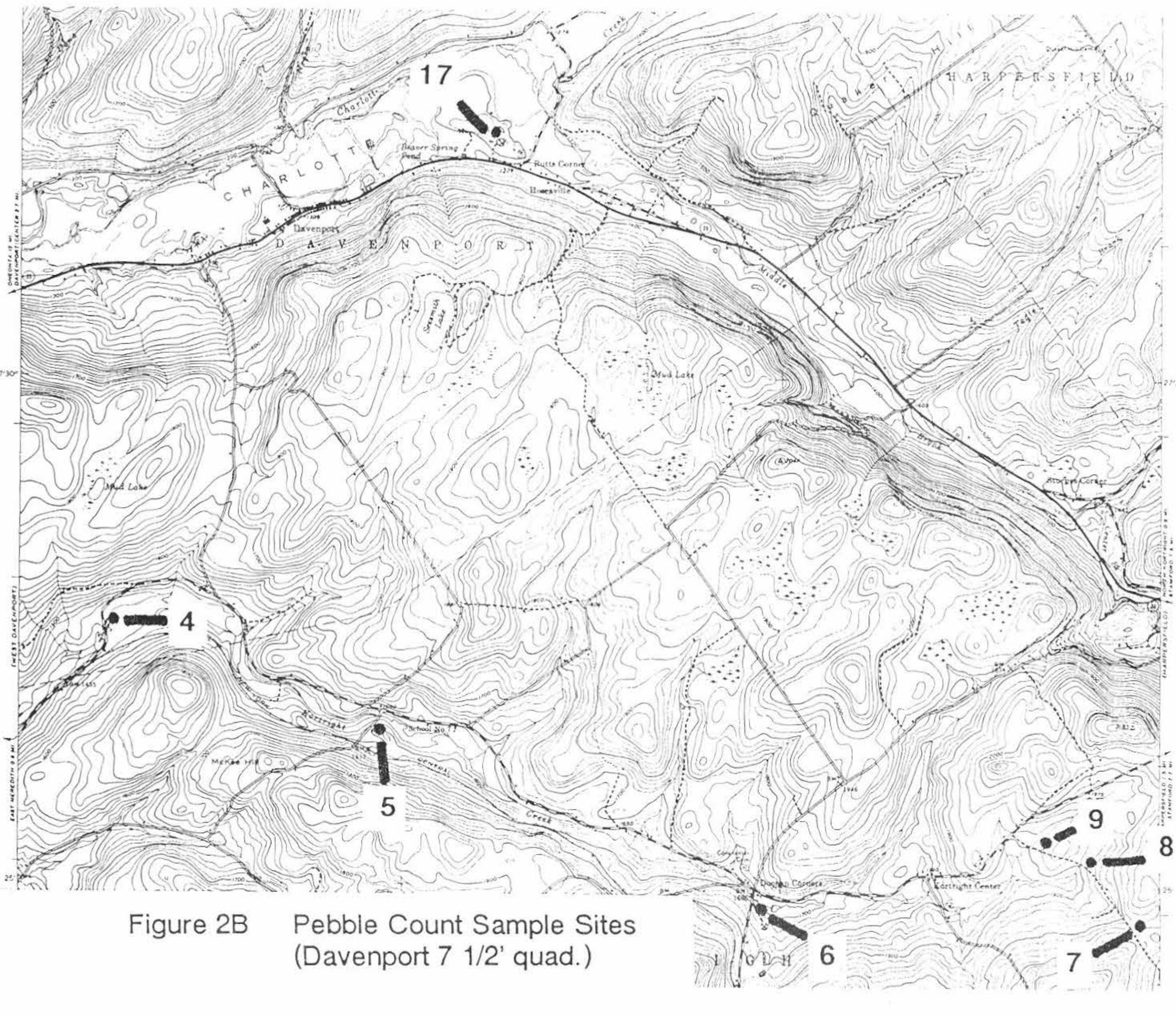

Figure 2B Pebble Count Sample Sites (Davenport 7 1/2' quad.)

West Davenport Moraine

Samples 12, 13, and 14 were taken along a north-south traverse across the moraine at West Davenport and sample 15 from a road cut at the Charlotte Valley Cemetery in West Davenport.

Dead-ice sink complex

Sample 16 is the only source of data from the dead-ice sink complex. It was collected at McMinn Cemetery west of Pine Lake.

Butts Corner

A large delta/fan complex at the confluence of Middle Creek with Charlotte Valley (sample 17) was sampled 30-40 feet below the top of the landform in an active working face of the borrow pit.

SAMPLING AND ANALYSIS

Samples of approximately 100 pebbles were sought. Care was taken to avoid contamination from human activities. At borrow pit locations, samples were simply scooped from the exposed face into a bag. In all other locations, samples were obtained from holes dug in the surface. Samples were passed through a half inch sieve and washed. Pebbles were split to expose a fresh surface. The color and lithology of each pebble were noted under a binocular microscope. A summary of all data is shown in Tables 1 and 2.

DISCUSSION

Kortright Creek

Data for samples 1-9 (Fig. 4-10) show little change in color or lithology from the head of the valley to the bedrock knick point at East Meredith. No pebble-sized exotics were counted, yet exotic boulders were observed within the drainage basin.

Pitted hanging delta

Data for the delta seem to show two different suites of lithologies but in each a low percentage (2-3%) of exotic pebbles were counted.

Charlotte Valley

Data from two samples taken at Butts Corner represent a single lithologic suite, which differed from data obtained on the moraine and dead-ice sink complex.

Comparison of data

Lithologic suites from the head of Kortright Creek valley compare favorably with those found on the moraine, but differ from those at Butts Corner. Data on color characteristics vary from place to place and do not appear to show meaningful changes.

CONCLUSIONS

The data collected neither proves nor disproves the postulated inwash source for drift in the western part of the Charlotte Creek valley. The dominant lithologies in all samples are local. No specific lithology can be used to indicate provenance. However, similarity of lithologic suites suggest inwash from Kortright Creek provided material to the Dead-ice Sink Complex. A different suite at Butts Corner, may indicate still another source for upvalley deposits.

TABLE 1. PERCENTAGE OF PEBBLES BY COLOR

Color	Sample No. 1	2	3	4	5	6	7	8	9	Avg 1-9
Red	39	30	39	37	30	33	53	31	31	36
Gray	33	29	25	16	36	26	31	37	38	30
Dark Gray	1	7	5	5	2	2	1	4	7	4
Tan	24	34	25	41	23	36	17	26	23	44

Color	Sample No. 10	11	Avg. 10-1	12	13	14	15	Avg. 12-15	16	17
Red	25	10	18	16	16	23	19	19	29	8
Gray	45	29	37	31	36	22	20	27	31	1
Dark Gray	14	33	24	24	21	23	31	25	18	49
Tan	16	26	21	29	26	32	27	29	45	30
Exotic	2	2	2	0	0	0	2	1	5	2

Sample Locations: #1-9, Kortright Creek; #10-11, Hanging delta; #12-15, Moraine; #16, Dead-ice sink; #17, Butts Corner

TABLE 2

PERCENT OF LITHOLOGIES

Lithology	Sample No. 1	2	3	4	5	6	7	8	9	Avg. 1-9
Sandstone	33	37	28	32	34	31	41	23	32	32
Siltstone	45	48	55	47	40	51	45	55	50	48
Mudstone	22	15	17	21	26	18	14	21	18	19
Exotic	0	0	0	0	0	0	0	1	0	0

Lithology	Sample No. 10	11	Avg. 10-11	12	13	14	15	Avg. 12-15	16	17
Sandstone	21	27	24	25	9	26	29	22	22	8
Siltstone	44	52	48	43	47	53	25	42	42	28
Mudstone	34	19	27	31	44	21	44	35	34	62
Exotic	1	2	2	0	0	0	2	1	2	2

Sample Locations: #1-9, Kortright Creek; #10-11, Hanging delta; #12-15, Moraine; #16, Dead-ice sink; #17, Butts Corner

Case #4 Implications of pebble count data; confluence of Unadilla River and Tallette Creek, Columbus Quarters, New York

Jim Yuchniewicz and P. Jay Fleisher, Earth Sciences Department, SUNY-Oneonta

INTRODUCTION

Classical debates concerning contrasting lithologic suites within the Appalachian Plateau drift (MacClintock and Apfel, 1944; Meritt and Muller, 1959; and Moss and Ritter, 1962) indicate that upland drift has a lower percentage of erratic pebbles then through valley stratified drift. This led to consideration of different transport mechanisms for each type of drift. Evenson and Clinch (1987) emphasize the significance of glacio-fluvial transportation and the importance of inwash as a sediment source in the glacial environment.

Tallette Creek is an upland tributary to the Unadilla River approximately four miles north of New Berlin along Route 8. It drains an area of approximately ten square miles on the west side of the Unadilla Valley. As with all major through valleys, the Unadilla shows conspicuous valley asymmetry, with larger tributaries on more gentle east-facing slopes.

This study considers inwash as a possible sediment source for drift in the Unadilla Valley. Pebble counts of "bright" and "drab" drift along Tallette Creek Valley are compared to those taken from within a kame field at it's confluence with the Unadilla Valley. By definition, "bright" drift is composed of 30-50 percent erratic lithologies, whereas "drab" contains 0-15 percent (Randall, 1973).

SAMPLING PROCEDURE, PREPARATION, AND PEBBLE COUNTS

Random pebble samples were taken from various locations along Tallette Creek Valley and from within the kame field (Figure 1). Gravel pits offered access to some locations, but in many cases pebbles were obtained from dug holes. Some samples were taken from mounds next to woodchuck holes. Samples were taken approximately every half mile within the valley, alternating between the valley floor and upper slopes, as well as from convenient spots from within the kame field.

Twenty samples were returned to the laboratory, placed in a 12.5 mm (1/2 inch) sieving screen, washed, and individual pebbles broken for binocular microscope identification and determination of roundness.The average number of pebbles counted per site was 118 with a range between 57 and 222. The pebble grain size averaged between 15 and 25 mm, but did include a few cobble size clasts.

LITHOLOGIC ANALYSIS

As anticipated, local lithologies dominated all samples, with an average of 91 percent mudstones and siltstones from the Panther Mountain Formation. Bright

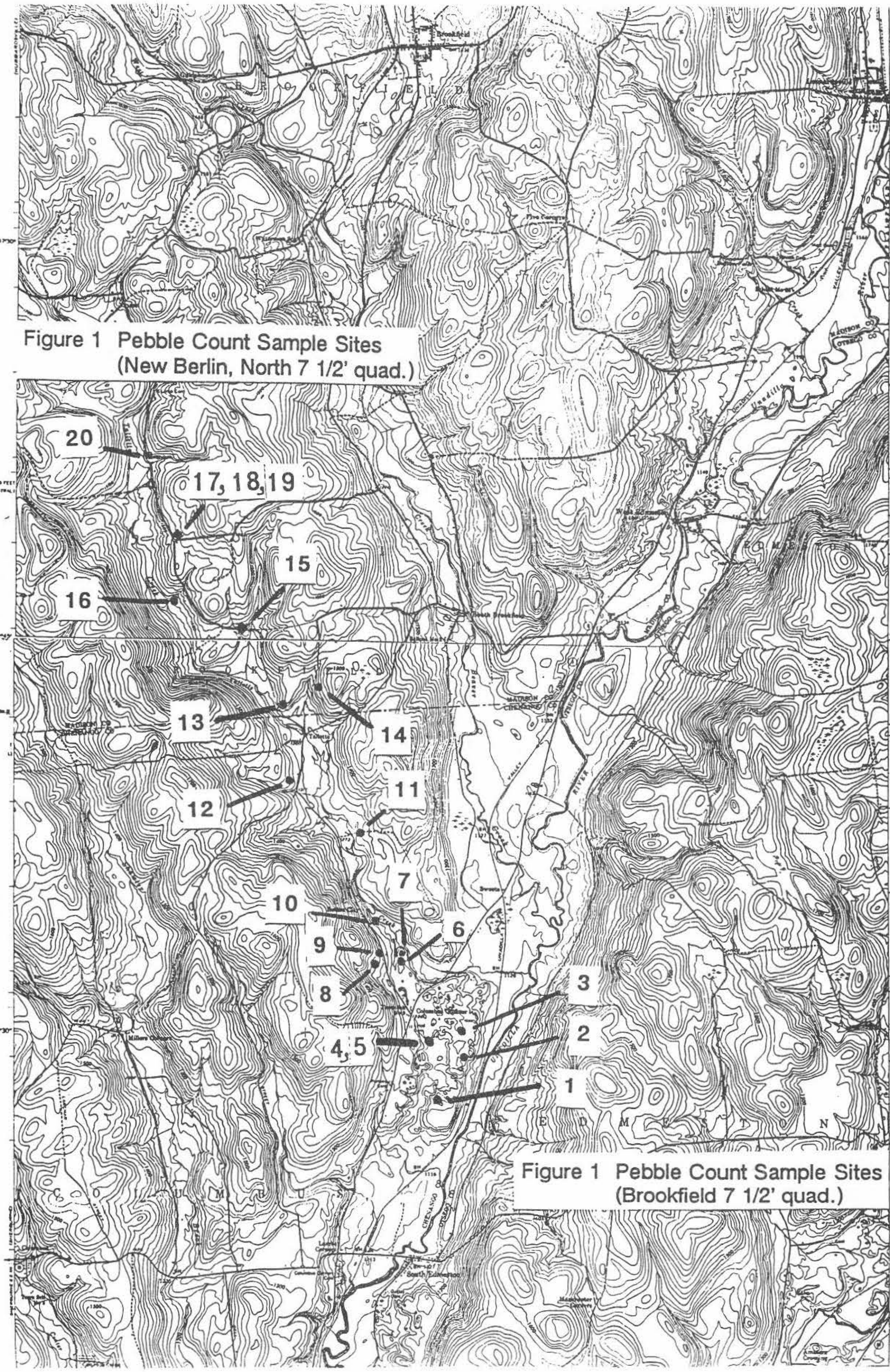
Figure 1 Pebble Count Sample Sites
(New Berlin, North 7 1/2' quad.)
Figure 1 Pebble Count Sample Sites
(Brookfield 7 1/2' quad.)
20
17, 18, 19
15
16
13
14
12
11
10
7
6
9
8
3
2
4, 5
1

lithologies varied between 0 and 27 percent. Kame field samples contained an average of 18 percent bright lithologies and ranged from 5 to 27 percent. Upland drift contains from 1 to 10 percent brights, with an average of 5 percent. The average along Tallette Creek Valley is 4 percent, with 3 percent along the stream channel.

Five types of bright lithologies were noted. Quartz sandstone and chert dominate, with decreasing amounts of limestone, quartzite and gneiss. Additional lithologies observed in the field, but not counted are anorthosite, megagabbro, amphibolite gneiss, and granitic gneiss.

Figure 2 summarizes all pebble count data. Samples from the upper slope at sites 12, 14, and 17-20 yielded relatively higher percentages of bright pebbles, which decreased gradually along the valley floor of Tallette Creek. At sites 6 and 7, an abrupt increase in the bright lithologies occurs. Samples taken from sites 4 and 5 contained even higher percentages bright lithologies, but the percentage decreased again along the eastern margin of the kame field (samples 1-3).

Figure 2. Summary of All Pebble Counts

SAMPLE NUMBER	LOCAL BEDROCK	QUARTZ SANDSTONE	CHERT	GNEISS	QUARTZITE	LIMESTONE
1	85.5	5.6	8.0	0.9	0.0	0.0
2	94.7	4.0	1.3	0.0	0.0	0.0
3	80.3	11.1	6.8	0.0	1.7	0.0
4	73.3	10.3	9.0	0.0	0.7	6.9
5	74.6	8.8	7.0	0.0	0.9	8.8
6	81.7	7.5	6.7	2.5	1.7	0.0
7	85.9	9.4	4.7	0.0	0.0	0.0
8	93.8	4.9	1.4	0.0	0.0	0.0
9	100.0	0.0	0.0	0.0	0.0	0.0
10	94.6	5.4	0.0	0.0	0.0	0.0
11	96.5	0.0	0.0	0.0	0.0	0.0
12	89.5	5.7	4.8	0.0	0.0	0.0
13	97.1	2.4	0.5	0.0	0.0	0.0
14	93.8	4.9	1.4	0.0	0.0	0.0
15	98.6	0.0	0.0	0.0	1.4	0.0
16	99.1	0.0	0.0	0.0	0.9	0.0
17	96.3	2.5	1.3	0.0	0.0	0.0
18	94.3	1.4	1.4	0.0	2.8	0.0
19	95.1	1.8	1.8	0.0	1.2	0.0
20	94.0	2.4	3.6	0.0	0.0	0.0

DEGREE OF ROUNDNESS

By definition, roundness refers to angularity of particle edges and corners (Powers, 1951). Most clasts are assumed to begin angular and become more rounded with increased distance of transport.

The roundness of pebbles from Tallette Creek Valley was compared with samples from the kame field. Along the upper slopes, pebbles were consistently sub-angular to angular, whereas along Tallette Creek's lower gradient and the kame field, they were predominantly sub-rounded to rounded. The change occurs at the location of sample 11, where Tallette Creek flows on a more gentler gradient and across a wider flood plain.

In addition, separate roundness studies were made for each prominent bright lithology (limestone, quartz sandstone, chert). As might be expected, limestone pebbles were sub-rounded to well-rounded (possibly by solution). Quartz sandstone pebbles seemed to become slightly more rounded, while chert remained angular.

CONCLUSIONS

1. The highest concentrations (14 to 27 percent) of bright pebbles are found within the kame field, whereas values along Tallette Creek Valley are significantly lower (1 to 10 percent) (Figure 3). Therefore, drift within the kame field is "bright" relative to the "drab" drift of Tallette Creek Valley.

2. Pebbles transported down Tallette Creek show some degree of increased roundness.

3. The drift within the kame field is from two sources: (1) reworked upland drift transported as alluvium down Tallette Creek after the uplands became ice free (deposited as inwash on the Unadilla Valley ice-tongue), and (2) supraglacial meltwater transportation along the western edge of the valley ice-tongue led to deposition at the mouth of Tallette Creek Valley. Therefore, the kame field contains pitted outwash that has been supplemented by inwash from Tallette Creek

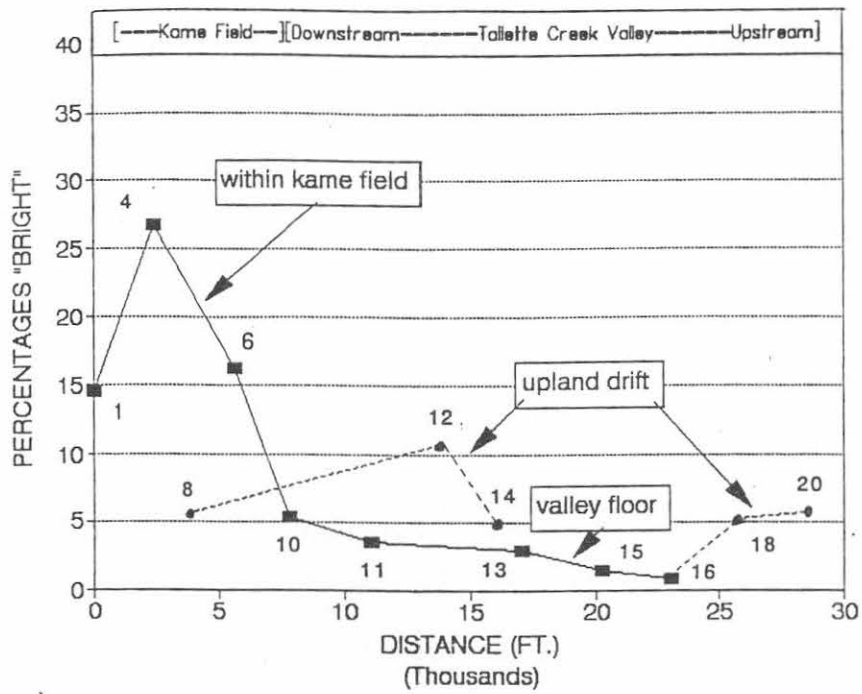

Figure 3. Changes in bright lithologies* along Tallette Creek. Plot of data (Figure 2) indicates the percentage of exotic lithologies is greatest within the kame field (14.6 - 26.7) and least along the valley of Tallette Creek (0.9-5.4). Values from upland drift vary between 4.3 and 10.5%. *All non-local rock type.

Road Log begins at intersecton of State Routes 7 and 205, West End, Oneonta.

Miles from last point	Cumulative Miles	
0	0	Intersection of Rt. 7 and 205, proceed north on Rt. 205.
1.1	1.1	Rt. 23 enters from right; continue north on Rt. 205 and Rt. 23.
.7	1.8	Bear left on Rt. 23.
9.2	12.9	Intersection of Rt. 23 and 51 in Village of Morris; continue straight through intersection on County Rt. 13 to New Berlin
7.2	20.1	Dead-ice sink on right
.8	20.9	Unadilla River
.2	21.1	Intersection of County Rt. 13 and Rt. 8; turn left (south) on Rt. 8
.4	21.5	Pull off to the right in gravel excavation and walk up dirt road (Angell Hill Road) to upper level of quarry

STOP # 1 - New Berlin: Through valley landforms - deltaic valley train and dead-ice sink. Large-scale deltaic foreset beds exposed here and within a currently inactive quarry in the center of the valley contain clast supported, very coarse gravels and pebbly, coarse sands; leached and reprecipitated carbonate (limestone) form partial cement in some units. This valley train is breached by the Unadilla River that flows from a dead-ice sink upvalley to a lacustrine plain downvalley. Landforms indicate detachment of large ice mass during active ice retreat.

		Turn around; proceed north on Rt. 8 through Village of New Berlin
.7	22.2	Intersection of Rt. 8 and 80 (traffic light), proceed north on Rt. 8 and 80.
1.3	23.5	Continue north on Rt. 8
.5	24.0	Road rises on gravel terrace above lacustrine plain
1.4	25.4	High-level terrace (1280') on west side of valley above road (continues for 1.5 miles) is of unknown origin.

2.2	27.6	Road rises on pitted planar gravel and kame field
.9	28.5	Columbus Quarters at intersection of Rt. 8 and Chenango County Rt. 41.

STOP # 2 - Columbus Quarters: kame field and pitted plain in through valley. These landforms, breached by the Unadilla River at its confluence with Tellette Creek, may have served to temporarily dam a local lake upvalley. Pebble counts from upland drab drift and kame field bright drift suggest tributary inwash from the Tellette Creek drainage basin was a partial source of sediment (see Case # 4 for discussion of pebble count data).

		Turn around, proceed south on Rt. 8.
1.9	30.4	At Lambs Corners turn left (east) onto Chenango County Rt. 25 (which changes to Otsego County Rt. 20 at Unadilla River), proceed east across lacustrine plain
4.2	34.6	Road descends onto the pitted and discontinuous valley train of Butternut Creek.
.6	35.2	Intersection of County Rt. 20 and Rt. 80, proceed east on Rt. 80.
3.0	38.2	Rt. 51 enters from the right, continue east on Rt. 80.
.4	39.3	Rt. 51 turns left, continue east on Rt. 80.
6.7	46.0	Rt. 205 enters from the right, continue east on Rt. 80 and 205.
1.7	47.7	Road descends onto the kame and kettle topography of the moraine at Oaksville.
.2	47.9	Intersection with Rt. 28, turn right, proceed south on Rt. 80 and 28.
.5	48.4	Road crosses the crest of the moraine and continues on and off the moraine for next 1.5 miles through the villages of Oaksville and Fly Creek.
1.7	50.1	Turn right (south) in Fly Creek on County Rt. 26, then bear left at fork.

1.2	51.3	Turn left onto gravel road, park and walk 1/4 mile into quarry.

STOP # 3 - Fly Creek: Valley train and dead-ice sinks in through valley. Sand and gravel valley train with classic assemblage of glaciofluvial sedimentary structures. What criteria may be used to distinguish outwash from inwash?

		Return to County Rt. 26, continue south.
2.3	53.6	Pull off on right shoulder, park and walk to crest of moraine.

STOP # 4 - Cassville-Cooperstown moraine at Index: This classic through valley landform (moraine and associated valley train) represents active ice deposition. It consists of stratified sand and gravel that grade upvalley into silt 120+ feet thick (see Case #1, Figure 2). Is it an end moraine, as proposed by Krall, 1977, or a recessional moraine? Although high enough, it did not dam Glacial Lake Cooperstown. Landforms and stratigraphy between here and Otsego Lake suggest detachment of a marginal-ice cleat and development of a dead-ice sink (see Case #1, Figure 4).

.2	53.8	Proceed to the intersection with Rt. 28, turn left (north).
.2	54.0	Dead-ice sink behind Cassville-Cooperstown moraine occupies the valley for the next mile.
1.7	55.7	Junction Rt. 80 and 28, proceed east on Rt. 80.
.3	56.0	Traffic light intersection with Main St., Cooperstown; continue straight through intersection east on Rt. 80.
.1	56.1	Stop sign, junction Lake St. with Rt. 80, turn right onto Lake St.
.3	56.4	Park at Intersection of Lake and River St.; stairway to Council Rock to the left.

STOP # 5 - Council Rock Park on Doubleday Ice Margin: Well data indicate 180 feet of bouldery silt beneath a moraine that plugs the valley and dams Otsego Lake. Subtle hummocky terrain in the village of Cooperstown can be traced westward to the golf course. While ice-cored, this landform dammed Glacial Lake Cooperstown at an elevation of 1250'. The view northward reveals conspicuous cross-sectional valley asymmetry, which is an important aspect of sediment source for Glacial Lake Cooperstown (see Case #1, Figures 5-9).

		Turn around and return to Rt. 80.

.3	56.7	Turn left onto Chestnut St. (Rt. 28 and Rt. 80 West).
.4	57.1	Junction Rt. 80 west, proceed south on Rt. 28.
2.1	59.2	Highway traverses Cassville-Cooperstown moraine.
1.0	60.2	Hamlet of Hyde Park; from here south for several miles, planar gravels appear related to upland tributary inwash sources.
5.3	65.5	Junction Rt. 166 at village of Milford (traffic light), continue south on Rt. 28. Valley floor consists of lacustrine clays from here south to a moraine at Portlandville at 68.9.
.9	71.6	Turn onto entrance road to new development in Milford Center
.2	71.8	Turn around, park.

STOP # 6 - Goodyear Lake overview at Milford Center: Pitted valley train and dead-ice sink. These landforms indicate local stagnation of the valley ice-tongue during active retreat. However, the valley fill consists of 60+ ft. of ice-contact stratified drift over 300+ ft. of silt interpreted to be of lacustrine origin. Does the stratigraphy indicate two stades (readvance) or is there a single-stade environment that accounts for both stratigraphic units? Aggradation over ground ice islands is suggested.

.3	72.1	Return to Rt. 28, turn south (right) onto Rt. 28.
1.8	73.9	Junction with Rt. 7, continue south on Rt. 28 to I-88 interchange.
.8	74.7	Rt. 7 overpass, dead-ice sink right and left.
.5	75.2	Turn right onto I-88 west.
2.5	77.7	Take Exit 16 (Emmons/Davenport Center)
.3	78.0	Stop sign, turn left toward Davenport Center on County Rt. 47.
.3	78.3	Cross Susquehanna River.
l.6	79.9	Delaware County line (Otsego County Rt. 47 becomes Delaware County Rt. 11).

I.0	80.9	Turn left in Davenport Center and continue past the Post Office on left. Road parallels Charlotte Creek through the moraine on the right.
1.9	82.8	Turn right into parking lot for Hartwick College's Pine Lake Camp. Walk downhill to pavilion.

STOP # 7 - Davenport Center: Pine Lake Dead-ice Sink Complex. These kames and sinks formed in an ice-cored terrain at the mouth of Charlotte Creek where the valley was clogged by remnant ice. Case # 3 considers the origin of landforms here. Also see Case # 3, Appendix A for the results of a pebble count study that tested an inwash origin for these materials.

		Return to road, turn right.
.4	83.2	Right hand fork goes downhill and across the lacustrine plain of Glacial Lake Davenport.
.4	83.6	Intersection with Rt. 23, proceed straight on Delaware County Rt. 10.
.2	83.8	Turn left into Town of Davenport Transfer Station, county gravel excavation

STOP # 8 - Davenport Center: pitted hanging delta. Current and past exposures contained well-developed, large-scale deltaic foreset beds in a hanging delta marked by kettles 40+ ft. deep. This indicates progradation onto and across grounded ice in Glacial Lake Davenport. See Case # 3 for discussion.

		Backtrack to Rt. 23.
.2	84.0	Turn left (west) onto Rt. 23.
1.9	85.9	Highway traverses moraine at West Davenport.
2.3	88.2	Terrain to the left is what may be the only lateral moraine in this area.
.7	88.9	Road descends to Susquehanna valley lacustrine plain.
1.1	90.0	Highway enters the breach of the Oneonta moraine (significantly altered by urban development).
1.3	91.3	Traffic light intersection I-88, 28 N/S, 23W; continue straight on 28 south.

.6	91.9	Turn right toward I-88 west.
.3	92.2	Turn left onto I-88 west.
2.0	94.2	Take Exit 13 (Morris and Route 205).
.3	94.5	Turn right (north) onto Route 205.
.2	94.7	This brings you back to the start of this log at Junction of Rts. 7 and 205.

OPTIONAL - continuation up non-through valley of Otego Creek.

0	0	Start at fork of Rt. 23 west (to West Oneonta and Morris) and Rt. 205 (Laurens) north, proceed north on Rt. 205.
4.2	4.2	Winnie Hill Road joins Rt. 205 from the right. Pull over and walk up Winnie Hill Road 0.1 mile for downvalley view south across local kame field.
.7	4.9	Turn left on County Rt. 11A (unmarked) toward Laurens.
.3	5.2	Turn left on County Rt. 11 (Maple Street) through village of Laurens.
.4	5.6	Bear left at fork (sign to Oneonta).
.8	6.0	Turn left into landfill entrance at cemetery, park.

OPTIONAL STOP # 9 - Village of Laurens landfill: ice-contact stratified drift. At various times, excavation has exposed tilted and truncated sequences of interstratified gravel and pebbly sand with fluvial sedimentary structures and local silt layers of lacustrine origin. This is interpreted to indicate semi-continuous aggradation of ice-cored terrain. Tributary inwash from the west is a possible sediment source.

		Return to road, turn right (north), backtrack through village of Laurens.
.8	6.8	Junction 11A and 11 (Maple Street), continue north on Maple Street.
2.1	8.9	Planar gravel, valley train remnant for next half mile.

.4	9.3	Turn right on unnamed road; sign indicates bridge closed.
.5	9.8	Turn left (.1 mile short of bridge) on unpaved gravel entrance road.
.3	10.1	Park and walk to various exposures.

STOP # 10 - Otsego County sand and gravel quarry: planar gravel remnant. This landform, typical of many others within the Otego Creek valley, is a discontinuous, planar gravel deposit interpreted as a remnant of a surface of aggradation that included semi-continuous ice masses as vestiges of a collapsed ice tongue. Ice-contact fluvial and lacustrine facies are common.

.2	10.3	Return to paved road (County Rt. 11).
.4	10.7	Turn right onto County Rt. 11, proceed north.
.9	11.6	Junction Rts. 11 and 11B, continue straight on 11. Road rises onto pitted planar gravel remnant.
1.0	12.6	Junction Rts. 11 and 15; bear right continuing on Rt. 11, road traverses terrain of ice-cored origin including a small esker out of sight to the right.
.3	12.9	Turn right onto Angel Road.
.6	13.5	Junction Angel Road and Rt. 205; turn left (north) onto Rt. 205 toward Hartwick.
.6	14.1	Pull off on left shoulder.

STOP # 11 - Overview of meltwater channel through discontinuous planar gravel remnant and terrain of ice-cored origin.

.5	14.6	Road descends to a local lacustrine plain (dead-ice sink?)
.7	15.3	Gravel excavation to the left removed a landform thought to be a proto-esker.
2.5	17.8	County Rt. 45 enters from the right, pull off on right shoulder.

STOP # 12 - Overview of discontinuous, planar gravel surface and terrain of ice-cored origin bordering a local, upvalley lacustrine plain.

		Proceed north on Rt. 205.
1.2	19.0	Village of Hartwick, Junction Rt. 205 and County Rt. 11, proceed north on Rt. 205.
1.3	20.3	Pull off on right shoulder.

STOP # 13 - Overview of headward area in non-through valley. Valley floor contains a few isolated kames. Unlike through valleys, the valley floor gradually rises to upland slopes across which ice flow diminished during glacial retreat resulting in downvalley ice-tongue starvation and collapse.

END OF ROAD LOG

Continue on Rt. 205 (north) 4 miles to intersection with Rt. 80 or return on Rt. 205 (south) 16.5 miles to the Junction of Rts. 23 and 205 where this optional trip began.