

TRIP A3

Glacial regime and depositional environments along the retreating Laurentide Ice Sheet, northeastern Appalachian Plateau, New York

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ABSTRACT

Deglacial environments along the retreating Laurentide Ice Sheet throughout of the upper Susquehanna region were primarily controlled by an ice front configuration draped across the high relief, Appalachian Plateau terrain. Valley ice tongues extended 20 km beyond upland positions commonly terminated in ice-contact lakes. Active ice flow nourished by the ice sheet persisted within through valleys, where as ice tongues in non-through valleys became detached and stagnated. Temperate conditions and hydrologic connection with the ice sheet supported highly turbid englacial and subglacial tunnel discharge directly into through valley ice-contact lakes. Subaerial streams favoring positions lateral to ice tongues transported sand and gravel outwash to prograding deltas. Thus, bottomset and toeset sediments interfinger with lacustrine silt and sand. Post-glacial dam incision followed by lake emptying led to the modern landscape. A finer facies of thick silt and sand lies beneath the modern floodplain and is typically bound by deltaic terraces (commonly called kame terraces) consisting of the coarser facies sand and gravel.

In contrast, downwasting over non-through valley headward divides led to loss of nourishment and near separation of the ice tongue from the ice sheet, thus causing the combined effect of hydrologic disconnection and stagnation. Consequently, non-through valley environments of deposition were less uniform, characterized by a) local ponding behind ephemeral ice-cored dams, b) kilometer-size, detached ice masses partially covered by outwash and lake sediments, and c) meltwater sources limited to the volume of the detached ice tongue. These collectively describe an environment significantly different from through valleys. Non-through valley sediments are far more diverse, less well sorted, and lack the laterally uniform subsurface conditions characteristic of through valleys. In addition, the landform assemble also differs through the lack of a continuous lacustrine plain, segmented terraces graded to local base levels, isolated small esker-forms, kame and kettle topography unrelated to systematic moraines (called kame fields), and mega-kettles that span the valley floor (known as dead ice sinks).

The Bering Glacier, Alaska, depicting sedimentary conditions in ice-contact lakes and along peripheral drainage systems is cited as an analog for the retreating Laurentide Ice Sheet in central New York.

INTRODUCTION

The Susquehanna River and its tributaries occupy glacially modified valleys that form an asymmetric, dendritic pattern of considerable antiquity on the coarsely dissected Appalachian Plateau (Figure 1). Major north-south valleys are oriented parallel to the general direction of overriding ice movement thus forming broad, U-shaped troughs separated by wide, low-relief divides. Many valley segments show the effects of glacier flow in oversteepened slopes, and several have asymmetric cross-

Figure 1. Index map.



sectional profiles, with steeper, west-facing and less steep, east-facing slopes, the cause of which appears unrelated to glacial erosion. Furthermore, the gently dipping (<10 degrees SW) Middle Devonian bedrock strata only have subtle expression in the topography, thus explanation for asymmetry remains elusive. Valleys oriented orthogonal to glacier movement show lesser effects of glacial erosion, although many have steep slopes assumed inherited from pre-glacial fluvial incision (Fleisher 1977). Local relief ranges between 250-300 m, but valley water wells (Randall, 1972) frequently penetrate 100-125 m of Quaternary sediment, thereby indicating that bedrock relief is considerably greater, averaging 300 m and reaching 400 m in some locations. The ice-marginal configuration of a temperate glacier draped across terrain was strongly influenced by this topographic relief. Depositional environments at the snout of ice salients within each valley were in significant contrast with those on adjacent uplands. The resulting landforms quite different from classic mid continent features formed along regionally extensive ice lobes.

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 effect 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 Melia (1975) and Fleisher (1977a). Fleisher (1984, 1985), used driller's logs (water wells) and landform expression as indicators of proglacial ice-contact lakes, proposed a late glacial lake chain throughout the upper Susquehanna drainage basin. Interpretation of chronology is hampered by the lack of datable materials. However, Krall (1977) interpreted the Cassville-Cooperstown Moraine (5 km south of Otsego Lake) to be topographically correlative with moraines in the Hudson Valley, thereby suggesting a readvance circa 14,000 years BP. However, Ridge et al. (1991) tentatively correlated the Cassville-Cooperstown Moraine with the West Canada Readvance of Late Wisconsinan, pre-Valley Heads age (circa 15.5 ka) based on elevation projections of ice-marginal deposits in the Mohawk Valley and the limit of readvance inferred from subsurface stratigraphy (Fleisher, 1986).

Coates (1974) defines the "Through Valley Section" of the Appalachian Plateau to be characterized by the valley of Otsego Lake at the head of the Susquehanna drainage basin. In this context a through valley may be depicted as a scoured glacial trough, oriented parallel to the regional ice flow direction, and "open-to-the-north" as a hanging valley on the northern plateau escarpment. Unlike non-through valleys that conventionally rise to upland divides, the head of the through valley drainage basin is on the valley floor where it separates streams flowing north out of low, poorly drained valley bogs and swamps from those that leave the same bogs and swamps but flow southward. The open through valley trough extends northward beyond the limits of the water divide. The valley walls do not converge, but rather remain on opposite sides of the valley where they gradually separate to join the escarpment slopes that define the northern boundary of the Appalachian Plateau (Figure 1).

As Laurentide ice moved southward from the southern Adirondack terrain and across the Mohawk Valley, it passed from crystalline terrain to an Appalachian Plateau substrate dominated by fine-grained, lower Paleozoic siltstone and shale. The combined influence of high topographic relief and a deformable substrate led to the development of valley ice tongues 20 km long (Fleisher, 1993).

MacNish and Randall (1982) used water well and boring log data to establish Quaternary aquifer properties, and diagrammatically illustrate generalized subsurface information and landforms to interpreted rate (rapid versus slow) and mode of retreat (active versus stagnant). Fleisher (1991 a, 1991b, 1993, 1986a) established a link between glacial regime and physiographic settings by recognizing that landforms produced by downwasting and ice-lobe collapse within non-through valleys differ significantly from those formed during active ice backwasting in through valleys, thereby supporting the MacNish and Randall association of retreat regime with landforms and subsurface stratigraphy.

The primary purpose of this paper is to present potential sediment sources, transport mechanisms, environments of sediment accumulation, and landform development using analog conditions of contemporary processes at selected glacial ice margins in Alaska.

VALLEY ICE-LOBE MODEL

The traditional view of the deglacial history of this portion of the northern Appalachian Plateau suggest regional backwasting of a more or less continuous ice margin of the Laurentide Ice Sheet during pre-Valley Heads (~18-15.5 ka) recession, with isolated areas of downwasting (Krall, 1977) . The ice front configuration is portrayed to consist of relatively short "ice-lobes" in pre-existing valleys (Cadwell, 1972; Fleisher and Cadwell, 1984; Fleisher, 1986a). Originally proposed by Moss and Ritter (1962), the ice-lobe model illustrates an ice margin in the form of topographically-controlled, short, valley lobes (a few kilometers in length) extending from upland recessions (Figure 2).

Using a combination of driller's logs from water wells and test borings and glacial landforms, MacNish and Randall (1982) developed diagrammatic illustrations of ground water distribution that combined rate of retreat (rapid or slow) with mode of ice-lobe activity (backwasting or downwasting). Fleisher (1991a, b) recognized that landforms produced by downwasting and ice-lobe collapse within non-through valleys differ significantly from those formed during active ice backwasting in through valleys, thereby supporting the MacNish and Randall association of retreat regime with landforms and subsurface stratigraphy.

Shortcomings of the short ice-lobe model

The early ice-lobe model has been useful in the development of ideas and concepts related to the origin of glacial landforms and environments of deposition. However, it fails to account for several puzzling aspects of regionally extensive deposits and does not portray sufficient details pertaining to depositional environments. For example, it does not address the following observations; 1) different landform assemblages in different valleys, 2) kame-moraines lose topographic expression on valley walls and across divides, which defies confident correlation from one valley to the next, and 3) moraines predominantly consist of crudely-sorted and stratified sand and gravel (some till) that interfinger with thick lacustrine silt, yet well-developed strandline features are uncommon. Furthermore, the model fails to identify sediment sources and transport mechanisms, and does not offer a basis for interpretation of anomalous landforms and stratigraphy. For these reasons, Fleisher (1993) proposed a modified ice-tongue model that takes into consideration the influence of a deformable bed on the ice surface gradient, and consequently the length of the ice tongue.

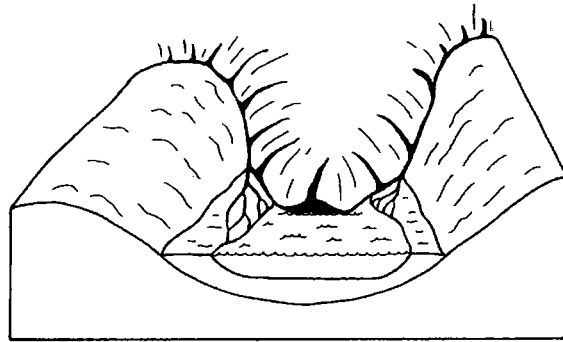


Figure 2. Short Ice-lobe model.

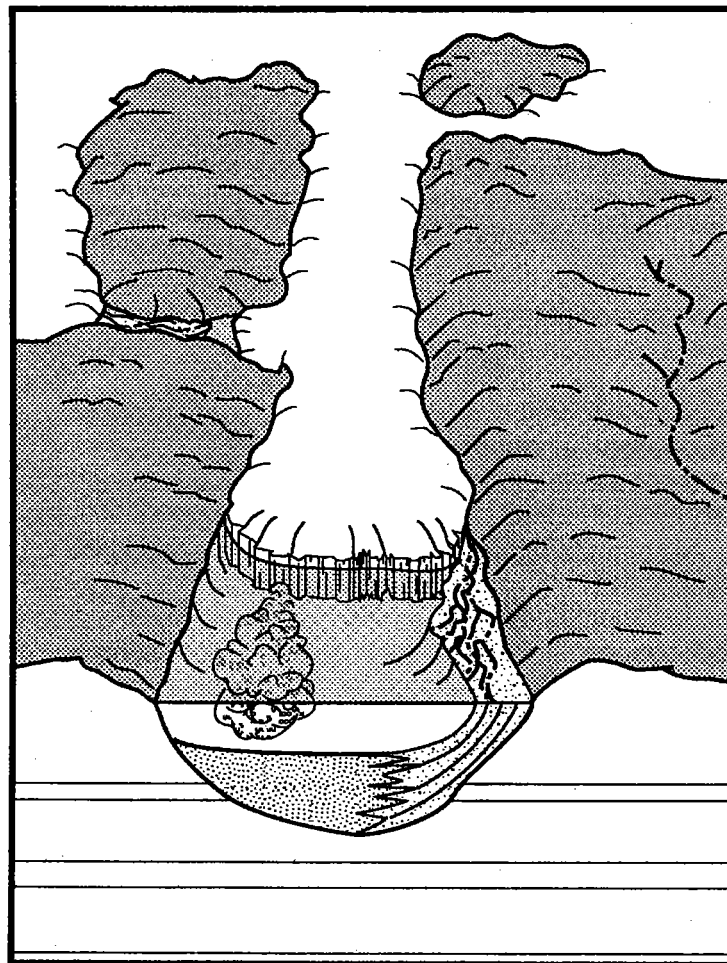


Figure 3. Modified ice-tongue model.

Extended ice-tongue model

As Laurentide ice advanced southward from the crystalline terrain of the southern Adirondacks, it crossed the Mohawk Valley and ascended the north-facing escarpment of the Appalachian Plateau where 450 to 600 m of Ordovician and Silurian shale and siltstone (60-

70% of the stratigraphic column) are exposed. Within the eastern Susquehanna drainage basin, lower and middle Devonian lithologies contain an additional 485 m of shale and siltstone that dip gently to the south-southwest (Rickard and Zenger, 1964).

Boulton and Jones (1979) present evidence to suggest "the glacier profile is related to the hydraulic and strength properties of potentially deformable bed material". Theoretical modeling predicts the strain response of a glacial substrate as critical shear stress exceeds sediment strength and subsequent initiation of sediment deformation. Shear stress and confining pressure are resisted by sediment strength (i.e., cohesion and frictional strength) (Benn and Evans, 1998; Boulton and Hindmarsh, 1987; Alley, 1989; Tulaczyk, 1999).

Water conditions at the sole of temperate glaciers provides maximum pore-water pressure and diminished intergranular friction in the presence of maximum shear stress. For fine-grained sediment, from which water does not readily escape, pore-water pressure builds, thus causing diminished strength. At a critical strain rate, dilation leads to ductile flow and pervasive deformation (Alley, 1989; Benn and Evans, 1998). Conversely, coarse, well-drained sand and gravel facilitate water escape thus reducing or precluding the potential for deformation (Boulton et al., 1974; Boulton and Hindmarsh, 1987).

The glacial substrate on the Appalachian Plateau 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 into the Susquehanna drainage basin. As modeling indicates, with low permeable materials, water pressure builds and bed deformation follows. Thus, the development of positive pore pressure within saturated bed material facilitated deformation and the development of a lower equilibrium profile (flatter ice surface gradient). Consequently, the ice-tongues may have reached lengths in excess of 20 km (Figure 3) (Fleisher, 1993). Calculations leading to this assume specific quantitative conditions that may be more precise than accurate for the purpose of developing a conceptual glacial model. Therefore, the most important aspect of the modified model is not the finite length of ice tongues, but rather the presence of valley ice adjacent ice-free upland slopes on which inwash process may have been active.

GLACIAL REGIME AND LANDFORM ASSEMBLAGES

Glacial regime

The Susquehanna River drains the eastern Appalachian Plateau in an asymmetric, dendritic pattern of glacially enlarged north-south oriented valleys and less well developed east-west tributaries (Figure 1). Gently undulating upland divides separate broad through and non-through valleys with local topographic relief of 200-330 m. Regional water well data (Randall, 1972) indicate that valley floors are underlain by stratified drift that commonly exceeds 100 m in thickness, which means the bedrock relief approaches 400 m. This magnitude of relief influenced the configuration of the retreating Laurentide ice margin, thus forming 20 km long ice-tongues in valleys oriented parallel to the ice flow direction (NNE-SSW). The dynamics of long ice-tongues in this terrain would favor stagnation and the development of detached remnant ice masses where nourishment from the ice sheet was restricted or limited. Evidence of this may be found in a) through valleys where active ice flow persisted during backwasting, in b) non-through valleys where nourishment to ice-tongues was restricted by thinning ice across upland divides, thus leading to downwasting, and in c) transverse valleys deprived of ice tongues by virtue of their orientation. This regime yielded two distinctly different landform assemblages, one representative of active ice retreat in through valleys and another depicting widespread stagnation and downwasting in non-through valleys and transverse valleys.

Through valley flow regime and landform assemblage

Observed detachment of ice masses along steeply rising internal structures (taken to be deep seated thrusts or shears) within the termini of several Alaska glaciers (Mulholland, 1982; Fleisher, 1993) lead to the concept of an ice-marginal "cleat" upon which active ice ramps, thus serving as a mechanism for ice mass separation within active ice retreat. It is suggested that similar ice-marginal cleats measuring ~70 m thick, 1 km wide and 3 km long formed semi-continuously during backwasting (Figure 4). Subsequent burial by both outwash and inwash would yield remnant ice masses and ice-cored valley train.

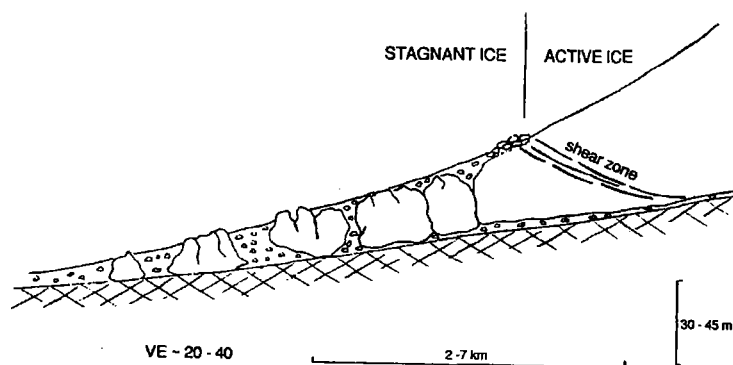


Figure 4. Detached and stagnant ice. (modified from Mulholland, 1982)

Through-valley landforms primarily consist of kame moraines, pitted and dissected valley train, kame terraces, kame fields, lacustrine plains, and occasional dead-ice sinks (Figure 5). Kame moraines typically consist of poorly-sorted and crudely-stratified, matrix-supported silty gravel, with ice-contact collapse structures. Boulders are well rounded and lack the effect of glacial abrasion. Data from wells and borings indicate that moraine gravels interfinger with lacustrine silts on the stoss side of moraines, which suggest that moraines served as dams for proglacial lakes. Moraines rise 25-30 m above the floodplain and modern streams that breach them. Their conspicuous kame and kettle topography typically reaches 10-15 m of local relief and occupies the full width of the valley floor. However, their use as morphostratigraphic units in regional correlation is limited because their topographic expression fades significantly on valley walls and is absent from uplands.

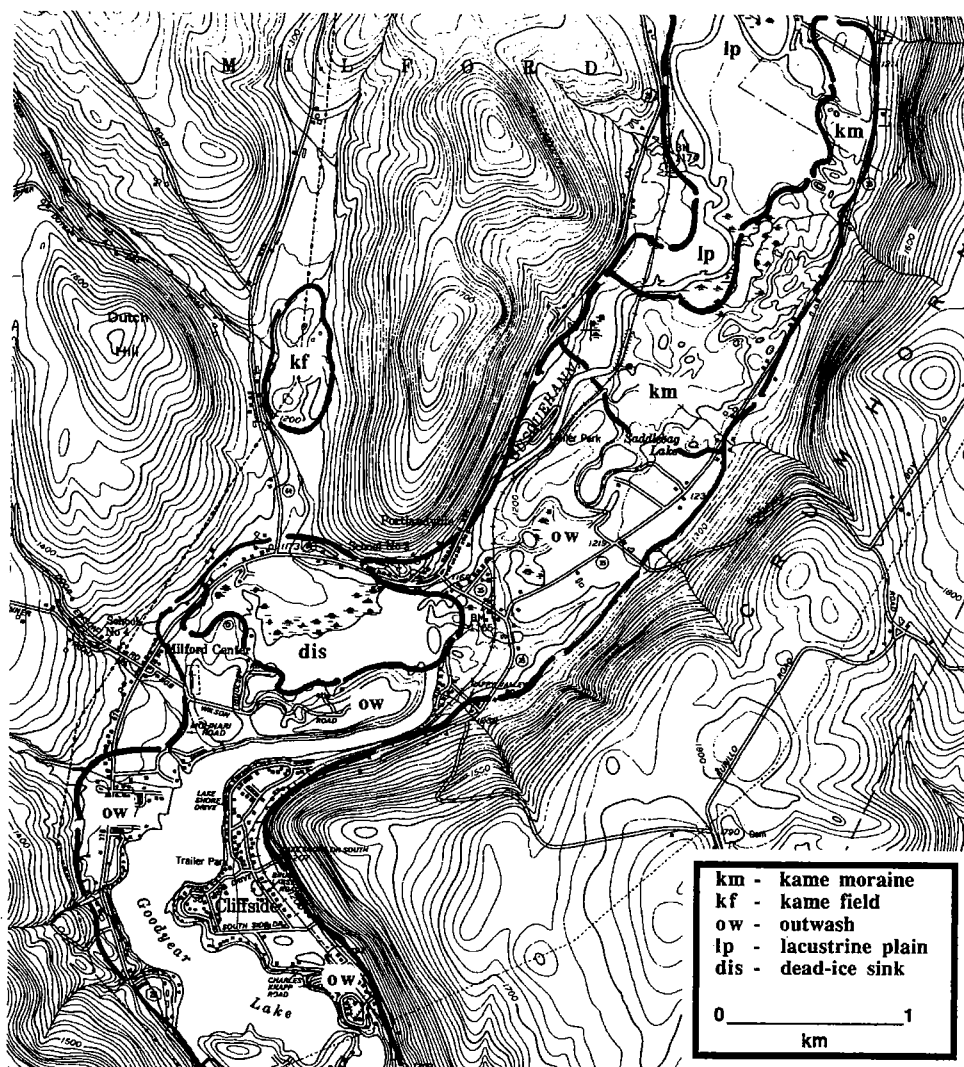


Figure 5. Through valley landform assemblage.

Pitted and dissected valley trains are often graded to kame moraines, but are also known to exist independently. They are found 20-25 m above the modern floodplain and represent the highest planar deposits in the valleys. Therefore, they have been referred to as high outwash as a means of distinguishing them from similar deposits at lower elevations which are called low outwash (Fleisher, 1986a). Most outwash is well-sorted, clast-supported, coarse, sandy gravel interstratified with layers and lenses of less well-sorted, matrix-supported silty gravel.

Kame terraces stand 25-30 m above the modern floodplain in paired and non-paired segments and may be topographically indistinguishable from dissected valley train. They are

typically less than 2 km long and commonly contain well developed deltaic internal structure, although associated lacustrine plains are not consistently present. Topsets are rather poorly-sorted and may show only the most subtle indication of imbrication, channel structures, and cross bedding. Deltaic foresets are massive in scale, occupy the full height of the terrace beneath topsets, and indicate progradation toward the axis of the valley. Texture ranges from coarse, pebbly sand to lenses of very coarse, matrix-supported gravel. A sandy and silt-rich toeset facies is occasionally exposed in borrow pits. Kame terrace segments found at tributary mouths are at grade with dissected hanging deltas.

Portions of most main valley floodplains contain broad and extensive, low gradient reaches across which exaggerated meanders flow. This topographic setting characterizes lacustrine plains throughout the region. Water well data indicate the terrace sands and gravels interfinger laterally at depth with 50-120 m of sand, coarse silt (locally called quicksand) and lesser amounts of clay, all of which are assumed to be of lacustrine origin (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 beneath extensive lacustrine plains.

The term dead-ice sink has genetic significance by simultaneously implying a condition of stagnation (e.g. dead-ice) and subsequent sediment accumulation (e.g. sink). It is a valley-floor feature characterized by an anomalously broad floodplain confined, upvalley and down, by valley train terraces. It originates in much the same way as a kettle, by collapse over buried ice. However, in this case, the buried ice is sufficiently massive to occupy the entire width of the valley, thus the dead-ice sink is a large-scale-kettle. Unlike a conventional kettle that forms a depression within a landform, the dead-ice sink is large enough to be a landform. Formation involves the initial detachment and burial of a large, remnant ice mass (several kilometers in diameter and upwards of 100 m thick) beneath an insulating veneer of outwash, which serves to retard melting during the deglacial processes (Figure 6). As pointed out by Mulholland, ice-block detachment may be a normal part of the backwasting mechanism. Should complete melting of the detached ice occur prior to retreat from the Appalachian Plateau, the developing dead-ice sink would be filled by outwash sediment as it develops, thus precluding the formation of a surface depression. However, buried ice masses that survive retreat from the Appalachian Plateau and the diversion of meltwater to the Mohawk drainage continue to slowly melt, leading to gradual subsidence and collapse without the addition of significant overlying sediment. The resulting valley-floor depression serves as a sediment sink for late glacial and post-glacial accumulation, which is reflected by subsurface data in the logs of water wells and test borings (Fleisher, 1986a).

Non-through valley flow regime and landform assemblage

Fleisher (1986) proposed stagnation of an entire non-through valleys ice-tongue by the progressive development of a negative ice budget due to restricted flow in thinning ice on headward divides. Figure 7 illustrates how ice-tongue starvation developed in non-through valleys, whereas active ice movement within through valleys (open-to-the-north) was sustained.

Non-through valleys contain a significantly different landform assemblage dominated by kame fields, isolated kames, and segments of discontinuous, remnant gravel plains, eskers, and dead-ice sinks, as noted along Otego Creek (Figure 8). The topographic expression of a kame field is similar to that of a kame moraine, but with significant differences. A kame field has limited lateral extent and is typically found to

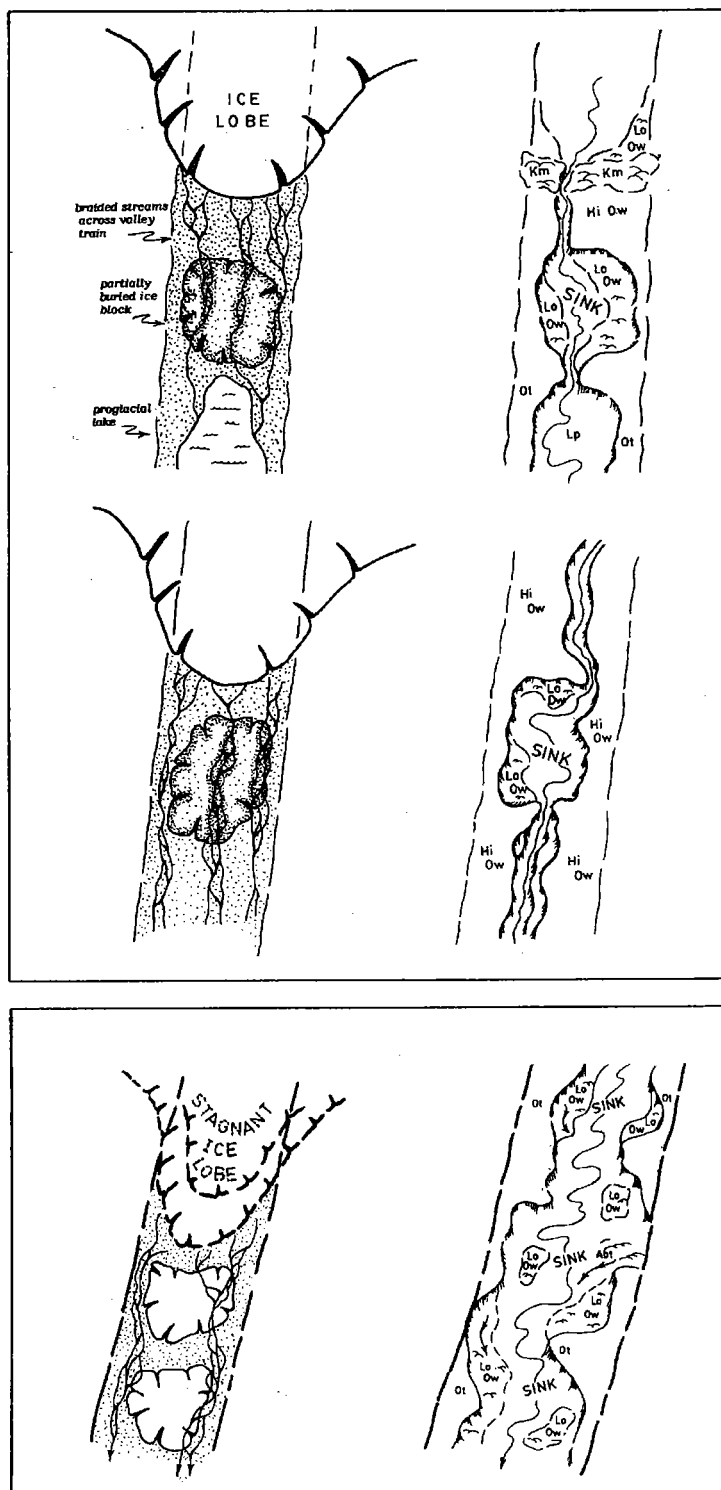


Figure 6. Development of dead-ice sinks.
 A - active ice in through valley.
 B - stagnant ice in non-through valley.

be lacking association with a valley train. Although kame fields are far more common in non-through valleys, they also occur in through valleys. They consist of poorly-sorted, silt-rich, matrix-supported gravel interstratified with lenses of diamict, well sorted gravel and sand beds, which are often disturbed by small scale collapse structures (Fleisher, 1984, 1986a). Kame fields are commonly found at the confluence of upland tributaries and main valleys, as are other landforms of ice-cored origin.

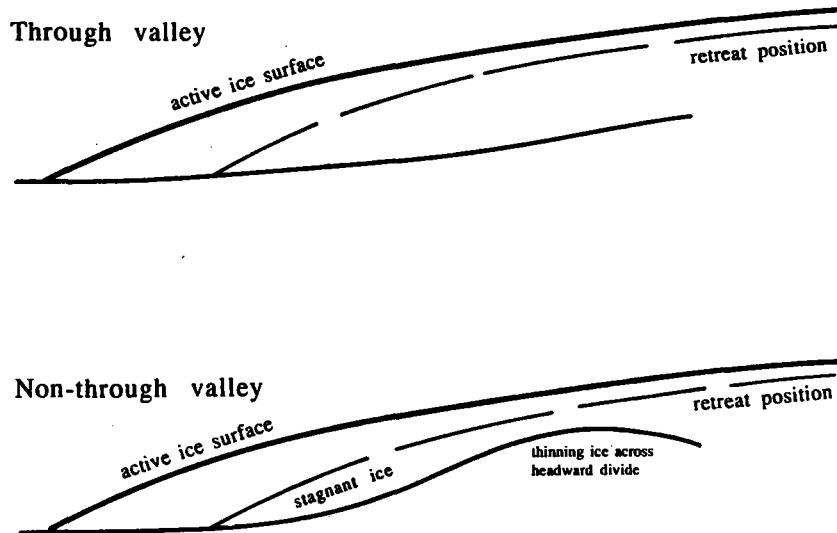


Figure 7. Schematic longitudinal profiles in through valley and non-through valley.

Transverse valley landform assemblages

Valleys of this type are recognized by their transverse to sub-transverse orientation to the regional ice flow direction. They are prominent components of the regional stream pattern, but show no evidence of having been fed by active ice-tongues. Although some tend to be significantly smaller than through and non-through valleys, they have comparable bedrock relief and are significantly larger than upland tributaries.

Deglacial thinning over up-glacier divides would cause depleted nourishment and eventual separation of ice masses hundreds of meters thick in transverse valleys. Burial beneath outwash and inwash would retard melting and lead to the formation of various late glacial and early post-glacial landforms of ice-cored origin, without additional sedimentation by proglacial outwash.

Three different landform assemblages have been identified in transverse valleys. They are 1) kames, kame fields and limited gravel plain segments, amidst small lake plains (short-lived, local base level) and dead-ice sink complexes, as noted along the West Branch Delaware River between Delhi and Hobart, and the valley of Schenevus Creek. 2) large lake plains, small hanging deltas, alluvial fans and occasional kames have been mapped along Charlotte Creek between Davenport Center and Butts Corner, and 3) till plugs, isolated kames, small lake plains and discontinuous pitted terraces as may be seen in the valley of Ouleout Creek.

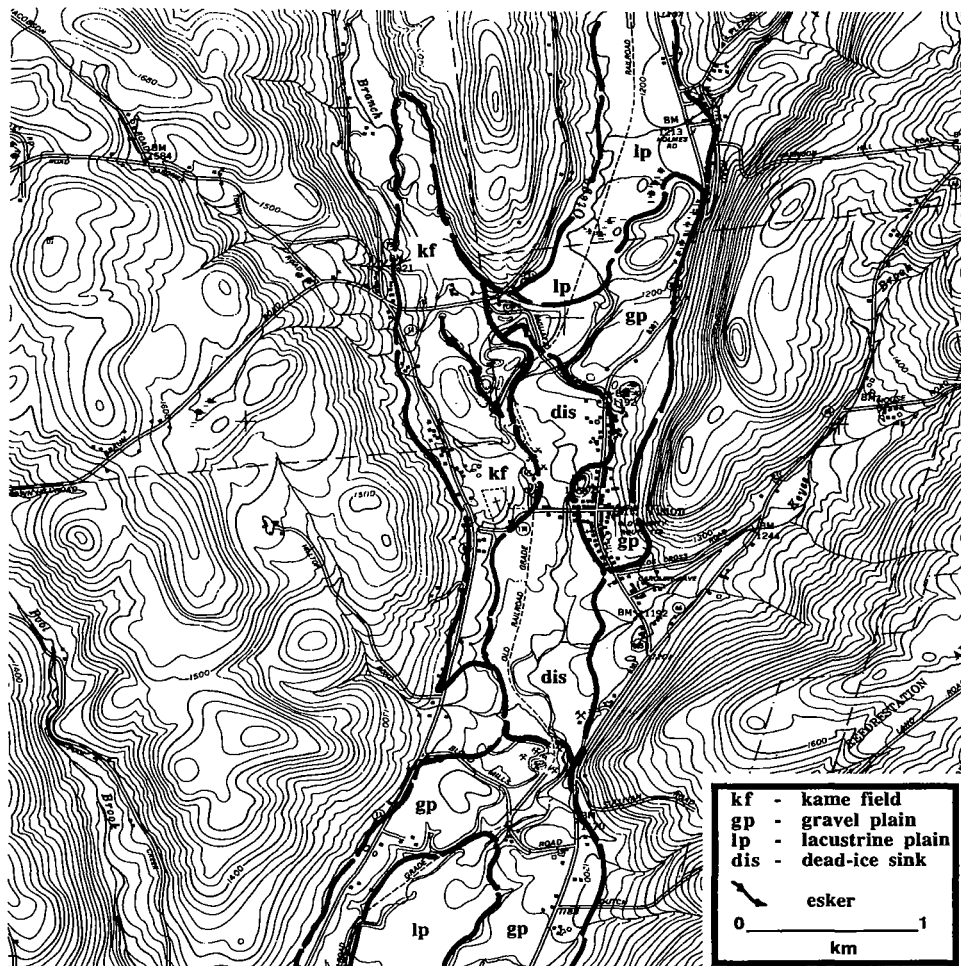


Figure 8. Non-through valley landform assemblage.

Pebble count Analysis - Method

Samples from borrow pits in kame terraces were taken by random selection of pebble and cobble-size clasts from multiple stratigraphic units at depth. Upland samples were obtained from hand-dug holes. After the samples were washed, screened, a representative number (in most cases more than one hundred) were broken, lithologic identification was made under a binocular microscope. The lack of a universally applied, standard sampling procedures may result in diverse pebble count data from multiple authors (MacClintock and Apfel, 1944; Merritt and Muller, 1959; Moss and Ritter, 1962; Coates, 1963; Denny and Lyford, 1963; Randall, 1978), thus limiting comprehensive interpretation. For example, in this study limestone and chert are technically not exotic materials because they are exposed in the headward reaches of

most through valleys. Yet, they are grouped with crystalline exotics because both indicate glacial transport from remote areas. Although the lack of uniform sampling and analytical procedures diminishes the value of pebble count data, they do lend themselves to basic generalizations.

Summary of pebble count data

A summary of pebble count data arranged by sample site location appears in Table 1. Rock type categories represent contrasting lithologic suites. Percentage values vary broadly and include anomalous highs that tend to skew averages, but not beyond useful limits,

	Through Valleys			Transverse Valley	
	Unadilla Valley	Susquehanna Valley	Oaks Creek Valley	Cherry Valley	Charlotte Creek Valley
number of sites sampled*	(12)	(12)	(6)	(7)	(8)
	%	%	%	%	%
local lithologies*	73.5	85.9	51.5	80.0	98.0
quartz-rich lithologies	9.0	1.1	4.8	0.8	—
limestone and chert	17.0	12.3	40.6	17.5	1.0
crystalline lithologies	0.5	0.7	3.1	1.7	1.0
	Non-through Valleys		Uplands		
	Butternut Valley	Otego Creek Valley	Unadilla Upland	Charlotte Creek Upland	
number of sites sampled*	(7)	(6)	(13)	(9)	
	%	%	%	%	
local lithologies*	95.9	93.1	95.6	99.8	
quartz-rich lithologies	1.2	2.1	2.9	—	
limestone and chert	2.0	1.5	1.3	0.1	
crystalline lithologies	0.9	3.3	0.0	0.1	

a minimum of 100 pebbles were counted at each sample site
 * includes shale, siltstone, mudstone, and graywacke

Generalizations derived from pebble count data

Several generalization may be derived from these data.

1. Through valley stratified drift contains a conspicuously higher percentage of exotic lithologies than drift in all other locations. This is primarily due to the greater abundance of limestone/chert, and light colored crystalline rocks, thus giving rise to the term "bright" drift. This is in contrast to non-through valleys and uplands, where "drab" drift prevails due to the concentration of clasts from local sources.
2. Data from through valleys show greater variation than non-through valleys and uplands.
3. Drift within non-through valleys is only slightly less "drab" than upland drift, containing a significantly higher proportion of local lithologies than brighter, through valley drift.
4. Virtually all upland drift appears derived from local bedrock sources.
5. The lithologic suites of transverse and non-through valleys are similar to upland drift.

Pebble count anomalies

Limestone outcrops at the heads of through valleys had a significant influence as a sediment source. This is demonstrated by exceptionally high pebble count values (27.2% and 56.4%) within 15-20 km of Onondaga Limestone outcrops in the valleys of Oaks Creek and the Susquehanna River. However, conspicuously fewer pebbles of limestone and chert in Cherry Valley suggest that factors other than distance from sediment source are also significant. Similar variations in pebble count data are also noted in Unadilla Valley near New Berlin. Five pebble counts from two sites depict a "bright" valley train (31.7% exotics) in the vicinity of New Berlin, whereas seven counts from the kame field at the mouth of Tallette Creek (a major tributary 6 km upvalley) are significantly less bright (8-16% exotics) (Yukinevicz Master's thesis). Local variations in data are also noted at the confluence of Oaks Creek and the Susquehanna River south of Cooperstown where exotics vary by 24% at two sample sites a few kilometers apart on the same moraine in adjacent valleys. Pebble count values for exotics in a few isolated locations along the Susquehanna River and Otego Creek are anomalously high. And, local lithologies in Charlotte Creek Valley (transverse valley) occur to the virtual exclusion of exotics.

Provenance

Pebble count data from the Cooperstown area of the upper Susquehanna Valley provides evidence in support of glacial transport during kame moraine formation. Here, the Cassville-Cooperstown moraine, situated 5 km south of Cooperstown, contains 41.5 % limestone (Melia, 1975). The nearest limestone source is 20 km upvalley where outcrops of Onondaga Limestone are exposed at an elevation 100 m higher than the moraine and 175 m above the bedrock valley floor beneath the moraine. With the ice-tongue margin at the moraine and an assumed basal shear stress between 0.5 and 1.0 bar, the surface elevation of the glacier would be 250-400 m higher than the limestone outcrops, which precludes a superglacial source. Neither downvalley transport by meltwater flow nor post-glacial fluvial processes could account for these deposits because sediment movement through the Otsego Lake basin would not have been possible following retreat from the moraine. Two possible alternative are 1) limestone plucking followed by transport as basal load to the ice front where rising shears incorporated it in supraglacial debris at the moraine site or 2) transport by pressurized water moving through englacial and subglacial conduits exiting the glacier at the moraine site. The high concentration of limestone exotics in the Cassville-Cooperstown moraine not only indicates that bright drift originates from upvalley sources, as suggested by Moss and Ritter (1962), but also demonstrates the importance transportation of glacier sourced material within the ice.

Consistent with observations by Moss and Ritter south of the Valley Heads Moraine, data obtained from drift along Oaks Creek reveals a rapid downvalley decrease of limestone pebbles, which indicates that nearly all (97%) glacial transport is limited to distances of 10-15 km from the bedrock source, as suggested by Halter et al., (1984). However, 1-3 % exotics within upland drift

throughout the region (including crystallines that must have originated 40 to 100 km or more to the north) were either carried much farther or derived from re-worked older drift. Holmes (1952) relates distance of transport to lithologic resistance to abrasion and crushing, noting the virtual lack of shale pebbles and cobbles beyond 6-7 km of a bedrock source, whereas well lithified sandstones remain as conspicuous till components 130 km from their outcrops.

Additional sediment source information is derived from a kame field and associated remnants of a pitted surface that occupy the Unadilla Valley at the confluence of Tallette Creek, a major tributary 5 km north of New Berlin. The lithologic suite of Tallette Creek is typical of upland, drab drift, which is in sharp contrast with bright valley drift. Here, and at many tributary confluences throughout the region, the kame field is thought to have developed from downwasting of ice-cored inwash. The drab drift of Tallette Creek yields abruptly to bright valley drift at the kame field, which indicates its upvalley source. However, the normally high percentage of exotics found in two sites 6 km down valley appears diluted by drab inwash from Tallette Creek, which is consistent with an inwash origin for the kame field material.

Additional evidence in support of the inwash mechanism as a source of drift is noted in a hanging delta at the mouth of Kortright Creek, a primary upland tributary to Charlotte Creek. Here, collapsed topset and foreset beds and large kettles in the delta surface indicate a near-ice or ice-contact origin. In contrast with pebble count data from through valleys of the Susquehanna system, exotic lithologies are very sparse and occur no more frequently than in upland drift. This unique landform must consist of tributary inwash that was prograded across grounded ice in Glacial Lake Davenport (Fleisher, 1991b). An associated kame field, with a similar lithologic suite, could not have derived sediment from outwash sources in the Susquehanna and is also thought to be of inwash origin.

With the exception of valleys oriented semi-parallel to the ice margin, from which ice tongues would have been excluded, it appears that most coarse valley drift consists of different proportions of two basic components; 1) re-worked drab drift off the uplands and b) re-sedimented outwash from upvalley ice-tongue sources. The relative proportion of each would be a function of specific depositional conditions in each individual valley. For example, Cherry Valley data includes an abrupt contrast in limestone (from 34.4 to 8.1 %) in sample sites from opposite ends (upvalley and down) of a lacustrine plain. This suggests an interruption of glaciofluvial transport by a proglacial lake basin, similar to that proposed north of the Cassville-Cooperstown moraine. Likewise, anomalous data from another areas, such as Charlotte Creek valley, indicate that local processes had defining influence on the source of sediment.

In summary, pebble counts reveal that non-through valley drift consistently contains a lower percentage of exotics, which supports the notion that upland sources dominated. High exotic values are typical of bright drift thought to have been derived from upvalley glacial and outwash sources within through valleys. Very high values are restricted to areas within 10-15 km of a limestone bedrock source, but local exceptions include a few distant, isolated occurrences in both through and non-through valleys.

DISCUSSION

Transportation mechanisms and sediment sources

Fundamental to the modified ice-tongue model are 20 km long extensions of the ice front within all through valleys and large non-through valleys, with adjacent ice-free upland slopes. Under these conditions, Evenson and Clinch (1987) suggest that there are two fundamental sediment sources; 1) materials overridden and picked up by the glacier (glacier sourced) and 2) material brought to the glacier by fluvial and slope processes from adjacent ice-free upland sources (inwash sourced). The glacier is the transport agent for material incorporated as bed load, as well as that which accumulates as superglacial debris. All remaining material is transported by fluvial

processes within englacial and subglacial conduits (Gustavson and Boothroyd, 1988), subsequently accumulating in a variety of depositional environments at the glacier snout. Lawson (1979) maintains that only 5 % of glacially derived sediment is actually deposited from the ice. The remainder is attributed to a process called re-sedimentation involving transportation and deposition by various forms of mass movements and fluvial processes (Lawson, 1979; Evenson and Clinch, 1987).

Landform mapping and water well logs from the upper Susquehanna region indicate that sand and gravel are concentrated within moraines, kame fields, valley trains and kame terraces, whereas fine sand and silt occur beneath lacustrine plains. Although fine and coarse sediments interfinger laterally and are the result of single-stade deglaciation (Fleisher, 1987), they represent two distinctly different environments involving more than one transport mechanism and several possible sources. Although definitive conclusions may not be drawn solely from existing data, systematic and uniform differences in lithologic suites suggest that the dominant transport mechanisms in through valleys differ from those in non-through valleys. For example, active ice-tongue flow would favor glacial transport accompanied by high meltwater discharge. Mechanical attrition of poorly consolidated local, drab lithologies by glaciofluvial transport, as speculated by Holmes (1952), combined with enrichment of exotics from upvalley sources would brighten through valley drift.

However, in non-through valleys, reduced ice flow over northern divides led to ice-tongue collapse, local stagnation, diminished meltwater discharge, and reduced downvalley transport. In effect, non-through valleys were ultimately isolated from bright drift source areas. As deglaciation progressed, inwash from ice-free uplands became increasingly significant.

Sources of Silt

The silt-rich upland drift must be considered as a potential source of lacustrine silt found within most valley fill. The total volume of valley fines was calculated from well data and compared with the area of adjacent uplands to determine the degree of upland denudation that would have been required to supply an equivalent volume of silt. The main valleys of the entire upper Susquehanna contain approximately 14 cubic miles of fine sediment fill, which when distributed as a uniform blanket over the adjacent uplands (1720 square miles) would add an average of 16 m to the existing upland drift mantle. No field evidence exists to support this degree of general dissection. Furthermore, winnowing by upland runoff would have produced abundant residual boulder lag, which is also lacking. Yet uplands must have been subject to late glacial and early postglacial erosion as indicated by incision along primary tributaries. Although slope wash, mass wasting, and re-sedimentation by tributary inwash certainly account for some of the fines, most of the silt must have been derived from the only remaining source - the glacier.

Boulton and Hindmarsh (1987) argue that because fine-grained, deformable, subglacial sediment would not be capable of draining a constant meltwater influx, subglacial channels and conduits form to transmit excess water. As water discharge increases, so does the piezometric gradient, which in turn raises water pressure values within the sediment to equal the ice overburden pressure. As Boulton and Hindmarsh suggest, this leads to sediment liquefaction in the glacier terminal zone, which, in turn, causes a "flow of liquefied sediment into the proglacial environment". Liquefaction of fine sediment in combination with subglacial migration of the saturated substrate under pressure would force fines into the subglacial hydrologic system of conduits that exit at the glacier snout, and in this case directly into ice-contact lakes. This is proposed as the primary source and transport mechanism of lacustrine silt (see Figure 3).

Sources of sand and gravel

Well data indicate sand and gravel represent only 25 % of the total valley fill, the remainder being silt and quick sand. Although dark, coarse sands are known from exposures thought to be at depths near or at the spring high water table, most sand and gravel consists of interbedded bright sand and very coarse, clast-supported gravel. Based on pebble count data reported here and by

Moss and Ritter (1962), this drift is much too bright to have had an upland source, and is, therefore, thought to have originated from upvalley sources that ultimately derived debris directly from glacier ice.

Sedimentary structures (ripple marks, cross-bedding, graded-bedding, cut-and-fill) and particle properties (size, shape, sorting) indicate fluvial transport was fundamental to downvalley movement of bright sand and gravel. Aggradation in the form of deltaic kame terraces confirm discharge of meltwater streams directly into proglacial lakes via lateral channels along ice-tongue margins (see Figure 3).

Re-sedimentation of ice-cored glacial debris is seen in matrix-supported and poorly sorted silty, sand and gravel of moraines and kame fields. However, the limited occurrence of well developed kame moraines and associated valley train suggest few stable ice-marginal positions existed during retreat. Yet, bright drift in kame moraines must have been derived from upvalley bedrock sources and remote regions, which indicates glacial flow was an essential debris transport mechanism and re-sedimentation accompanied depositional processes. Although landform distribution suggests kame fields received inwash from upland sources, pebble count data show a strong affinity of bright drift to upvalley sources. Many valleys contain a few anomalous pebble counts that deviate from general trends, but the distinction between upland and valley drift is clear. In summary, the data suggest more than one sediment source for sand and gravel, and downvalley movement involved multiple transport mechanisms.

ANALOG ENVIRONMENTS OF SEDIMENTATION

The Bering Glacier paradigm

Conditions peripheral to Bering piedmont glacier, Alaska, include examples of depositional environments that existed during late glacial retreat in central New York State. Fed by the Bagley Ice Field, Bering Glacier spreads on a coastal lowland where it coalesces with the Steller Glacier. Combined, they form a 30 km wide piedmont lobe that fronts in several ice-contact lakes. Tsvat and Tsiu lakes basins flank Weeping Peat Island along the eastern sector (Figure 9). As a large, warm-based glacier it simulates lobate conditions along the retreating margin of the Laurentide Ice Sheet, and embraces environments of deposition analogous to conditions of ice tongue retreat from the Appalachian Plateau.

Suspended sediment

A well established subglacial conduit system discharges, highly turbid meltwater directly into ice-contact lakes. Multi-year measurements of suspended sediment load and rates of sedimentation derived from annual bathymetric surveys and stratigraphic evidence provide a comprehensive data base for evaluating comparable conditions in the late Pleistocene glacial lakes of central New York State (Gardner, et al., 1993; Casamento, et al., 1997; Dell and Fleisher, 1998; Fleisher, et al., 1998; Fleisher, et al., 2000; Fleisher, et al., 2003). Furthermore, Bering Glacier is known to surge periodically (Post, 1972; Muller and Fleisher, 1995), most recently in 1993-95. Among the many abrupt and noteworthy changes brought about by the surge was a six-fold increase in suspended sediment (Figure 10). This is taken to indicate the closing of meltwater tunnels forcing broader meltwater distribution at the sole of the glacier and increased access to subglacial sediment. A mid surge outburst (jokulhlaup) marked the re-establishment of conduits to convey meltwater, thus turbidity began to return to pre-surge values.

Bathymetry

Bathymetric surveys conducted during the 1991-2000 decade in Tsivat and Tsiu Lakes captured changes in lake basin morphology, including net accumulation of sediment from which rates of sedimentation are derived. Cumulative bathymetric changes include two pre-surge years (1991-1992), followed by two years involving the surge (1993-1995), and ending with four post-surge years (1996-2000). Distinctly different sedimentary conditions prevailed within each lake basin and from year-to-year, thus leading to relatively rapid changes in lake basin morphology. These changes are attributed to four main causes: 1) ice-front advance, 2) fluctuation in suspended sediment load, 3) migration of meltwater vents, and 4) retreat.

Excluding delta aggradation by heavily loaded inflowing subglacial streams, vertical settling is assumed to be the primary cause of sediment accumulation and bathymetry change. Therefore, changes in water depth detected from sequential bathymetric surveys are used as a proxy for the amount of accumulated sediment. Rates of sedimentation are interpreted from this information, as summarized in Table 2.

Table 2 contains a summary of processes that had an influence on the rate of sediment accumulation. Several relevant generalizations are: 1) uniform conditions did not prevail from year-to-year, 2) sedimentary processes within each lake changed with time, 3) rates of sediment accumulation due to vertical suspension settling vary with time, 4) rates of glaciolacustrine sedimentation related to vertical settling of suspended sediment increased significantly from 0.6-1.2 m yr⁻¹ prior to the surge to 3.1 - 3.3 m yr⁻¹ during the surge, where they remained for five years following the surge, 5) post-surge turbidity returned to pre-surge values, yet the rate of sedimentation remained high, thus suggesting adjustment related to the re-establishment of equilibrium conditions involving the redistribution of sediment already in the system.

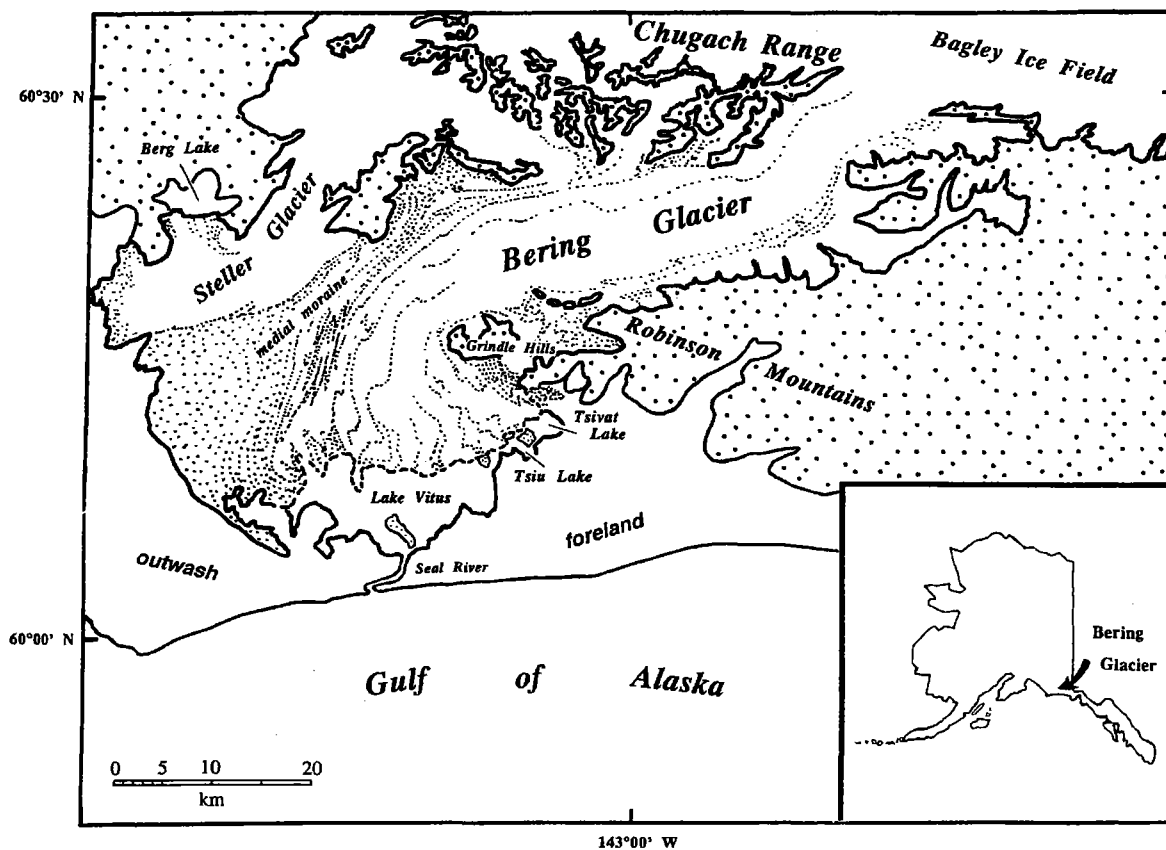
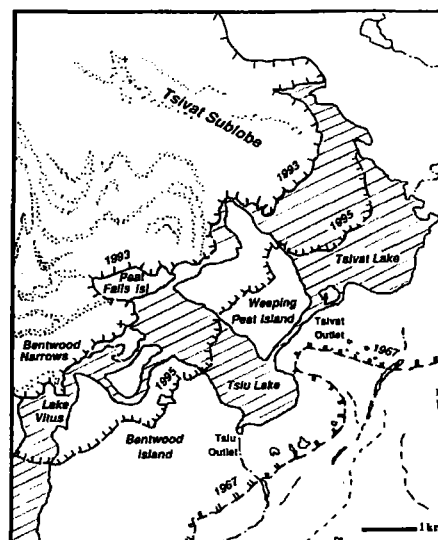


Figure 9. Bering Glacier location map and eastern sector lakes and islands.

Ice-contact Tsiat and Tsiu Lakes flank Weeping Peat Island.



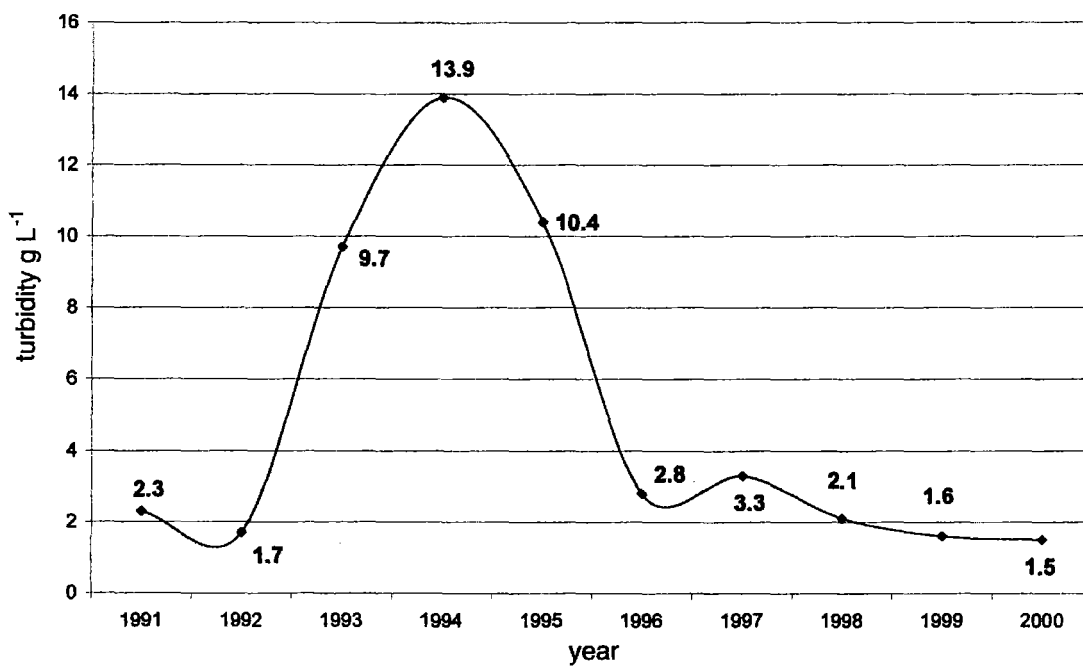


Figure 10. Average turbidity, Tsiu Lake, 1991-2000.

TABLE 2. TIMING, EVENTS, PROCESSES AND RATES OF SEDIMENTATION

	Timing of significant Events and processes	Glaciolacustrine rates Suspension settling
Pre-surge 1990/91	Upwelling vents on Tsivat ice front provide increased suspended sediment; also sediment from englacial and subglacial portals	0.6 to 1.2m yr ⁻¹ increased to 9.7 m yr ⁻¹ with input delta growth
Early surge 1993	Subglacial tunnels close; highly turbid leakage from base of advancing ice front; continued sediment settling	No data
Full surge 1994/95	Outburst into sandur forms in Tsivat Lake basin; six-fold increase in turbidity; accelerated sediment settling; push moraine forms on Tsiu Lake floor.	2.2 to 3.1 m yr ⁻¹ average from 5-year total.
Early post- surge 1996/97	Outburst into Tsivat Lake basin; rapid growth of Tsiu Delta; reduction of basin volume; turbidity reduced	3.1-3.3 m yr ⁻¹ with minimal input from ice front or delta growth.
Late post-surge 1997/98	Expansion of outburst sandur; sand fills Tsivat Lake basin; sediment bypassing to Tsiu Lake increases rate of Tsiu Delta growth.	Data from all sites affected by input from ice front and/or delta growth.
Post-surge 1999/2000	Tsiu Delta occupies 60% of Tsiu Lake basin. Growth rate exceeds rate of ice front retreat. Pervasive input by delta rate growth.	3.0 m yr ⁻¹ from single site.

Stratigraphic evidence

A semi-continuous aerial photo record (1978 to 1991) documents retreat positions of the eastern Bering piedmont lobe and the development of a shallow embayment of Tsiu Lake on Weeping Peat Island during the summer of 1986. An observed breakout in 1989 abruptly dropped lake level 17 m causing water to withdraw from the embayment, thereby exposing the net accumulation of three years of lacustrine sedimentation that accumulated within 100 m of the ice front. Three observation trenches were excavated to expose a short stratigraphic column containing three annual couplets with a total thickness of 54 cm. This continuous record of accumulation during a documented 3-year period yields an average, net accumulation rate of 18 cm/year. However, turbidity flows and undercurrents known to be common in this type of environment may have on occasion interrupted sedimentation by scouring, as indicated by notable diastems. Therefore, varve thickness here is taken to represent minimum values for annual accumulation.

The lower portion of each annual couplet averages 11.5 cm in thickness and consists of fine, gray silt laminae, each approximately 0.7 mm thick. These grade upward into interlaminated light and dark gray silt and tan, very fine sand beds, each about 0.5 cm thick, with an average cumulative thickness of 6.5 cm. The sand is commonly cross bedded and contains subtle graded bedding. Because the uppermost interlaminated unit was the last to be deposited prior to the breakout, it must represent summer-season sediments. This interpretation is consistent with intermittent higher summer discharge and associated currents that periodically introduce sand and hold silt in suspension. Lower energy conditions beneath frozen lakes during winter months favor quieter water and yield a thicker accumulation of uniform silt.

Application to Susquehanna Drainage

Applying these elevated rates of sedimentation to similar deposits beneath the Susquehanna River floodplain, and other central New York valleys, has interesting implications.

For example, within the valleys of the Susquehanna Valley it is common to find thick lake silts (100-125 meters) associated with sand and gravel terraces. Bathymetric information from ice-contact lakes at Bering Glacier indicate rates of sediment accumulation on the order of 0.6 to 1.2 m/year. Applying these rates to the Susquehanna Valley suggests that lakes at the front of the retreating Laurentide Ice Sheet would have been relatively short-lived, existing for approximately 100 to 200 years. Using an average annual rate of 0.18 m/year derived from measured varves formed at the ice front on Weeping Peat Island, the duration of accumulation would have been approximately 700 years. Regardless of which analog rate is used, it is amply clear that Late Pleistocene lakes on the Appalachian Plateau persisted for a much shorter period of time than might be implied by applied annual rates based on varves a few centimeters thick.

These conditions changed significantly as the ice pulled back from the northern drainage divide of the plateau and meltwater was diverted from the Susquehanna to the Mohawk Valley (Ridge et al., 1991). Subsequently, reduced fluvial discharge, fed only by meteoric runoff, carrying a severely diminished sediment load to remaining lake basins. Such was the case in Otsego Lake where Holocene deposition was determined by Yuretich (1981) to have been less than 8 m. The eventual failure of most lake dams and events associated with subsequent release of water are virtually unknown.

SUMMARY

1. The margin of retreating Laurentide ice consisted of valley ice tongues 20 km long and adjacent ice-free uplands.
2. The landform assemblage produced by active ice backwasting in through valleys differs from that in non-through valleys, where stagnation and downwasting occurred.
3. The sediment sources, transport mechanisms and formative processes for moraines and kame fields are different. Kame moraines mark the position of active ice margins, whereas kame fields indicate the concentration of inwash on stagnant valley ice and downwasting.
4. The primary source of lacustrine silt, which constitutes 75 % of all stratified drift, was subglacial meltwater flow through conduits that discharged saturated, fine sediment directly into proglacial, ice-contact lakes. Pebble count data indicate that most outwash sand and gravel was originally transported by glacier movement from upvalley bedrock sources, then reworked by meltwater flow at or near the ice margin.
5. The exotic pebble content of through valley drift is primarily responsible for its bright appearance, whereas inwash-derived drab drift from upland sources is a significant component of non-through valley deposits.
6. The through valley landform assemblage and subsurface stratigraphy have been interpreted to depict rapid retreat of active ice tongues from a proglacial, ice-contact lake environment. Such conditions would favor subglacial conduit discharge of highly turbid water and subsequent rapid lacustrine deposition. In addition, surface meltwater streams entering lakes from positions lateral to valley ice tongues would account for active deltaic aggradation. Indeed, well data indicate the interfingering of delta foresets with penecontemporaneous lake sediments, which means that lacustrine sedimentation and subaerial deposition were synchronous, and both occurred very rapidly. Analog conditions of rapid deposition at the margin of Bering Glacier, Alaska, suggest that the accumulation of glaciolacustrine fine sand, silt and clay may have been on the order of 0.6 to 1.2 m/year. At these rates, the average glaciolacustrine sequence of 100-125 m would have accumulated in just 100 to 200 years. At a rate of 0.4 m/year, as derived from exposed varves, the duration of accumulation may have been 700 years, which is much more rapid than varves of conventional thickness would suggest. Consequently, lakes in the Susquehanna Valley during Laurentide retreat were relatively short-lived.

7. Remnant ice and associated dead-ice sedimentation were common during regional deglaciation. Entire ice tongues stagnated and downwasted in non-through valleys due to restricted flow across headward divides. Local stagnation of detached ice masses is recognized to have accompanied backwasting in through valleys.

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As is customary for NYSGA field trips, a Road Log of planned stops is prepared months in advance. However, the ephemeral nature of Quaternary exposures generally leads to changes made only days prior to the trip. Therefore, the stops described below may not correspond to stops made during this field trip.

ROAD LOG

This Road Log begins at the I-88, Rt. 23, Rt. 7 interchange at Exit 13, west of Oneonta.

<u>Miles from last point</u>	<u>Cumulative Miles</u>	
0.0	0.0	Proceed west on I-88 from Exit 13. I-88 parallels the Susquehanna River for the next 2.4 miles. Active floodplain aggradation mantles lake sediments of Glacial Lake Otego that was dammed during retreat by the Wells Bridge Moraine, our first stop.
5.6	5.6	The highway rises above the valley floor and provides a good view of the modern flood plain and the abrupt change in valley trend that is a remnant of a preglacial engrown meander. A bedrock promontory on the horizon to the right (north) protrudes into the valley along the inside of the meander bend.
0.2	5.8	STOP 3 is on the left (south), but we won't stop now. Access to this area is possible from County Rd. 48 (locally referred to as the Otego-Wells Bridge Rd.), which we will take on our return to the Oneonta area from Wells Bridge.
3.3	9.1	Continue west on I-88 past Rt. 7 & Otego exit.
1.3	10.4	Good view to the west of the valley plug formed by the Wells Bridge Moraine.
1.7	12.1	View to the right (north) across the valley includes the back of the Wells Bridge Moraine and an associated kame terrace. The next 1.2 miles provides excellent overviews of the moraine and the gap cut by the Susquehanna River.
1.2	13.3	Rest Area Exit.

STOP 1. Wells Bridge Moraine (Franklin and Unadilla Quadrangles). The hummocky relief of this moraine is common for other moraines in the upper Susquehanna drainage basin. This moraine completely blocked the valley following glacier retreat thus forming the dam for Glacial Lake Otego that extended about 25 km upvalley to Oneonta. Logs of water wells from

the floor of the valley penetrate more than 400 feet of silt without encountering bedrock. Rates of sedimentation from similar ice-contact lakes at Bering Glacier, Alaska, related to vertical settling of suspended sediment between 0.6 and 1.2 m/yr. Therefore, Glacial Lake Otego may have been relatively short-lived.

The configuration of the Laurentide ice front during retreat is commonly thought to have consisted of valley ice tongues. Early work portrayed relatively short ice tongues, where as more recent evidence suggests that they may have reached 20 km in length. This raises implications of sediment source areas and transport mechanisms, while opening the topic of mechanisms of moraine formation. Considering that most moraines in the region consist of stratified drift and lack topographic expression other than in valleys, alternatives to conventional moraine formation may be entertained.

The Wells Bridge Moraine is assumed to have been emplaced about 15,000 years BP and breached about 14,000 years BP. We will consider the field evidence for an 1140 feet lake level at stop 3. Field work in the Unadilla and Sidney areas indicates that the Upper Susquehanna Lake Chain has greater downvalley extent than will be covered in this road log.

		Return to I-88.
1.5	14.8	Cross Ouleout Creek.
0.5	15.3	Leave I-88 at exit for N.Y. 357, Franklin and Unadilla. Turn right on Rt. 357-West.
1.2	16.5	Cross Susquehanna River and turn right on Rt. 7- East. Highway parallels the river for one mile.
1.7	18.2	Railroad overpass. Highway climbs onto outwash terrace near mouth of Sand Hill Creek.
1.8	20.0	Highway drops into Sand Hill Creek incision of outwash and immediately climbs to follow the contact of the Wells Bridge Moraine and outwash .
0.6	20.6	Crest of moraine on the left, breach on the right. It was within the breach of this moraine that archeological excavations uncovered charcoal in silt covering a buried river point bar. This was dated at 13,000 to 14,500 years BP, thus providing limiting age of Lake Otego.
0.4	21.0	Village of Wells Bridge. Turn right (south), cross Susquehanna and turn left (east) at the end of the bridge on Otego-Wells Bridge Rd., which becomes Otsego County Rd. 48. Road parallels river for 0.6 miles before rising onto an outwash terrace. A correlative terrace can be seen across the valley to the north at an elevation of about 1140 feet.

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| 3.0 | 24.0 | Cross over I-88. White house across the valley to the north is situated on a terrace at about 1120 feet. Other planar landforms can be found along the valley that suggest a second level for Lake Otego below 1140 feet. Entering Otego Quadrangle. |
| 0.6 | 24.6 | Turn right into parking lot of Gus's Diner. |

STOP 2. Massive red sandstone outcrop on south side of parking lot is covered with glacial abrasional features on joint faces that parallel the valley wall. Note that striae and grooves are inclined in the downvalley direction, thus indicating topographically controlled ice flow consistent with valley ice tongue model. Basal flow of overriding ice would not have produced such features. Several joint block surfaces that do not parallel the valley trend are also polished, but not by ice flow. These very smooth and polished surfaces are the product of abrasion by highly turbid water flow and support the notion of a highly charged subglacial hydrologic system.

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| 0.7 | 24.7 | Access to I-88 on left. Continue straight ahead. |
| 0.6 | 25.3 | Fork in road, bear left. |
| 0.3 | 25.6 | Pass under I-88. |
| 0.1 | 25.7 | Intersection at end of bridge, turn right remaining on County Rd. 48. |
| 0.2 | 25.9 | Gravel excavation on right contained deltaic foreset and topset beds indicating a downvalley current direction. |
| 0.2 | 26.1 | Similar exposure in excavation on the left. |
| 0.6 | 26.7 | Road drops to modern flood plain, which is superimposed on Lake Otego lacustrine plain. Lacustrine plain continues to the left and right. |
| 1.1 | 27.8 | Kame on the right. Prior to construction of I-88 a bit smaller feature could also be seen to the northeast. |
| 0.2 | 28.0 | Stop is situated to the right, across I-88, on the lower valley wall below "treeline". Proceed to I-88 overpass. |
| 0.5 | 28.5 | |

STOP 3. Park on the right beyond the overpass and walk south between I-88 and the forested slope. About 300 meters south of overpass and upslope from the fence, at an elevation of approximately 1120-1140 feet (2/3 the way up the forest-free slope), pebbly coarse sand, fine sand, silt, and a few clay seams were exposed during construction. Fluvial channel structures with small scale foreset beds inclined into the slope and down valley rested

upon finely laminated, rippled and cross bedded silt and fine sand. These sands are interpreted to have formed along the strandline of Lake Otego by wave generated currents that moved into this valley wall alcove. The sand about 1140 feet lacks pebbles, is considered to be of eolian origin, blown up slope from the beach.

1.3	29.8	Continue east on County Rd. 48. Road drops back down to the lacustrine plain at an elevation of 1060 feet. Entering Oneonta Quadrangle.
1.5	31.3	Turn left on access to I-88, cross river and I-88, and proceed to intersection with Rt. 7. This is where the Road Log began.
0.5	31.8	Go straight through traffic light at Rt. 7 intersection. The highway traverses an outwash/alluvial bench between 1100 and 1120 feet at the confluence of Otego Creek and the Susquehanna River.
0.8	32.6	Continue straight through traffic light.
0.3	32.9	Junction of Rt. 23 from the right. Continue straight on Rt. 205-23.

Take note that the Road Log mileage tally starts anew at the I - 88, Exit 13 intersection with State Routes 7 and 205, West Oneonta, so that this segment of the trip may be run independent of the preceding Road Log.

<u>Miles from last point</u>	<u>Cumulative Miles</u>	
0	0	Intersection of Rt. 7 and 205, proceed north on Rt. 205.
1.1	1.1	Traffic light at Rt. 23 / 205 intersection.
.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 sharply left (south) on Rt. 8
.4	21.5	Pull off right into gravel quarry operation, that has been inactive for several years

STOP 4 - Through valley landforms (New Berlin South Quadrangle)- high terrace (1180') consisting of deltaic valley train marking the ice margin (no moraine) at the head of which is a dead-ice sink. Although excavation in this quarry has been inactive for several years, large-scale, deltaic foresets

are still expressed through the colluvium. The nearest base level to which these deposits may be graded is 22 km downvalley at Rock Wells Mills (Guilford Quadrangle), where kame and kettle, morainic topography may have served as a drainage plug in a constricted segment of the valley. A partially preserved hummocky, kame terrace at White Store (16 km downvalley) also suggests ice burial.

Earlier exposures here included interbedded, very coarse, clast supported boulder gravel and pebbly, coarse sand. A carbonate cement in some units is related to limestone clast weathering, leaching, and reprecipitation at depth, thus indicating the "bright" nature of this outwash.

The Unadilla River is incised within a valley train and flows from a dead-ice sink upvalley to a lacustrine plain downvalley. These landforms and their internal features are indicative of active ice retreat with glaciofluvial sediment transport into an ice-contact lake. Partial burial of a detached, large remnant ice mass caused retarded melting until after meltwater sedimentation ceased, thus allowing collapse and topographic preservation of a dead-ice sink. Water well logs within the sink verify the lack of subsurface, stratigraphic continuity.

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	A high-level terrace at 1280' to 1300' on west side of valley continues for 1.5 miles west of South Edmeston. It appears to be an older kame terrace with no apparent downvalley base level, thus suggesting ice served that purpose.
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 5 - Inwash sediment source and topographic expression of kame field, Columbus Quarter (New Berlin North Quadrangle). The kame field and associated pitted plain may have served as a local, temporarily dam (1220') for a lake upvalley. Pebble counts from upland "drab" drift now resting on the "bright" drift of the kame field suggest tributary inwash from Tallette Creek as a source of sediment (see plot of pebble count data, after Yuchniewicz, 1996).

Turn around, proceed south on Rt. 8.

1.9	30.4	At Lambs Corners turn left (east) onto Chenango County Rt. 255 (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 Oaksville Moraine.
.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	Quarry operation in sand and gravel valley train with classic assemblage of glaciofluvial sedimentary structures on the left (east). The road continues along the floor of a dead-ice sink.
2.3	53.6	Pull off on right shoulder, park and walk to crest of moraine.

STOP 6 - Cassville-Cooperstown moraine, at Index (5 km south of Cooperstown) (Cooperstown Quadrangle). This classic through valley landform assemblage (a moraine linked with an extensive valley train) represents active ice deposition. Old excavations in a nearby borrow pit exposed stratified sand and gravel. Test borings indicate that the gravel of the moraine interfingers with silt upvalley, toward Cooperstown, before yielding entirely to silt (120+ feet thick, Fleisher, 1992). The classic kame and kettle expression of this moraine loses all topographic expression as it rises to the uplands, as is characteristic of moraines in central New York State. This has implications for a moraine forming mechanism that would involve sediment transport dominated by glaciofluvial processes. Based on topographic criteria, Krall (1977) correlates the Cassville-Cooperstown Moraine with moraines in the Hudson Valley, thereby suggesting a readvance circa 14,000 years BP. Ridge, et al., (1991) makes a tentative correlation with the West Canada Readvance of Late Wisconsinan, pre-Valley Heads age (circa 15.5 ka) based on elevation projections of ice-marginal deposits in the Mohawk Valley and the limit of readvance inferred

from subsurface stratigraphy (Fleisher, 1986).

Although high enough, this moraine did not dam Glacial Lake Cooperstown, although landforms and stratigraphy suggest that a proto-Lake Cooperstown may have extended to the moraine by "swamping" a dead-ice sink

.2	53.8	Proceed eastward over the moraine to the intersection with Rt. 28, turn left (north).
.2	54.0	Dead-ice sink between Cassville-Cooperstown Moraine and Cooperstown occupies the valley floor 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.
.2	56.3	Park at Intersection of Lake and River St.; stairway to Council Rock to the left.

STOP 7 - Lake Front Park on Doubleday Ice Margin (Cooperstown Quadrangle). Well data from the crest of the moraine (Bassett Hospital) indicate that the morainic dam for Otsego Lake consists of 180 feet of bouldery silt. Subtle hummocky terrain in the village of Cooperstown may be traced northwestward, where it is well expressed in a golf course at the lake shore. Hanging deltas along the western lake shore and lake clay on the northwestern shore indicate that while the moraine dammed Glacial Lake Cooperstown at an elevation of 1250'. The view northward includes a conspicuous cross-sectional valley asymmetry, which played an important role in determining the primary sediment source for Glacial Lake Cooperstown.

A seismic reflection survey shows that Otsego Lake occupies a rock basin that has been eroded as much as 132 m (433') lower than present lake level and infilled with up to 88 m (289') of sediment. Sediment thickness is greatest in the southern half of the lake basin and thins by ~50% to the north where maximum water depths occur. As an alternative to the traditional view of deglaciation, which depicts continuous backwasting of an active ice margin with short valley lobes, subsurface and land-based evidence suggests backwasting retreat of 20 km long ice-tongues that remained in the valleys, subject to collapse while adjacent upland regions were ice-free. Proglacial sediments were largely transported to the south end of the lake basin as alluvium from ice-free western tributary streams (Fleisher, et al., 1992).

Return to Lake Street, turn right (west)

.5	56.8	Turn left onto Chestnut St. (Rt. 28 and Rt. 80 West).
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.5	57.3	Junction Rt. 80 west, proceed south on Rt. 28.
2.3	59.6	Highway traverses Cassville-Cooperstown moraine.
1.1	60.7	Hamlet of Hyde Park; from here south for several miles, planar gravels appear related to upland tributary inwash sources.
5.4	66.1	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.
6.0	72.1	Turn right on Clover Hill Road in Milford Center and park

STOP 8. Goodyear Lake overview at Milford Center (Milford Quadrangle). 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, sand and gravel outwash 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.

Continue south on 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	Pass under I-88 and continue straight on Gersoni Road (unmarked)
1.0	76.2	Stop sign, turn right on Dead End road leading to Seward's Gravel Quarry.
.4	76.6	Approval for entrance to quarry must be obtained in Office. Mileage into, around and out of quarry not included in road log.

Stop 9. Moraine at margin of Dead Ice Sink (West Davenport Quadrangle). Many of the landforms that characterize the through valley assemblage may be found in the vicinity of this location. They include a kame moraine and associated outwash features, dead ice sink, and lacustrine plain. During the many years of operation, a variety of ice-contact features have been exposed. Upper- most units are flow tills (debris flows) that randomly interfinger with interbedded sand and medium gravel, frequently including silt and sand of local significance. Blocks of lodgment till have been observed in piles with sub-meter boulders that come off screening equipment. Massive foreset beds of gravel that range in size from bouldery gravel, through cobble and

pebble gravel, to pebbly, coarse sand are commonly found beneath the overlying flow till at an intermediate, vertical position through the operation. Well sorted, laminated lake sand is very common on the downvalley, lower margin of the quarry.

The hummocky, kame and kettle topography common to the quarry area fades downvalley to join with the surface of extensively developed outwash terraces. Over the years, quarry operations in these terraces have uniformly exposed massive gravel foreset beds indicative of meltwater sedimentation in an ice-contact lake that expanded upvalley to follow the retreating ice front. MacNish and Randell (1982) describe such a setting to depict slow, active-ice retreat.

These extensive paired terraces formed as "lateral deltas" into the expanding ice-contact lake, thus growing in the upvalley direction as space became available during retreat. Episodes of slow retreat would accommodate growth until delta terraces from either side of the valley coalesced to fill the valley, thus forming a valley train with internal deltaic structure.

.4	77.0	Leave the quarry and return to the stop sign intersection. Turn right (west) onto Hemlock Road (and unmarked continuation of Gersoni Road)
.4	77.4	Entrance to Broe Pit, Cobbelskill Products quarry. Time permitting, we will drive in to view massive foresets of delta terrace.
1.2	78.6	Stop sign intersection with County Rt. 47. Turn left
1.2	79.8	Delaware County line (Otsego County Rt. 47 becomes Delaware County Rt. 11).
.9	80.7	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	82.7	Turn right into parking lot for Hartwick College's Pine Lake Camp. Walk downhill to pavilion.

STOP 10 -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 this transverse valley was clogged by remnant ice. These landforms are typical of valleys in which the stagnated ice was subject to inwash burial and many local ponds and lakes existed. Here, the Davenport Moraine dammed Charlotte Creek valley at 1280 ft. Many shoreline landforms along the valley are graded to this lake.

Return to road, turn right.

.3	83.0	Right hand fork goes downhill and across the lacustrine plain of Glacial Lake Davenport.
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.6	83.6	Turn left at intersection with Rt. 23 in Davenport Center.
.2	83.8	Town of Davenport Center quarry on right (no longer active). Past exposures contained well-developed, large-scale deltaic foreset beds graded to 1280 ft. water level. Forty foot deep kettles indicate progradation that incorporated grounded ice.
3.7	87.5	Town of Davenport. Proceed east on Rt. 23.
1.5	89.0	Clark Company Stone Products. Mileage into, around and out of quarry not included in road log.

Stop 11. Hanging delta at the confluence with Middle Brook (Davenport Quadrangle). Although quarrying activity continues, much of this original hanging delta into Glacial Lake Devenport (1280 to 1300 ft.) has been excavated. Remnants of evidence supporting this interpretation may still be seen. Exposed at the upper quarry level are several meters of well stratified, strongly imbricated topset gravel. Discontinuous exposure to lower positions eventually lead to multiple packages of lacustrine sand and silt interspersed with small-scale deltaic infill and cut and fill structures.

Return to Rt. 23.

.2	89.2	Turn left onto Delaware County Rt. 9 to Fergusonville.
.3	89.5	Road crosses lacustrine plain
1.0	90.5	Entering Fergusonville.
.7	91.2	Turn right on Olive Branch Road (Dead End) at Brandow's Trailer Sales. Access by permission of Brandow Family. Mileage into, around and out of quarry not included in road log.

Stop 12. Very localized hummocky topography reaching elevations of 1360 ft. Recent excavations exposed several, 2-3 m thick, alternating units of cross bedded coarse sand, tan laminated lake sand, and sand and fine gravel, each with unconformable contacts. The entire sequence was overlain by 2-3 m of moderately well sorted pebble gravel. Isolated from potential tributary sources, this ice-contact landform appears to have originated as meltwater infill of localized ponding on ice-cored terrain, much as would be anticipated in contact with a remnant ice masses in a transverse valley.

2.2	93.4	Backtrack to Rt. 23. Turn right (west) onto Rt. 23.
1.9	100.9	Highway traverses moraine at West Davenport.
2.3	103.2	Terrain to the left is what may be the only lateral moraine in this

area.

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| .7 | 103.9 | Road descends to Susquehanna valley lacustrine plain. |
| 1.1 | 105.0 | Highway enters the breach of the Oneonta Moraine (significantly altered by urban development). |
| 1.3 | 106.3 | Traffic light intersection I-88, 28 N/S, 23W; continue straight on 28 south. |
| .6 | 106.9 | Turn right toward I-88 west. |
| .3 | 107.2 | Turn left onto I-88 west. |
| 2.0 | 109.2 | Take Exit 13 (Morris and Route 205). |
| .3 | 109.5 | Turn right (north) onto Route 205. |
| .2 | 109.7 | This brings you back to the start of this trip at intersection of Rt. 7 and Rt. 205. |

END OF ROAD LOG

OPTIONAL - field trip up the non-through valley of Otego Creek.

Miles from last point	Cumulative Miles	
0	0	Start at traffic light intersection of Rt. 23 west (to West Oneonta and Morris) and Rt. 205 (Laurens) north, proceed north on Rt. 205.
4.0	4.0	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	4.7	Turn left on County Rt. 11A (unmarked) toward Laurens.
.3	5.0	Turn left on County Rt. 11 through Village of Laurens
.4	5.4	Bear left at fork (sign to Oneonta).
.8	5.8	Turn left into Town of Laurens Quarry adjacent to Maple Grove Cemetery. Mileage into, around and out of quarry not included in road log.

OPTIONAL STOP 1. Village of Laurens Quarry (Vision Quadrangle). This is a good example of inwash sourced aggradation associated with a stagnant, downwasting ice tongue that occupied this non-through valley and other like it. 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. The basic feature is an ice-contact, hanging delta (Mt. Vision Quadrangle). Landform expression and large-scale foreset and topset beds indicate deposition into an ice-contact lake. Initial aggradation here was into a lake graded to a moraine dam at West Oneonta, 5 km downvalley. A local inwash source provided adequate sediment for additional topset, alluvial aggradation above lake level. Otego Creek passes through a narrow breach that interrupts the continuation of this landform at the same elevation across the valley floor. The localized concentration of accumulated sediment here is attributed to inwash from Lake Brook. Similarly, other tributaries fed sediment from ice-free slope to local lakes and ponds distributed along the valley and gravel plains are graded to these water bodies.

Return to road, turn right (north), backtrack through Village of Laurens.

.8	6.6	Junction 11A and 11, continue north on Maple Street.
2.1	8.7	Gravel plain remnant for next half mile
1.3	10.0	Junction Rts. 11 and 11B, continue straight on 11. Road rises onto pitted planar gravel remnant.
1.0	11.0	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.

- .1 11.1 Park on right shoulder and walk through the pasture to the crest of an esker on the near, right horizon.

OPTIONAL STOP 2. Esker in dead-ice terrain (Mt. Vision Quadrangle). Once again we see concentrated aggradation at a stream confluence, here where West Branch joins Otego Creek. Landforms typical of a non-through valley assemblage include a kame field (differs from a kame moraine only in that the kame and kettle topography is not linked with planar outwash, as would be anticipated at an active ice margin), gravel plain, dead-ice sink, and an esker. Eskers are rarely preserved, partly because conditions favoring formation were not common, but also because they are excellent sources of sand and gravel, thus they are typically removed by gravel mining. Here, the esker is indicative of stagnant ice conditions. Lack of excavation precludes investigation of internal features.

Proceed on Rt. 11.

- .2 11.3 Turn right onto Angel Road that traverses kame field.
- .6 11.9 Junction Angel Road and Rt. 205; turn left (north) onto Rt. 205 toward Hartwick.
- .6 12.5 Pull off on left shoulder.

OPTIONAL STOP 3. Overview of meltwater channel through discontinuous gravel plain remnant and terrain of ice-cored origin.

- .5 13.0 Road descends to a local lacustrine plain (dead-ice sink?)
- .7 13.7 Gravel excavation to the left removed a landform thought to be an esker.
- 2.5 16.2 County Rt. 45 enters from the right, pull off on right shoulder.

OPTIONAL STOP 4. 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 17.4 Village of Hartwick, Junction Rt. 205 and County Rt. 11, proceed north on Rt. 205.
- 1.3 18.7 Pull off on right shoulder.

OPTIONAL STOP 5. 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 due to thinning during glacial retreat, thus resulting in downvalley ice-tongue

starvation and ultimate stagnation.

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.