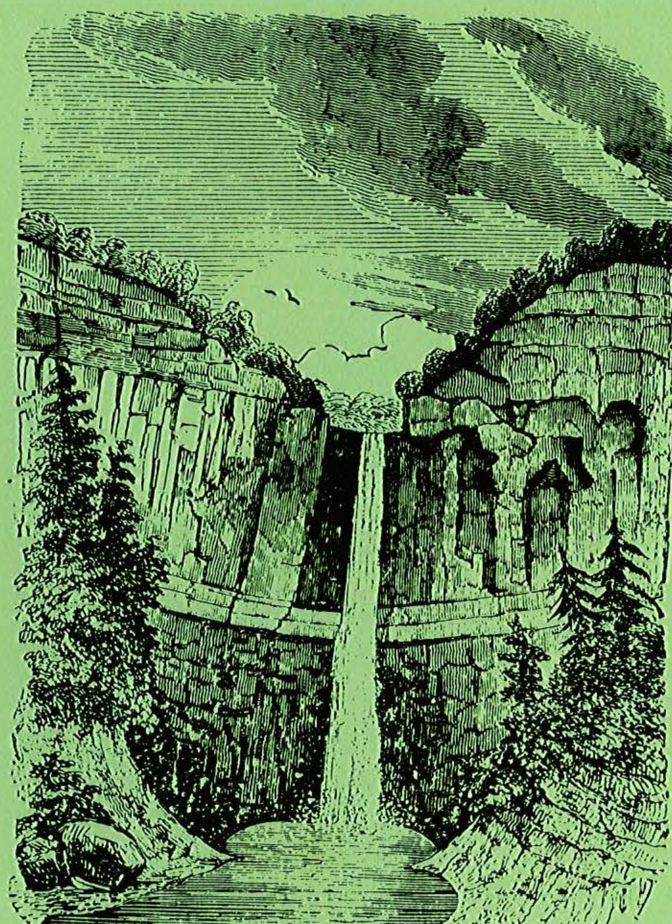
NEW YORK STATE
GEOLOGICAL ASSOCIATION

GUIDEBOOK FOR FIELD TRIPS IN SOUTH-CENTRAL NEW YORK



Taghannuc falls, Tompkins county. From a sketch by Mrs. HALL.

53rd Annual Meeting
State University of New York
at Binghamton
18-20 September 1981

NEW YORK STATE GEOLOGICAL ASSOCIATION

53RD ANNUAL MEETING 18-20 SEPTEMBER 1981

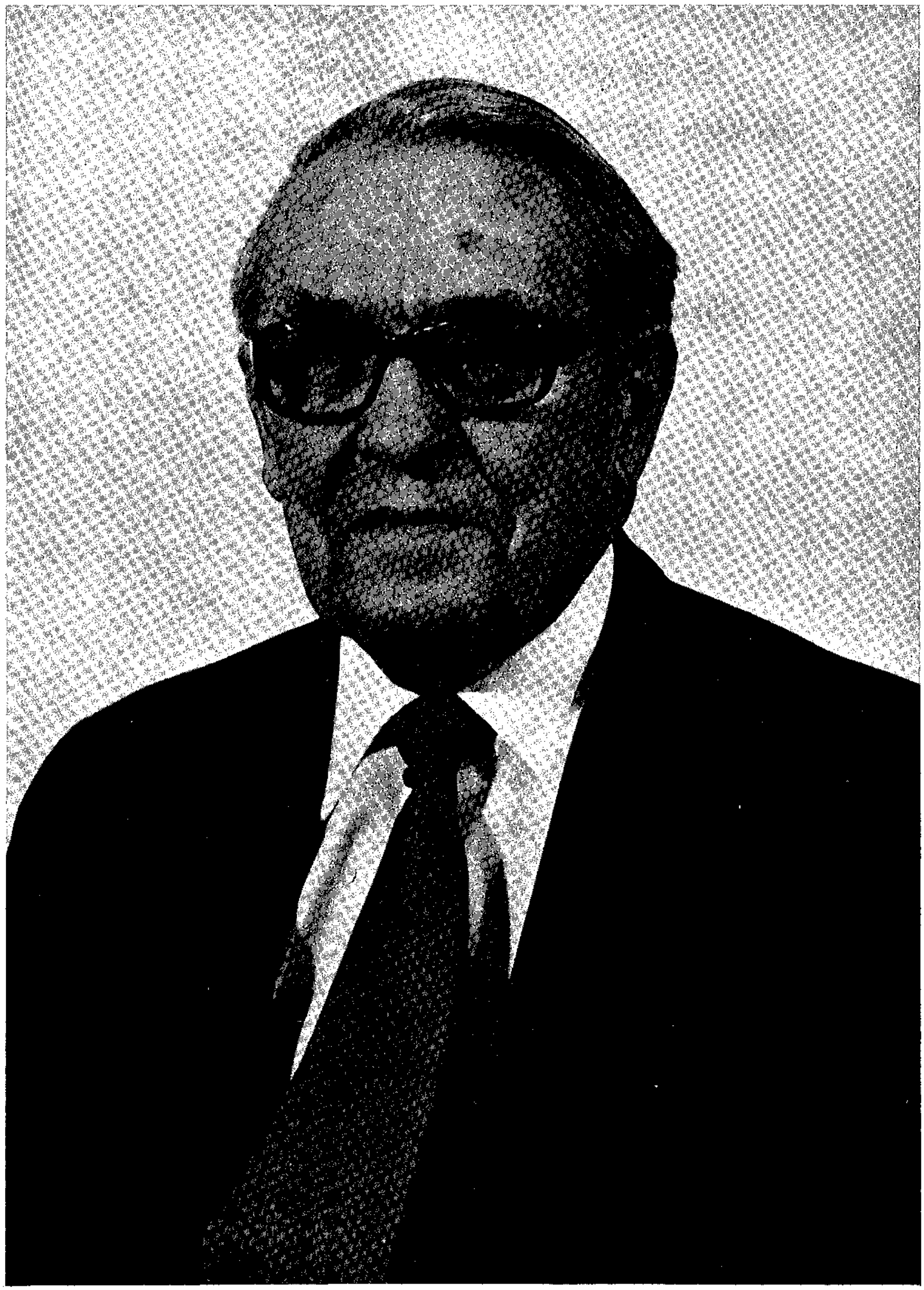
BINGHAMTON, NEW YORK

GUIDEBOOK FOR FIELD TRIPS IN SOUTH-CENTRAL NEW YORK

PAUL ENOS, EDITOR

STATE UNIVERSITY OF NEW YORK AT BINGHAMTON

Published by the New York State Geological Association. Guidebook available from the executive secretary: Manfred P. Wolff, Department of Geology, Hofstra University, Hempstead, New York, 11550.



DEDICATION

GLENN G. BARTLE (1899-1977)

When Binghamton last hosted the N.Y.S.G.A. meetings in May of 1963, an unusual feature of our campus was that we had a geologist as president, Glenn Bartle. Since that time, other institutions (such as Cornell) have discovered the value of such a choice.

Our campus probably is one of the few in the world that can claim to have been almost single-handedly built by a geologist, the same Glenn Bartle, deceased in September of 1977. In 1946, Glenn took a job as dean of the two-year extension of Syracuse University in Endicott, New York. When the numbers of WW II veterans subsided and Syracuse decided to close their extension, it was Glenn Bartle who prodded and cajoled local business and professional people to support the retention and expansion of the former Syracuse facility into a four-year college, eventually to become Harpur College, the liberal arts college of the fledgling State University of New York system. Glenn masterminded (and manipulated) this college from temporary barracks onto a site of his choosing and supervised building the physical plant you are now visiting. Even more importantly, he guided the recruitment of the faculty for one of the fine public liberal arts colleges during the early 1960's. He also successfully guided Harpur College towards its emergence as one of the four graduate centers of the State University of New York. This done, Glenn "retired"to his geological profession and worked avidly as petroleum geologist until his death.

Glenn built well, in part because of affection and loyalty he inspired in his colleagues. He was a joy to be with, and we, and "his boys" in the Geology Department at SUNY-Binghamton, have lasting and affectionate memories of him.

It is to his memory that this guidebook is dedicated.

-----JAMES E. SORAUF-----

COVER: Sketch of Taughannock (Taghannoc) Falls, west side of Cayuga Lake, by Mrs. James Hall, to illustrate the Geology of New York, Part IV, by James Hall, 1843, p. 377. This locality will be visited by the field trip on "Middle and Upper Devonian shales and adjacent facies of south-central New York."

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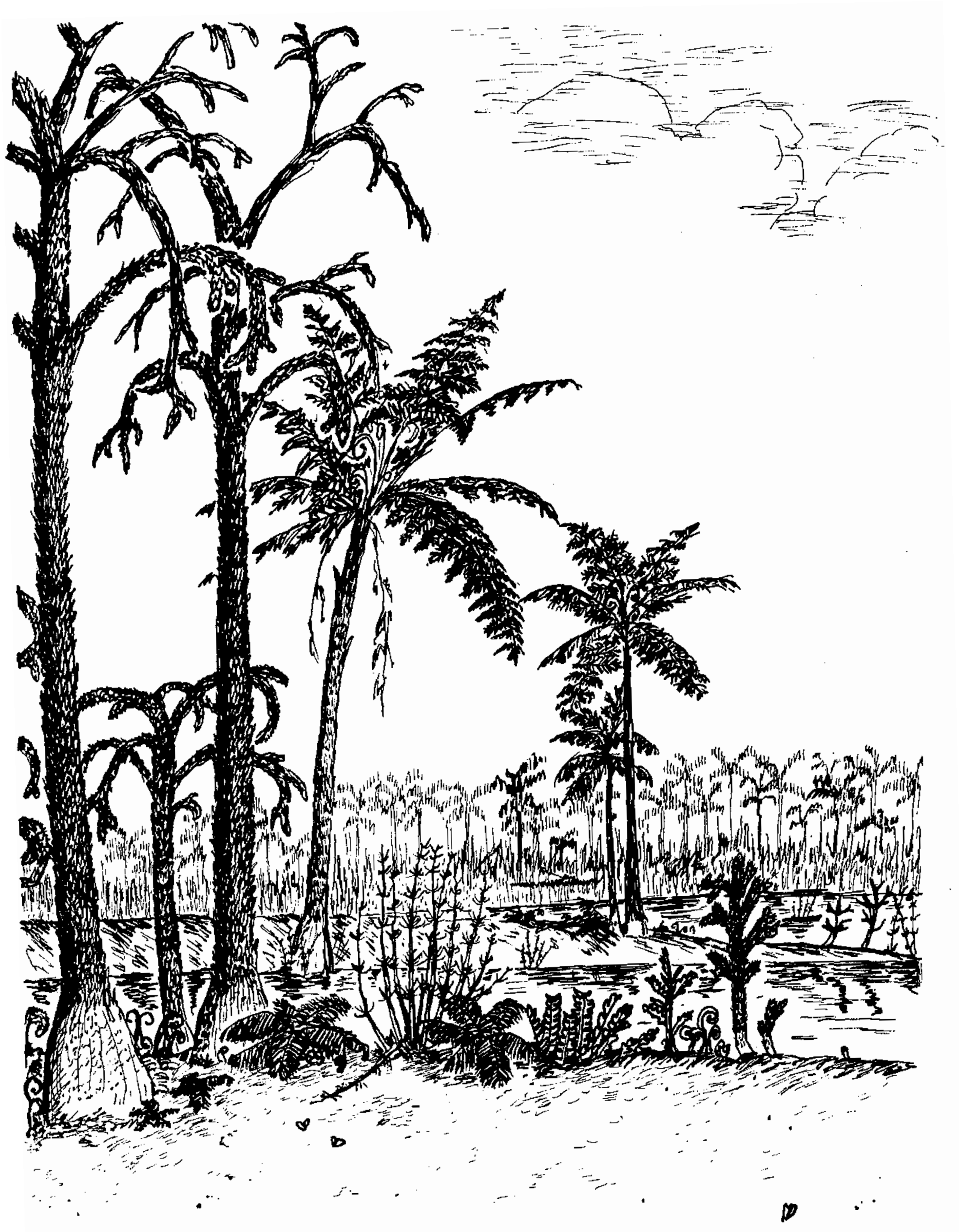


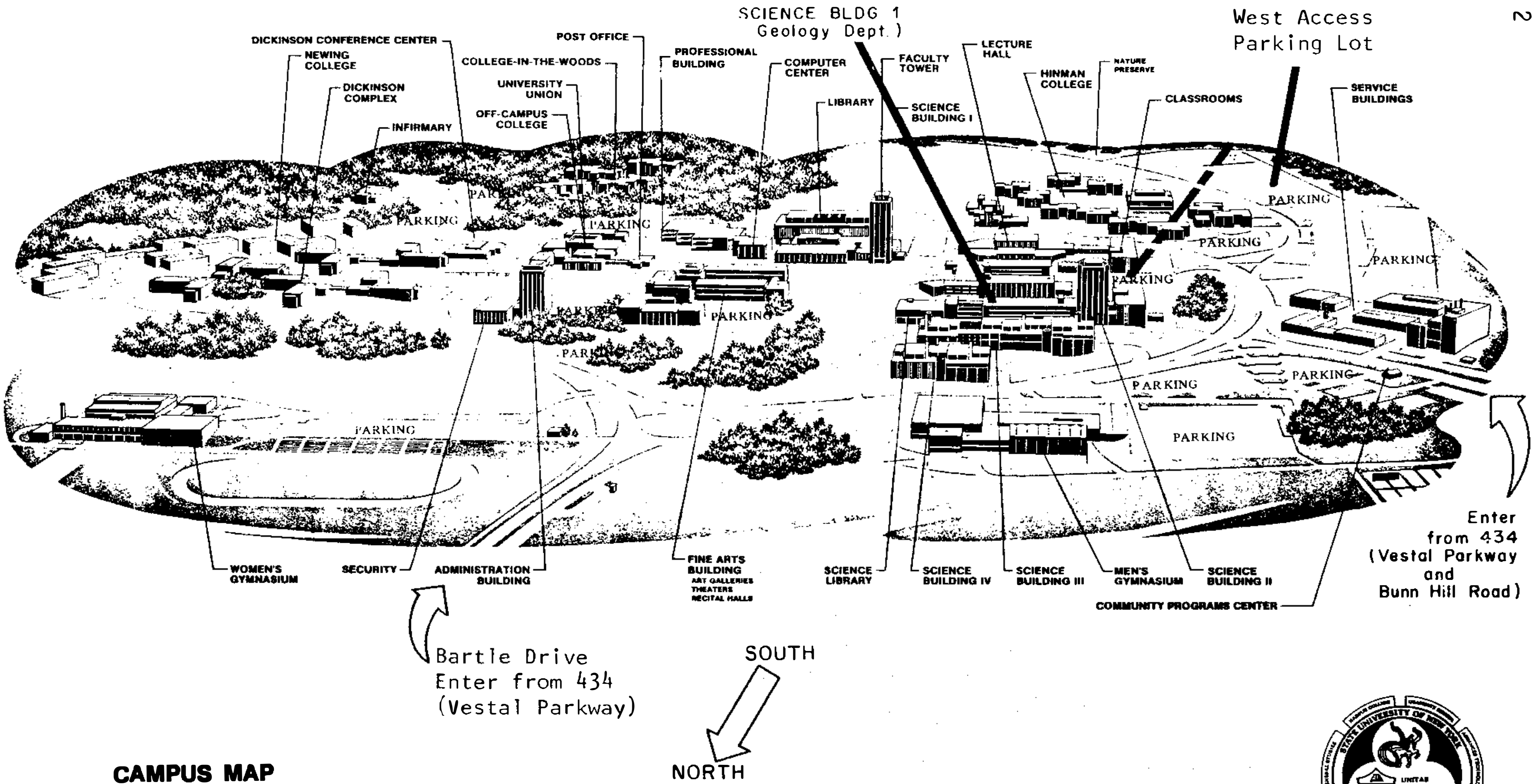
The Devonian Gilboa forest from the Catskill Mountains. From R. H. Dott, Jr. and R. L. Batten, 1981, Evolution of the Earth, p. 280. Redrawn by Kevin Enos.

GREETINGS

Donald R. Coates
President, New York State Geological Association

It is with much pleasure that we once again host the New York State Geological Association annual field conference after an 18-year hiatus. During these nearly two decades, much has been learned and published about the south-central New York region. This is in sharp contrast to the literature base that existed prior to 1963. In the preface to the 1963 guidebook, I commented on the scarcity of information, and this was reflected by the low number of field trips, four, for the occasion. Today, however, we are able to offer 11 trips which should whet the appetite of most geologists, with the exception of the unfortunate lack of local "hard rock" or structurally deformed terranes. Some of the increased geologic communications result from the location of the State University of New York at Binghamton campus in the heart of the region. Faculty and students at the University have added to the geologic information base of the surrounding area. Far from being provincial, however, 8 of the 18 field-trip leaders or co-leaders are "invaders" invited to guide you through the wonders of the region. This blend of homegrown and imported talent should provide a healthy mixture for obtaining greater understanding of south-central New York. We are also pleased that two trips visit our close neighbor, Pennsylvania, one under the guidance of a member of the Pennsylvania Geological Survey. Indeed, participants at this field conference are lucky; for the price of one admission they may sample something of two states.

So sit back, enjoy your bus ride and excursions through our region, and read on through this guidebook about the new frontiers that are being revealed and some old favorites that are being revisited.



CAMPUS MAP

STATE UNIVERSITY OF NEW YORK AT BINGHAMTON

THE WEST FALLS GROUP (UPPER DEVONIAN) CATSKILL DELTA COMPLEX: STRATIGRAPHY, ENVIRONMENTS, AND SEDIMENTATION

JAMES R. EHRETS
University of Rochester
Rochester, New York

INTRODUCTION

Since the last meeting of the NYSGA in Binghamton (Coates, 1963), significant advances have been made in the refinement of correlation and environmental interpretation of Upper Devonian strata in south-central New York. One purpose of this article is to summarize the updated stratigraphy in this region and to review how the stratigraphy has been developed. The main theme of the article is the result of recent study in the Binghamton area (Fig. 1). By tracing thin units of dark-gray shale to the east of Binghamton, a stratigraphic framework was developed which permitted further analysis of environmental and sedimentologic relationships of the Catskill delta complex in the area. Through the demonstration of the dark-shale framework and discussion of deltaic environments, a few of the interpretations are presented. The accompanying field trip is designed to illustrate a variety of deltaic environments and how they fit into the stratigraphic framework.

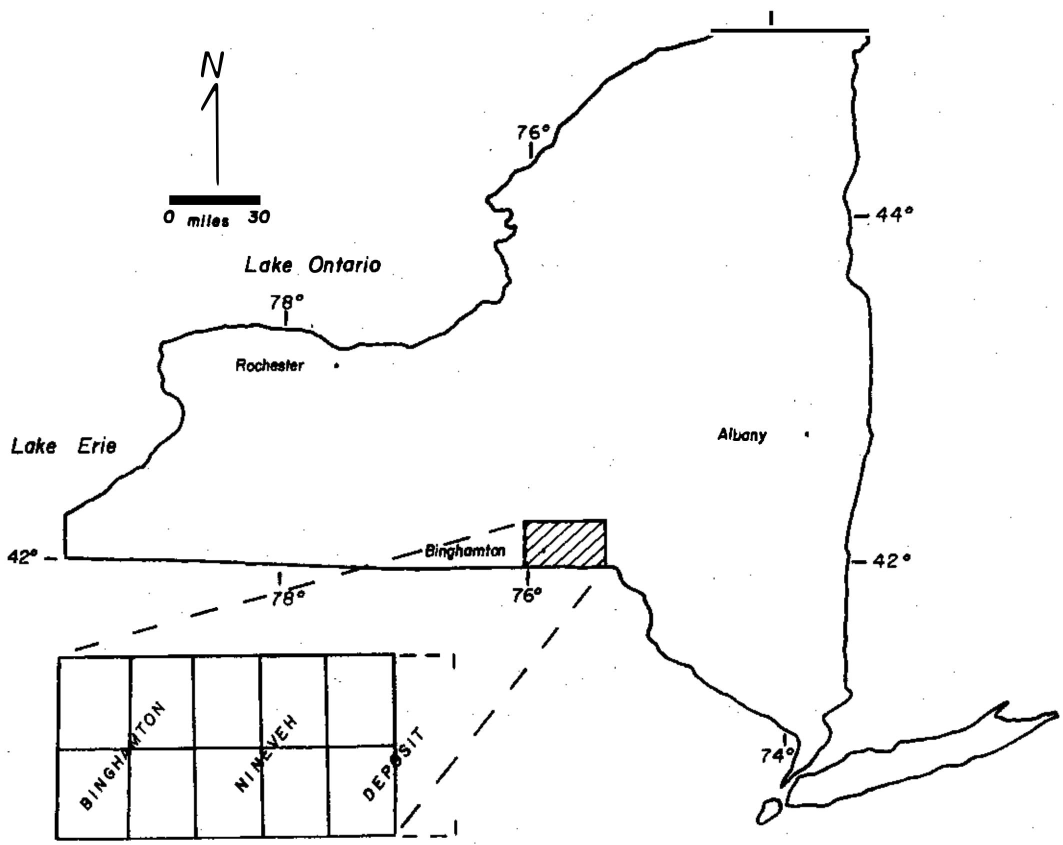


Figure 1. Study area location; inset shows included 15-minute quadrangles.

STRATIGRAPHY

Beginning about 1950, the use of dark shales for correlation in the Upper Devonian of New York has gained significant interest due to the discovery that certain units of dark shale parallel biostratigraphic zones. This has not only aided in the solving of some of the problems involved in correlating strata characterized by numerous facies changes but it has also facilitated the study of the interrelationships of the facies themselves. At present, four major dark-shale tongues are recognized in western New York: the Geneseo black shale, which defines (in part) the base of the Genesee Group as well as the base of the Upper Devonian; the Middlesex black shale, which defines the base of the overlying Sonyea Group; the Rhinestreet black shale, which defines the base of the next highest West Falls Group; and the Dunkirk black shale, the highest of the major shale tongues, which defines the base of the overlying Canadaway Group. These tongues, as well as a few intervening and overlying minor tongues, extend from the Ohio black shale which lies to the west of New York State (Fig. 2).

In a stratigraphic and structural study of the Dryden and Harford quadrangles which are located northwest of the Binghamton area, Sutton (1959) recognized two thin units of dark-gray shale which parallel the faunal zones he was tracing through the region. Sutton determined that the dark-shale units were more suitable for mapping purposes due to their persistence in the field. Following methods established by Sutton, Humes (1960), Nugent (1960), and Woodrow (1960) traced thin units of dark shale in the Ithaca, Watkins Glen, Elmira, Owego, and Waverly quadrangles located to the west and northwest of Binghamton. Two of the units were determined to be eastern equivalents of the Middlesex black shale: the Montour Member and the Sawmill Creek Member. Three such units were found to be equivalent to the overlying Rhinestreet black shale: the Moreland Member, the Dunn Hill Member and the Roricks Glen Member. Sutton (1963) later determined that a fourth shale member—the Corning Member—also extended eastward from the Rhinestreet. As the shale members were traced to the east the nomenclature for the intervening strata developed as new facies were encountered. The nomenclature at that time was summarized by Woodrow and Nugent (1963).

During later reconnaissance, Sutton (1964) and Woodrow (1964) determined that although the lithologic character differed from their respective type sections to the west, the shale members continued eastward and in fact intertongued with the nonmarine rocks of the Catskill facies. Shale members from the underlying Middlesex and Geneseo black shales were also shown to extend eastward into nonmarine strata (Sutton and others, 1970; Thayer, 1974). The most recent editions of the New York State Devonian correlation chart (Rickard, 1975) and the geologic map of New York State (Fisher and others, 1970) both include the dark-shale framework as a major feature in conjunction with biostratigraphic correlation.

With the exception of a small percentage of Sonyea Group strata which crop out in the valleys of the area, rocks within the Binghamton area are in the West Falls Group. A lithostratigraphic framework is

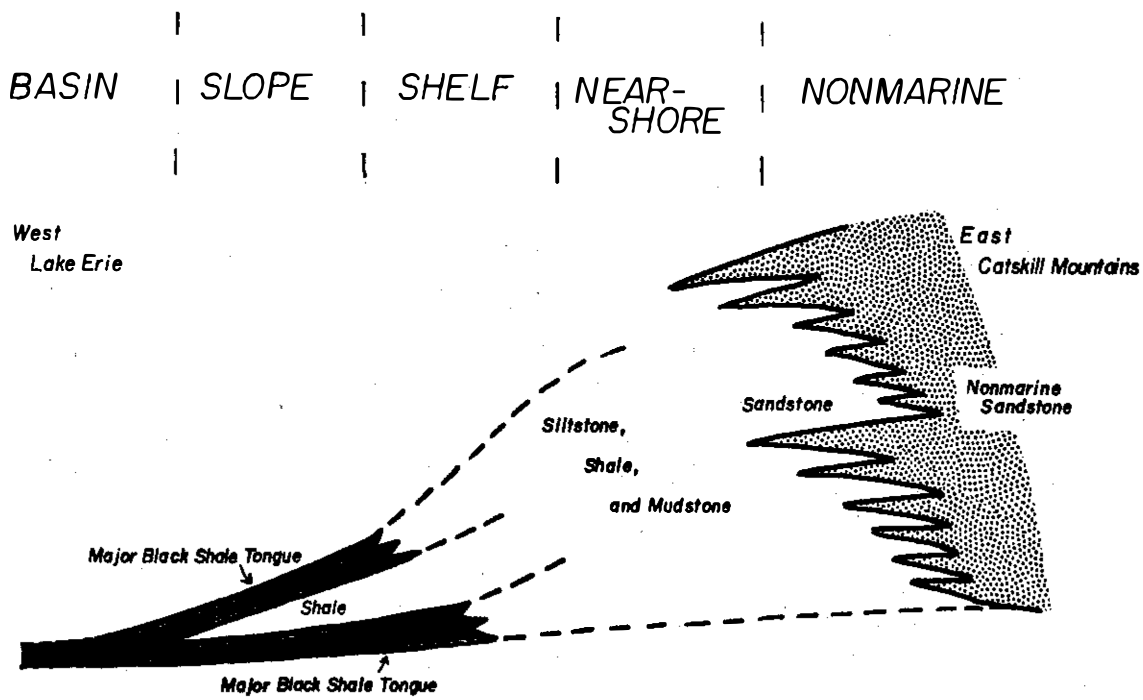


Figure 2. Generalized stratigraphic and environmental relationships of two consecutive tongues of black shale.

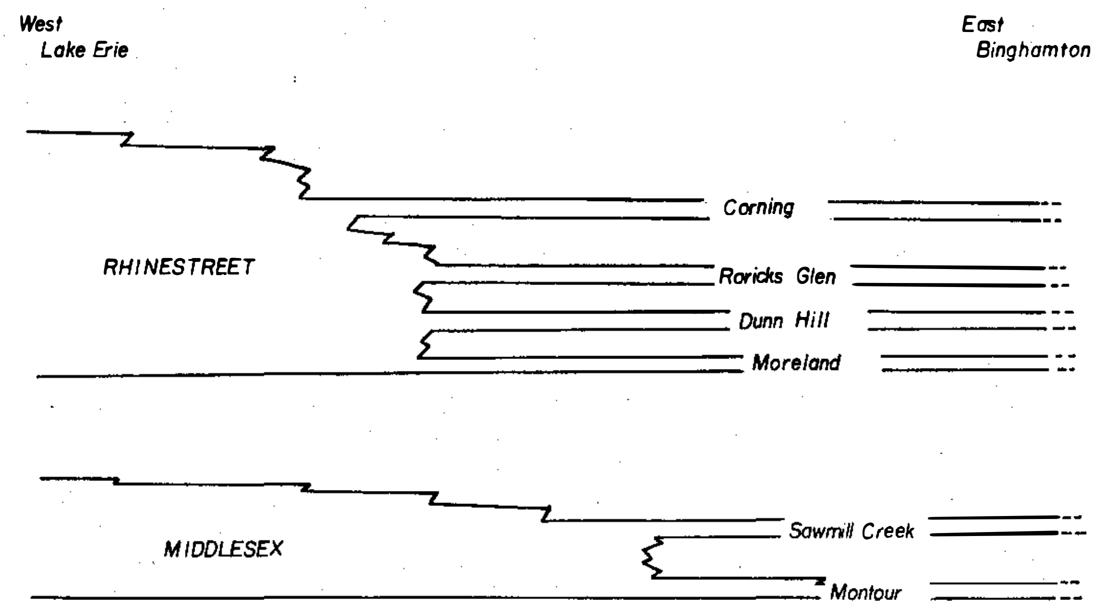


Figure 3. Time-stratigraphic relationship of dark-shale members of the Middlesex and Rhinestreet black shales. (after, Rickard 1975)

provided by the Moreland, Dunn Hill, Roricks Glen and Corning Members of the Rhinestreet Formation (Fig. 3). The nomenclature for the stratigraphic intervals between the shale members is established for rocks to the west of the Binghamton area (Fig. 4). The names Meads Creek, Beers Hill, and Millport, which have in the past been considered to define members of the Rhinestreet Formation in the Binghamton area, are now confined to strata within the Portage facies to the west of Binghamton (Rickard, 1975). Although new names are needed for rocks of the Chemung facies in the Binghamton area, no new names are proposed here. For discussion purposes, a specific interval will be referred to by the shale members which bound it above and below. For example, the rocks of the lowermost interval are referred to as the Moreland-Dunn Hill interval. Rocks occurring below the Moreland Member belong to the Glen Aubrey Formation of the Sonyea Group. Rocks overlying the Corning Member belong to the Gardeau Formation of the West Falls Group. All of these intervals and formations are composed of marine rocks of the Chemung facies as described by Rickard (1975). The marine rocks within the Binghamton area intertongue with nonmarine rocks of two formations within the Catskill facies. The Moreland-Dunn Hill and Dunn Hill-Roricks Glen intervals are correlative with the Upper Walton Formation while the remaining overlying rocks (Roricks Glen-Corning interval and the Gardeau Formation) are correlative with the Slide Mountain Formation (Fig. 4).

RECENT STUDY

Introduction. With the essential stratigraphic framework established by the dark-shale members of the Rhinestreet, the area between Binghamton and Deposit has been recently examined in detail in an effort to refine the framework and to develop an environmental and sedimentologic interpretation of the Catskill Delta complex for the West Falls Group (Ehrets, 1981). The Binghamton area provides excellent exposure of rock in the form of numerous road cuts and quarries for such detailed study. Of over 100 exposures in the area, 59 were chosen which were incorporated into a cross-sectional framework (Fig. 5). The most important aspect of the study centered around the tracing of the individual dark-shale members of the Rhinestreet due to their importance in the stratigraphic framework. The dark-shale members were traced through consideration of the stratigraphic and geographic relationships of scattered exposures of dark shale in the study area (see symbols, Fig. 5).

Structure in the Study Area. As a knowledge of possible structural complications is necessary to ensure the correct tracing of a particular shale member, it is important to briefly consider the geologic structure of the study area. Wedel (1932) determined the essential structural feature in south-central New York to be a gently flexed monocline which dips "practically due south at the rate of approximately 40 feet to a mile." The flexures in the monocline are the expressions of anticline (better described as series of aligned domes) and synclines which become more prominent to the southwest. Wedel traced the "Nichols syncline" (southernmost), the "Elmira anticline", and the "Horseheads syncline" (northernmost) east to Binghamton, observing that their existence east of Binghamton was questionable (Fig. 6). Woodrow (1968) added support to

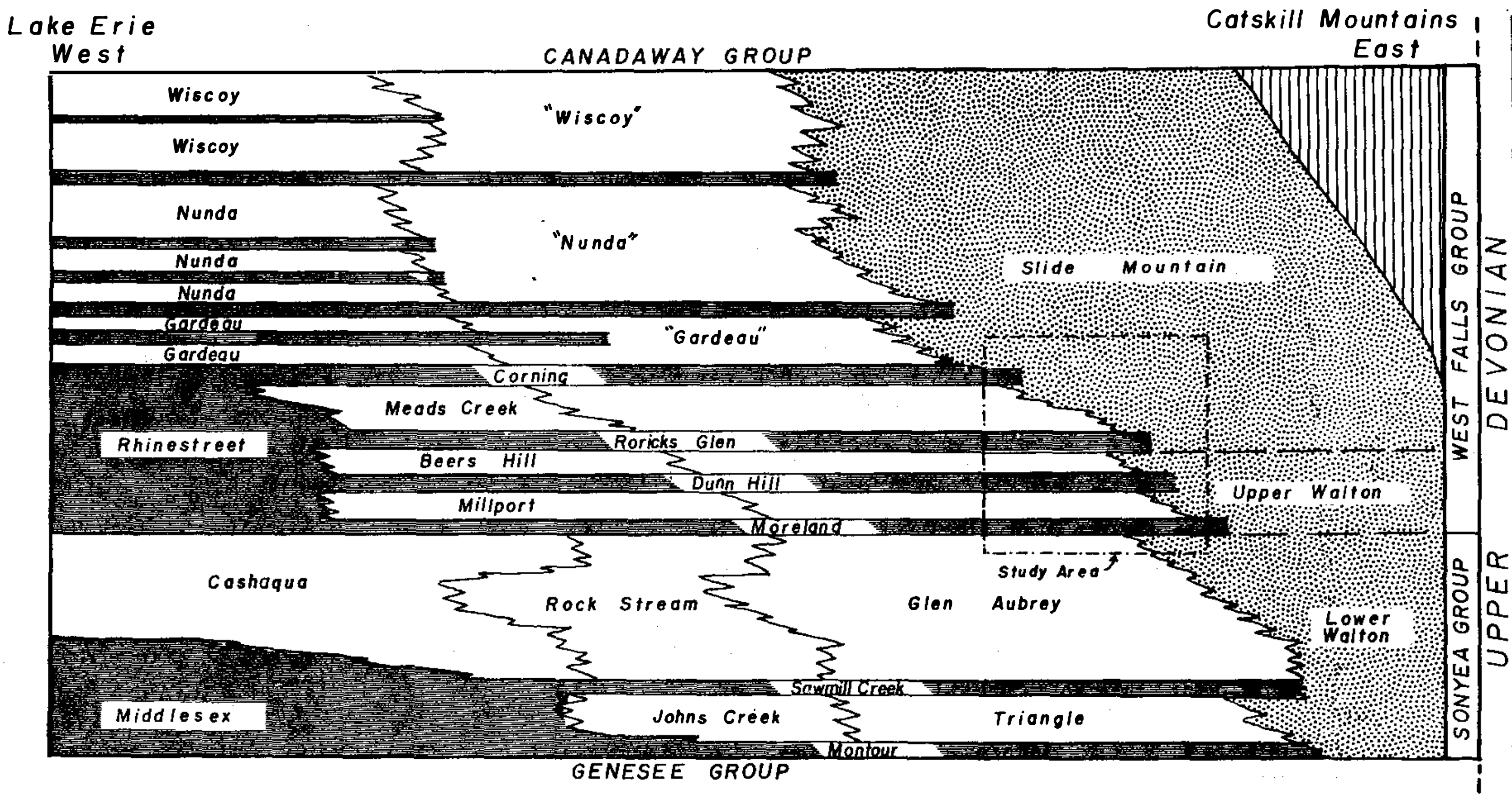


Figure 4. Summary of stratigraphic nomenclature and time-stratigraphic relationships for the West Falls and Sonyea Groups. Names in quotes are used informally. Closely ruled lines denote dark-shale members and black shale-formations; stipple denotes Catskill facies. (after Rickard, 1975)

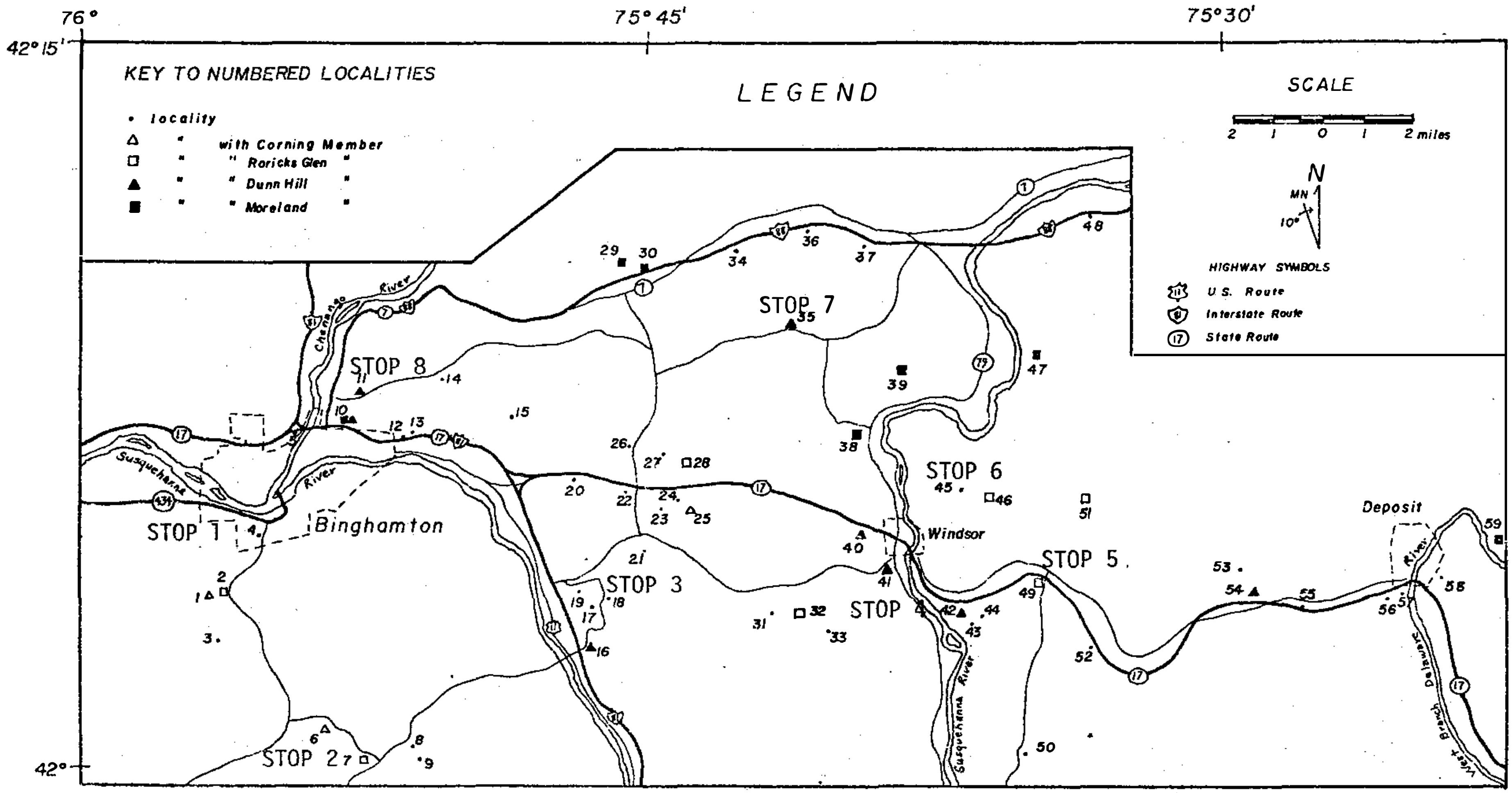


Figure 5. Geographic distribution of the 59 study area localities. Arrows indicate field trip route.

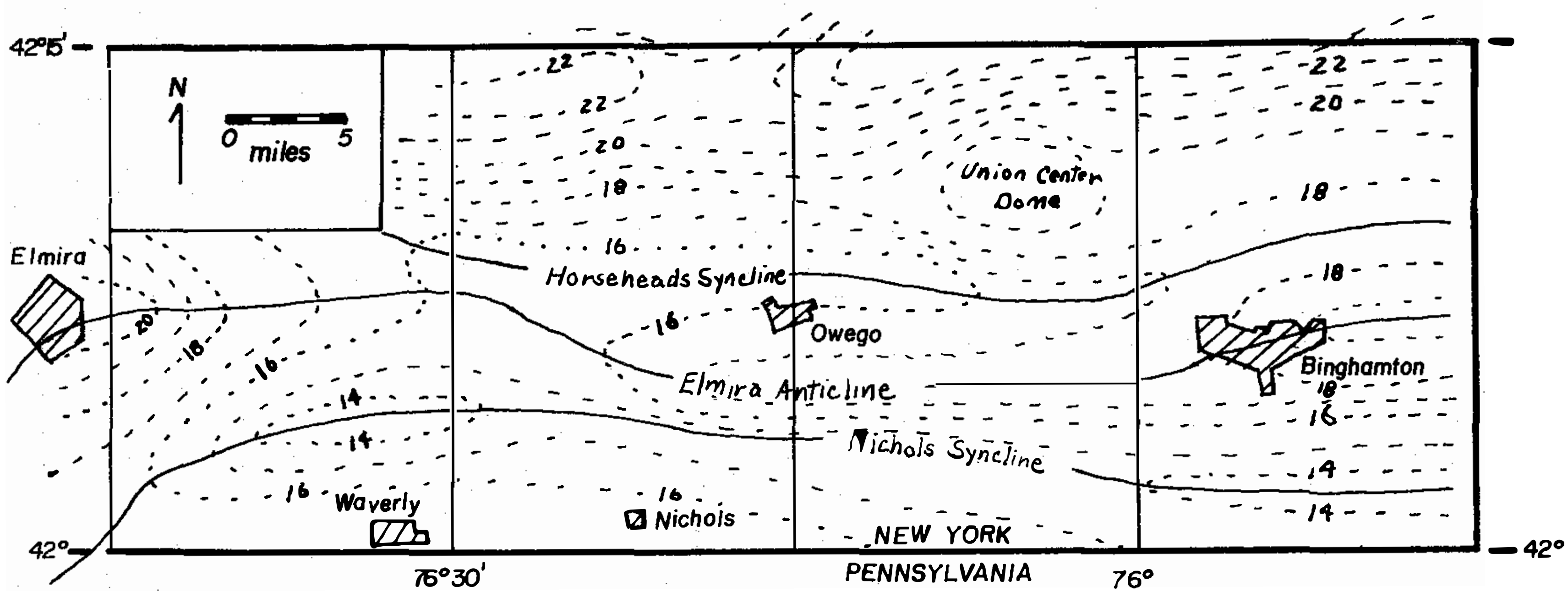


Figure 6. A portion of Wedel's (1932) structure-contour map of south-central New York showing the major structural features near and within the study area. Contour intervals are in hundreds of feet.

Wedel's work by observing that structural features in Bradford County, Pa. (southwest of Binghamton) indeed became more difficult to trace towards the northeast. Wright (1973) compiled deep-well data for much of New York and northern Pennsylvania from which a structure-contour map of the Tully Limestone (Middle Devonian) was compiled. General surface trends reported by Wedel and Woodrow are reflected in this subsurface map.

Using the elevation of exposures of dark shale, the writer employed simple construction methods to determine if the "surfaces" defined by the assumed exposures of individual dark-shale members showed inconsistancies which might either suggest the presence of structure or the incorrect tracing of the shales. The elevations of five assumed exposures of the Roricks Glen Member were found to define a nearly planar surface over a broad area in the central portion of the study area. The attitude of this constructed surface is N 77° W with a dip to the south of 47 feet per mile. This attitude agrees well with that reported by Wedel (1932). With the establishment of the Roricks Glen datum plane, the positions of the remaining dark shales were correlated with it. The results revealed that the other three shale members, within an east-to-west trending belt, defined planes essentially parallel to the Roricks Glen. Three anomalies were discovered, one in the west-central portion of the study area, where elevations of dark shale exposures ranged up to 50 feet higher than expected; and in the northwestern and southwestern areas, where dark shale elevations were on the order of 50 feet lower than expected. These anomalous areas coincide with the trends of the anticline and synclines which Wedel reported to extend into the Binghamton area. It was not determined how far east these features exist; however, due to the low magnitude of structural variation from the regional dip, it was concluded that no errors were made in the tracing of individual shale members.

With the geographic positions of individual shale members established, the 59 study area localities (Fig. 5) were projected onto a cross-sectional plane in a manner similar to that of Sutton and others (1970) and Thayer (1974) using exposures of dark-shale members as the basis from which all other localities were projected (Fig. 7). Since sedimentary-structure data indicates the general direction of sediment transport to be due west (Fig. 8), an east-west trend was chosen for the plane of projection, as this should best represent an average cross section of rock within the Catskill complex.

THE CATSKILL COMPLEX

Before describing the deltaic environments interpreted for the study area, it is important to briefly outline a few general environmental aspects which set apart the Catskill complex from modern deltas. Primarily, the physical nature of the complex as a network of coalesced alluvial plains and fluvial depocenters which paralleled the trend of the rising Acadian landmass is unequaled in modern physiography. In this light, the term "Catskill delta" is misleading as it implies a single river system. Individual delta "lobes" which existed contemporaneously have been recognized (Willard, 1939; Dennison and de Witt, 1972; Sevon, 1979), each of which can be interpreted as constituting a delta. The proximity of several fluvial systems to one another resulted in the lateral blending of

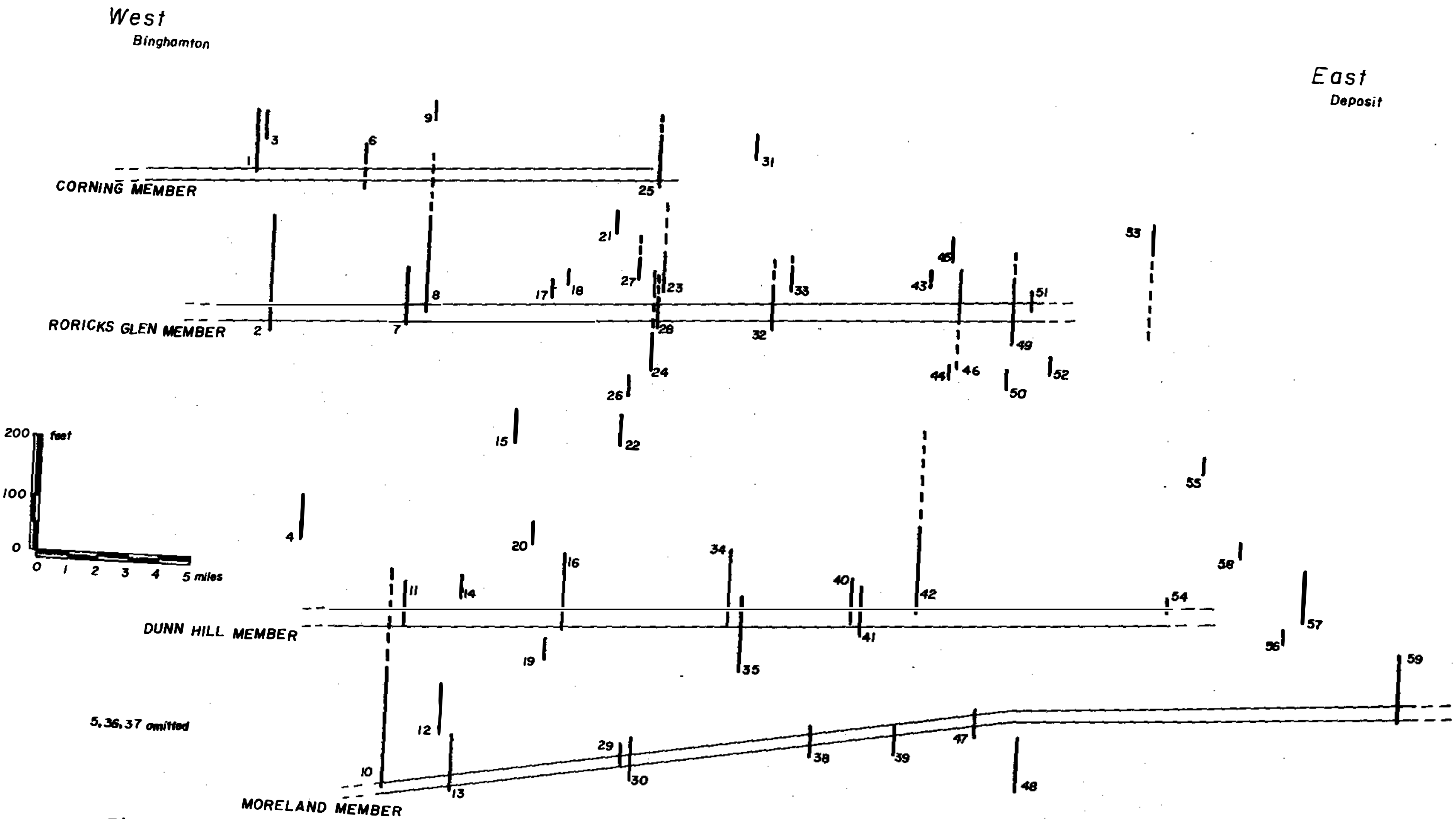
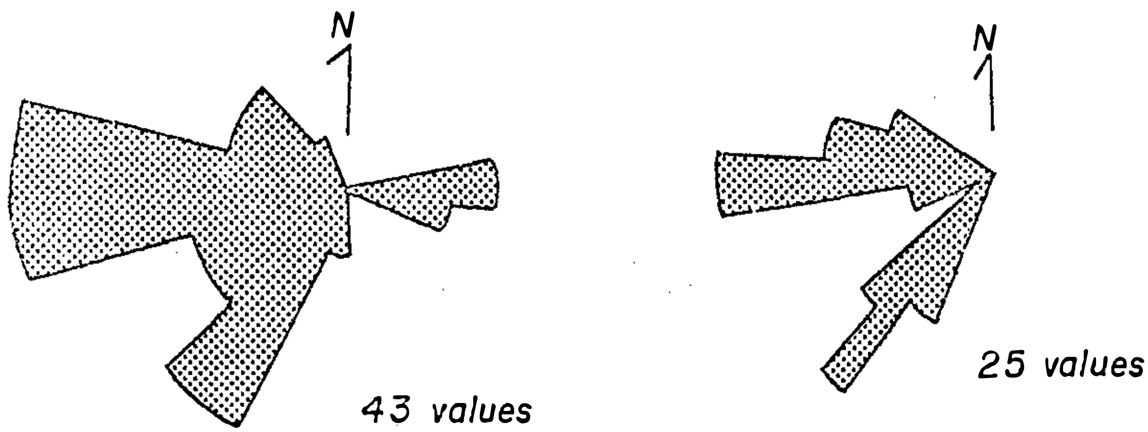


Figure 7. Cross-sectional summary of the stratigraphic and geographic distribution of the study-area localities. The cross section trends N 90° W. Vertical lines represent extent of exposure (dashes indicate partially covered intervals).



(a) all sedimentary structures *(b) current ripples*

Figure 8. Summary of sedimentary structure data. The circular histograms were constructed using 10° class intervals.

their various environments along the prograding shoreline. In areas where fluvial systems were limited in number or size, "interlobe" or"nondeltaic" areas resulted. Thus extensive areas along the shoreline were entirely deltaic, entirely nondeltaic, or transitional in character.

Water depth, which is directly related to the width of the continental shelf in front of an advancing delta, strongly controls deltaic morphology (Kolb and Lopik, 1966; Coleman and Wright, 1975). Large deltas which build into deep water, as well as small deltas, in general, develop recognizable bottomset, foreset, and topset beds of the Gilbert-type delta (Moore, 1966; Morgan, 1970). However, large deltas building into water which is not directly underlain by a rigid basement are characterized by regional downwarping and local sediment compaction (Morgan, 1970). This results in the accumulation of a thick sediment pile which obscures the various beds normally associated with the term "delta". The Mississippi delta, the most commonly used modern example, is this type of delta. In addition, there is a significant difference between the modern Mississippi delta (Balize lobe) and its several ancient lobes. The ancient lobes were formed by progradation into shallow water and are termed "shoal-water deltas" (Gould, 1970). Distributaries of these deltas frequently plugged with sand which resulted in numerous bifurcations and the eventual formation of continuous "sheet sands" around the front of the deltas. In contrast, the Balize lobe is building out into relatively deep water near the edge of the continental shelf. The result is that this lobe differs drastically in size, configuration, and in sediment distribution from the ancient lobes. The Catskill complex is much more similar in character to the ancient lobes, thus making it impractical to use the modern Mississippi delta as a Catskill analogue.

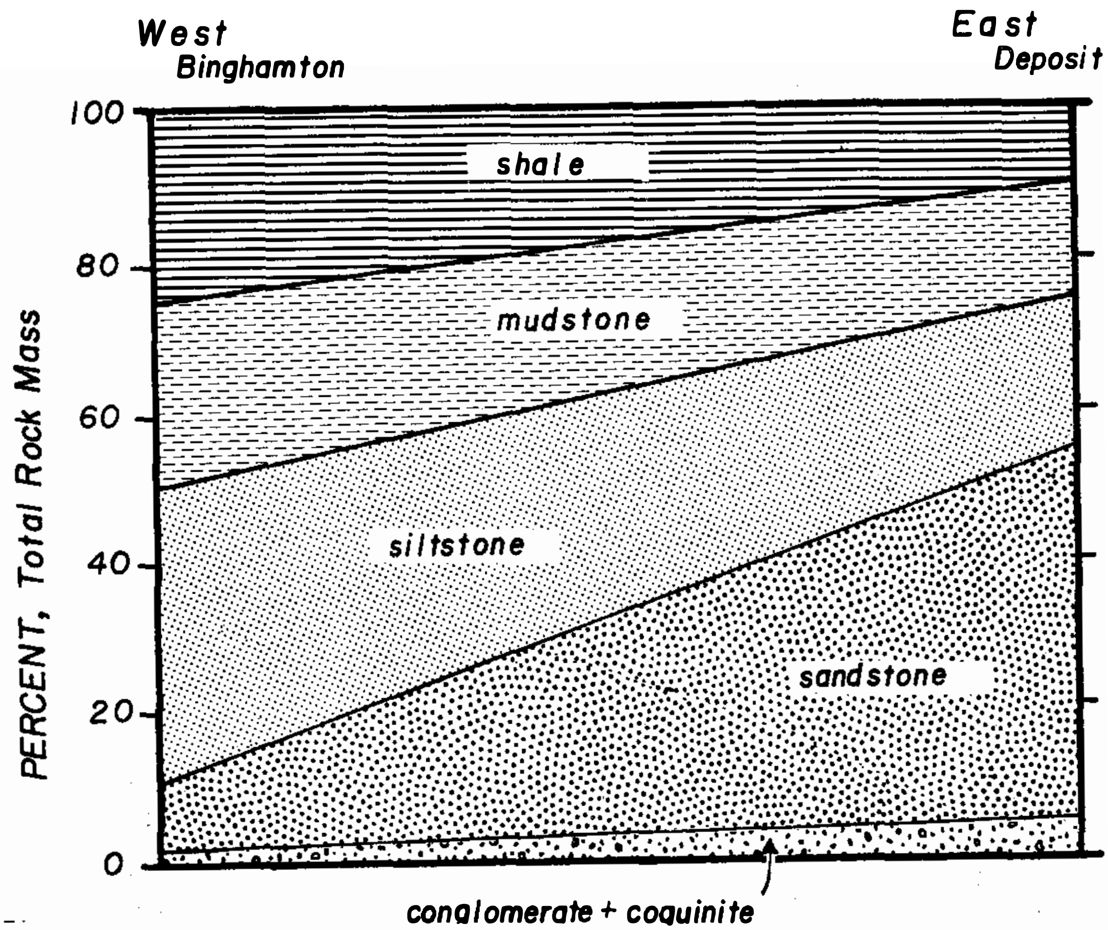


Figure 9. Relative distribution of major lithologies in the study area.

Drainage-basin relief and climate combine to play the dominant role in determining deltaic depositional facies and physiography (Coleman and Wright, 1975). As the Mississippi delta borders a low-relief coastal plain well removed from mountains of active tectonic character, the sediment is predominantly mud and silt with sand representing but a small percentage of total sediment. In contrast, the relative proximity of an active mountain system resulted in a sediment component in the Catskill complex which was dominantly sand (Friedman and Johnson, 1966; Fig. 9). The Catskill alluvial plain, described as being composed of fan, stream, and lacustrine deposits, had a climate which was probably similar to the modern tropical, wet-and-dry savanna of near-equatorial position (Woodrow and others, 1973). The relatively low density of vegetation, especially with respect to grass, resulted in numerous, short-lived floods which provided for the transport of an abundance of coarse sediment to the actively building delta complex. It is the exception rather than the rule for large deltas to have sand as the primary constituent. This is an important factor to consider as sediment grain size plays an important role in the development of environments.

As a modern analogue to the Catskill complex is lacking, it is necessary to reconstruct the Catskill environments from consideration of several delta systems both modern and ancient. Although the modern Mississippi is the best studied of all modern deltas, the writer concluded that information on its ancient lobes (Kolb and Lopik, 1966; Gould, 1970) is much more suitable for comparison. For modern deltas, the Niger, Rhone, and Orinoco River deltas provide excellent comparative information. Leblanc (1975) summarizes references on various aspects of deltaic sedimentation and should be first consulted for specific studies.

Deltaic Environments

Sutton and others (1970) used descriptions of modern deltas to delineate several marine-shelf environments in the Sonyea Group which directly underlies West Falls strata in the Binghamton area. Nonmarine, marsh, estuarine, distributary-mouth bar, delta platform, delta-front sand, prodelta, and open-shelf environments were described. Through the study of Sonyea environments in conjunction with additional specific information on deltaic sedimentation, the writer recognized deltaic environments in the West Falls group. Proceeding from nonmarine to marine, the three major West Falls environments are: the subaerial delta plain, the subaqueous delta plain and the delta platform. These and their sub-environments, diagrammatically summarized in Fig. 10, are considered in detail below.

Subaerial delta plain. The strictly nonmarine environments are grouped into a general environment termed the subaerial delta plain. The modern counterpart of this area is characterized by a large, low-lying floodplain which is dissected by meandering rivers and streams. In the study area, the subaerial delta plain is dominated by channel deposits composed of coarse-grained sandstone which displays a variety of crossbedding, ripple marks, and cut-and-fill features. A much smaller percentage of the strata is composed of floodplain deposits which are characterized by siltstone, silty mudstone, and shale displaying burrows and ripple marks. The presence of mud cracks indicates that portions of this environment were periodically exposed to the atmosphere. In some places thin red beds are associated with these deposits although in general only a very small percentage of the Catskill facies is actually red. Interbedded channel and floodplain deposits are observed to define fining-upwards cyclothems.

Closely associated with a few channels are poorly sorted, coarse-grained and conglomeritic deposits which are interpreted to be levees. These deposits are similar to levee deposits encountered in borings in the Mississippi River delta plain (Kolb and Lopik, 1966) with the well-indurated character and red-orange color of the sediment produced during periods of exposure and desiccation. Abundant organic debris is a common constituent of both Mississippi delta plain and West Falls levee deposits. The extensive exposure in study-area locality 46 provides excellent examples of the channel, floodplain, and levee sub-environments of the subaerial delta plain.

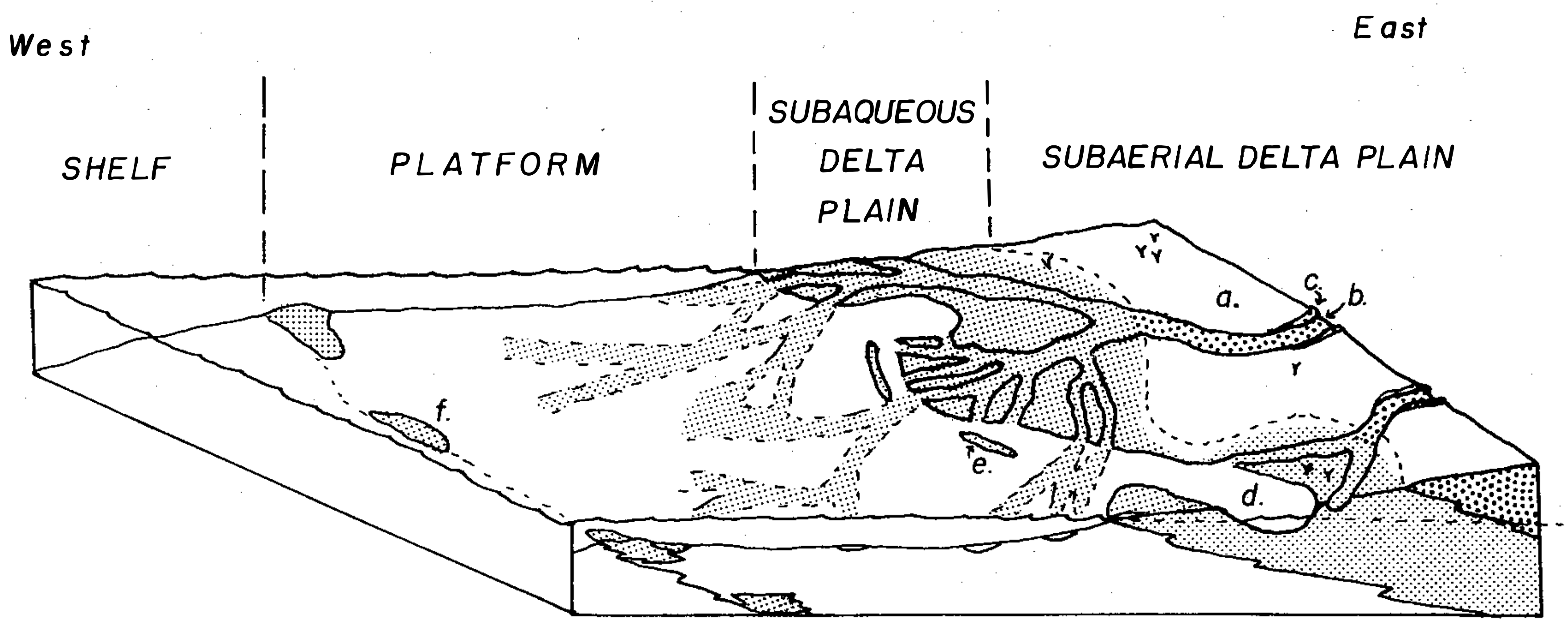


Figure 10. Aerial distribution of West Falls deltaic environments. (a) floodplain, (b) channel, (c) levee, (d) bay, (e) distributary-mouth bar, (f) delta-front sand. Large stipple denotes coarse-grained sand deposits; small stipple denotes fine-grained sand deposits. Sand deposits on the platform are subaqueous continuation of distributaries.

Subaqueous delta plain. Before entering the sea, the rivers of most modern deltas bifurcate into numerous distributaries over a large area which is commonly flooded by tides and river discharges. It is characterized by channels surrounded by marshes and tidal flats which support mangroves, grasses, and other plants. The vegetation tends to stabilize the interdistributary areas. Arms of the sea sometimes reach inland in this portion of the delta in the form of bays. The bays collect silt and clay carried in suspension by the distributaries.

In the study area, large expanses of fine-grained sandstone form a transitional zone between nonmarine and marine environments. This area is termed the subaqueous delta plain. Here, numerous shallow distributaries supplied fine-grained sand to the submerged interdistributary areas. These deposits are now seen as flatbedded and low-angle, cross-bedded sandstones (flagstone). Large channels are rare, indicating that most sediment supplied to this portion of the delta and beyond was carried by the shallow distributaries. These smaller channels frequently plugged with sand and changed course, thus accounting for the large extent of the subaqueous delta-plain deposits. With the exception of oriented plant fragments on a few bedding planes, organic debris is relatively sparse. The lack of stabilizing plant material in this environment would suggest that this would have been an extremely dynamic portion of the delta complex.

Shallow-water marine deposits are also recognized within the subaqueous delta plain. These deposits are interpreted as bays and are characterized by lenses of sandstone interbedded with siltstone, shale, and mudstone. The sandstone lenses formed as a result of channel splaying in the bay vicinity. Marine fossils are rare in the splay sands but are present in significant numbers in the finer-grained bay sediments. This relationship suggests that the splay sands were deposited by streams of relatively fresh water while the remaining bay environment was sufficiently saline to support a marine biota. Locality 1 best displays sub-environments of the subaqueous delta plain.

Delta platform. For modern deltas, the delta platform is described as a broad, terrace-like structure several miles in width which extends from the shoreline to a break in slope at the delta front. This break in slope occurs within water depths between 30 and 60 feet thus defining the platform as a shallow-water marine environment. This area is actively affected by tides and currents which winnow out the finer grained portion of the fluvial-supplied sediment.

As in the Sonyea delta platform, the West Falls platform is dominated by fine sandstone and mudstone. For discussion purposes, the platform is divided into proximal and distal portions. The proximal platform is characterized by numerous lenses of crossbedded sandstone interpreted to be extensions of distributaries into the shallow sea. The water velocity in the distributaries decreased as it moved onto the platform,resulting in repeated bufurcation of the distributaries. In some areas, distributaries broadened and merged with each other to form sheets of sand. Graded beds are common, indicating rapid deposition from turbid water. Accumulations of well-sorted, cross-bedded sandstone are interpreted as distributary mouth bars formed as a result of the reworking and concentration of distributary sand by currents and waves. Current ripples are very common in the proximal platform as are a variety of pillow structures.

Grain size and sandstone bed thickness progressively decrease moving out towards the distal portion of the platform. Sandstone lenses decrease in abundance. Distal platform sedimentary structures include cross lamination, current and wave ripples, and very large pillow structures. Graded beds are still present but are much thinner than those in the proximal platform. Deposits of fossiliferous, crossbedded sandstone occur on the distal platform. These "delta front sands" accumualted by the reworking of sand by currents and waves near the edge of the platform.

As the platform is the largest of the West Falls deltaic environments, this is the environment most frequently seen in outcrop in the Binghamton area. An excellent example of the proximal platform can be seen at locality 42 while the distal platform is best displayed at locality 13.

Dark-mud environment. The four dark-shale members define the only non-deltaic environment observed in the study area. Periodic marine transgression resulted in the accumulation of these relatively fine-grained deposits over the entire range of deltaic environments. The presence of siltstone in the shale members indicates that rivers continued to supply plumes of turbid water to this environment. The presence of only a few fossils in the dark-shale members indicates that this environment was of generally poor quality for fauna.

Cross section of environments. With the environmental interpretation of individual field localities established, the stratigraphic framework of Figure 7 is easily transformed into a cross section of environments in the study area (Fig. 11). This cross section is a summary of data which is the basis for the remaining discussion.

SEDIMENTARY AND ENVIRONMENTAL PATTERNS

Sedimentary Cycles

The distribution of environments in Figure 11 reveals three complete sedimentary cycles associated with the dark-shale framework. The cycles are asymmetrical in configuration with each cycle considered to consist of an interval of regression which is abruptly terminated by a transgressive interval represented by a dark-shale member. (For this discussion, the Moreland Member is not considered as a part of any cycle. In addition, a fourth cycle, post-Corning, was initiated but not completed in the study area due to erosional limitations). During transgression, deltaic sedimentation was essentially halted resulting in the accumulation of the dark muds now represented by the dark-shale members. The net effect of transgression was the displacement of deltaic environments to the east.

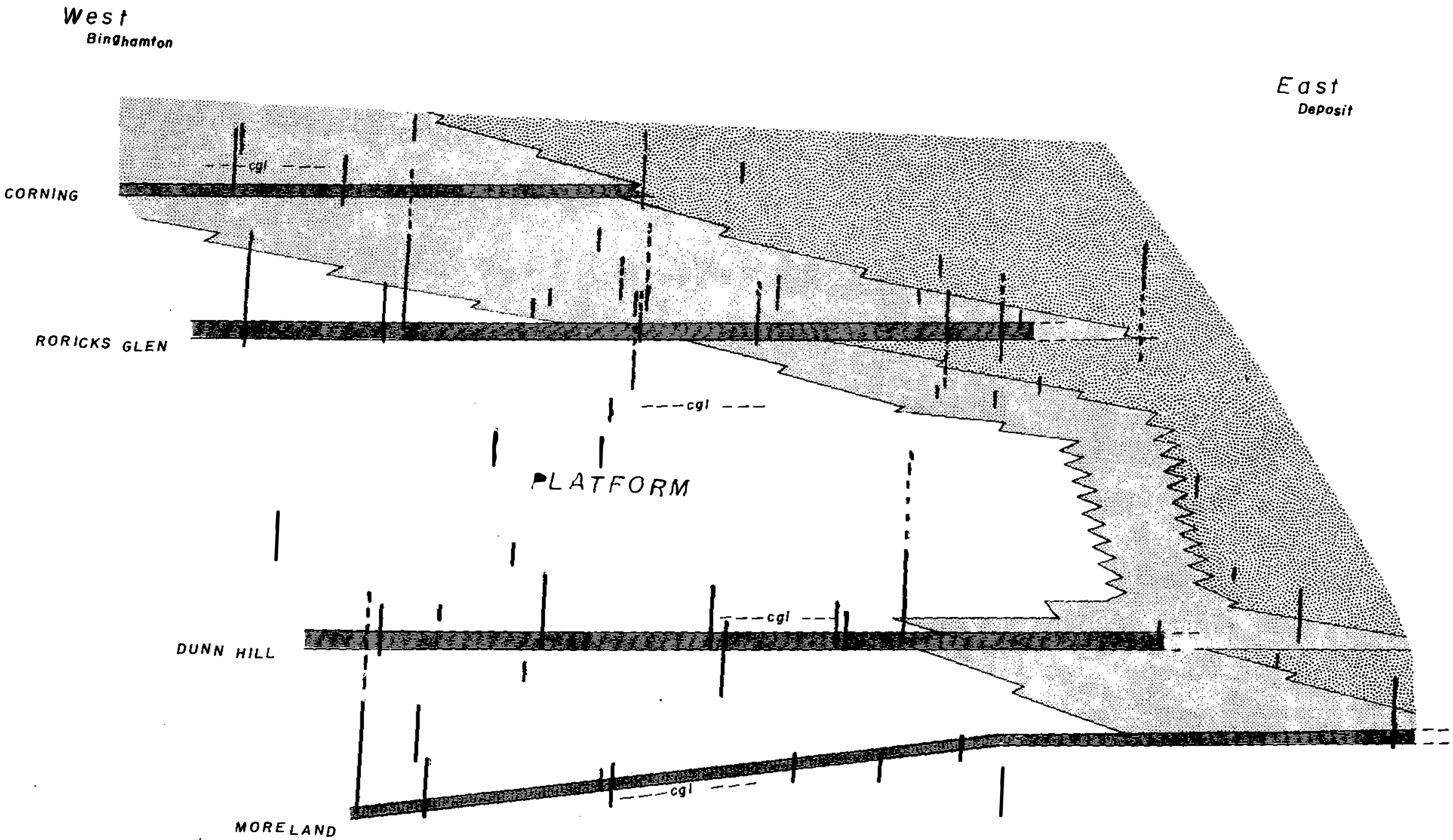


Figure 11. Cross section of West Falls deltaic environments: subaerial delta plain (coarse stipple), subaqueous delta plain (fine stipple), delta platform (blank). Positions of quartz-pebble conglomerates indicated by dashes.

Following transgression, the marine environments redeveloped close to their pre-transgressive positions and continued to prograde westward.

Sonyea time came to an end after a period of rapid regression which extended the width of the delta platform in the Glen Aubrey Formation to approximately 38 miles (Sutton and others, 1970). The prodelta environment which was present in earlier Sonyea time was essentially eliminated due to the rapid westward advance of the delta complex. This final Sonyea regression was brought to a halt by a marine transgression represented by the Moreland Member. As in the Sonyea delta complex, the rate of deltaic progradation in the West Falls delta complex was not uniform. In the Dunn Hill-Roricks Glen interval, progradation virtually reached a standstill in the Binghamton area. Possible explanations for such a standstill include a shift in course of a major river or a gradual subsidence in the nearshore region of the complex at a rate which nearly equalled the supply of sediment. Before completion of the associated cycle, progradation resumed and the deltaic environments rapidly developed westward. Progradation continued at a rapid rate through the last complete cycle. After the deposition of the transgressive Corning Member, regression suddenly resumed resulting in the deposition of the sandstones of the Gardeau Formation which extend several tens of miles west of Binghamton. The obvious erosion of the Corning Member associated with this regressive event is a unique feature with respect to all other shale members in the study area which, with the exception of small amounts of shale-chip conglomerate, conformably grade into overlying rocks.

Significant Features

Figure 11 reveals that the stratigraphic positions of four thin stringers of quartz-pebble conglomerate correlate with periods of rapid deltaic progradation. Nugent (1960) observed three stringers of this type of conglomerate in similar stratigraphic positions to the west of the study area. Woodrow (1968) correlated two of these stringers with conglomerates observed in the Gardeau Formation in Bradford County, Pa. Thus it appears that extensive areas were affected by these rapid pulses of progradation. Uplift of the source area is one possible explanation for the source of energy required for the transport of the pebbles across the platform. An alternative energy source could be provided by intense storms. During storm events, streams and rivers would have altered their courses resulting in the erosion of delta-plain deposits. Previously deposited conglomerates could then be exumed and transported by torrential waters to and across the platform. Downcutting by streams as a result of uplift could have exumed conglomerates in a similar manner; thus, the source area for the quartz pebbles need not be as far east as the Acadian highlands regardless of the source of energy. The high degree of rounding of the pebbles supports the hypothesis that the conglomerates may have been reworked several times before their final deposition. Whatever the mode of deposition, the potential importance of these conglomerates as stratigraphic tools as well as clues to sedimentary mechanisms should be considered in future analysis.

Graded beds observed on the platform are interpreted as proximal turbidites as suggested by Woodrow and others (1973). Catskill climatic conditions allowed for short-lived but intense storms which moved large amounts of relatively coarse-grained sediment across the delta plain and onto the platform. The associated currents winnowed away unconsolidated fine-grained sediment leaving behind shell material as lag deposits. These lag deposits, preserved as coquinites, formed the basal portions of the graded beds deposited as the current velocity diminished. A similar mode of origin for Sonyea coquinites is suggested by Bowen and others (1974) who demonstrated that the coquinites had not been transported very far before their final deposition.

Orientations of current ripples and other sedimentary structures (Fig. 8) imply that two currents were operating on the West Falls delta platform. The westerly current direction is interpreted to be the dominant trend associated with fluvial sedimentation and the transport of sediment across the platform by subsqueous continuation of river currents. A second current direction, approximately S 45° W, is attributed to longshore currents which were operating subparallel to the trend of the shoreline. In addition, interference ripples displaying both directions are common platform-environment sedimentary structures.

At the edge of the delta platform (distal platform environment), crossbedding of delta-front sands show an average orientation of about N 60° W (Fig. 12). This orientation fits a pattern of current-direction change from the platform to the shelf and slope environments where strata thin in a direction of N 60° W. A proportionally larger sediment supply to the region south of New York State could account for this clockwise rotation of the direction of sedimentary-structure orientations.

Factors Controlling Sedimentation Patterns

Transgression. Using the Roricks Glen as a model for dark-shale members, the nature of the West Falls transgressions is now considered. Close examination of the Roricks Glen in the field revealed that the marine portion of this dark-shale member gradually intertongues with rare occurrences of relatively thick floodplain (overbank) deposits. Thick overbank deposition associated with transgression has been shown to exist elsewhere in the Devonian (Johnson and Friedman, 1969). For the Roricks Glen, the thick overbank deposits indicate that drowning of the delta plain occurred which was accompanied by the raising of the base level of the streams and rivers within the delta plain. If the value of 3 feet per mile (Woodrow and others, 1973) is used as an average landslope on the delta plain, shoreline displacement during the Roricks Glen transgression suggests an increase in water depth of approximately 30 feet. As overbank deposition proceeded on the delta plain in response to flooding, the dark muds of the Roricks Glen accumulated on the sediment-starved platform and shelf to the west. By the time the shoreline regained its pre-transgressive position, a minimum of 25 feet of mud (the thickness of the Roricks Glen) had accumulated on the platform. This thickness of mud compensated for the deepening associated with

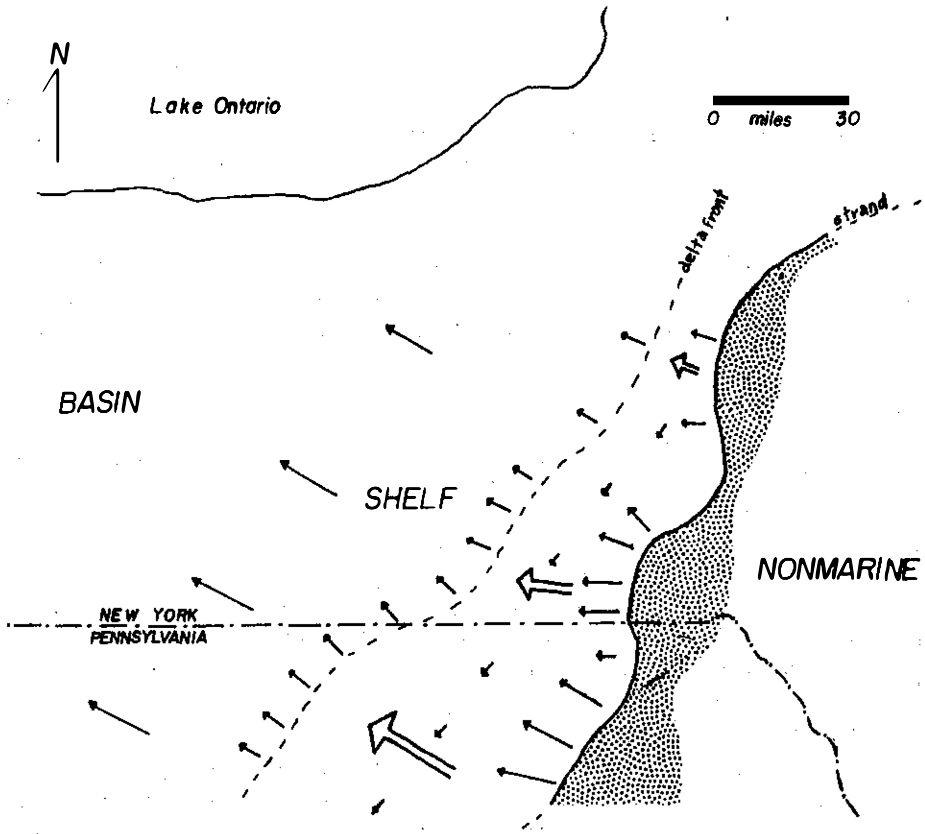


Figure 12. Current directions in the West Falls deltaic and basinal environments. Magnitude of sediment supply indicated by size of open arrows. Stippling denotes the delta plain.

transgression thus allowing the platform to redevelop close to its pre-transgressive position.

Regression. Preservation of the nonmarine correlatives of the Roricks Glen is evidence that a relative lowering of sea level was not responsible for the return of normal deltaic sedimentation on the platform. The effect of a relative sea-level drop would have been the destruction by erosion of overbank deposits in response to the lowering of stream base level. Instead, nonmarine channel and subaqueous delta plain deposits conformably overlie the Roricks Glen. The return of regression is best explained by the completed filling of the flooded delta plain.

Mechanism of cyclic sedimentation. Consideration of both the transgressive and regressive characters of the Roricks Glen model leads to the following conclusions involving the mechanism responsible for the cycles of sedimentation observed in the West Falls Catskill delta complex.

1) Transgression was not caused by an actual rise of sea level.

2) Transgression is best explained by relatively rapid basin-margin subsidence including a large portion if not all of the platform and delta plain. Dark mud accumulated on the platform and shelf as a result of overbank deposition on the delta plain and sediment starvation of environments to the west.

3) Relative lowering of sea level was not responsible for resumed deltaic progradation. Regression resumed after overbank deposition filled in the subsided portion of the delta plain.

Large-scale trends. With the mechanism of transgression and regression established for the model dark-shale member, it is possible to extend this model to the other dark-shale members and briefly consider trends in the evolution of the delta complex on a larger scale. Within the Dunn Hill-Roricks Glen interval, deltaic progradation intensified and resulted in the rapid westward advance of the delta platform onto the shelf. Evidence for this is the presence of the sandstones of the Gardeau Formation far west of the study area. In Gardeau time, a large portion of the shelf appears to have been replaced by platform and subaqueous delta-plain environments which developed in response to an increased supply of sand from the eastern nonmarine environments. It is significant to note a similarly increased sediment supply was determined to have been responsible for the elimination of the prodelta environment at the close of Sonyea time (Sutton and others, 1970). The relatively coarse Gardeau sediments, the correlative Slide Mountain conglomerates to the east of the study area, and evidence for the widespread erosion of the Corning Member suggest that a drastic change in the general pattern of sedimentation occurred in this portion of West Falls time. For the Roricks Glen dark-shale model to remain valid, a relative sea-level drop is necessary to explain the erosion of the Corning. Considering that actual sea-level changes are not evident in pre-Gardeau strata, it appears that uplift of a portion of the basin margin is necessary.

The replacement of the prodelta environment by the delta platform in late Sonyea time may represent the first major effect of the general process of basin filling in New York. Further blending of the delta platform with the marine shelf later in West Falls time could then be interpreted as the next major step in the filling process; the net result is the modification of the various deltaic and basinal environments in response to the shallowing conditions.

To conclude this evaluation of sedimentary mechanisms and trends, the writer suggests that the pattern of subsidence-produced transgressions and uplift later in West Falls time could be extended to define larger cycles of transgression and regression in the Upper Devonian. Such cycles might be represented by the major tongues of black shale and the intervening strata which characterize a large portion of Upper Devonian stratigraphy in New York.

The writer acknowledges Dr. Carlton Brett and Dr. Robert Sutton for reviewing drafts of this manuscript and the thesis from which it was derived.

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ROAD LOG FOR WEST FALLS GROUP DELTAIC ENVIRONMENTS

CUMULATIVE MILEAGE	MILES FROM LAST POINT	ROUTE DESCRIPTION
0.0	0.0	Glen G. Bartle Drive, main entrance SUNY Binghamton; turn right onto Route 434 east.
2.7	2.7	Bear right, Pennsylvania Avenue exit.
2.8	0.1	Stop sign; turn right onto Pennsylvania Avenue.
3.9	1.1	Gillen Road intersection; pull off on right side of Pennsylvania Avenue (unpaved parking area); walk south to road cut on Pennsylvania Avenue. STOP 1.

STOP 1. DISTAL PLATFORM (Field locality 4).

The relatively thin beds of fine-grained sandstone interbedded with gray shale are characteristic of middle- and distal-platform environments. Sandstone beds of approximately one foot in thickness have basal coquinites 1-2 inches thick; thicker sandstone beds have basal coquinites up to one foot thick. The coquinites are typically composed of disarticulated and articulated brachiopod shells with significant numbers of crinoid columnals. Some sandstone beds are graded coarse-to-fine. The shales are generally not very fossiliferous, but localized areas of bedding planes can produce quite a few well-preserved brachiopods and molluscs. A zone of large pillow structures is distinctly visible near the top of the exposure. The zone is laterally persistant for the length of the exposure. These pillows (denoted as type I) are composed of fine-grained sandstone; the distorted matrix of shale in which they are situated suggests that these structures formed as a result of vertical foundering. Smaller, solitary pillows are present at the base of the exposure near the parking area. These are denoted as type II pillows and are probably formed in a similar manner as type I pillows.

This portion of the platform was well-removed from the effects of rivers and streams. Sand which reached this area was supplied in the form of "sheets" from distributaries, which had merged with each other closer to the shoreline. Grading in the sandstones as well as the presence of basal coquinites indicates a "proximal turbidite" origin.

		Continue south on Pennsylvania Avenue; this road is called Hawleyton Road to the south.
8.9	5.0	Church on right; turn left (east) onto Saddlemeir Road.
9.3	0.4	Turn right onto Brady Hill Road.
11.5	2.2	Pull off on right side of road; walk to outcrop on right side of road. STOP 2.

STOP 2. DARK MUD AND PROXIMAL PLATFORM ENVIRONMENTS (Field locality 7).

The dark-gray and rust-brown mudstones and shales of the Roricks Glen Member are exposed in the basal portion of this exposure. Outcrops of the Roricks Glen on the delta platform are all about 20 feet thick and are composed predominately of mudstone with smaller amounts of shale. Thin-bedded lenses of fine-grained sandstone and siltstone are present in the dark-shale member; fossils are few in number. Flagstones and sandstone lenses near the top of the exposure indicate a return to normal deltaic conditions. The large block of cross-bedded sandstone at the very top of the bank (although slightly dislodged) is of channel origin in what is inferred to be the subaqueous delta plain. Cross-bedded sandstone is present in the form of scattered blocks on the hillside above the roadside exposure. The Roricks Glen will be seen again in a similar environmental association later in the field trip.

		Continue east on Brady Hill Road.
11.8	0.3	Turn left onto Brinkman Road.
13.0	1.2	Turn right onto Conklin Forks Road (unmarked).
16.2	3.2	View of Susquehanna River Valley.
17.3	1.1	Interstate 81 underpass; continue straight. This road becomes Cedarhurst Road which bends to the left and continues up the hill.

17.7	0.4	Brink Road intersection; bear right and continue on Cedarhurst Road.
19.7	2.0	Pull off on right side of road and enter roadside quarry. STOP 3.

STOP 3. SUBAQUEOUS DELTA PLAIN (Field locality 18).

The flagstone deposits in the quarry are typical of the subaqueous delta-plain environment. The large-scale crossbedding in the south wall of the quarry is evidence that this portion of the delta plain is gradational with a distributary-mouth bar within the delta platform. This interpretation is supported by a nearby shale pit which exposes normal delta-platform deposits. Current and wave ripples, low-angle cross-bedding and parting lineations are common sedimentary structures at this locality. Orientations of the structures indicates a southwesterly average current direction. With the exception of oriented plant fragments on a few bedding planes, fossils are rare.

		Continue north on Cedarhurst Road.
20.1	0.4	Turn left onto Zimmer Road (unmarked).
21.8	1.7	Turn left onto Trim Street.
22.3	0.5	Turn right onto Route 11 (northbound).
24.3	2.0	Bear right onto Crescent Drive (signs to Route 81 and Route 17).
24.5	0.2	Turn left onto Francis Street.
24.7	0.2	Turn right onto Court Street (unmarked); continue straight (this road becomes Route 17 east).
36.9	12.2	Exit 80 (Damascus, Lanesboro); bear right onto off ramp.
37.2	0.3	Turn left onto State Line Road; continue 0.1 miles, pull off on right side of road just before Route 17 underpass. Walk uphill on road which parallels Route 17 to road cut on right side of road. STOP 4.

STOP 4. PROXIMAL PLATFORM (Field locality 42).

The numerous lenses of sandstone exposed here are subaqueous extensions of distributaries which flowed westward from the subaqueous delta plain. Near the base of the exposure on this road is a zone of organic-rich sandstone that is interpreted to be the result of a rapid regressive (storm) event which interrupted normal progradation of the deltaic environments. A thin stringer of quartz-pebble conglomerate is associated with this sand deposit. This conglomeritic horizon has been found to exist elsewhere in the study area to the west. In addition to a variety of ripple marks which can be observed on the terraces in the upper portion of the outcrop, a lens of large pillow structures and a variety of crossbedding are among the more obvious sedimentary structures.

		Return to entrance ramp to Route 17 east and proceed on eastbound lane.
39.9	2.7	Exit 81 (East Bosket Road); bear right onto off ramp and pull over on right side of ramp. STOP 5.

STOP 5. PROXIMAL PLATFORM AND SUBAQUEOUS DELTA PLAIN (Field locality 49).

The interbedded flagstones, cross-bedded sandstones, and shales in this exposure illustrate the variety of lithologies which coexist at the margin of the subaqueous delta plain. The beds and lenses of sandstone are small distributary-mouth bars formed as a result of the reworking of sand at the fronts of streams. The relatively dark shale was deposited behind the bars in restricted waters of small bays. Cross-bedded, organic-rich sandstone lenses are the remains of distributaries which migrated laterally in this region. The bar deposits gradually give way to subaqueous delta-plain channel deposits which are composed of more poorly sorted, cross-bedded sandstone. Higher up in the exposure is a 10-foot-thick bed of dark shale and mudstone which is correlative with the Roricks Glen Member. This fine-grained deposit grades upwards into drab-colored overbank deposits of the subaerial delta plain.

		Continue up ramp to top of hill.
40.3	0.4	Turn left onto East Bosket Road (overpass).
40.4	0.1	Turn left onto Old Route 17 (westbound).
42.5	2.1	Hard right turn onto unmarked road; sign to Forest Hill Park.
43.8	1.3	Bear left at "Y" intersection.
44.8	1.0	Top of hill; pull off on right side of road; walk up dirt road to the right up to the quarry. STOP 6.

STOP 6. SUBAQUEOUS AND SUBAERIAL DELTA PLAIN (Field locality 46).

The extensive exposure in this quarry (formally called the Ostrander quarry) displays a variety of flagstone and cross-bedded sandstone of stream origin within the delta plain. Ripple marks and scour marks are very common on the surface of terraces in the quarry. Burrowed and mud-cracked mudstone forms the floor on the second terrace (west face of mountain). These and similar deposits in the quarry are of floodplain origin. Red-orange, conglomeritic debris scattered around the area are levee deposits which can be seen in place as lenses associated with channel sandstone in the quarry walls. A thin red bed is visible about two thirds up the sheer face on the western wall. Just above the red bed (one of only a few in the study area) is a dark-shale bed that is correlative with the Roricks Glen Member. A thin coquinite at the base of the overlying sandstone bed contains shell material left as a lag deposit resulting from winnowing of the very top of the Roricks Glen.

		Return to Old Route 17.
47.1	2.3	Turn right onto Old Route 17.
49.6	2.5	Bridge over Susquehanna River.
49.8	0.2	Flashing red light; turn right onto Route 79 (northbound); Village of Windsor.
53.3	3.5	Sign to Nathanial Cole Park; turn left onto Ouaquaga Road.
53.9	0.6	Turn right onto unmarked dirt road.
55.1	1.2	Yield sign; continue straight (road becomes paved).
55.7	0.6	Stop sign, turn left onto Farm-to-Market Road (unmarked), sign for Nathanial Cole Park.
57.1	1.4	Turn right onto dead-end road.
57.2	0.1	Pull off on right side of road; walk up dirt road to the left to quarry. STOP 7.

STOP 7. DISTAL PLATFORM (Field locality 35).

The conspicuous zone of extremely large pillow structures in the lower section of the quarry is a feature which is characteristic of the distal platform environment. These pillows are referred to as type III and are interpreted to have formed due to repeated foundering of relatively coarse platform sediments which were continually supplied to submarine topographic lows by the reworking effects of waves and currents at the edge of the platform. The mudstone and siltstone matrix of the pillows contains a random distribution of small fossil material consisting mostly of crinoid columnals. The matrix as a whole displays a weak cross lamination in places. The number of lenses and thin beds of sandstones increases upsection indicating the westward progradation of the shoreward environments. Basal coquinites are common in these sandstones. (The top of this section, which is exposed along Farm-to-Market Road at the top of the hill, contains a horizon of dark shale within the Dunn Hill Member).

		Return to Farm-to-Market Road.
57.3	0.1	Turn right onto Farm-to-Market Road.
60.4	3.1	Turn right onto Sanitaria Springs Road.
60.8	0.4	Church on left; turn left onto Old State Road.
65.1	4.3	Intersection with Stratmill Road; continue straight on Old State Road.
68.2	3.1	Pull off on right side of road just after "Fallen Rock Zone" sign. STOP 8.

STOP 8. DISTAL PLATFORM AND DARK MUD ENVIRONMENTS (Field locality 11).

Once again, the presence of extremely large (type III) pillow structures indicates the distal platform environment. Here, the pillows are resting upon and within the Dunn Hill Member. The pillows are compositionally similar to those at STOP 7. Currents on the delta platform were active during the deposition of the Dunn Hill as evidenced by these relatively "coarse-grained" structures as well as lenses of sandstone which display crossbedding. Although not shown here, deposits of fossiliferous, cross-bedded "delta-front sands" can also be fairly well developed only a few feet above a given dark-shale member. Thus, a normal deltaic environments rapidly redeveloped after transgression.

		Continue on Old State Road (downhill).
69.0	0.8	Turn left onto Route 7 (southbound).
69.7	0.7	Bear right onto entrance ramp for Route 17 (Elmira) and Route 81 (Syracuse).
70.3	0.6	Bear left at "Y" on Route 17 West (Elmira).
73.8	3.5	Exit 70 S; bear right onto exit ramp; signs for Route 201 S; Johnson City.
75.1	1.3	Traffic circle; bear right onto Route 201 (sign for SUNY).
75.8	0.7	Follow signs for Route 434 E and SUNY.
76.4	0.6	Turn right; main entrance, SUNY Binghamton. END OF TRIP.

THE MIDDLE AND UPPER DEVONIAN CLASTIC WEDGE IN NORTHEASTERN PENNSYLVANIA

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INTRODUCTION

Throughout most of the Paleozoic, much of what presently constitutes the eastern half of North America was part of an inland sea which intermittently received clastic sediment from an eastern source area. The Appalachian basin was the central focus of this sedimentation.

The largest integrated wedge of clastic sediment in this basin was deposited by the Catskill delta system during the Middle and Late Devonian. The purpose of this paper is to present an overview of Late Devonian sedimentation, with particular reference to Pennsylvania. The purpose of the field trip is to examine some of the rocks from which interpretations about the Catskill delta system are made. The information presented here derives from the literature as well as my own work. The term "delta system" as used here refers to multiple contiguous deltas operating in the same sedimentary basin at approximately the same time. The Catskill delta system is also a tectonic delta complex in the sense defined by Friedman and Johnson (1966, p. 185-186) for New York State and used by Humphreys and Friedman (1975, p. 369-370) in Pennsylvania: "a deltaic complex built into a marine basin contiguous to an active mountain front and dominated by orogenic sandstone derived from the nearby tectonic highland."

GEOLOGIC SETTING

Source Area

Sedimentary rocks of Paleozoic age occur at the surface or in the subsurface throughout the length of the eastern part of North America. Exposures of rock comprising the total sequence occur in the Appalachian Mountains and extend from central New York to Kentucky. The Cambrian and Ordovician rocks are dominantly carbonates although a moderate quantity of Lower Cambrian clastics occur. The source for these sediments appears to have been to the west. The remainder of the Paleozoic rocks are dominantly clastics and paleocurrent, isopach, and lithofacies data indicate derivation from an eastern source area.

The dramatic change from a western to an eastern source area and the apparent absence of an eastern source has only recently been satisfactorily resolved by the development of plate-tectonic models for the eastern margin of North America (Bird and Dewey, 1970; Schenk, 1971; Dietz, 1972; Hatcher, 1972; Dewey and Kidd, 1974; Rankin, 1975; and Van Houten, 1976). The general plate-tectonic model for the eastern North American continental margin is shown in Figure 1 and a more detailed model for the Devonian is shown in Figure 2.

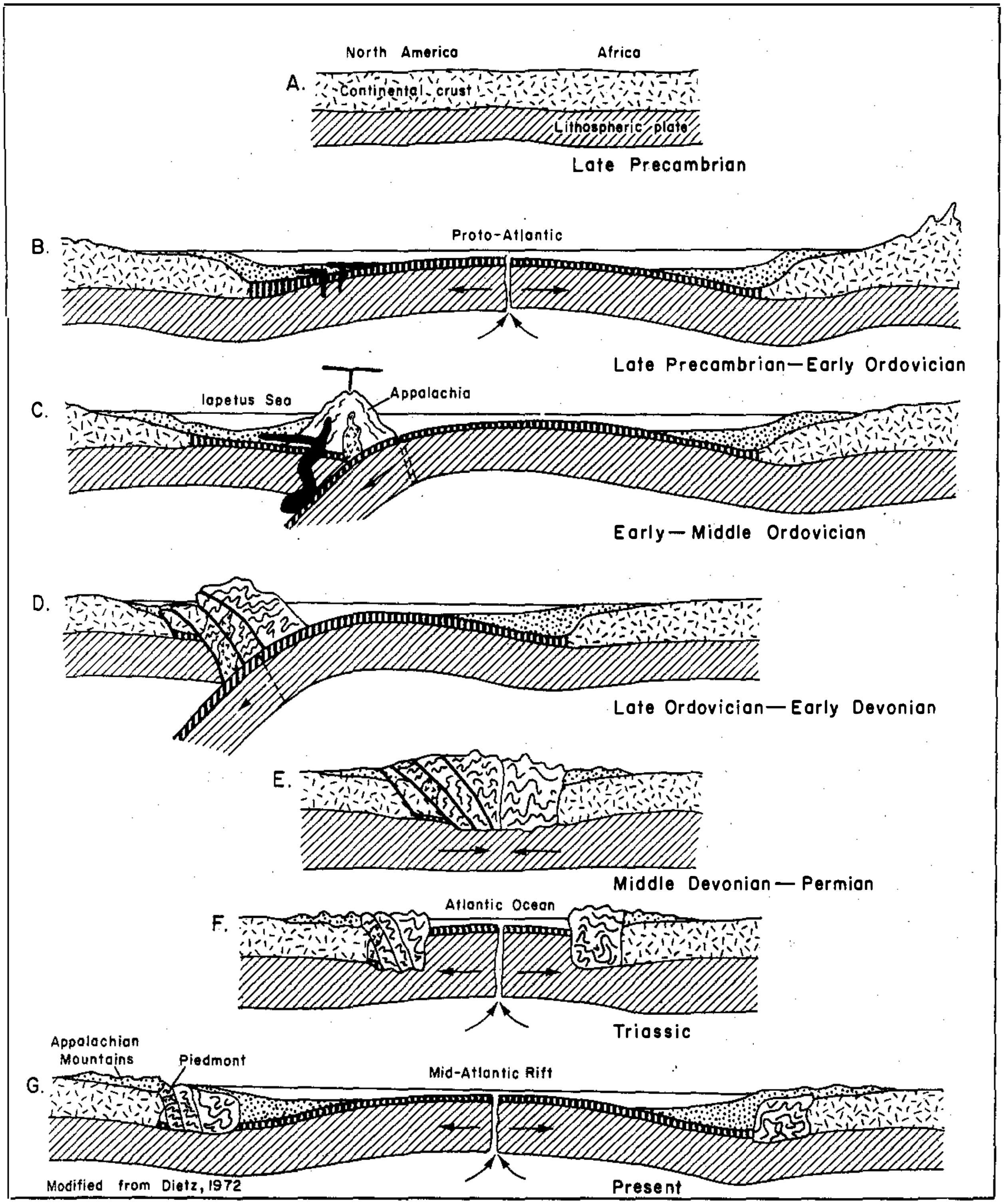

Figure 1. Diagrammatic model of the plate tectonic history of the central Appalachian Basin.

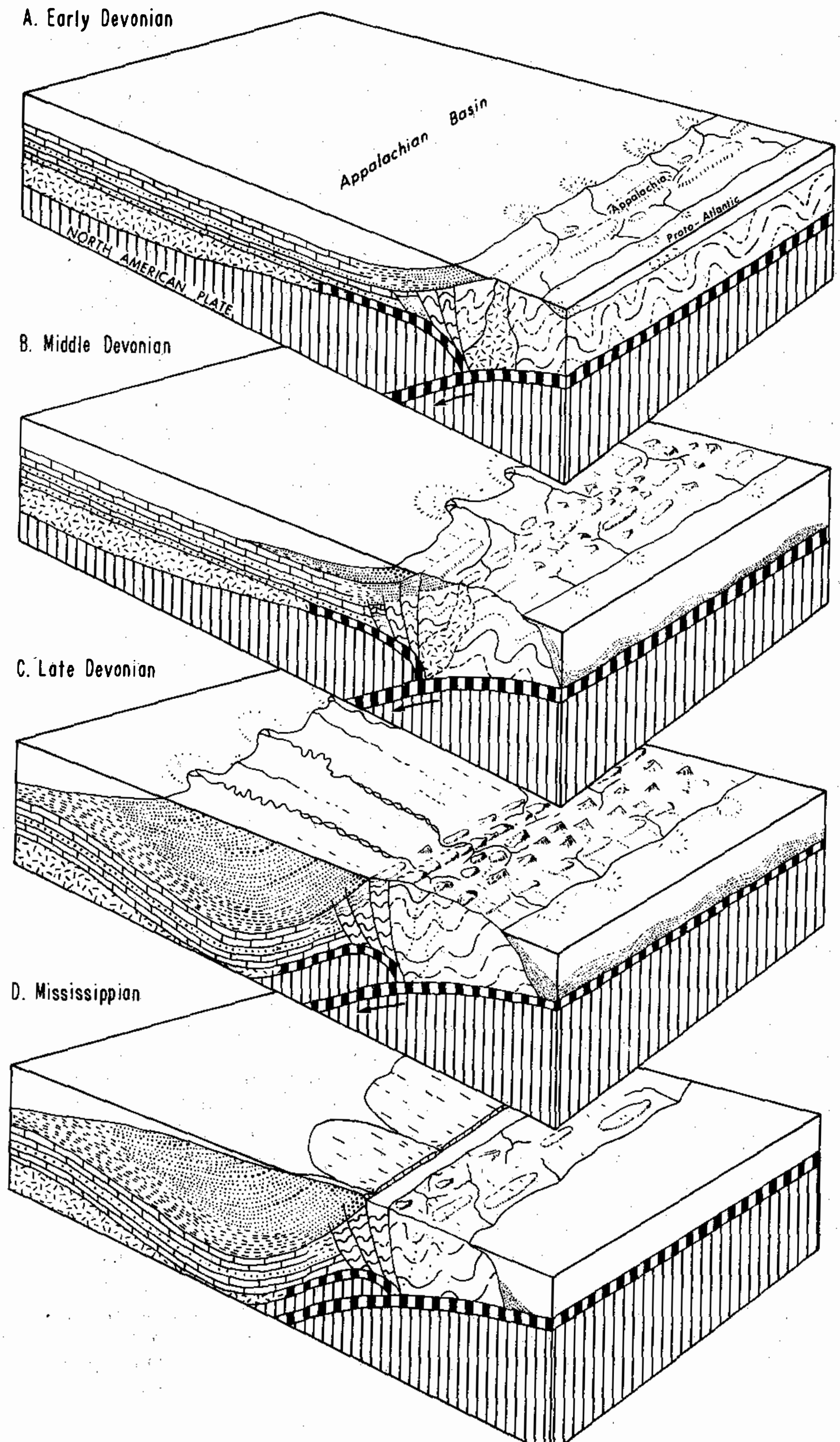

Figure 2. Diagrammatic model of Devonian history of the central Appalachian basin.

During the Late Precambrian through the Early Ordovician the eastern margin of the North American continent was a miogeocline in which some clastics and much carbonate were deposited (Fig. 1B). Sediment entered the basin from the west. Sometime in the Ordovician, North American, South American, and African plate divergence stopped and convergence commenced. Convergence continued following the formation of an island-arc system during the Ordovician (Fig. 1B), and a continental land mass, Appalachia, was developed by Middle Devonian time (Fig. 1E). Appalachia comprised uplifted and metamorphosed Precambrian(?), Cambrian, and Lower Ordovician miogeoclinal sediments as well as volcanics and instrusives. This composition has been confirmed by identification of coarse clasts (Barrell, 1914; Mencher, 1939; Sevon, 1969; Perry and deWitt, 1977; Sevon and others, 1978; Seaman, 1979; and Kirby, 1981) and thin-section petrography (Mencher, 1939; Lucier, 1966; Sulenski, 1969; Kramers and Friedman, 1971; Humphreys and Friedman, 1975; and Sevon, unpublished data) of Catskill rocks. These studies indicate a mixed terrain of low-grade metamorphic and sedimentary rocks and a general absence of feldspar. Quartzites are particularly common as coarse clasts and sericite-chlorite-rich rock fragments are common in thin section.

Estimates of the position of Appalachia during the Devonian and Mississippian range from 40 km (Lucier, 1966) to 204 km (Pelletier, 1958) east of the present outcrop in Pennsylvania and New York. Its actual position is not known, but estimates of 50 to 100 km seem reasonable.

Appalachian Basin

After the development of Appalachia, the eastern part of North America became an elongate inland sea, the Appalachian basin, with a central focus of sediment accumulation in New York and Pennsylvania (Fig. 3; Colton, 1970; Cook and Bally, 1975) and farther east. The large quantities of sediment contributed by Appalachia caused marked subsidence in the eastern part of the basin (Fig. 2C), and presumably even greater thicknesses of Middle and Upper Devonian sediment, now lost to erosion, were deposited east of the present outcrop margin.

Local tectonic activity contemporaneous with sedimentation in the Appalachian basin has been identified primarily in the present Appalachian Plateau part of the former depositional basin, but also occurred farther east. The presence of deep-seated faults along which recurrent movement occurred during Paleozoic sedimentation is discussed by Bradley and Pepper (1938), Woodward (1963), Kelley and others (1970), Harris (1975), Wagner (1976), and Root (unpublished, Pennsylvania Geological Survey, Harrisburg). Growth of folds and their effect on coal deposition in western Pennsylvania is discussed by Kent and Gomez (1971), Williams and Bragonier (1974), and McCulloch and others (1975). Growth of folds during Devonian sedimentation in eastern Pennsylvania and New York has been suggested by Fletcher (1964) and Fletcher and Woodrow (1970), and in north-central Pennsylvania by Woodrow (1968). The Wyoming-Lackawanna Basin in northeastern Pennsylvania was tectonically active during the Mississippian and may have been active during the Late Devonian (Woodrow and Fletcher, 1967; Glaeser, 1974).

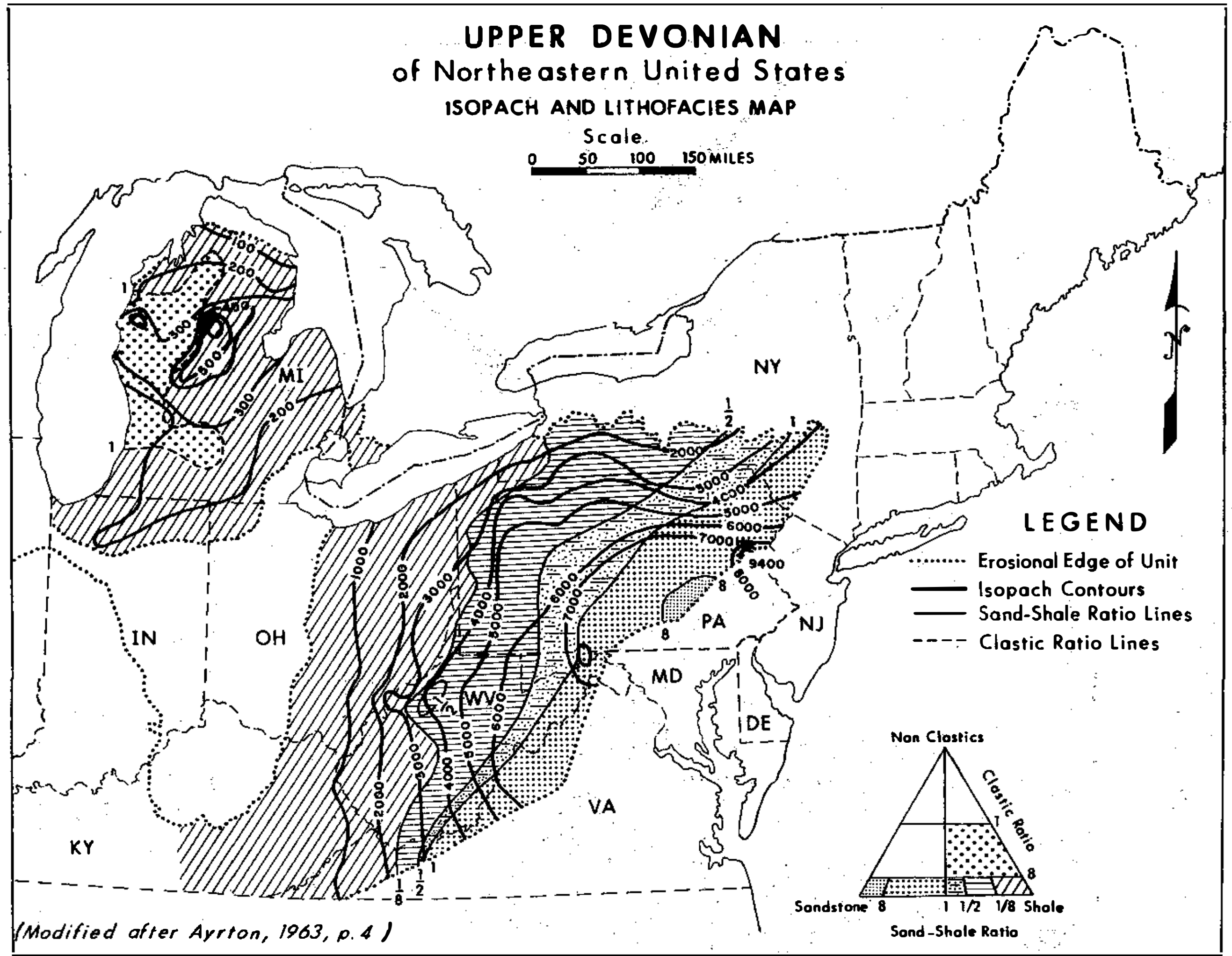

Figure 3. Isopach and lithofacies map for the Upper Devonian of northeastern United States.

Paleogeography and Climate

Available paleoclimatic data combined with some paleomagnetic data allows reconstruction of Devonian world paleogeography (Woodrow and others, 1973; Heckel and Witzke, 1979; Ziegler and others, 1979; and Bambach and others, 1980) and a reconstruction for the Late Devonian is presented in Figure 4. In this reconstruction, the Catskill deltaic system would have developed in an equatorial belt affected by easterly

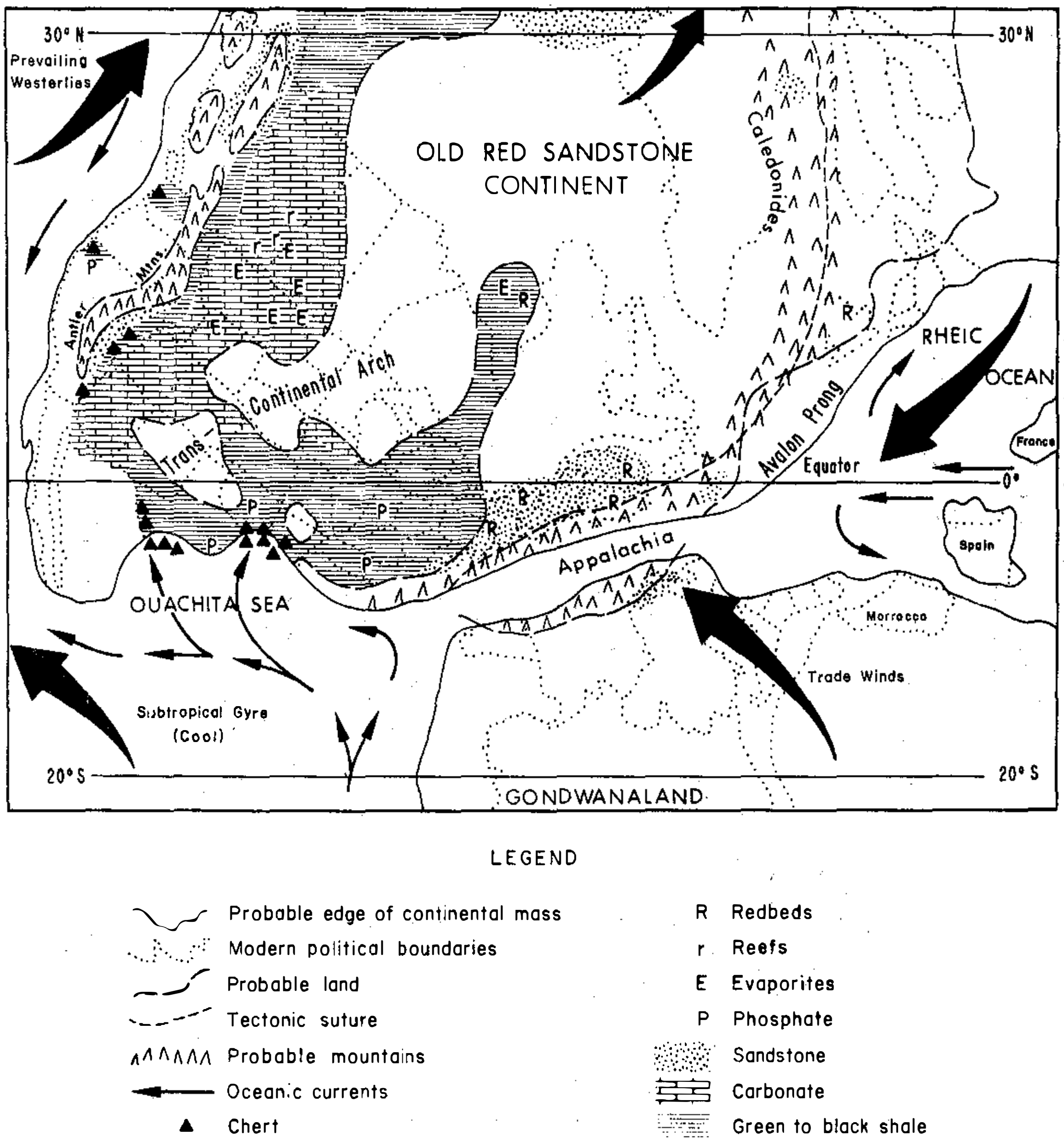

Figure 4. Late Devonian paleogeography and lithofacies for North America. Diagram simplified and modified from Ettensohn and Barron (1980, p. 18).

trade winds. Because of the wind barrier created by Appalachia, the Appalachian basin would have existed in a rain shadow. Seasonal aridity and warm-to-hot temperatures prevailed in the basin and presumably in the western part of Appalachia. Vegetation was sparse to lush on the depositional plain, but its seasonal fluctuations are not known. The extent to which vegetation may have existed in Appalachia is likewise not known.

Modern Analogue

The spatial and tectonic relationship between the New Guinea island arc and the Australian craton (Fig. 5) may be the modern analogue for Appalachia and the North American craton during much of the Paleozoic (Dott and Batten, 1971, p. 295). The equatorial position is not the same in this model as during the Late Devonian, but wind circulation appears comparable and climatic conditions may be similar.

THE CATSKILL DELTAIC SYSTEM

Introduction

Sediments deposited in the Appalachian basin during the Late Devonian are often attributed to the "Catskill delta" with the implication that a single delta of unspecified nature was responsible for all of the sediments. The real situation was well stated by Barrell in 1913 (p. 466):

> "The uniformity in character of the delta from northeast to southwest, its development marginal to the uplands, and somewhat rapid gradation from gravel to sand and clay on leaving the mountains suggests the presence of a number of comparatively short streams which build flat coalescing fans rather than the debouchement of one or two great continental rivers."

Elaboration of this concept of Barrell is the object of the remainder of this paper.

Sediment-Input Systems

Willard (1934) was the first to attempt to define the number and position of the rivers which brought sediment to the Appalachian basin. He named and sited 3 delta lobes in Pennsylvania: Fulton (south central), Snyder (central), and Wyoming (northeast). These lobes defined the hypothetical position of the early Chemung shoreline (the base of Chemung is marked by the first appearance of the brachiopod, *Spirifer disjunctus*). Caster (1938) illustrated the position of 3 arcuate deltas in northwestern Pennsylvania, but says nothing about them. I recently identified the position of the axes of 8 sediment-input systems for the Upper Devonian Appalachian basin (Fig. 6; Sevon and others, 1978; Sevon, 1979a). Locations of axes of sediment-input systems in New York are derived from Burtner (1964) and McCave (1968), and those in Virginia-West Virginia from Dennison (1970) and Dennison and deWitt (1972).

Facies Progradation

At any particular instant of time during development of the Catskill clastic wedge, a variety of specific subaqueous and subaerial environments of deposition coexisted. As progradation and subsidence occurred, one

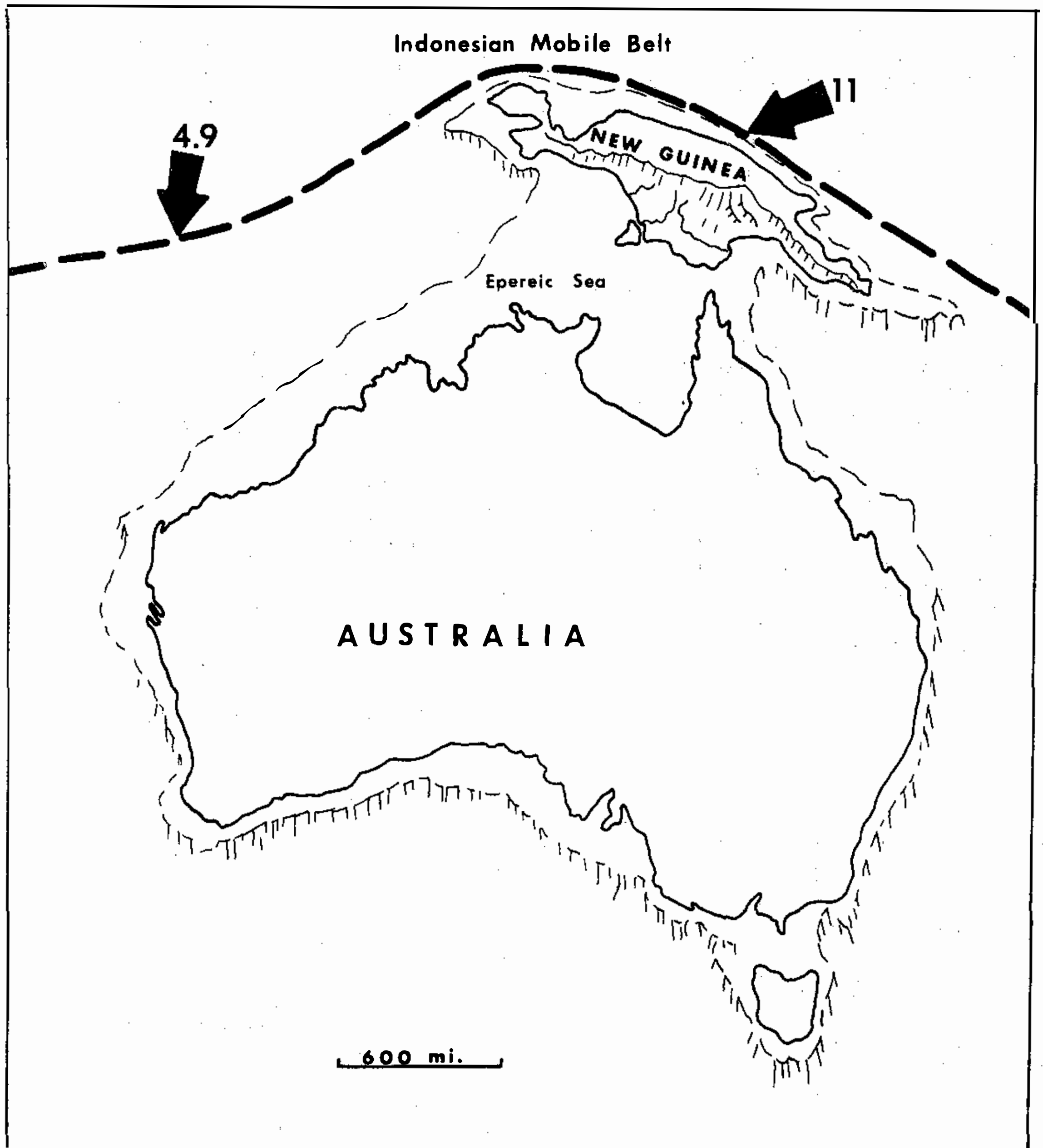

Figure 5. New Guinea and Australia - a modern analogue for Appalachia and the North American craton (Dott and Batten, 1971, p. 295). Direction of tectonic compression indicated by arrows and rates given in cm/year (LePichon, 1968).

depositional environment was succeeded by another and the resulting vertical sequence of sediments exemplified Walther's (1884) law of facies: under the controls of transgression or regression, facies which coexisted laterally will be preserved vertically in stratigraphic sequence. Generally the succession of environments was an orderly progression from farthest offshore to farthest onshore, and the vertical sequence preserved today is, as a whole, one of upward coarsening. These lateral and vertical relationships are shown in Figure 7.

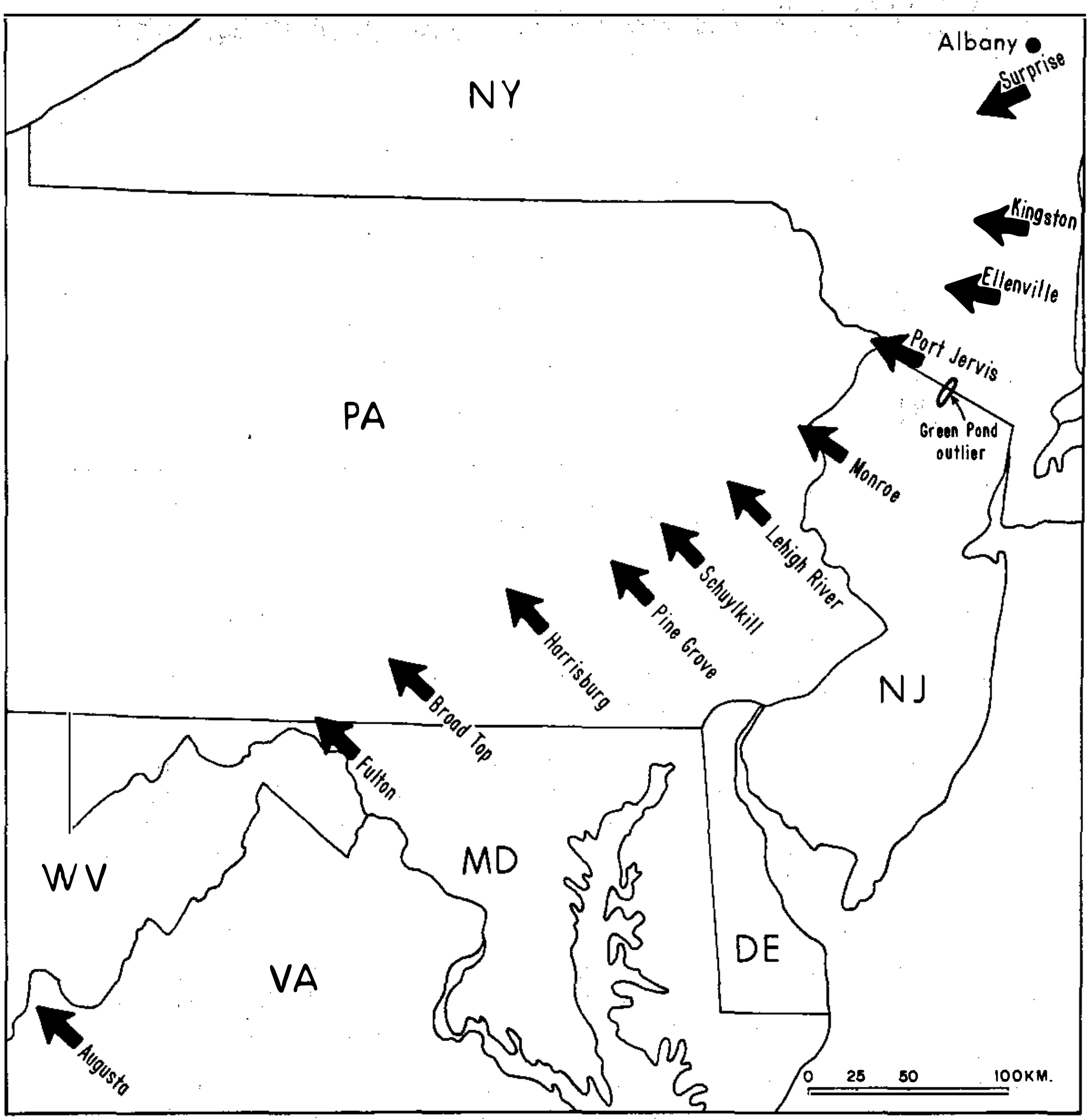

Figure 6. Interpreted axes of sediment-input systems entering the central Appalachian basin during the Devonian.

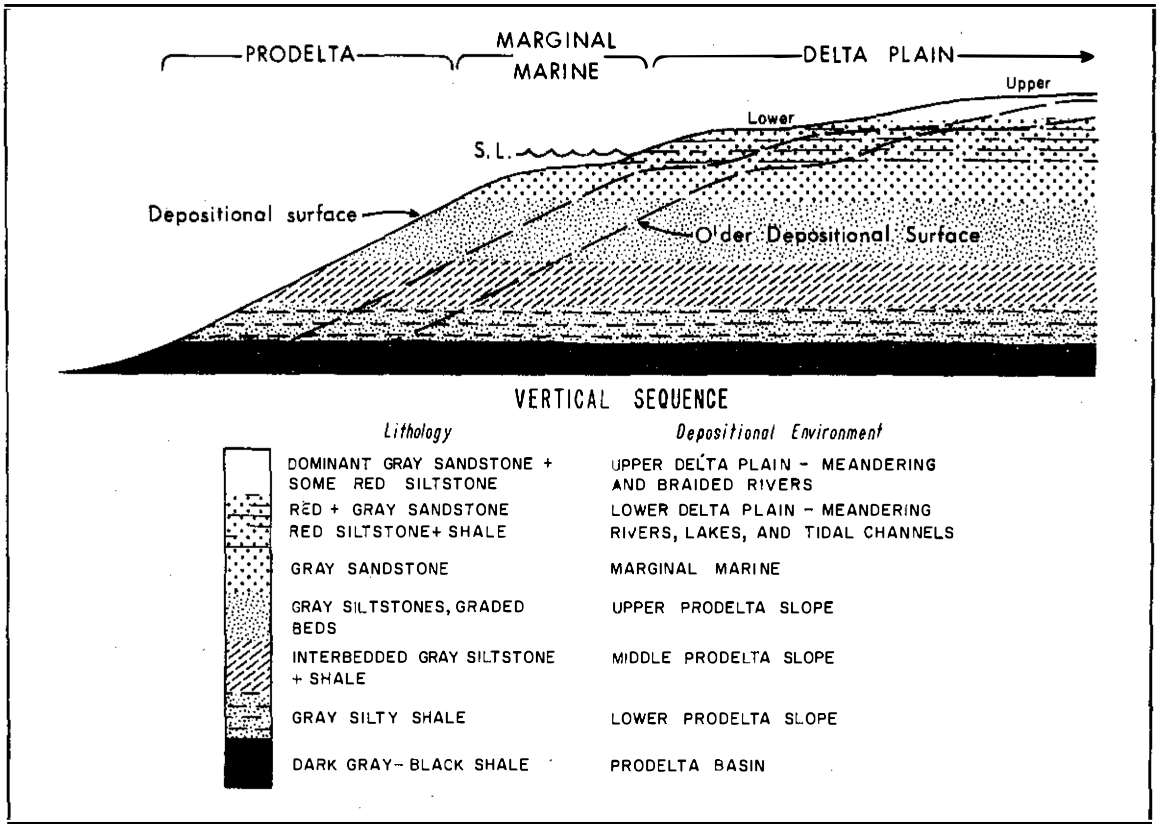

Figure 7. Idealized Middle and Late Devonian Catskill delta progradation model. Modified from Glaeser (1979, p. 347).

Although an orderly progression of rocks representing successive depositional environments occurs uninterrupted in some places, interruptions of orderly progression are common and represented by repetitive sequences. Examples of such sequences are:

1. The Irish Valley "motifs" (repeated facies sequences) in the Irish Valley Member of the Catskill Formation, east-central Pennsylvania (Walker and Harms, 1971; Walker, 1972) which comprise numerous repeated sequences of gray-to-green marine sandstone and/or shale passing upwards into non-marine red mudstone.

2. The Walcksville, Beaverdam Run, and Long Run Members of the Catskill Formation in northeastern Pennsylvania (Epstein and others, 1974; Berg, 1975; Berg and others, 1977) in which the deposits of the Beaverdam Run represent a major marine transgression onto the lower delta plain.

3. Changes in depositional environment and resulting rocks are correlated with major basin-wide sea-level variations by Sutton (1963), Sutton and others (1970), and Dennison and Head (1975). Sutton recognizes regionally correlative black to dark-gray shales which he interprets to

be the deposits representing extensive marine transgression onto the Catskill clastic wedge. Dennison and Head recognize regional lateral migration of terrigenous clastic environments and near-shore carbonate environments in response to sea-level change. Ettensohn and Barron (1981) present a model for the cyclic alternation of black shales and coarser clastics in the Catskill clastic wedge.

4. Glaeser (1974) recognized both regular environmental succession and numerous interruptions in sequence in rocks of the Catskill Formation preserved in the subsurface of northeastern Pennsylvania.

Delta Model

Although the term, delta, has been applied many times to the origin of the progradational deposits of the Catskill delta, a specific model has never been established. In fact, both Allen and Friend (1968) and Walker and Harms (1971) rejected the concept of deltaic deposition. Allen and Friend suggested that Catskill sedimentation occurred in a vast alluvial coastal plain characterized by barrier islands, tidal flats, and lagoons at its western margin, and by meandering and braided streams in its eastern parts. Walker and Harms argued that, at least in south-central Pennsylvania, deposition occurred along a quiet, muddy, prograding coastline which received sediment via longshore currents from a distant source.

In reality, however, the overall progradational character of the Middle and Upper Devonian rocks is consistent with the definition of a delta given by Ferm (1970, p. 247-248):

> "Recent marine deltas form when sediments, carried by rivers into relatively large bodies of open water, accumulate at the river mouth until the surface of the sediment pile reaches sea level. The emergent portion comprises the subaerial expression of the delta.... Delta growth continues as sediment-laden streams pass over the emergent surfaces and deposit sand, silt, and clay over the frontal delta slope. As this process of building new land at the delta margin continues, the delta is said to prograde, and the product of progradation can be thought of as the typical delta sequence."

The formation of the Catskill clastic wedge fits this definition of a delta, but no single delta model can be applied to the whole because the numerous sediment-input systems varied in several ways which affected the manner in which each interfaced with the Appalachian basin:

1. The intensity and timing of orogeny was laterally variable resulting in differential uplift in Appalachia and differential sediment supply to the basin.

2. The length of each of the numerous sediment-input rivers varied through time in relation to itself and the other rivers. This variation

resulted in differences in gradient, grain size of the sediment transported, and, possibly, sediment quantity. These differences in turn caused variations in the character of the depositional environments associated with each stream.

3. The interaction between adjacent sediment-input systems was variable.

The result of the large potential for variation during deposition of the Catskill clastic wedge is that the rocks comprising that wedge are extremely variable in both the vertical and horizontal dimensions. Thus, the interpretations of both Allen and Friend (1968) and Walker and Harms (1971) are correct for specific rocks in specific places, but those interpretations do not necessarily apply to other rocks in other places.

Friedman and Johnson (1966) pointed out that the Catskill deltaic complex in New York State differs considerably from the modern Mississippi River delta (a frequently used model at that time), but is deltaic in nature. Manspeizer (1969) outlined the physical character and dimensions of a single delta complex in south-central New York and north-central Pennsylvania, but, unfortunately, he did not apply a specific model to the rocks nor did he publish the details of the study. The only satisfactory fitting of a delta model to Devonian rocks in Pennsylvania is the work of Kaiser (1972) who established the suitability of the Rhone River delta model for the Middle Devonian Montebello Member of the Mahantango Formation in south central Pennsylvania. Much work remains to be done before even a general model of the whole progradational complex can be generated.

Depositional Environments

Rocks of the Middle and Upper Devonian clastic wedge have long been attributed to sediment deposition in both marine and non-marine environments. However, it has only been in the last 25 years that specific depositional environments have been recognized and described in detail for these rocks. Table 1 presents a summary of the various environments which have been described for the clastic wedge. There is adequate literature available describing the characteristics of these environments and they are not elaborated upon here. Figure 7 illustrates the lateral and vertical position of the broad categories in which more specific environments occur.

Demise of the Catskill Deltaic System

The rocks in Pennsylvania representing latest Devonian and earliest Mississippian time indicate that the following events occurred: (1) sedimentation on the Catskill alluvial plain stopped, (2) erosion occurred in areas nearest to Appalachia and was greatest at the centers of the axes of sediment-input systems, (3) marine waters transgressed onto the alluvial plain (Dennison and Head, 1975), (4) polymictic diamictite, a unique lithology in the Paleozoic rocks of Pennsylvania, was deposited in the areas of greatest erosion (Sevon, 1979b) and (5) the Wyoming-Lackawanna basin actively subsided. The following hypothesis is presented as an explanation.

Subaerial sedimentation at the end of the Devonian was remarkably uniform throughout the Appalachian basin and comprised mainly deposition by meandering streams. During the period when the northwestern South American plate pulled away from the northern Appalachian plate (Dewey and Kidd, 1974), rifting occurred in either Appalachia or the most proximal part of the Catskill alluvial plain. This rifting beheaded the sediment-input rivers and dammed waters draining Appalachia. The cessation of sediment-input caused by the rifting resulted in a sedimentation-subsidence imbalance and marine transgression occurred in the distal part of the alluvial plain. The dammed waters eventually crested the rift dam and began downcutting of the proximal alluvial plain. Erosion may have been amplified in northeastern Pennsylvania because of subsidence in the Wyoming-Lackawanna basin. Transgression apparently encroached far enough onto the alluvial plain to drown some of the newly eroded valleys (Fig. 2D).

At this point one or more flood events of enormous proportions flushed a heterogeneous mixture of debris from Appalachia and resulted in deposition of the polymictic diamictite in the drowned valleys. Regression occurred and fluvial deposition was eventually restored throughout most of the basin.

The diverse lithologies present in the polymictic diamictite provide the final and most intimate information about the composition of Appalachia. Thereafter, throughout the remainder of the Paleozoic, sediment brought into the Appalachian basin from the east was derived by erosion of the most proximal parts of the former alluvial plain.

Paleontology

The Appalachian basin had abundant life during the Middle and Late Devonian. Marine invertebrates flourished wherever environmental niches were available. Fish were apparently abundant in the rivers flowing from Appalachia, and lungfish survived the dry seasons by burrowing into fluvial muds. A variety of air-breathing creatures left abundant tracks and trails in the alluvial muds. Land plants were at least seasonally abundant on the delta plain, but their presence or absence in Appalachia is conjectural.

The fossil remnants and fossil traces of the various forms of life vary from abundant in rocks of marine origin to absent in many rocks of non-marine origin. Good entries into the literature of the marine faunas associated with the clastic wedge are: McGhee and Sutton, 1981; Thayer, 1974; Bowen and others, 1974; Willard and others, 1939; and numerous papers in House and others, 1979. The distribution of trace fossils in the Catskill in New York was recently reviewed by Miller (1979). Berg (1977) discusses bivalve burrow structures and Woodrow (1968) documents aestivation burrows of Devonian lungfish. The world of Devonian plants can be entered through papers by Banks (1966) and Chaloner and Sheerin (1979).

Stratigraphy

The Pennsylvania Geological Survey maps rock-stratigraphic units. These map units represent, for the most part, reasonably homogenous lithologic entities with more or less distinct boundaries. Although genesis

Table 1. Environments of deposition identified in rocks of the Middle and Upper Devonian Catskill clastic wedge.

Environment Source	Geographic Area	Rock Unit
Alluvial fan		
Sulenski, 1969	GPO*	Skunnemunk Fm.
Kirby, 1981	GPO	Skunnemunk Fm.
Braided rivers		
Lucier, 1966	SE NY	Kiskatom-Kaaterskill Fms.
Buttner, 1968	SE NY	Genesee Gp.
Friedman, 1972	SE NY	Catskill Fm.
Glaeser, 1974	NE PA	Sawmill Run
Epstein & others, 1974	NE PA	Berry Run-Clarks Ferry Mbrs.
Humphreys & Friedman, 1975	NC PA	Catskill Fm.
Buttner, 1977	SE NY	Genesee Gp.
Sevon & others, 1978	NE PA	Duncannon Mbr.
Rahmanian, 1979	C PA	Duncannon Mbr.
Kirby, 1981	GPO	Skunnemunk Fm.
Meandering rivers		
Allen, 1965	NE PA	Catskill Fm.
Woodrow & Fletcher, 1967	SE NY-NE PA	Catskill Fm.
McCave, 1968	SE NY	Moscow-Ludlowville Fms.
Johnson & Friedman, 1969	SE NY	Tully Fm.
McCave, 1969	SE NY	Catskill Fm.
Sulenski, 1969	GOP	Bellvale Fm.
Friedman, 1972	SE NY	Catskill Fm.
Glaeser, 1974	NE PA	Duncannon Mbr.
Epstein & others, 1974	NE PA	Duncannon Mbr.
Humphreys & Friedman, 1975	NC PA	Catskill Fm.
Buttner, 1977	SE NY	Catskill Fm.
Rahmanian, 1979	C PA	Sherman Creek Mbr.
Dune		
Johnson & Friedman, 1969	SE NY	Tully Fm.
Delta plain		
Glaeser, 1974	NE PA	Walcksville-Long Run Mbrs.
Epstein & others, 1974	NE PA	Walcksville-Long Run Mbrs.
Kirby, 1981	GPO	Bellvale Fm.
Marsh		
Sutton & others, 1970	SC NY	Sonyea Fm.
Humphreys & Friedman, 1975	NC PA	Catskill Fm.
Interdistributary bay		
McCave, 1968	SE PA	Moscow-Ludlowville Fms.
Tidal deposits		
Woodrow & Fletcher, 1967	SE NY-NE PA	Catskill Fm. (PA)and Laurens and Manorkill Fm. (NY)
McCave, 1968	SE NY	Moscow-Ludlowville Fms.
Johnson & Friedman, 1969	SE NY	Tully Fm.
Friedman, 1972	SE NY	Tully Fm.
Humphreys & Friedman, 1975	NC PA	Catskill Fm.
Rahmanian, 1979	C PA	Irish Valley Mbr.

Table 1. (Continued)

Environment Source	Geographic Area	Rock Unit
Distributary mouth bars		
Sulenski, 1969	GPO	Bellvale Fm.
Sutton & others, 1970	SC NY	Sonyea Gp.
Krajewski & Williams, 1971a	NE PA	Catskill Fm.
Epstein & others, 1974	NE PA	Towamensing Mbr.
Kirby, 1981	GPO	Bellvale Fm.
Tidal channels		
Sutton & others, 1970	SC NY	Sonyea Gp.
Kirby, 1981	GPO	Bellvale Fm.
Estuaries		
Sutton & others, 1970	SC NY	Sonyea Gp.
Beach		
Lucier, 1966	SE NY	Kiskatom-Kaaterskill Fms.
Krajewski & Williams, 1971a	NE PA	Catskill Fm.
Delta front		
Glaeser, 1974	NE PA	Towamensing Mbr.
Delta platform		
Sutton & others, 1970	SC NY	Sonyea Gp.
Nearshore shallow marine		
McCave, 1968	SE NY	Moscow-Ludlowville Fms.
Offshore bar		
Johnson & Friedman, 1969	SE NY	Tully Fm.
Krajewski & Williams, 1971a	NE PA	Catskill Fm.
Distal bar		
Kirby, 1981	GPO	Bellvale Fm.
Lagoon		
Lucier, 1966	SE NY	Kiskaton-Kaaterskill Fms.
Johnson & Friedman, 1969	SE NY	Tully Fm.
Friedman, 1972	SE NY	Hamilton Gp.
Prodelta		
Sulenski, 1969	GPO	Bellvale Fm.
Sutton & others, 1970	SC NY	Sonyea Gp.
Glaeser, 1974	NE PA	Trimmers Rock Fm.
Epstein & others, 1974	NE PA	Trimmers Rock Fm.
Slope shales		
Walker, 1972	SC PA	Irish Valley Mbr.
Open shelf		
McCave, 1968	SE NY	Moscow-Ludlowville Fms.
Sutton & others, 1970	SC NY	Sonyea Gp.

*GPO = Green Pond outlier, see Figure 6.

is not considered a part of the definition of a rock-stratigraphic unit, it controls the composition of rock sequences and the uniformity or diversity of their lithologic components. Thus, the orderly progression of progradational lithologies shown in Figure 7 can be easily subdivided into mappable rock-stratigraphic units in part, but not completely.

Those rocks which originated in the prodelta basin and prodelta environments are generally relatively uniform in lithology and comprise good map units although their boundaries are commonly transitional.

Rocks which originated in the subaerial part of the clastic wedge are characterized by a diversity of lithologies created by (1) multiple depocenters, (2) multiple depositional environments, and (3) variable distance from the source area. This diversity complicates rock-stratigraphic subdivision for these rocks. As a result, the approach in Pennsylvania has been to map assemblages of heterogeneous rocks which generally have arbitrary boundaries. These subdivisions may be well-defined in one area (e.g., Carbon County, PA; Epstein and others, 1974), but lack lateral persistence and require redefinition (e.g., Poplar Gap Member, Berg, 1975). In general, useful subdivisions of the Catskill Formation have been erected wherever detailed mapping (scale 1:24,000) has been done and the lateral relationships of these subdivisions has been established. Figure 8 presents some of the Middle and Upper Devonian stratigraphy currently used in Pennsylvania. Sutton (1963) has utilized regionally persistent black and dark-gray shales as stratigraphic marker horizons in New York State, but their presence and utility in Pennsylvania has not been demonstrated.

Economic Products

A variety of economic products derive from the diverse rocks of the Middle and Upper Devonian clastic wedge in Pennsylvania. Gas and oil have been produced in western Pennsylvania for over 120 years and are closely related to sandstones of the Upper Devonian clastic wedge (Kelley, 1967). Large quantities of flagstone have been produced from lower Catskill Formation sandstones in northeastern Pennsylvania and southeastern New York (Glaeser, 1969; Krajewski and Williams, 1971a & b; Sevon, 1978). Some uranium minerals occur at various stratigraphic levels in northeastern and north central Pennsylvania (Sevon and others, 1978), but the economic potential of these occurrences is currently unknown. Many Middle and Upper Devonian siltstones and sandstones have been utilized as sources for good-quality crushed rock and riprap, and Catskill Formation red shales are frequently quarried for use as random fill and base course for low-use secondary roads.

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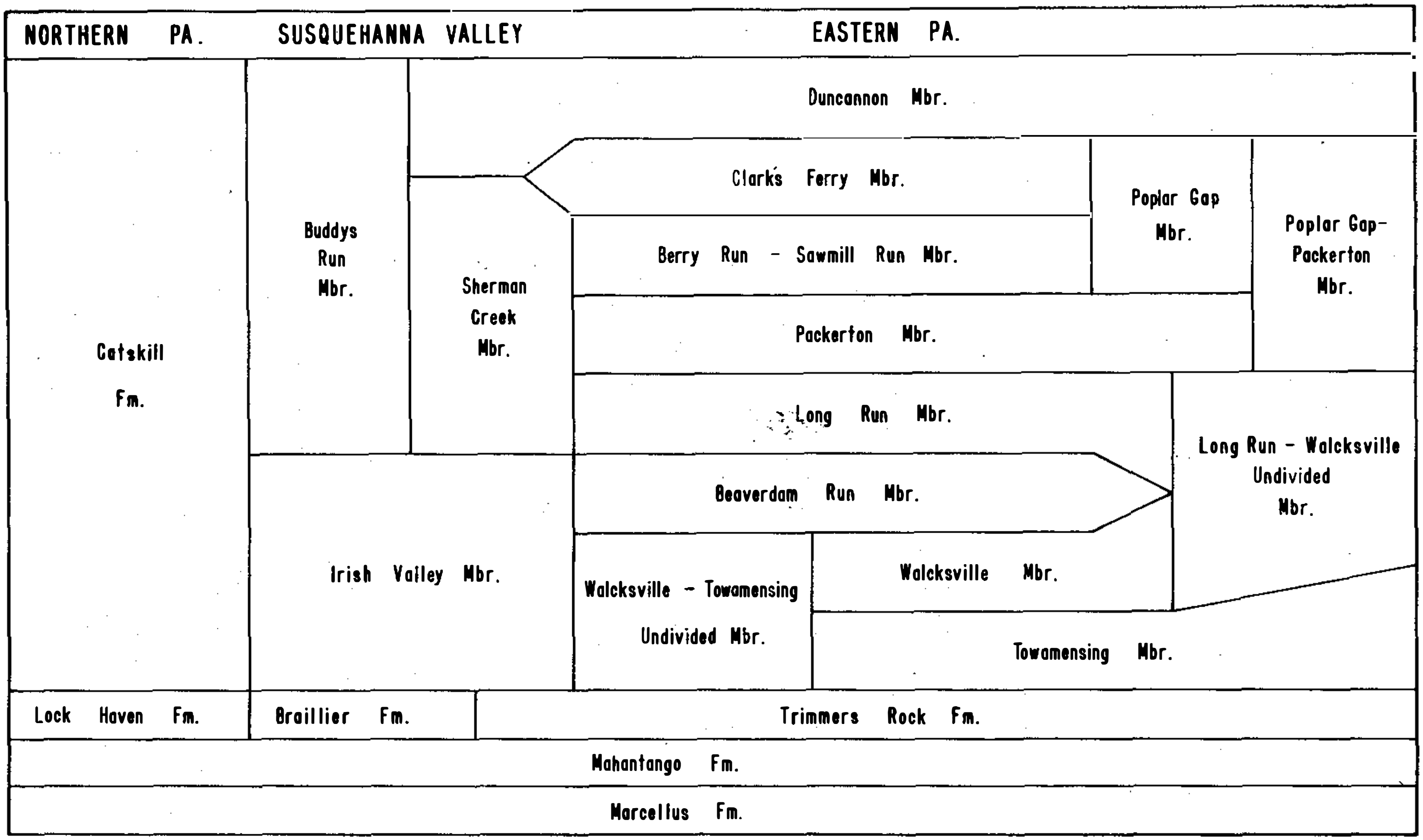

Figure 8. Stratigraphic correlation chart for the Devonian of Pennsylvania (from Berg and others, 1980).

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ROAD LOG AND STOP DESCRIPTIONS FOR CATSKILL CLASTIC WEDGE IN NORTHEASTERN PENNSYLVANIA

Road log starts at junction of Bartle Drive, SUNY Binghamton campus main exit, and New York Route 434

CUMULATIVE MILEAGE	MILES FROM LAST POINT	ROUTE AND STOP DESCRIPTION
0.0	0.0	Turn left onto NY Route 434 west.
0.2	0.2	Turn right onto NY Route 201 to Johnson City.
1.0	0.8	Enter traffic circle.
1.2	0.2	Exit traffic circle on NY Route 201 north.
1.6	0.4	Bear left to NY Route 17.
1.9	0.3	Exit right to NY Route 17 east.
5.1	3.2	Bear right onto Interstate 81 south.
14.1	9.0	Exit right at Exit 1, Kirkwood.
14.3	0.2	Turn right at yield sign to NY Route 7.
15.2	0.9	Turn left at stop light onto NY Route 7 south.

16.9	1.7	Turn right following NY Route 7 south.
18.1	1.2	PA-NY state boundary. Now following PA Route 29 south.
19.0	0.9	Brookdale limits.
19.6	0.6	Road intersects on right.
19.9	0.3	STOP 1. Pull off and park at wide berm area.

STOP 1. PRODELTA ENVIRONMENT

Rocks at this stop are exposed in an outcrop along the west side of the road and along the west side of the stream below road level. Stream elevation is 970 feet.

The rocks exposed here comprise dark-gray shales and siltstones of the Lock Haven Formation. The siltstones have sharp bases, overlie shales, and fine upwards into shale. Burrow traces and fossils of marine invertebrates occur. Lateral continuity of beds is good within the extent of the small outcrop.

This is a good problem outcrop. If examined as an isolated entity, how much information can be gained from the outcrop itself? How much does it help to know the relative position of the outcrop in the vertical sequence of Figure 7? How important is context in the study of (1) rock lithology, (2) depositional environment, (3) stratigraphy, and (4) geologic history?

Leave Stop 1 and proceed south on PA Route 29.

22.7	2.8	Lawsville Center limits.
22.9	0.2	Turn left onto Franklin Hill Rd. (State 075).
24.0	1.1	Stone Crop Rd. (Twp. 735) forks to left, keep right on Franklin Hill Rd.
24.5	0.5	Turn left onto gravel lane. Gray house of G. Perkins to left on rise above road. Proceed up hill through gate.
24.7	0.2	STOP 2. Park on flat at quarry entrance.

STOP 2. MARGINAL-MARINE ENVIRONMENT

The flagstone quarry at this site was opened into rocks of the lower part of the Catskill Formation. The flagstone derives from the planar-bedded sandstones exposed in the floor of the quarry at elevation 1660 feet. The sequence of lithologies exposed here (Fig. 9) is interpreted to represent deposition in an offshore bar environment and the following information about the stop is from Krajewski and Williams (1971b, p. 13-15).

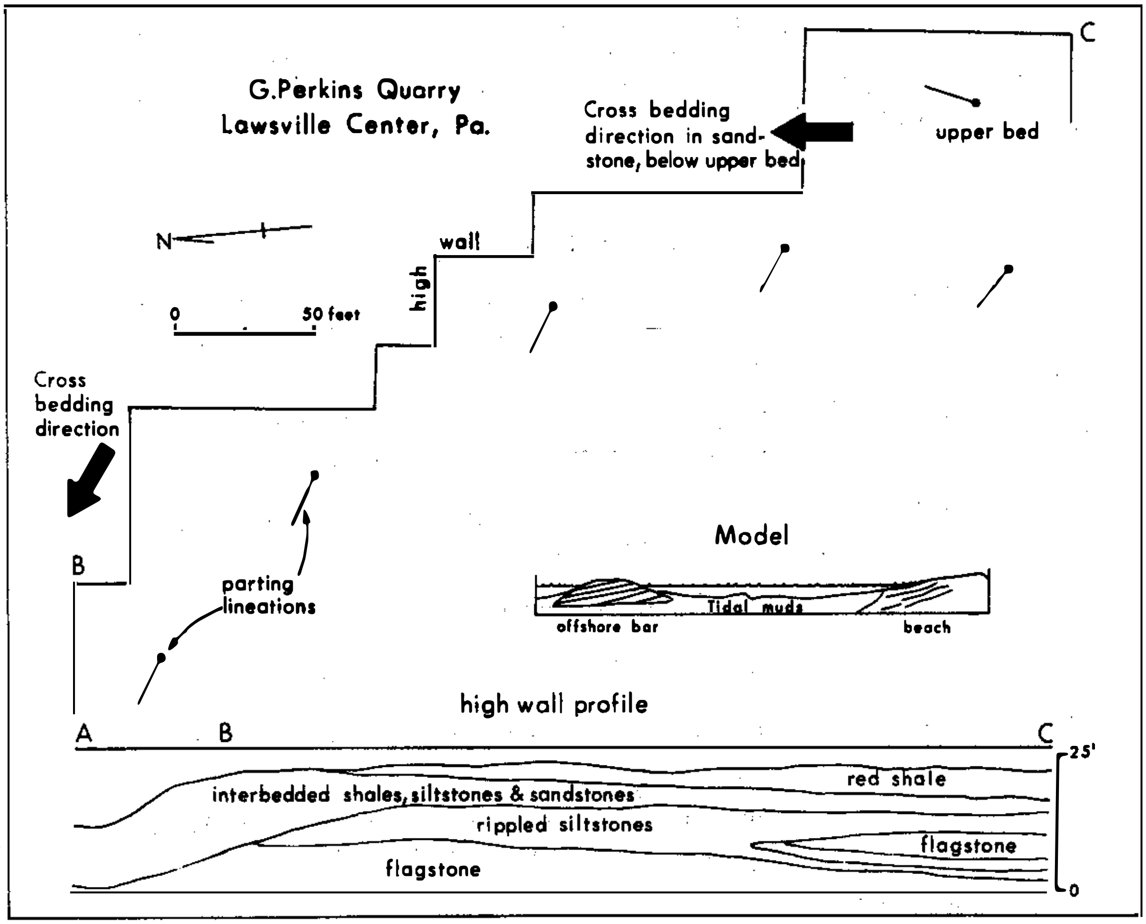

Figure 9. Plan and profile of rocks exposed in flagstone quarry at Stop 2. Rocks are part of lower Catskill Formation. Modified from Krajewski and Williams (1971b, p. 14).

"Characteristics of this type of quarry include:

a) a convex upper surface;
b) parallelism of the parting lineation orientation and the down-dip direction of the cross-bedding;
c) a steep front in the seaward direction;
d) replacement seaward (northwestward) by darker, fossiliferous, marine shales;
e) replacement landward (southeastward) by siltstones, small-scale rippled sandstones, and red shales; and,
f) uniformness and regularity of the joint systems (the high walls of the quarry parallel the joint systems).

"Two offshore bars are visible in the diagram, the larger lower one, and another smaller one in the south end of the quarry. The long direction, or paleogeographic trend, of these bars would be perpendicular to the parting lineation shown in the diagram, or northeast-southwest and east-west for respectively the lower and upper bars. The rock sequence represents a marine transgression onto the tidal flats and shore zone."

Examine this stop with the following questions in mind: (1) How easily can the depositional environment be determined from the physical properties of the rocks? (2) Does the interpretation of this outcrop aid in the interpretation of the rocks at Stop 1? (3) What information can be gained about the source area from this stop?

Return to Franklin Hill Rd.

24.9	0.2	Turn left onto Franklin Hill Rd.
27.1	2.2	Turn left onto paved road to New Milford at cross roads near bottom of hill at Franklin Corners.
31.0	3.9	Turn right at stop sign at Tingley after going under railroad overpass.
31.6	0.6	Turn right onto US Route 11 at stop sign, New Milford limits.
32.2	0.6	Turn left onto PA Route 492 east.
33.2	1.0	(Jackson St.) to Interstate Route 81.
33.7	0.5	Turn right to Interstate Route 81 south.
36.8	3.1	STOP 3. Pull off and park on wide berm at south end of road cut and examine the rocks along the terrace above the road.

STOP 3. FLUVIAL ENVIRONMENT

The rocks at this stop are exposed in the southernmost of three large roadcuts on the west side of Interstate Route 81, all of which expose similar rocks. Elevation here is 1640 feet.

Exposed here are a variety of sandstones, both gray and red, which contain lag gravels, cross beds, planar beds, and ripples (Fig. 10). Interbedded red siltstones are locally burrowed. Lag gravels comprise predominantly calcium-carbonate nodules. Cross-bed forsets are oriented mainly to the southwest while parting lineations are to the northwest. Lateral and vertical relatonships are complex, but the whole can be interpreted in terms of a fining-upward cycle formed by a meandering stream (Allen, 1970).

What features of this outcrop are critical to the interpretation of its depositional environment? How many levels of context are involved in studying this outcrop? What information about the source area can be gained from this outcrop?

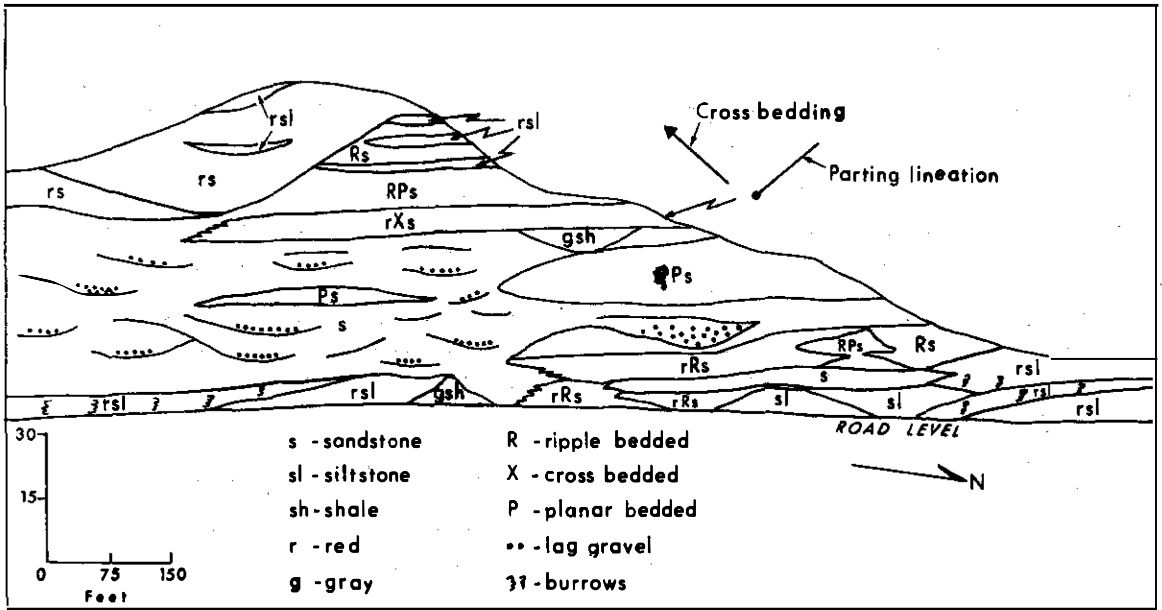

Figure 10. Sketch of rocks exposed in roadcut on west side of Interstate Route 81 south of New Milford (Stop 3). Rocks are part of lower Catskill Formation. Sketch modified from Krajewski and Williams (1971b, p. 67).

Leave Stop 3 and proceed south on Interstate Route 81.

65.4	28.6	Outcrops of Spechty Kopf, Pocono, and Pottsville Formations in road cuts. Entering northwest side of Wyoming-Lackawanna basin.
67.5	2.1	Scarlift reclamation on right.
69.7	2.2	Bear right on Interstate Route 81 south at junction with Interstate Routes 84 and 380.
72.6	2.9	Exit right to Moosic St. and PA Route 307 at Exit 52.
72.7	0.1	Turn right onto PA Route 307 at stop sign.
73.0	0.3	Turn right onto PA Route 307, US Route 11, and Harrison Ave. at stop light.
73.3	0.3	Turn right onto Mulberry St. at stop light.
73.6	0.3	Enter Nay Aug Park.
73.7	0.1	Turn left into parking lot near Brooks Model Coal Mine. Park.

STOP 4. NAY AUG PARK - LUNCH

This stop is principally for the purpose of ingesting food, but some interesting things can be seen.

The Brooks Model Coal Mine area has the entrance to an underground anthracite mine and some old coal cars. The sandstones exposed throughout the park are part of Llewellyn Formation of Pennsylvanian age. This formation contains a number of anthracite coal seams which have been mined, principally by underground methods, in the Wyoming-Lackawanna basin (the northern anthracite field). Exposures along the railroad track along the east side of the park allow evaluation of the depositional environment of the rocks exposed here.

Leave Stop 4 and retrace route proceeding north on Mulberry St.

74.1	0.4	Turn left onto PA Route 307, US Route 11, and Harrison Ave. at stop light.
74.4	0.3	Turn left following PA Route 307 south at stop light.
74.6	0.2	Turn right onto Meadow Ave. at stop light. Follow signs to Interstate Route 81.
74.8	0.2	Turn left at T intersection following signs to Interstate Route 81.
75.0	0.2	Turn left onto Interstate Route 81 north.
77.8	2.8	Bear right to Interstate Routes 84 and 380.
78.7	0.9	Outcrops on both sides of road of basal Pottsville Formation conglomerate. Outcrop to left is upper part of Stop 7.
79.4	0.7	Outcrop to left is Spechty Kopf Formation and site of start of Stop 7. To right across Roaring Branch is dip slope underlain by Spechty Kopf Formation sandstones. Shale exposed below sandstone.
81.5	2.1	Exit left onto Interstate Route 84 east.
82.2	0.7	STOP 5. Park on wide berm to right.

STOP 5. FLUVIAL ENVIRONMENT

The rocks at this stop are exposed in cuts on both sides of Interstate Route 84 east at the point where entrance ramps from Interstate Route 380 east and west join. Park on wide berm on the south side of the road and examine the rocks in the cut on the north side of the road. Cross the interstate with care!

The rocks exposed here are near the uppermost part of the Poplar Gap-Packerton Member of the Catskill Formation (Fig. 8). The sequence (Fig. 11) comprises mainly gray sandstones with some red mudstones. Sandstone and siltstone contacts are sharp. The gray sandstones are cross bedded, contain some lag gravels, and are locally calcareous (calcareous cement weathers brown). Scattered red-shale clasts are common near the bases of some sandstone bed sets. The lag gravels contain some quartz pebbles, but mainly calcium-carbonate nodules, many of which look like oncolites. Root traces occur locally in the sandstones.

The lower red mudstone is burrowed in part and contains abundant calcium-carbonate nodules. Note the thinly laminated character of some of the mudstone enclosing the nodules.

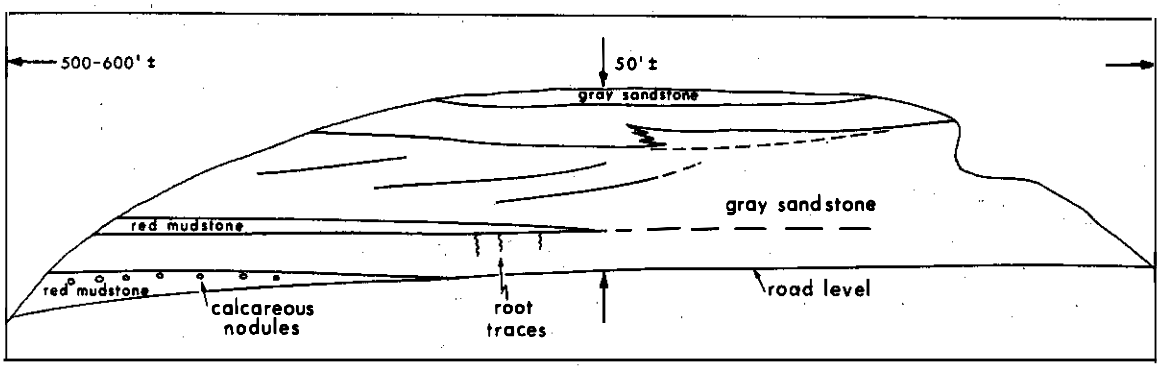

Figure 11. Sketch of rocks exposed in roadcut on north side of Interstate Route 84 just east of junction with Interstate 380 (Stop 5). Rocks are in uppermost part of the Poplar Gap-Packerton Member of the Catskill Formation.

There are many questions to be asked about these rocks. Would information on context be useful in making an environmental interpretation? What information can be gained about the source area? Is source-area information at this outcrop more or less abundant than at Stop 3? What is the significance of the presence of calcium carbonate? How were the calcium carbonate nodules formed in the red mudstones and in the gray sandstones? What is the evidence for a fluvial origin for this sequence? What specific fluvial environment best fits these rocks?

Leave Stop 5 and proceed east on Interstate Route 84.

86.0	3.8	Exit right to Mt. Cobb.
86.2	0.2	Turn left at stop sign.
86.4	0.2	Turn left onto PA Route 348 west at stop sign.
86.7	0.3	STOP 6. Park on wide berm on right.

STOP 6. FLUVIAL ENVIRONMENT

The rocks exposed here are near the lowermost part of the Duncannon Member of the Catskill Formation (Fig. 8). The rocks comprise one complete and two partial fining-upward cycles (Fig. 12). Two cycles have a sharp base with gray sandstone overlying red mudstone. Above the base there is an upward decrease in grain size from sand to clay and change to red color.

The lower sandstone has some calcium-carbonate cement, a few root traces, and bedding changes upward from crossbedding to planar bedding. The overlying red mudstones contain repeated zones with abundant burrows and root traces. The mudstone between these zones is generally thinly laminated. The upper third of the lower major cycle is a smaller cycle which has a grayish red sandstone above a sharp base. Calcium-carbonate nodules are abundant in the mudstone below the sharp base of the upper major cycle. Several large channel cuts occur in the gray sandstone of the upper cycle.

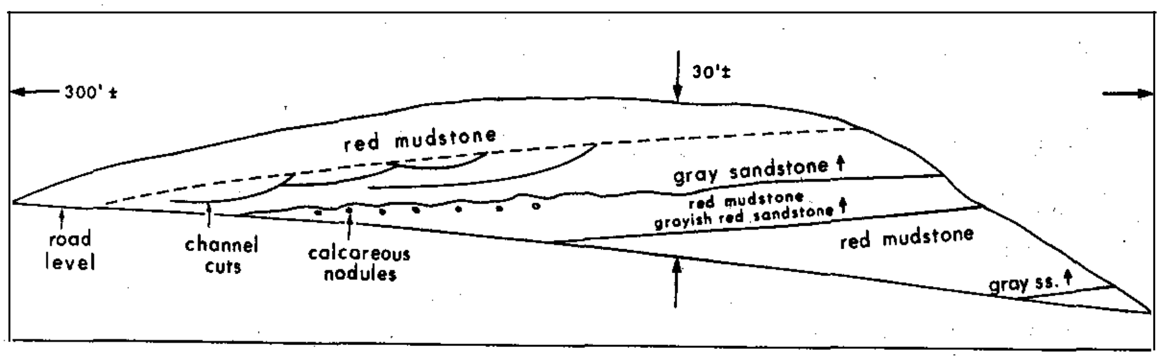

Figure 12. Sketch of rocks exposed in roadcut on north side of PA Route 348 just west of Mt. Cobb (Stop 6). Rocks are in lowermost part of the Duncannon Member of the Catskill Formation.

What is the probable origin of the excellent fining-upward cycles displayed in this exposure? How do these cycles compare with that at Stop 3? What is the significance of the alternating burrowed and non-burrowed zones in the mudstone? How do the calcium-carbonate nodules in the red sandstone compare with those at Stop 5? What differences in depositional environment are suggested by the differences in the rocks exposed here and at Stop 5? Is context of particular value in interpretation of the depositional environment of these rocks? What information about the source area do these rocks provide?

Leave Stop 6 and proceed west on PA Route 348.

89.2	2.5	Bear right to PA Route 435 north.
89.4	0.2	Bear right onto PA Route 435 north at stop sign.
92.2	2.8	STOP 7. Park well off road on berm to right before guard rail starts and just after junction of entrance ramp to Interstate Route 380 west. Bus will proceed 0.8 farther to end of upper outcrop for pickup.

STOP 7. SPECHTY KOPF FORMATION AND LOCAL SUBSIDENCE

This stop occurs at roadcuts on the east side of Interstate Route 380 just south of Dunmore and consists of two parts: a long, high roadcut just north of the entrance ramp (lower outcrop) and a long,low roadcut about 0.7 miles farther north (upper outcrop). The exposures are separated by a covered interval.

The identification of the rocks at this stop is different from that indicated on the 1960 edition of the geologic map of Pennsylvania (Gray and others, 1960), but is in agreement with the 1980 edition of that map (Berg and others, 1980). The changes result from the work of Sevon (1969) and subsequent reconnaissance mapping. The Pocono Formation occurs in the covered interval between the lower and upper exposures.

The lower outcrop (Fig. 13) consists of dark shales and sandstones of the Mississippian-Devonian Spechty Kopf Formation. The shales are here about 300 feet thick and are underlain (in exposures along Roaring Branch) by polymictic diamictite. Interbedded thin siltstone layers exposed in the lower third of the shale in the former Nay Aug quarry across Roaring Branch have bottom-surface flow structures and upper-surface ripples which indicate

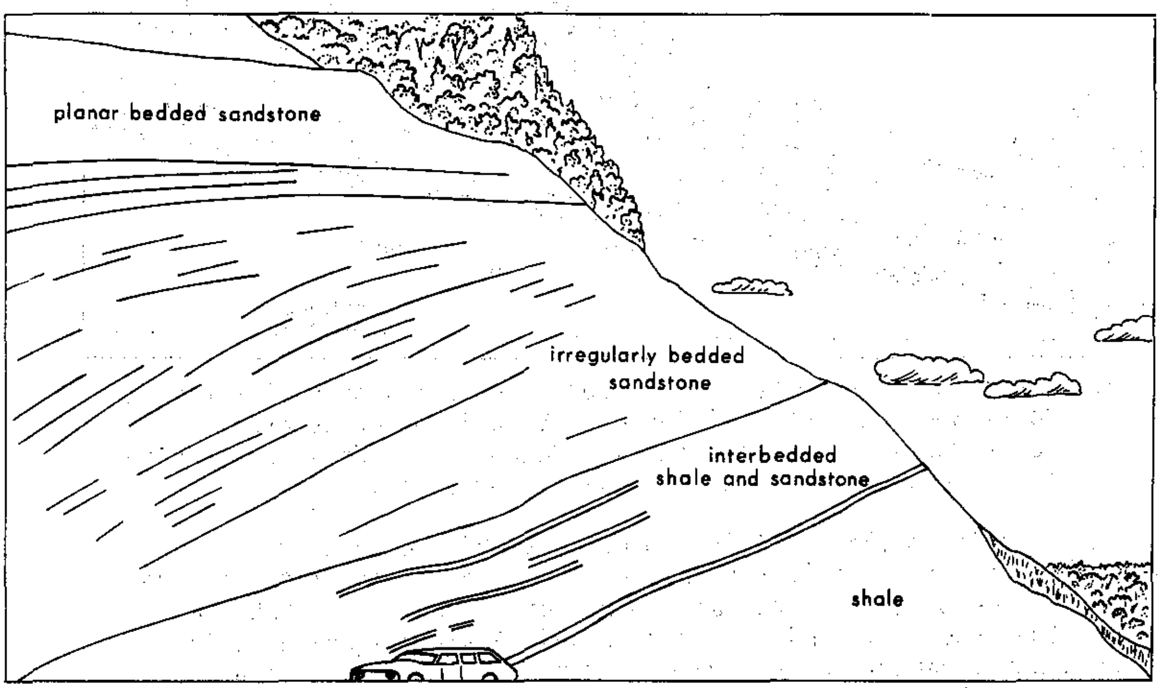

Figure 13. Sketch of rocks exposed in roadcut on east side of Interstate Route 380 just north of entrance ramp from PA Route 435 (Stop 7, lower outcrop). Rocks are part of the Spechty Kopf Formation.

current flow of N12E. Very large and relatively small slump structures occur in the upper third of the shale and are very well exposed here. The contact of the shale and sandstone is gradational through a zone of interbedding. A zone of load casts occurs just below the first sandstone interbed.

The sandstone comprises relatively uniform and moderately well-sorted grain size throughout, is very light gray in color, weathers tan, and is about 160 feet thick. Bedding in the lower part defines wedge-to-irregularly lens-shaped bed sets which dip northwestward at a steeper inclination than the overlying planar beds. The planar bedding is remarkably uniform and persistent. Some beds have ripples, but most are apparently structureless.

What are the environments of deposition of the rocks exposed in this outcrop? The rocks underneath the basal polymictic diamictite are those of the Duncannon Member seen at Stop 6. Does this aid or complicate the interpretation? Also of importance is the fact that 5 miles to the north on the other side of the basin, this sequence is compressed to 110 feet in thickness and is totally absent along all margins of the basin within 20 miles of this locality. The unusual thickness of this sequence and its lateral variation is interpreted to indicate local subsidence at the time of deposition. The regional aspects of these facies (Sevon, 1969; 1979b) are interpreted as part of the demise of the Catskill delta (see earlier text).

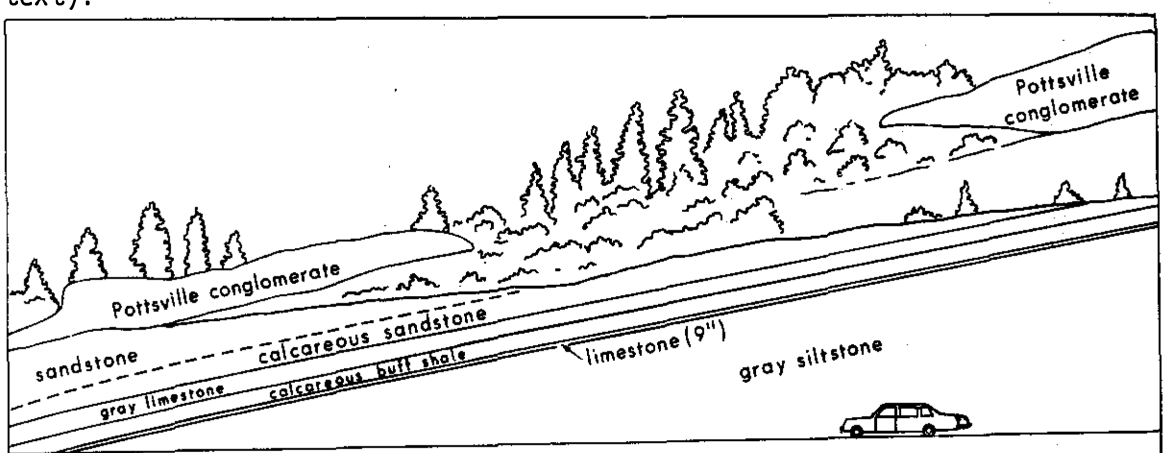

Figure 14. Sketch of rocks exposed in roadcut on east side of Interstate Route 380 just south of Dunmore (Stop 7, upper outcrop). Lower rocks are correlated with Mauch Chunk Formation. White conglomerate is base of Pottsville Formation.

The upper outcrop (Fig. 14) is capped by a white conglomerate of the Pottsville Formation of Pennsylvanian age. The conglomerate is underlain by a sequence of siltstone, sandstones, and limestones which are unlike any other rocks at similar stratigraphic positions anywhere in eastern or central Pennsylvania. The sequence is correlated with the Mississippian Mauch Chunk Formation on the basis of stratigraphic position, but the rocks are very dissimilar from the red siltstones and sandstones which characterize the Mauch Chunk as near as 20 miles to the southwest. However, known facies changes along the southeastern margin of the Wyoming-Lackawanna basin and the abundance of calcium carbonate suggests that such correlation is valid. What environments of deposition may be represented by these lithologies--particularly the thin limestone? Do these rocks, including the Pottsville conglomerate, tell much about the source area?

93.0	0.8	Leave Stop 7 and proceed west on Interstate Route 380.
93.7	0.7	Straight ahead onto Interstate Route 81 north.
98.3	4.6	Pottsville-Spechty Kopf sequence.
139.1	40.8	PA-NY State boundary.
151.9	12.8	Bear left on NY Route 17 west.
155.6	3.7	Turn right onto exit 70 S: Johnson City, NY Route 201 south, and SUNY.
156.3	0.7	Join Riverside Drive.
156.7	0.4	Traffic circle. Exit right immediately.
157.7	1.0	Join NY Route 434 east.
158.0	0.3	Bartle Drive, entrance to SUNY Binghamton. END OF TRIP.

MIDDLE AND UPPER DEVONIAN SHALES AND ADJACENT FACIES OF SOUTH-CENTRAL NEW YORK

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GENERALIZED STRATIGRAPHY OF MIDDLE AND UPPER DEVONIAN SEDIMENTS OF NEW YORK

The Devonian section in New York is the standard reference section for eastern North America. Early workers recognized that major facies changes occur in the Middle and Upper Devonian sequence as a result of deposition of the Catskill delta. Consequently, older units originally mapped as formations or groups in a particular area were recognized as portions of major facies (Catskill, Chemung, Portage) cutting across time lines. Later workers concentrated on finding thin, widespread key beds within the thick deltaic sequence that are unaffected by rapid facies changes characteristic of rocks above and below. These key beds, which are believed to represent time planes, divide the deltaic sediments into several time-stratigraphic units.

The major facies, groups, formations, members, and key beds are illustrated on the Devonian correlation chart of New York (Rickard, 1975), a segment of which is shown in modified form in Figure 1. The field-trip participants, especially those not familiar with the Devonian stratigraphy of New York, will find it very useful to refer to this correlation chart to visualize facies changes, both vertically and horizontally, of units to be observed on this trip. Keep in mind that the vertical scale of the chart represents time, not thickness. Sedimentation rates are highly variable within the clastic units and show a general decrease from east to west within any given unit. The total thickness of Middle and Upper Devonian sediments included from the base of the Hamilton Group to the top of the West Falls Group decreases from over 5,000 feet at the Syracuse meridian to less than 1,000 feet at the eastern shore of Lake Erie. Figure 1 represents a classic example of the law of correlation of facies, which essentially states that within a given sedimentary cycle, the same succession of facies that occurs laterally also occurs in vertical succession.

The oldest time-stratigraphic unit which we will see on this trip, the Hamilton Group, is divided by 3 thin, persistent limestones (Stafford, Centerfield, and Menteth-Portland Point) into 4 formations: the Marcellus, Skaneateles, Ludlowville, and Moscow. Major east-west facies changes occur within the clastic interval above each key limestone bed. To the east the upper part of the black-shale facies (Marcellus and Skaneateles) is replaced by the lower part of the younger gray-shale/siltstone facies (Ludlowville and Moscow).

Thin limestones mapped as key beds in other intervals of the clastic wedge include the Tichenor Limestone (upper Ludlowville), the Tully Limestone, the Lodi and Genundewa Limestones (Genesee Group), and the Parrish Limestone (Sonyea Group). However, for the most part, the thick clastic wedge that developed from east to west during the Late Devonian can be more

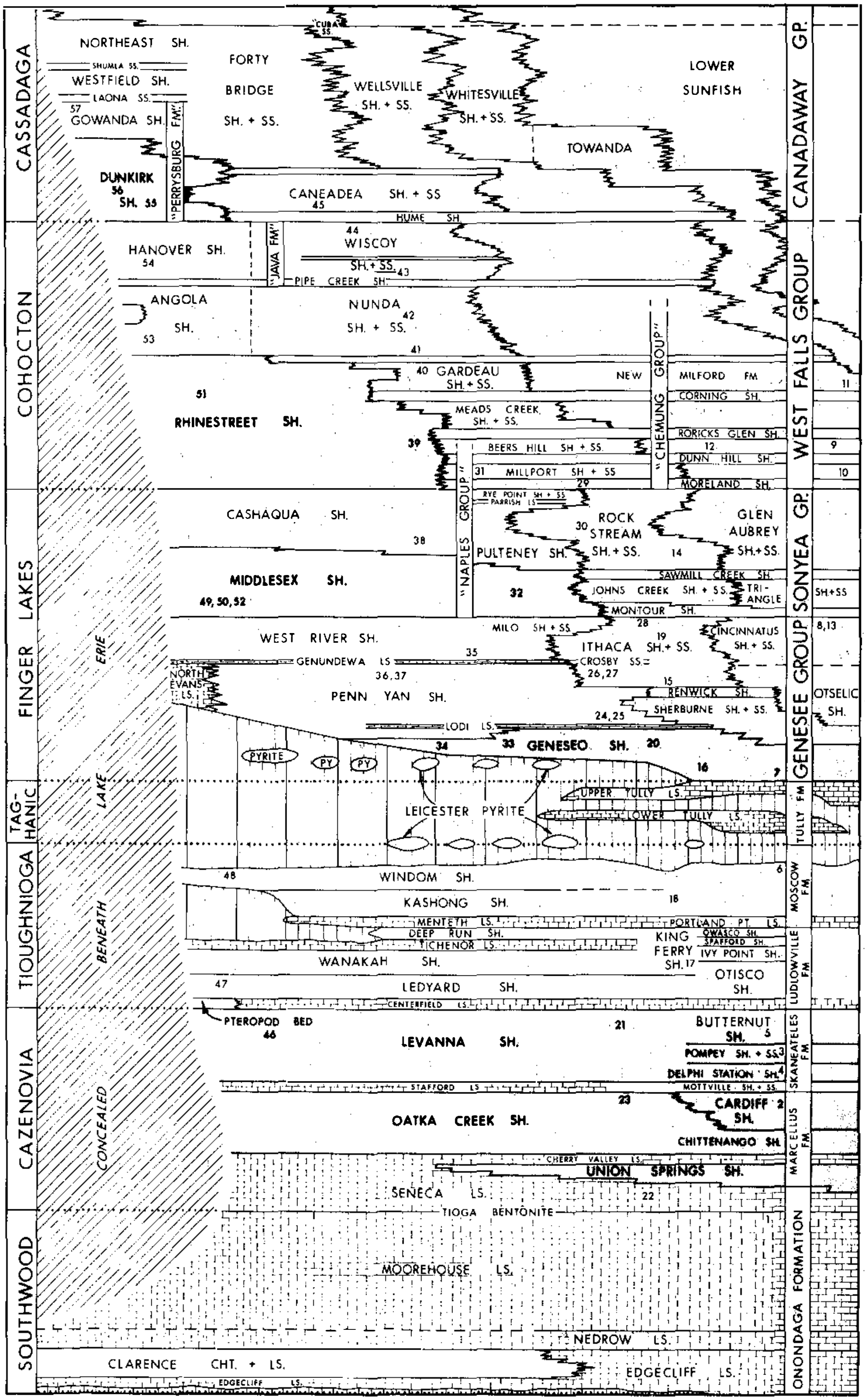

Fig. 1. Middle and Upper Devonian correlation chart for central and western New York (Modified from Rickard, 1975)

effectively divided across facies boundaries by tracing black-shale-tongues, and their eastward dark-gray-shale equivalents, across the state. The resulting stratigraphic framework has a black shale (Geneseo, Middlesex, Rhinestreet, Pipe Creek, Dunkirk) at the base of each thicker unit (Genesee, Sonyea, West Falls, Java, and Perrysburg, respectively). Although differences of opinion as to the rank of these major units (formation versus group) still exist, the concept of tracing black and dark-gray shales eastward into the clastic wedge appears to be accepted by most.

The base of the Genesee Group is defined by the black Geneseo Shale Formation, which extends from the Chenango valley northeast of Binghamton as far west as Erie County. Above the Geneseo Shale the clastics exhibit gross facies changes from east to west typical of most younger units as well. In the east, two units are mapped in the coarser facies: the Sherburne and Ithaca Formations, separated by the Renwick Shale. The Renwick merges westward with the Penn Yan Shale, which is the lower of two shales in the finer, westward facies. The West River Shale is separated from the older Penn Yan by the Genundewa Limestone.

The West River Shale of the Genesee Group is overlain in the west by the black Middlesex Shale of the Sonyea Group. The black-shale facies and a dark-gray eastern equivalent, the Montour Shale, define the base of the Sonyea. A second, younger dark shale, the Sawmill Creek Shale, defines the approximate eastward extent of the upper part of the Middlesex. Above the basal shale units,typical east-west facies changes occur, resulting in the mapping of an upper Rock Stream Siltstone and a lower Pulteney Shale in the east and a greenish-gray Cashaqua Shale in the west (Colton and deWitt, 1958).

The Cashaqua Shale is overlain in the west by the black Rhinestreet Shale of the West Falls Group. The Rhinestreet Shale defines the base of the West Falls from Lake Erie eastward to Seneca Lake. East of Seneca Lake, fingers of dark-gray shale define the eastern Rhinestreet equivalent (Sutton, 1963; Woodrow and Nugent, 1963).

The main purpose of this field trip is to observe the field relationships and subtle lithologic changes within and between the shales and adjacent facies in the "Catskill Delta" sequence and to interpret these characteristics in terms of the depositional environment. By the end of the trip, you should have a better understanding of why detailed mapping and identification of facies relationships remains complex and problematical in spite of exhaustive study by numerous workers for over a century.

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ROAD LOG FOR MIDDLE AND UPPER DEVONIAN SHALES AND ADJACENT FACIES OF SOUTH-CENTRAL NEW YORK

STARTING POINT: Intersection of Rt. 434 (Vestal Parkway) and Bartle Drive (main) exit of SUNY Binghamton campus.

CUMULATIVE MILEAGE	MILES FROM LAST POINT	ROUTE DESCRIPTION
0.0	0.0	Turn left onto Rt. 434 and get in right lane.
0.1	0.1	Enter Rt. 201 on right.
0.7	0.6	Cross bridge over Susquehanna River.
1.0	0.3	Enter traffic circle and remain on Rt. 201 (N).
1.3	0.3	Overpass: 10'10" clearance
1.6	0.3	Bear left toward "Harry L" Drive.
2.0	0.4	Exit for Rt. 17 (W) toward Elmira and remain on Rt. 17 (W).
5.7	3.7	Town of Vestal (road sign)
10.5	4.8	Tioga County (road sign)
19.7	9.2	Leave Rt. 17 at Exit #64 for Owego and Rt. 96.
19.9	0.2	Turn left at end of exit toward Owego and cross bridge.
20.3	0.4	Turn right at yield sign and follow Rt. 96.
20.4	0.1	Turn right, staying on Rt. 96. Cross bridge into Owego and go through traffic light.
20.8	0.4	Turn right at light and then turn left immediately at next light, staying on Rt. 96.
21.1	0.3	Overpass: 12'7" clearance
21.2	0.1	Turn right onto East Avenue at traffic light.
21.3	0.1	Keep left at intersection of East and Prospect Streets, then keep left again for East Beecher Hill Road.
22.0	0.7	Park along road at STOP 1.

STOP 1. EAST BEECHER HILL ROADCUT, OWEGO, N. Y.

(Description partially from Woodrow and Nugent, 1963, and Patchen and Dugolinsky, 1979).

Exposures of Beers Hill and Roricks Glen Members of the Rhinestreet Formation. Elevation at bottom of exposure is 950'.

The Rhinestreet Shale defines the base of the West Falls Group from Lake Erie eastward to Seneca Lake. East of Seneca Lake workers at the University of Rochester under the direction of Robert Sutton have traced fingers of dark-gray shale that define the eastern Rhinestreet equivalent. In the area from Elmira to Binghamton, four dark shales and three intervening lighter colored shales and siltstones (including the Beers Hill Member exposed here) are defined as members of the Rhinestreet Formation (Sutton, 1963; Woodrow and Nugent, 1963).

The Roricks Glen is represented by scattered, very dark-gray shales at the upper end of the exposure. However, many of the darker shale layers have been covered and are difficult to locate. Also, the land above the roadcut is posted and it is therefore recommended that you limit your observations to the exposure along the road. Most of the exposure is in the Beers Hill Member, with flow rolls well developed in the section near the bottom of the hill, just above a landslide scar. Fossils are scarce, but worm burrows and trails (<u>Fucoides graphica</u>) are often found, particularly on the base of the thick, resistant coarser-grained beds. Flute casts, cross bedding and other current indicators are also found.

		Turn vehicle around whenever convenient and begin mileage from last point, heading back down East Beecher Hill Road.
22.8	0.8	Turn right (north) on Rt. 96 at traffic light.
23.0	0.2	Outcrop of Beers Hill Member of Rhinestreet Fm. is on the right.
24.1	1.1	Junction of Rts. 96 and 38. Continue north on Rt. 38. Beers Hill Member is outcrop on right.
31.5	7.4	Newark Valley
32.6	1.1	Outcrop of Sonyea Group extends east.
37.9	5.3	Berkshire
42.0	4.1	Richford
42.2	0.2	Turn left (west) on Rt. 79.
42.4	0.2	Park on right in parking area. <u>With caution</u>, walk across road to small stream and <u>STOP 2.</u>

STOP 2. RICHFORD, N. Y.

The base of the Middlesex Shale Member of the Sonyea Group may be seen exposed in the stream bed across the road from the parking area. The basal few feet of stream exposure is West River Shale, followed by brownish-black shale and siltstone of the Middlesex. Only a few feet are clearly exposed at this location.

Well-developed groove casts and burrows may be found on the base of silty beds. Most of a steep roadcut on the same side of the road as the parking area is covered, but a little digging will expose the Middlesex.

The Middlesex Shale gets progressively finer grained and darker farther west, towards the type section. This facies relationship is generally true for all Upper Devonian shale units.

		Continue west on Rt. 79.
44.5	2.1	Outcrop of Sonyea Group is on right.
47.0	2.5	Tompkins County
47.6	0.6	Caroline (road sign)
50.3	2.7	Slaterville Springs (sign)
52.3	2.0	West Slaterville
54.5	2.2	Turn right (north) on Landon Road and proceed up steep hill.
55.3	0.8	Turn left (west) on Snyder Hill Road at yield sign.
57.2	1.9	Turn right (north) on Quarry Road by old brick house.
57.6	0.4	Park where convenient and walk into quarry on left, STOP 3.

STOP 3. FINGER LAKES STONE CO. QUARRY, QUARRY ROAD, ITHACA, N. Y.

(Descriptions from Chute, 1970 and Dugolinsky, 1972)

The entire quarry exposure is within the Rock Stream Formation (previously named the Enfield Formation) of the Sonyea Group. Layers of sandstone alternate with layers of shale and siltstone. The Rock Stream has been interpreted as representing a shelf deposit (300-500 depth), althouch much shallower depths are likely. Its westward equivalent, the Cashaqua Shale, represents a slope and basin environments in deeper water.

Many of the silty and sandy beds pinch out rapidly. Fossils and sedimentary structures are very common and well developed. Sole marks may be best seen on the base of silty layers. Worm burrows are common. Excellent sole marks, including flute and groove casts, are usually most accessible on quarried slabs.

The quarry is presently operated by the Finger Lakes Stone Company. Slabs of sandstone are obtained without blasting by drilling holes on seams and separating them with wedges. The slabs are cut to the desired sizes in the mill by diamond and wire saws. Pieces with good joint surfaces are used for special surface effects. The stone is laid in various patterns with different finishes, as illustrated by the exterior of the company's office at the quarry.

		Return to vehicle and continue north on Quarry Road.
57.8	0.2	Turn left (west) on Ellis Hollow Road.
60.2	2.4	City of Ithaca. Ellis Hollow Road becomes Mitchell Street.
60.6	0.4	Merge right onto Rt. 336 at third stop sign.
60.9	0.3	Bear right onto Rt. 79 (end of Rt. 336). Follow Rt. 79 (East State St.) through city.
61.3	0.4	Bear right, staying on Rt. 79. Cornell University is on right.
62.1	0.8	Junction Rts. 13 and 34. Turn right (north). Continue through three traffic lights to a 4-lane expressway.
63.7	1.6	Exit for Rt. 34 and Stewart Park. Leave expressway and turn left at stop sign. Drive under Rt. 13. Continue north on Rt. 34. Cayuga Lake is on left.
69.1	5.4	South Lansing
69.4	0.3	Turn left on Rt. 348 toward King Ferry.
70.1	0.7	Turn left on Portland Point Road (at Cargill Salt sign).
71.6	1.5	Park near abandoned cement plant and walk north along RR tracks to STOP 4-A.

STOP 4-A. RAILROAD CUT BETWEEN SALT COMPANY AND ABANDONED CEMENT PLANT, NORTH OF ITHACA

Exposure of the King Ferry Shale of the Ludlowville Formation.

About 25 feet of the King Ferry Shale is exposed here, with a noticeable dip to the south. Many concretions may be seen in the siltstone beds, which are lighter colored than the shales. The darker shales are commonly cross-laminated. A coquina (fossil-rich layer) is evident at the base and top of the siltstone unit.

Much of the unusual weathering characteristics of this exposure is thought to be due to salt contamination from passing railroad cars coming from the salt plant just to the north of the exposure.

Return to the road and walk a short distance into the gorge to the east. The bus can turn around in a wide area across the culvert and wait at the mouth of Gulf Stream, STOP 4-B.

STOP 4-B. GULF STREAM (SHURGER GLEN)

Exposured are the upper King Ferry Shale, the Portland Point Limestone, Moscow Shale and Tully Limestone.

The upper King Ferry Shale of the Ludlowville Formation may be seen in the exposures in and below the lower falls in Gulf Stream. The overlying Portland Point Limestone, near the lip of the lower falls, is very fossiliferous. Note the peculiar curved fracture pattern in the shales just below the falls.

The lower falls may be scaled on the left side (facing upstream) with care, where the approximately 5 feet of Portland Point Limestone may be examined. The Portland Point is coarsely crystalline and weathers brown. Slickenslides may be observed in the lower 8 inches of the member.

Float at the base of the lower falls includes samples of Tully Limestone, which is finer-grained and less fossiliferous than the Portland Point limestone.

The Moscow shales overlie the Portland Point and are roughly 100 feet thick in this area. They are well exposed upstream from the lower falls. A nearly complete exposure of the entire 100 feet may be seen about 5 minutes upstream. The overlying Tully Limestone is visible at the top of the cliff exposure and forms the lip of the upper falls which is about another 20-minute walk farther upstream. Dark shales (Geneseo) may be seen in the float below the upper falls and are probably exposed above the upper falls. There is excellent joint control in the stream at and immediately below the upper falls.

The gray silty shale and siltstones of the Ludlowville and Moscow demonstrate a high diversity fauna and represents sediments deposited in a subtidal shelf-delta platform (Grasso, 1978).

Return to vehicle and proceed back up hill.

73.1	1.5	Turn right (south) on Rt. 34B.
73.8	0.7	Turn right onto Rt. 34 at Rogue's Harbor Inn.
78.4	4.6	Ithaca
79.4	1.0	Junction Rts. 13, 34, and 79. Turn right onto Rts. 13 and 34, before bridge, and enter expressway.
81.3	1.9	Junction Rts. 13, 34, 79, 89, and 96. Continue straight ahead.
83.1	1.8	Turn left into Buttermilk Falls State Park, STOP 5.

STOP 5. BUTTERMILK FALLS STATE PARK

A recent study (deWitt and Colton, 1978) has revealed that 14 feet of Renwick Shale are exposed at the base of the lower falls, overlain by 478 feet of Ithaca shales and sanstones, which in turn are overlain by 135 feet of West River Shale. The Renwick Shale is easily accessible here, particularly for large groups. Note the darker color of the shale with the interbedded lenticular silty beds exposed. The precise contact of the Renwick with the overlying Ithaca is not easy to pick out.

The Renwick here separates coarser-grained clastics: the underlying Sherburne and the overlying Ithaca Formation. Westward, the Renwick merges with the Penn Yan Shale, which is the lower of two shales in the finer western facies. A travertine deposit may be seen on the vertical exposures at the left of the main swimming pool (looking upstream). It is assumed that these deposits are very recent (post-glacial) and result from the deposition of materials in groundwater.

		Return to vehicle and exit park. Turn right onto Rts. 13 and 34.
84.9	1.8	Jct. Rts. 13, 34, 79, 89, and 96. Turn left (west) onto Rts. 79, 89, and 96.
85.4	0.5	Turn hard right onto Rt. 89 north.
86.8	1.4	Exposure of Ithaca Formation is on left.
89.9	3.1	Town of Ulysses
95.7	5.8	Park on left (before bridge) in Taughannock Falls State Park. Walk into gorge, using trail on the left bank. STOP 6.

STOP 6. TAUGHANNOCK FALLS STATE PARK

(Descriptions from Cornell University Department of Geology, 1959, and from deWitt and Colton, 1978).

Note: At 215 feet, Taughannock Falls is the highest falls in North America east of the Mississippi River.

The main falls is at the head of a deep post-glacial gorge one mile long, with walls from 200 to nearly 400 feet high. The falls is determined partly by the superior resistance of the Sherburne sandstones and partly by the jointing pattern within the rock units. The upper 90 feet of the Geneseo Shale is exposed in the gorge walls from the level of the plunge pool to a point marked by a change in lithology and color. Much of the Geneseo is covered with talus in places. The entire Penn Yan (about 80 feet thick at this locality), the overlying Sherburne (also about 80 feet thick) and the Renwick Shale (about 35 feet thick) are exposed here, along with the lower portion of the Ithaca. At the mouth of the gorge, near the lake, is the lower falls (about 15 feet high), over the resistant Tully Limestone, (also about 15 feet thick here), between the weak basal Geneseo shales above and the shales of the Moscow Formation below.

Both the Sherburne and the Renwick facies are lost only about 15 miles west of this location, where they merge into the Penn Yan Shale.

Note particularly the interfingering nature of the contact between the Tully and the overlying Geneseo, easily seen in the stream bed above the lower falls. This indicates that although there may be an unconformity at the top of the Tully Limestone farther west, there is none here.

Joint control is well developed, particularly within the Geneseo Shale, and is best seen in the cliffs below the main falls and in the stream bed just below the main falls. Alonite dikes, each less than one inch thick, are present in the stream bed beneath the main falls.

We are now on the northern flank of the Fir Tree Anticline and a slight northern dip may be seen easily if you look at the water flowing over the Tully Limestone, following a bedding plane, just above the lower falls where the stream bed is wide. En echelon fractures are well developed in the Tully Limestone exposed in the stream bed.

This gorge was used in pre-World War II days for sets in "western" movies and some of the wooden shacks precariously perched on the face of some of the cliffs were used to house the cameras during filming.

		Return to vehicle and continue north on Rt. 89.
96.9	1.2	Seneca County line
97.6	0.7	Frontenac Road and gorge to right

108.7	11.1	Turn left (west) on county road 138, just past shale quarry on left.
110.6	1.9	Jct. with Rt. 96. Turn right on Rt. 96 at stop sign.
112.4	1.8	Ovid
112.5	0.1	Turn right (north) at flashing traffic light, staying on Rt. 96.
115.0	2.5	Jct. with Rt. 414. Stay right on Rt. 414.
121.3	6.3	Fayette
121.9	0.6	Turn left on Poorman Road by Rapini's Fayette Motel.
122.2	0.3	Park by gate on left and walk into old quarry, STOP 7.

STOP 7. FAYETTE TOWN DUMP

(Description partly taken from Cornell University Department of Geology, 1959).

This quarry was originally operated by the Town of Fayette for road material and is now used as a dump. The upper 40 feet of the Levanna Shale and lower 20 feet of the Centerfield Limestone are exposed here. The Levanna is the upper member of the Skaneateles Formation whereas the overlying Centerfield Limestone is the basal unit of the Ludlowville Formation of the Hamilton Group.

We have observed the upper of two general facies within the Hamilton Group at Stops 4-A and 4-B: The Ludlowville and Moscow Formations. Here, we can see a different gross facies represented by the dark, fissile Levanna Shale Member of the Skaneateles Formation, which represents an anerobic, distal basin (Grasso, 1978). The overlying Centerfield Limestone separates this facies from the lighter colored, silty Ludlowville.

This is a good fossil-collecting site, particularly near the base of the Centerfield Limestone. Most common fossils are brachiopods, pelecypods, gastropods, plant fragment and fish remains. Mineral-filled fractures are common and asphaltic-filled fractures are occasionally found. Pyrite is commonly seen, particularly in the dark Levanna Shale. Large septarian concretions are seen in several layers within the Levanna Shale.

Return to vehicle. Turn around and head back toward village of Fayette.

Clarence (Ozol, 1963; Oliver, 1966), is characterized by an anomalous abundance of chert. This member is not recognized east of Avon, New York and will receive no further attention in this paper.

Unless otherwise noted, the descriptions contained in this section are derived from Oliver's (1954, 1956a) work on the type localities of Onondaga members in the trip area. Representative thicknesses and spatial relationships of the members as seen in outcrop between Buffalo and the Helderbergs are schematically depicted in Figure 1.

Edgecliff Member

At its type locality, Split Rock (stop 2), the Edgecliff Member is a light-gray, very coarse-grained limestone. Beds range in thickness from 15 cm to in excess of 1.5 m. A basal section, 1 cm to 1.8 m thick, ranges from a quartz arenite to a sandy limestone. Quartz abundance rapidly decreases upward. Light-colored chert nodules are found throughout the member. The Edgecliff is characterized by an abundant fauna of rugose corals, tabulates, and large crinoid columnals. This fauna forms a coral biostrome throughout much of the state (Oliver, 1956b). Coral bioherms of the formation are restricted to this member.

Nedrow Member

In central New York, the Edgecliff Member is overlain by a 3-4.3 m unit of thin-bedded, very fine-grained argillaceous limestone, referred to as the Nedrow Member. The type section of this unit is Indian Reservation quarry (stop 1), just south of Nedrow, New York. Though chert is uncommon in the Nedrow, the upper beds locally contain some scattered medium- to dark-gray nodules.

The Nedrow may be recognized by its typical recessed weathering caused by numerous shaly beds within the matrix. In outcrop this feature makes the member easy to recognize; however, in a fresh cut (such as found in the Jamesville quarry, stop 3) it becomes difficult to differentiate between the Nedrow and Edgecliff strictly on lithologic criteria. For this reason, recognition of a Nedrow fauna is critical. The lower Nedrow bears a brachiopod-dominated fauna with several species of platyceratid gastropods and very few corals. Two characteristic coral species are *Amplexiphyllum hamiltonae* and a turbinate growth form a *Heliophyllum halli*. The upper Nedrow has a less diverse brachiopod fauna with only a few platyceratids.

Moorehouse Member

At its type locality in the Jamesville quarry (stop 3) the Moorehouse Member is a medium-gray, very fine-grained limestone with numerous shaly partings. This unit gradationally overlies the Nedrow Member, making their contact difficult to place. However, in the central New York region the top of the Nedrow is considered to be the uppermost shaly bed, usually separated by about 15 to 20 cm from the underlying shaly bed. Dark-gray chert is common throughout the Moorehouse. Chert increases in abundance in the upper half of the member, where it commonly forms beds or anastomosing networks. The upper half of the Moorehouse is also less shaly and more fossiliferous than the lower half. The entire member in central New York is strongly dominated by brachiopods; nowhere are corals abundant.

Seneca Member

The type section of the Seneca Member is at Union Springs, New York. There the basal part of the member is a fine- to medium-grained limestone, which overlies a greenish-gray to ochre-colored clay layer 15 cm thick. This clay layer, the Tioga Bentonite, separates the base of the Seneca from the lithologically similar upper Moorehouse Member throughout much of the state. Approximately 3 m above the bentonite layer occurs a zone of "Chonetes" aff. lineata (= Zone J of Oliver, 1954). The Seneca is a "muddy" limestone, highly argillaceous, and poorly fossiliferous except for the "Chonetes" Zone. The Seneca grades upwards into the Marcellus Shale. The gradation is represented by a 2+ m section of increasing shale content within the limestone and by alternating beds of shale and lime.

Formational Contacts

Over its exposure area, the Onondaga overlies several older formations which generally increase in age westwards. In eastern New York, the Onondaga conformably and gradationally overlies the Schoharie Formation. Between Cobleskill and Richfield Springs the Onondaga overlies the Carlisle Center Formation. There the contact is marked by phosphorite nodules and glauconite grains, and represents a minor unconformity. Within the field-trip area, the Onondaga unconformably overlies one or the other of the Lower Devonian Oriskany, Coeymans, or Manlius Formations. Further west erosional remnants of the Lower Devonian Bois Blanc and the Silurian Akron Formations underlie the Onondaga.

The Onondaga Limestone is overlain by the Marcellus Shale. West of Cherry Valley the Marcellus Formation rests on the Seneca Member of the Onondaga Formation. In central New York, the contact is both interbedded and gradational. The limestone-shale contact is not exposed in western New York. However, it appears to be more abrupt than in the type area. East of Cherry Valley and north of Catskill, the Marcellus Formation rests on the Moorehouse Member of the Onondaga Formation. Their contact is abrupt, marking a minor unconformity which is either erosional (Chadwick, 1927) or nondepositional (Cooper, 1930; Flower, 1936). Oliver (1956a) tended to support the latter conclusion. Subsurface data indicates that the Seneca is the uppermost Onondaga member south of Catskill (L. V. Rickard, pers. comm., 1980).

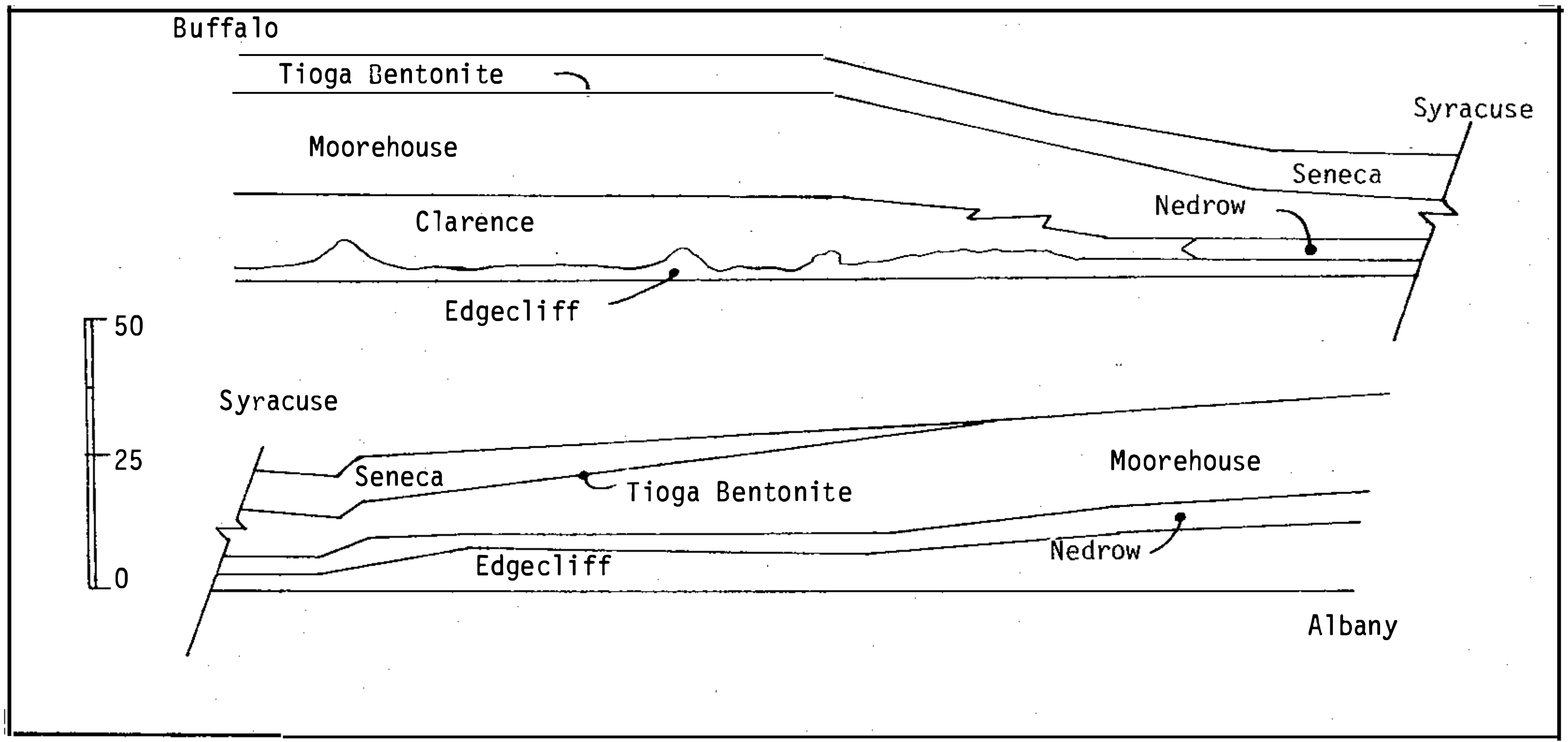


Figure 1. - Generalized cross section of the Onondaga Formation members as seen in outcrop between Buffalo and the Helderbergs, showing thicknesses and physical relationships. Adapted from Oliver (1954, 1956a). The scale is in meters.

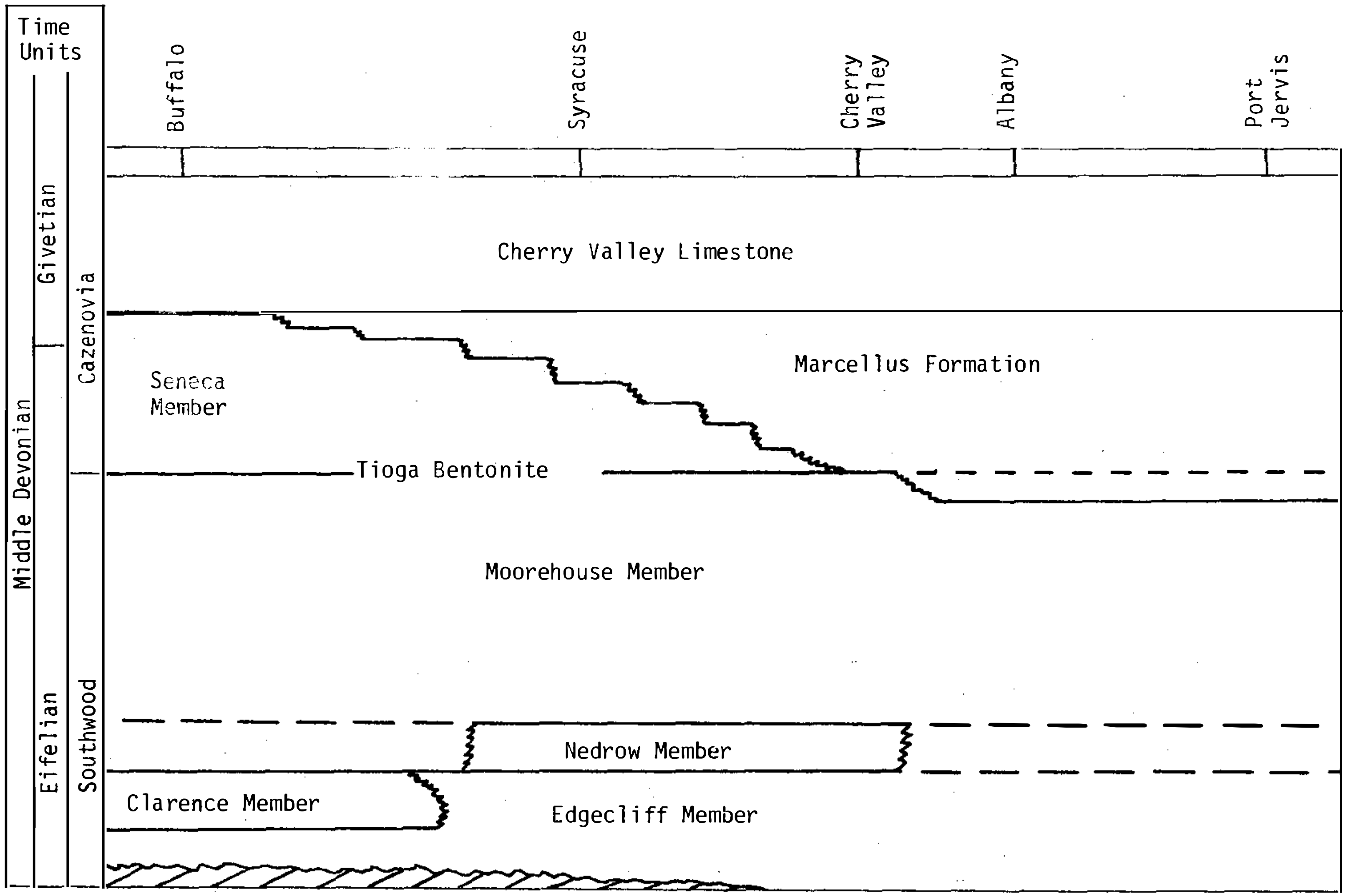


Figure 2. - Chronostratigraphic correlation of the Onondaga Formation throughout New York State. Horizontal dimension not to scale. Adapted from Oliver and others (1969) and Rickard (1975).

Age and Correlation

The Onondaga Limestone was deposited during early Middle Devonian time. Age determinations based on coral (Oliver, 1954, 1956a), cephalopod (Oliver, 1956c), and conodont (Klapper and others, 1971) faunas indicate that the Edgecliff, Nedrow, and Moorehouse members are of Southwood age (Rickard, 1975). The Seneca Member was deposited during Cazenovia time (Rickard, 1975). The formation is entirely correlative to the Eifelian of Europe. Intraformational chronostratigraphic relations are presented in Figure 2.

The basal Edgecliff Member marks the earliest record of the Middle Devonian throughout the state (Oliver, 1954, 1967). The Edgecliff is conformable with the Schoharie Formation in eastern New York and unconformable with the Bois Blanc Formation in the western area. Oliver (1967) concluded that the Bois Blanc and Schoharie Formations are age-equivalent, and that the base of the Onondaga is nearly, but not perfectly, isochronous.

The Tioga Bentonite, which is found at the Moorehouse-Seneca contact, delineates the end of Southwood (Onesquethaw) time (Oliver and others, 1967), placing the Seneca Member in the Cazenovian stage. Proceeding from west to east in outcrop, the Tioga and the top of the Seneca converge and become coincident just east of Cherry Valley. In eastern New York, outcrops north of Catskill, the Tioga presumably lies within the Marcellus Shale. In the subsurface of southeastern New York, the Tioga lies between the Moorehouse and Seneca members as it does in central New York outcrops (L. V. Rickard, pers. comm., 1980). Rickard has also found that in western New York the Tioga occupies three separate horizons, one of which is found at the Moorehouse-Seneca contact. Thus, the exact chronostratigraphic character of the uppermost Onondaga is not as straightforward as was previously thought. However, as seen in outcrop in the trip area, the top of the Onondaga is considered to be of early Cazenovia age (Rickard, 1975).

METHODS OF COMMUNITY ANALYSIS AND DESCRIPTIONS OF ONONDAGA PALEOCOMMUNITIES

In independent studies of Onondaga paleosynecology the authors generally applied Fager's (1963, p. 45) definition of paleocommunities as "recurrent organized systems of organisms with similar structures in terms of species presence and abundances". However, Feldman (1978, 1980) concentrated on the distributions of brachiopod genera while Lindemann (1980) stressed coral distributions and dealt with the remaining fauna bionomically or at high taxonomic levels. While these divergent approaches partly resulted from differences in research objectives and observational scales, both attempted to recognize and reconstruct the fossil equivalents of once-living communities of organisms. For this reason we feel that it is instructive to separately describe our analytic methods and the resultant communities.

Communities Recognized by Feldman

More than 50 localities within the outcrop belt of the Onondaga Limestone were studied, but only 30 were sampled in detail. Silicified outcrops were encountered mainly in southeastern New York, specifically in the mid-Hudson valley. Silicification was poor to nonexistent in the central region. The most productive outcrops were those which consisted of a large expanse of limestone in which single bedding planes were exposed laterally. In many cases, it was possible to identify the specimens in the field. However, if preparation of a specimen was necessary or if a particular specimen was needed for comparison and further study, it was often possible to crack out small slabs of limestone. This is especially true of the Nedrow Member. Collecting in the Jamesville quarry (stop 3) was relatively easy in this respect due to blasting and subsequent jointing. However, the fresh unweathered surfaces provided little in the way of good fossil material.

Nine brachiopod communities have been recognized in the Onondaga Limestone from Syracuse to southeastern New York (see Feldman, 1980, p. 31). Four of these, briefly discussed below, are found in the Syracuse area. The vertical distribution of these communities in the trip area is presented in Figure 3.

1) Leptaena-Megakozlowskiella community. This community is found in Nedrow-Moorehouse age rocks at the Onondaga Indian Reservation quarry (stop 1) in Nedrow, New York and the Jamesville quarry (stop 3). A mid-neritic environment of deposition is probable along with a moderately to highly argillaceous substratum. Leptaena and Megakozlowskiella are the dominant brachiopod genera (29.9 and 28.4 percent, respectively) of a total of 17 genera. Other taxa present include tabulate and rugose corals (common), gastropods (very rare), a cephalopod (Foordites, very rare), trilobites (rare), camerate crinoid columnals (rare), and bryozoan fragments (rare). The trophic nucleus of this community is one of low-level suspension feeders (Leptaena-Megakozlowskiella). Noticeably absent are the numerous platyceratid gastropods of the mid-Hudson valley.

Locality description: Onondaga Indian Reservation, Nedrow, New York (stop 1), southwest of the junction of Route 11 and Highway I-81. Almost the entire Onondaga Limestone is exposed at this locality. This is the type section of the Nedrow Member (Oliver, 1954). The Edgecliff Member is best observed on the quarry floow at the southwest end of the quarry. Large crinoid columnals and stems are visible. The Nedrow-Edgecliff contact is best seen at the south face of the quarry wall where interbedded shaly beds mark the Nedrow. The Moorehouse-Nedrow contact is at the topmost shaly bed. The Seneca-Moorehouse contact is represented by the Tioga Bentonite which forms a re-entrant on the face of the quarry wall. The top of the Seneca is eroded.

2) Pacificocoelia community. I have found evidence of this community at only one outcrop within the Nedrow Member. The limestone

Figure 3. - Biostratigraphy of the Onondaga Limestone in central New York showing distribution of Feldman's brachiopod communities

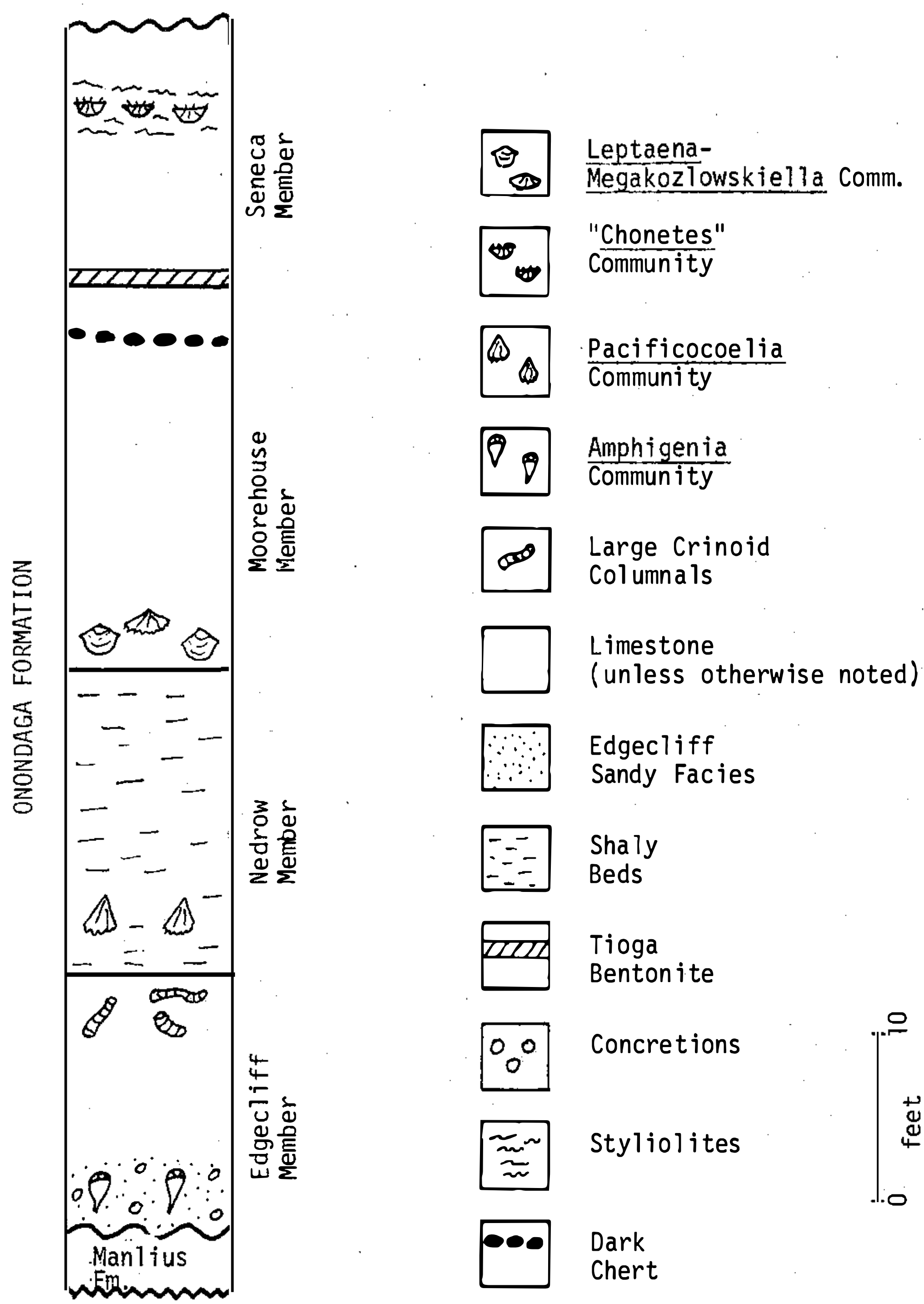
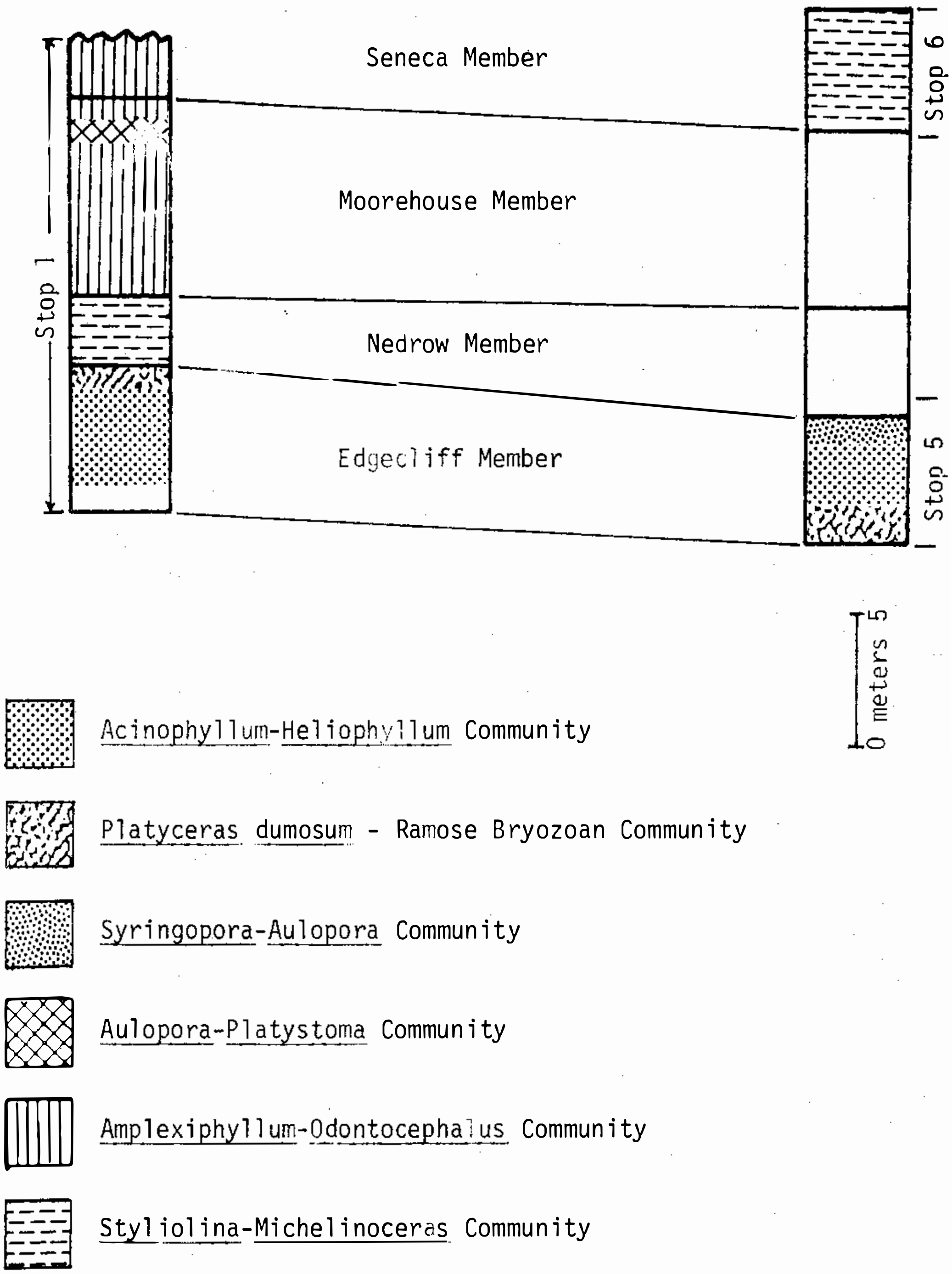


Figure 4. - Biostratigraphy of the Onondaga Limestone in central New York showing the distribution of Lindemann's communities.



is quite muddy and the brachiopods are packed on bedding planes which must be split with a chisel in order to extract them in quantity. A near-shore (inner-neritic) environment is postulated by virtue of the fact that the faunal composition of Johnson's (1974) acrospiriferid-leptocoelid biofacies (representing shallow water with abundant Acrospirifer, Howellella, and Pacificocoelia or Leptocoelina, with no Gypidula, Atrypa, and Schizophoria) is so similar to this Pacificocoelia community. Other brachiopod genera represented here are: Leptaena, Megakozlowskiella, orthotetacids, Dalejina, Pentamerella, Athyris, Megastrophia and Rhipodomella?. Pacificocoelia comprises 59.1 percent of brachiopod genera. Other taxa present include rare trilobite fragments.

Locality description: Most of the Nedrow Member is exposed along the southbound lanes of Highway I-81 about 0.5 miles north of exit 16, Nedrow, New York. The top of the Nedrow Member is missing. Also exposed is the Edgecliff Member including the basal sandy facies. The lower part of the Nedrow at this locality consists of about 1 m of medium-grained shaly limestone with alternating light and dark bands of gray to tan beds. Fossils are not common. The upper section consists of 2.1 m of medium- to fine-grained shaly limestone, tan to gray in color, medium- to thin-bedded. Chert nodules are commonly found within the bedding and locally weather out.

3) Amphigenia? community. This community is found in the basal sandy facies of the Edgecliff Member. This unit, similar to the sandy base of the Bois Blanc Formation, may represent one of three conditions: 1) reworked Oriskany Sandstone, 2) reworked "Springvale" Sandstone, or 3) unnamed sand of combined Oriskany and "Springvale" ages. The environment of depositon is considered inner-neritic. The only brachiopods found in this unit are specimens of Amphigenia? along with tabulate (Favosites?) and rugose (Acinophyllum, Heliophyllum, "Heterophrentis") corals. Preservation is fragmental and very poor.

Locality description: The Edgecliff Member, including the basal sandy facies, is exposed along Highway I-81, about 1.5 miles north of exit 16, along the southbound lanes, on the east side of the road. The lowermost unit consists of 30 cm of limy, coarse-grained friable sandstone, locally jointed. Nodules are locally present within the matrix. Above the basal unit is 1.2 m of sandy, brownish, coarse-grained, massive limestone gradational with the overlying unit. This unit is 3.2 m thick and consists of medium- to coarse-grained light-gray crystalline biostromal limestone with large amounts of crinoidal debris.

4) "Chonetes" community. This community is found only in the Seneca Member about 3 m above the Tioga Bentonite. The limestone contains a large amount of silt and is representative of outer-neritic conditions. The substratum was most likely a soft lime mud. "Chonetes" is the dominant brachiopod (99 percent) with a few specimens of Megakozlowskiella, Leptaena, Megastrophia, indet. orthotetacids, Athyris, Atrypa, and Pentamerella. Other taxa present include rugose corals, euomphalacean gastropods, trilobites, and crinoid columnals.

Locality description: This section is located in the Jamesville Quarry #3 pit. The quarry is actively mined and, therefore, there are few weathered surfaces. Collecting is often difficult. The Seneca here is a muddy, fine-grained, medium- to thick-bedded limestone with some chert nodules. The weathered surface is earth colored whereas a fresh surface is dark gray. Bedding planes are commonly wavy with numerous stylolites. A chert band 2.5 cm thick occurs at the top of the "Chonetes" zone.

Communities Recognized by Lindemann

Thirty-eight exposures within the outcrop belt of the Onondaga Limestone were selected for detailed measurement and study. At each site, faunal censes, hand specimens, and lithologic descriptions were obtained from single beds. Each sampled bed was divided into small quadrats and each quadrat meticulously examined with a large hand lens for detail and small specimens. Because of the nature of most exposures and the erosional characteristics of the formation, sampling was predominantly restricted to vertical sections of beds. As a result, the diversity of identifiable faunal elements was limited, and with the exception of corals, was usually restricted to high-level taxa.

A total of nine communities was identified in the Onondaga Formation in the area between Buffalo and Catskill. Six occur in the field trip area. The faunal compositions of these six are described below. Their stratigraphic distributions in central New York are shown in Figure 4.

1) Acinophyllum-Heliophyllum community. Corals dominate this community. With 25 genera constituting in excess of 50 percent of the fauna, they are nowhere more diverse or numerous. Acinophyllum and Heliophyllum are particularly abundant. Many genera such as the rugosans, Siphonophrentis and Cystiphylloides, as well as the massive tabulates, Emmonsia, Favosites, and Lecfedites, reach formation-wide abundance maxima. Cylindrophyllum and Pleurodictyum, which are usually restricted to Edgecliff reefs, are present. Stromatoporoids are also present. Camerate crinoid columnals measuring up to 2.5 cm in diameter and 15 cm in length attain extra-biohermal abundance maxima. The trilobite, Phacops, is present in small numbers as are the gastropods, Platyceras and Straparollus. Bryozoans are rare and never intact. Brachiopods are few in number.

Tubular vertical burrows dominate the ichnofauna. Three undulatory forms with diameters of 1 mm, 3 mm, and 7 mm, are present. The 3-mm form is most numerous. Conical vertical burrows 6 mm in length and 2 mm at the largest diameter are common. Cup-shaped epichnia 3 cm wide and 4 cm deep are present. Horizontal burrows and bioturbation are rare.

2) Platyceras dumosum-ramose bryozoan community. Bryozoans dominate this community. Ramose fistuliporid cyclostomes are abundant

and mostly restricted to this community. Fenestrate cryptostomes are also present, but far less common. Trepostomes are rare. Specimens of all bryozoan growth forms are preserved intact. Brachiopods are the second most abundant faunal group. Approximately 45 percent of the individuals remain articulated. The corals, which comprise less than 10 percent of the fauna, are represented by twelve genera. The rugosans and tabulates are respectively dominated by Heteophrentis and Coenites. Specimens of all coral genera occur intact and/or in presumed life position. Stromatoporoids and the lithistid sponge, Hindia, are present in small numbers. Trilobites are dominated by Phacops, with Odontocephalus also present. Cephalopods include rare specimens of orthoconic and gyroceraconic nautiloids. Gastropods, while not abundant with respect to the total fauna, are by no means few in number. Playtceras, Platystoma, and Pleurotomaria occur as juveniles and adults. Platyceras dumosum is common, ubiquitous, and essentially restricted to this community.

A well-developed ichnofauna is present. Two forms of horizontal burrows with diameters of 2 mm and 5 mm are present. The burrows occasionally convert to a curved concave-up orientation and terminate at bedding surfaces or prominent intra-bed laminations. These traces have been identified as forms of Chondrites by Adolf Seilacher (Lindholm, 1967, p. 66). Two additional horizontal burrows with diameters of 1 cm and 1.5 cm are also present. Three forms of straight, lined, mud-filled vertical burrows, having diameters of 4 mm, 1 cm, and 2 cm were also observed. The latter form terminates at upper bedding surfaces, and is a minimum of 15 cm in length. Undulatory vertical burrows occur in small numbers. These traces measure 2.5 mm in diameter and 4.5 cm in length. They invariably terminate in calcarenite-filled swales of irregular upper bedding surfaces. An epichnion measuring 3 cm wide and 4 cm deep is present. In plan view, these traces are discontinuous and cross one another at irregular intervals and angles.

3) Syringopora-Aulopora community. Dominance of this community is shared by brachiopods and corals. Approximately 40 percent of the brachiopods are articulated. Corals are represented by 11 genera. The tabulates, Syringopora and Aulopora, are the most numerous corals. The former attains its formational maximum abundance within this community. The rugosans are dominated by Heterophrentis and Heliophyllum. Coral preservation is excellent. Many specimens show no signs of post-mortem wear and some remain in presumed life position. Domal stromatoporoids of moderate dimensions are present. Gastropods, particularly Pleurotomaria, attain their maximum formational abundance. Numerous juveniles are present. Trilobites are represented by Phacops. Fenestrate bryozoans are few in number, but several specimens were found to have been preserved intact and nearly in life position.

Ichnofossils are abundant. Three common forms of Chondrites measure 2 mm, 5 mm, and 7 mm. Three forms of vertical burrows are present. The most common has a diameter of 2 mm and is lined. Another has an outer diameter of 1.5 cm, a lining 3.5 mm thick, and a living space of 8 mm in diameter. The third vertical trace is an "escape burrow", 3.5 mm in diameter. It contains sediment in which fossil material is oriented in a concave-down chevron pattern. This fossil is known from a single specimen.

4) Aulopora-Platystoma community. Brachiopods dominate this community. However, they constitute less than half the total fauna. Juvenile specimens are common. Overall, 33 percent of the adult brachiopods are articulated. This percentage is somewhat higher for the genera Atrypa and Athyris. Ramose cyclostome bryozoans constitute a large percentage of the fauna. Their zoaria are occasionally preserved intact and semierect. One specimen was preserved with its holdfast in place, having simply fallen over prior to burial. Trilobites, represented by Phacops, are relatively common. Corals are dominated by the tabulate, Aulopora, and the solitary rugosan, Bethanyphyllum. Aulopora reaches its formational maximum abundance in this community. The gastropods, Platystoma and Platyceras, occur in small numbers as juvenile, and adult individuals. The lithistid sponge, Hindia, is also present, though rarely common.

The ichnofauna of this community is abundant and diverse. The most common trace is a form of Chondrites 2 mm in diameter. This fossil is commonly present and locally pervasive within the strata. Two additional forms of Chondrites, with diameters of 0.5 mm and 4 mm, are much less common. Vertical traces are restricted to a single form of straight, lined, mud-filled domichnia measuring 4 mm in diameter. A form of epichnia measuring 3 cm in width and 5 cm in depth is also present. In plan view, these traces are discontinuous, arcuate, and locally coincident.

5) Amplexiphyllum-Odontocephalus community. This community is dominated by the trilobites, Odontocephalus, Dechenella, and Phacops. Brachiopods are the second most abundant group of organisms. Corals comprise less than 4 percent of the fauna and are of low diversity. They are dominated by the small solitary rugosan, Amplexiphyllum hamiltoniae. Heterophrentis and the tabulate, Aulopora, are also present. Cephalopods are represented by local occurrences of the gyroceraconic nautiloid, Halloceras.

Numerous signs of bioturbation indicate that detritivores were an important component of this community. However, only two types of recognizable ichnofossils are present. The more common trace is a Chondrites with a diameter of 2-3 mm. The other recognizable trace consists of straight, discontinuous epichnia measuring 2 cm in width and 5-50 cm in length. They occur only at the upper surfaces of beds, are filled with coarse fossil debris, and show no preferred orientation.

6) Styliolina-Michelinoceras community. The macrofauna of this community is overwhelmingly dominated by brachiopods. Many are juveniles and retain delicate shell structures. Styliolina, best seen in thin section, is of maximal abundance. It constitutes up to 95 percent of the fossil material seen in thin section and is therefore the actual faunal

dominant. Trilobites are represented by Odontocephalus and Phacops. Rugose and tabulate corals are rare, of low diversity, and primarily consist of juvenile and/or stunted individuals. Gastropods, primarily Platystoma, are also rare. The orthoconic nautiloid, Michelinoceras, is present, as is the goniatite, Foordites.

Bioturbation is prevalent. However, recognizable ichnofossils are uncommon. A Chondrites measuring 2-3 mm in diameter is present. A vertical trace measuring 5 mm in diameter is also present, but uncommon. The most common tract fossil is an epichnial groove 4 cm in depth and 2 cm in width. Each epichnion occurs at the top of the host bed, extending downward as a flat-bottomed trough of fossiliferous sediment.

ACKNOWLEDGEMENTS

We thank Drs. Richard A. Park of Rensselaer Polytechnic Institute, Lawrence V. Rickard of the New York State Museum and Science Service, and John J. Thomas of Skidmore College for their consultations and for reviewing various sections of that which became this paper. Special thanks go to Faith Grey Lindemann for proofing the final draft, driving during preparation of the trip log, and patiently coping with RHL during the production of this article.

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ROAD LOG FOR PALEOCOMMUNITIES OF THE ONONDAGA LIMESTONE

CUMULATIVE MILEAGE	MILES FROM LAST POINT	ROUTE DESCRIPTION
0.0	0.0	Leave SUNY Binghamton campus and heat east on Route 434 to Interstate Route I81. Take I81 north towards Syracuse. Log mileage begins at this point.
66	66	Leave Route I81 at Exit 16 and turn left onto U. S. Route 11 north.
66.7	0.7	Very shortly after passing beneath I81 turn left from Route 11 onto Quarry Road.
67	0.3	STOP 1. Park on the road shoulders and walk into the quarry which lies between Quarry Road and the interstate.

STOP 1. This is the Indian Reservation quarry described in connection with Feldman's Leptaena-Megakozlowskiella community.

67.3	0.3	Return to Route 11 and turn left (north) towards Nedrow.
70.1	2.8	Leave Route 11 turning left (west) onto Route 173.
75.2	5.1	Turn left onto Split Rock Road. The road sign may not be visible so watch for a black-on-white D.O.T. sign and a yellow-and-blue state historical marker.
75.9	0.7	STOP 2. Continue to the end of Split Rock Road and into the quarry entrance.

STOP 2. This is the type section of the Edgecliff Member which is named for Edgecliff Park, located just to the west. The member's full thickness of 2.3 m is exposed in the upper areas of the main quarry. Its erosional base is marked by the presence of phosphate nodules. In the southern quarry wall, 3 m of the Nedrow Member are exposed. The top of the Nedrow has been removed by erosion.

76.6	0.7	Upon leaving the quarry, return to Route 173 and turn right.
86.7	10.1	In Jamesville turn right (south) onto N.Y. Route 91.

87.5	0.8	STOP 3. Turn left into the broad flat quarried area adjacent to Route 91 and park. The Onondaga is exposed in the upper quarry. Walk up into it by way of the unpaved road which comes down to Route 91 just south of where you've parked.

STOP 3. A nearly complete section of the Onondaga is exposed in this quarry. Only the top of the Seneca is gone. Thicknesses of the members are given below. An excellent exposure of the "Springvale" can be found to the west of the large storage building seen from Route 91. Here the "Springvale" contains a sparse brachiopod-bryozoan fauna and phosphate nodules. The main quarry floor is in the Edgecliff Member which is best seen along the south wall; it is at your right as you enter. Most of the quarry wall consists of the Nedrow and Moorehouse Members. This is the type locality of the latter. Fracturing and jointing of the rock have made hand and foot holds insecure. Please avoid falling or dropping rocks (or yourself) on other trip participants. The Seneca Member can be closely observed above and to the east of the main quarry. You can get to it by walking around the quarry to the north.

Seneca Member	1.5 m
Moorehouse Member	7.0 m
Nedrow Member	4.4 m
Edgecliff Member	5.2 m
"Springvale"	1.5 m

88.3	0.8	Take Route 91 back to Route 173 and turn right (east) uphill toward the prison.
88.8	0.5	Turn right (south) onto Taylor Road.
89.6	0.8	STOP 4. Park along road shoulder.

STOP 4. East of Taylor Road in the clump of trees across from the first house on the left is exposed a small section of a reef in the Edgecliff Member. While exposure does not permit a three-dimensional study of the reef, the fauna which includes _Cylindrophyllum_, _Acinophyllum_, _Emmonsia_, and _Favosites_ indicates that what we are seeing is part of the core facies.

90.4	0.8	Return to Route 173 and turn right.
95.8	5.4	STOP 5. Drive through Manlius on Route 173. Going up hill and out of town there will be what appears to be a wooded and scrubby area on the left (north) side of the road. This is sharply broken by the school athletic field. Turn left into the western corner of the field and park near the storage shed. Frog Pond quarry is the scrubby area west of the athletic field. You can get into the quarry by way of any of the foot paths which lead in that direction.

STOP 5. The full 5.8 m of the Edgecliff Member are exposed here as is the lower 1 m of the Nedrow. The base of the Edgecliff is exposed on the middle of the quarry face which overlooks the frog pond in the southwest corner of the quarry. Rock weathering and extensive exposure of bedding surfaces make sampling and fossil collection here excellent.

114.1	18.3	Continue east on Route 173. At Chittenango get onto Route 5 headed east.
126.7	12.6	At the intersection of Routes 5 and 46 turn right (south) onto Route 46 towards Munnsville.
135.2	8.5	Proceed through Stockbridge and Munnsville and bear right on Phillips Road.
135.4	0.2	Turn right onto Phillips Drive.
135.8	0.4	Turn right onto Stockbridge Falls Road.
137.3	1.5	STOP 6. Proceed uphill until you see black shale to the right and Oneida Creek to the left. Look for and park in the wide road shoulder to your left as the road leans right. Watch out for eastbound traffic!

STOP 6. While virtually all of the Onondaga can be pieced together in the creek bed downstream from here, we will observe only the Seneca Member, 3.7 m of which are present in the stream. If water levels and recent sediment transport have been kind, we will be able to see a rare exposure of the contact between the Onondaga and the Union Springs Shale. For those who are interested, the full thickness of the Cherry Valley Limestone, about 1 m, is exposed on the north side of the road. There is no need to scramble up the shale for it, just trace the bed along until you meet it near road level.

138.8	1.5	Take Stockbridge Falls Road downstream to Pratts Road and turn right.
143.8	5	Turn right from Pratts Road onto Route 20.
174.8	31	At Lafayette get onto I81 headed south and return to Binghamton.
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GLACIAL GEOLOGY OF THE CHENANGO RIVER VALLEY
BETWEEN BINGHAMTON AND NORWICH, NEW YORK *

by

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REGIONAL SETTING

The field trip area (Fig. 1) is located in the southern part of the Chenango River valley, a tributary to the Susquehanna River drainage system. The region is included in parts of the Binghamton West, Castle Creek, Greene, Chenango Forks, Brisben, Tyner, Oxford, Norwich, and Holmesville 7½ minute U.S.G.S. topographic quadrangles. The area has a total relief of 341 m (840 to 1980 ft). The bedrock is predominantly Devonian shale, siltstone, and sandstone (Fisher et al., 1970).

GLACIAL GEOLOGIC SETTING

Brigham (1897) recognized the extent of the glacial sediments within the Chenango River valley between Binghamton and the Mohawk Valley. Tarr (1905) described the characteristics of the glacial deposits near the Finger Lakes. Fairchild (1932) named the thick drift units in the Finger Lakes region the Valley Heads moraine. He delineated other areas of drift deposits: the Olean at the terminal moraine in Pennsylvania and the Susquehanna Valley kames. MacClintock and Apfel (1944) used the term "Binghamton moraine" to describe the Susquehanna Valley kames of Fairchild, indicating that the drift was deposited during a separate advance. They suggested that the Olean was the oldest Wisconsin, Binghamton - middle Wisconsin, and the Valley Heads - youngest Wisconsin. Peltier (1949) correlated terraces along the Susquehanna River in Pennsylvania with pre-Wisconsin, Olean, Binghamton, Valley Heads, and Mankato ice advances in New York. Denny (1956) questioned the presence of the Binghamton advance in the Elmira region. He theorized that (1) the Binghamton border may be north of the Valley Heads border and therefore concealed, (2) the Binghamton border is incorporated within the Valley Heads border, and (3) the character of the Binghamton materials is completely changed between the type locality (Binghamton) and Elmira. Connally (1960, 1964) on the basis of heavy-mineral analyses indicated that the Binghamton is related to the Valley Heads advance. Moss and Ritter (1962) suggested that the Binghamton was not a separate advance, but a phase of the Olean. Coates (1963) suggested that a single ice sheet deposited the drift with the Olean as the upland facies and the Binghamton as valley facies. Hollyday (1969) showed that the drift in the valleys in the vicinity of the field trip ranges

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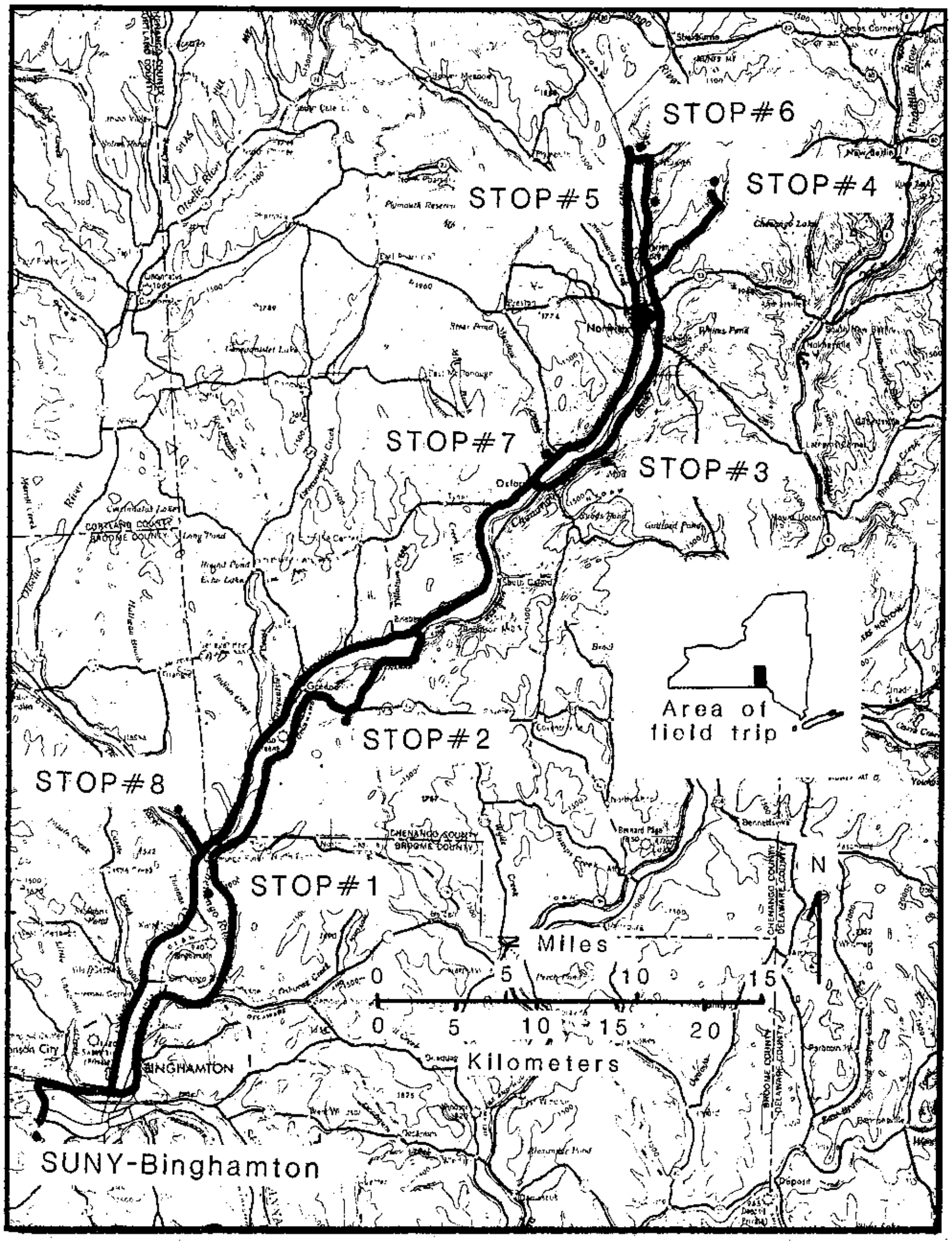

Figure 1. Field trip route and stops.

between 15 and 76 m (50 and 250 ft) thick. I have presented the idea of a single Woodfordian ice sheet that deposited the Olean and Binghamton deposits, with a miminum age of 16,650 ± 1800 radiocarbon years B.P. for wood at the base of a bog (Cadwell, 1972a, 1972b, 1973a, 1973b, 1974, 1975, 1978). Incorporated in this idea is the concept of "valley ice tongue" retreat. An ice tongue may have extended several miles down valley beyond the upland ice-margin position. Krall (1977) described the Cassville-Cooperstown moraine as a late Wisconsin recessional moraine. Aber (1978) described the erratic-rich drift in the Appalachian Plateau of N.Y.

THE PROBLEM OF ICE RETREAT

The number of glaciations in central New York is difficult to delineate. Denny and Lyford (1963) indicated that the earlier Wisconsin (Olean) ice did not build a prominent moraine at the drift border or construct any significant moraine south of the Valley Heads moraine. Cadwell (1978) described the bedrock control of retreating ice margin positions in the northern Chenango River valley. Figures 2 and 3 suggest the mode of formation as the ice retreats in the uplands.

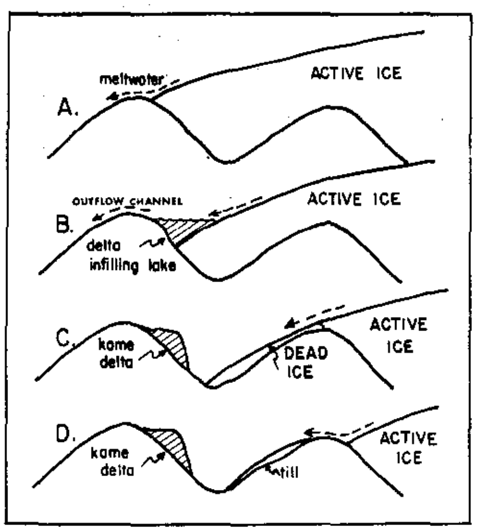

Figure 2. Diagram of profile of upland ice retreat.

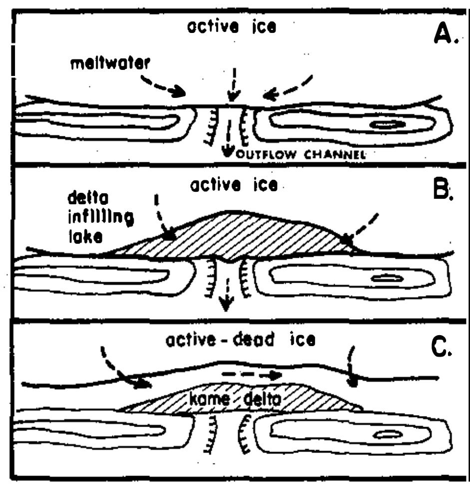

Figure 3. Sketch map of upland ice retreat.

Figure 2 illustrates a profile of upland ice retreat. Margin locations and associated deposits are governed by preglacial topography. Figure 3 is a sketch map of upland ice retreat. In Figure 3A the ice is against the mountain, with meltwater flowing through an outflow channel. In Figure 3B the hachured area represents a proglacial lake and delta. Meltwater continues to drain through the outflow channel. Figure 3C represents continued ice-margin retreat. Meltwater ceases to flow through outflow channel and remains within the stream valley. Deposition of a kame-terrace sequence may occur as meltwater adjusts to the present stream level. Criteria that led to the identification of these ice-margin positions include the following: (1) the surface morphology or shape of the upland hills, (2) the location of outflow channels in the uplands,

(3) the association of upland meltwater deposits, (4) the configuration of stratified drift around umlaufbergs (bedrock hills surrounded with outwash deposits), and (5) the sequence of valley meltwater deposits.

The valley meltwater deposits are the primary emphasis for this field trip. Of particular importance is the relationship between the retreating valley ice tongue and glaciofluvial and glaciolacustrine deposits. The sequence creates a depositional mosaic or suite of deposits associated with the retreating ice tongue (Figure 4). Field trip stops 3 and 5 examine kame deltas associated with an ice tongue near Oxford.

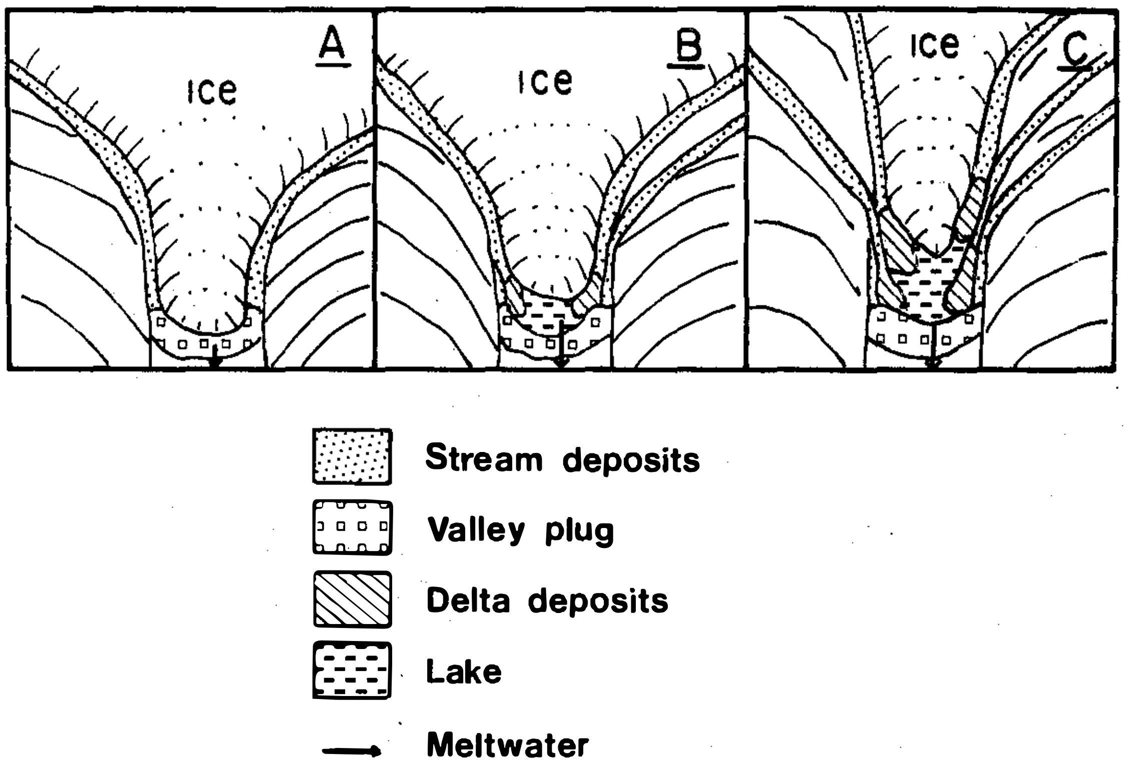

Figure 4. Generalized diagram of ice-tongue retreat with associated valley and upland deposits.

Deglaciation Chronology

The Woodfordian ice sheet retreated primarily by backwasting in the uplands. Stagnant ice deposits continuous from one valley to another across a divide are absent in the study area. This suggests that active ice tongues remained in the valleys. The size of the ice tongues remaining during retreat was governed by such factors as rate of upland retreat and rate of valley ice melting. In areas of rapid upland ice retreat, long tongues of ice could have remained in the valleys behaving in some ways similar to a valley glacier. The valley ice tongue retreated by both backwasting and downwasting.

Ice margins identified in the field trip area are indicated in Figure 5. During the development of each margin an ice tongue remained in the main Chenango River valley.

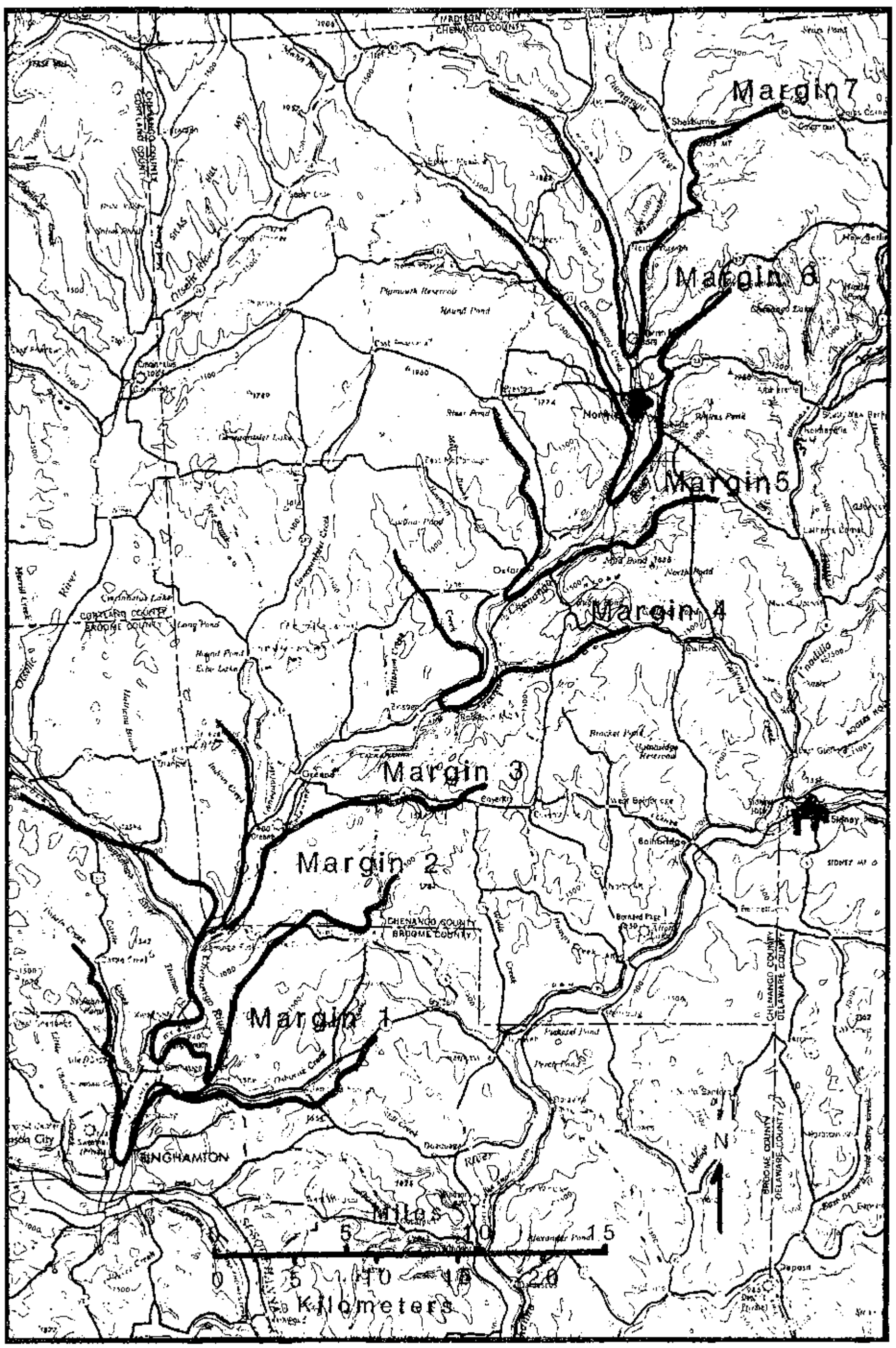

Figure 5. Retreatal ice margin positions in the southern Chenango River valley.

Margin 1. During development of this margin an ice tongue extended in the Chenango River valley south to Binghamton. The umlaufberg at Kattellville was exposed through the ice and meltwater carved the windgap in the upland.

Periglacial conditions existed south of the ice front. Evidences of these climate conditions are found at tor localities, at the top of Ingraham Hill, southwest of Binghamton. Meltwater flowed laterally to the ice tongue and, as the ice tongue diminished in size, kames and kame terraces were deposited along the east and west valley walls of the Chenango valley. This kame terrace can be seen at the 9.0 mile position on the field trip. Meltwater flowed out of high-level channels south of Port Crane during early stages of ice retreat.

Margin 2. During the development of this margin, streams flowed laterally to the ice tongue. Ice dammed the valley at Port Crane forcing meltwater to flow on the west side of the umlaufberg in the Kattellville valley (112.0 miles in field-trip description). The manner of retreat of the ice front changed, briefly, from backwasting to downwasting as meltwater deposited thick sands and gravels in the vicinity of the Chenango Valley State Park at Chenango Forks. Meltwater streams were unable to transport the large amounts of sand and gravel, and the valley became choked with debris. This debris (valley plug) controlled the local base level of deposits laterally to the ice within the valley to the north.

Large blocks of ice left within these deposits formed kettles, such as Lily and Chenango Lakes. An ice-channel filling is preserved where the debris was deposited between the blocks (See the topographic map, STOP 1, of Chenango Valley State Park in the field-trip guide). Numerous other kettles are preserved to the north of these lakes, similar perhaps to pitted outwash. A radiocarbon age date of 16,650 ± 1800 yBP (BGS-96) was obtained from wood at the base of peat in a kettle-hole bog.

Margin 3. The most notable feature associated with margin 3 is the Wheeler Creek esker complex. A topographic map of this area is included in the field-trip guide for STOP 2. This discontinuous esker is 6.4 km (4 mi) long with an average relief of 12 m (40 ft). The ice at the margin was very thin, barely covering the south end of the esker complex. An ice tongue remained in the main Chenango valley south to Chenango Forks. Meltwater flowed westward from the region of the esker to the ice tongue. With continued retreat meltwater flowed in the main Chenango valley adjacent to the ice and deposited the kame terrace noted at mile 20.2 in the road log.

Margins 4 and 5. The kame field and kame terraces northeast of Brisben were deposited as the upland ice retreated to margin 4. During retreat to margin 5, the terminus of the valley tongue was just south of Oxford Station. Large amounts of sand and gravel surrounded the stagnating blocks of ice and formed a valley plug that served as a dam and controlled base level for deposition of the Lynk delta (STOP 3) and the Emerson delta (STOP 7).

Margin 6. During the retreat of the ice to margin 6, the valley plug south of Oxford still dammed the valley, creating a large lake and the temporary base level for the streams lateral to the ice. The ice tongue extended several miles south of Norwich and a lake extended south to the valley plug. The presence of this lake is substantiated by well data, indicating greater than 91 m (300 ft) of clay with sand and gravel. A well for the Norwich Pharmaceutical Company reveals gravel between 0 and 6 m (20 ft) and silt and clay to bedrock at 93 m (305 ft). Figure 6, a longitudinal profile of the Chenango River valley, has the thickness of sand and gravel plotted for the field trip area. The R following a well depth indicates depth to bedrock.

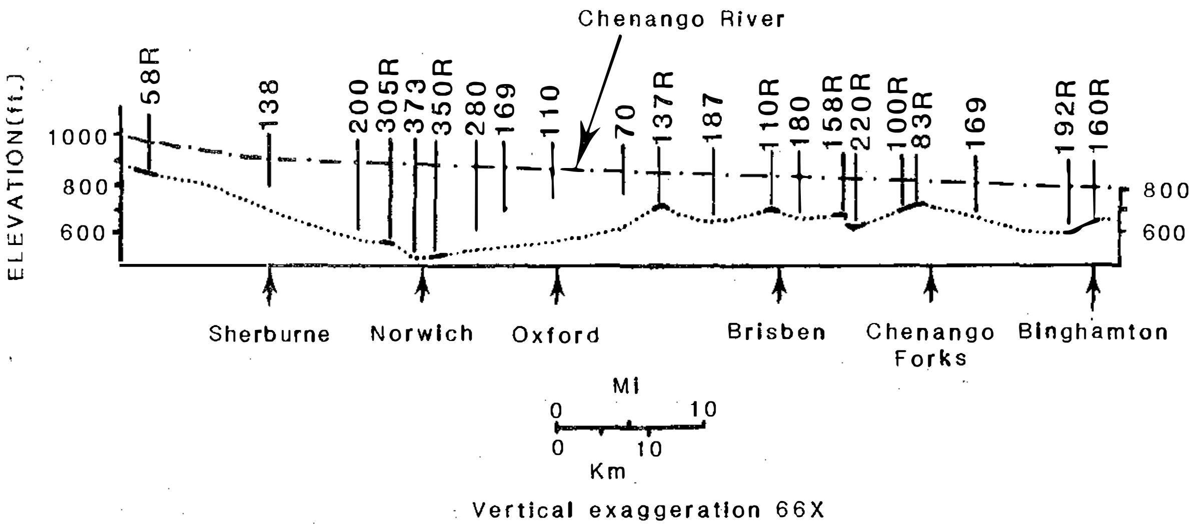

Figure 6. Longitudinal profile of the Chenango River. The dotted line is the inferred depth to bedrock in feet.

Margin 7. During the development of this margin,meltwater was flowing from the active ice margin through a zone of stagnating ice, to a moat or ice-dammed lake that developed at North Norwich. Major meltwater flow was along the western side of a nunatak, the prominent isolated hill north of Norwich, and transported fine-grained sands (1.93ϕ, 0.233mm) into a supraglacial ice-dammed lake. These sediments will be seen at STOP 6 in the field trip. It is suggested that an englacial empoundment of water was present beneath the supraglacial lake. The englacial pond drained, causing collapse of the supraglacial pond. This collapse caused dis-orientation of blocks of frozen (?) sediments as they dropped into the englacial void. Unconsolidated or unfrozen sediment was then deposited en masse, as a unit surrounding the blocks producing a massive (unstratified) unit. The lake refilled while remaining shallow, with formation of current ripples above the massive zone. For further discussion of the development of this lake,see Cadwell, 1974.

Deglaciation continued to the north of the Valley Heads Moraine, permitting some of the lakes to the northwest to drain through the Mohawk and Hudson Rivers. The ice then readvanced to the late Woodfordian Valley Heads moraine. Associated with this margin are thick valley-train deposits infilling the Chenango River valley.

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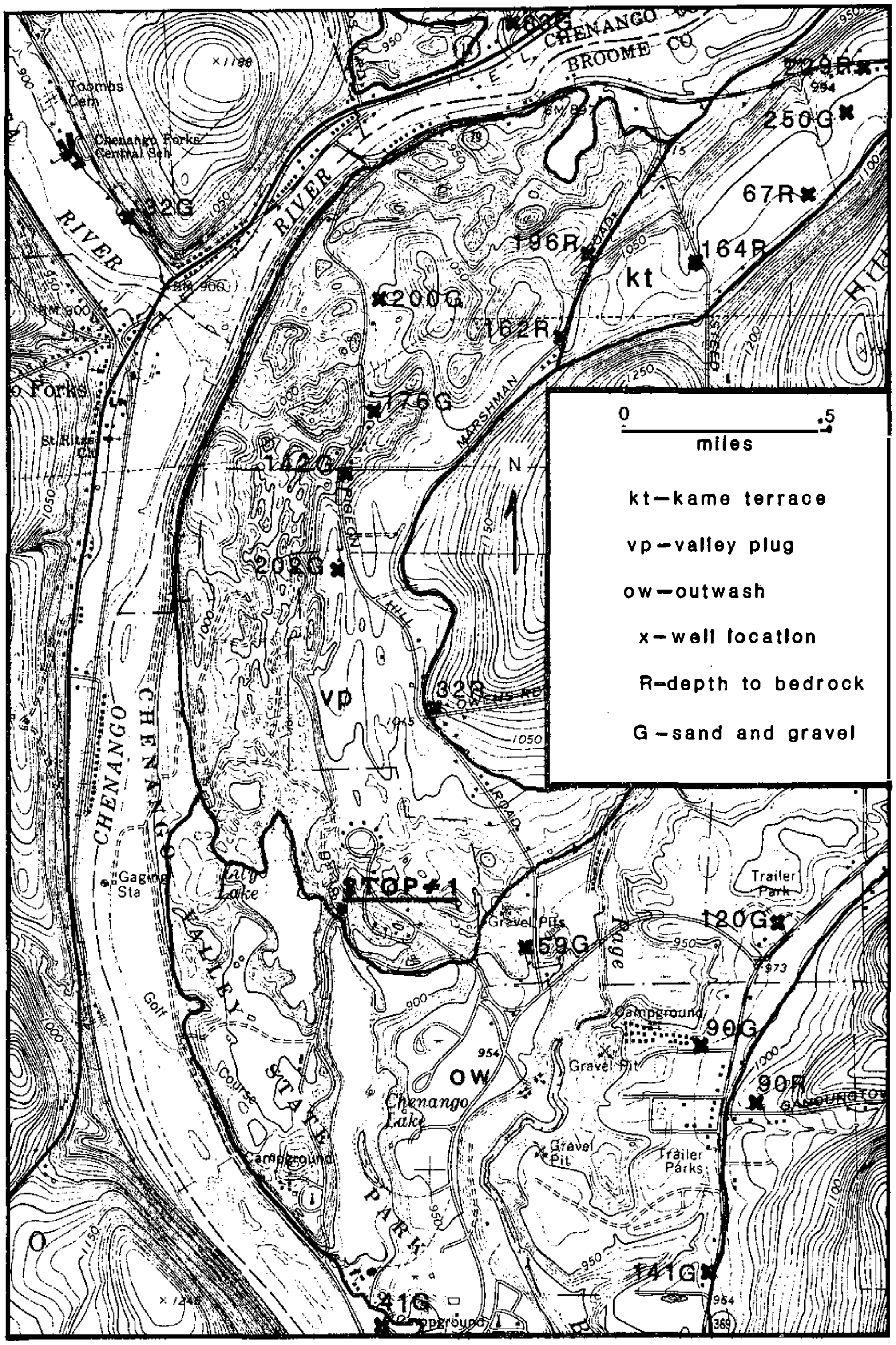


STOP 1. Chenango Valley State Park and vicinity.

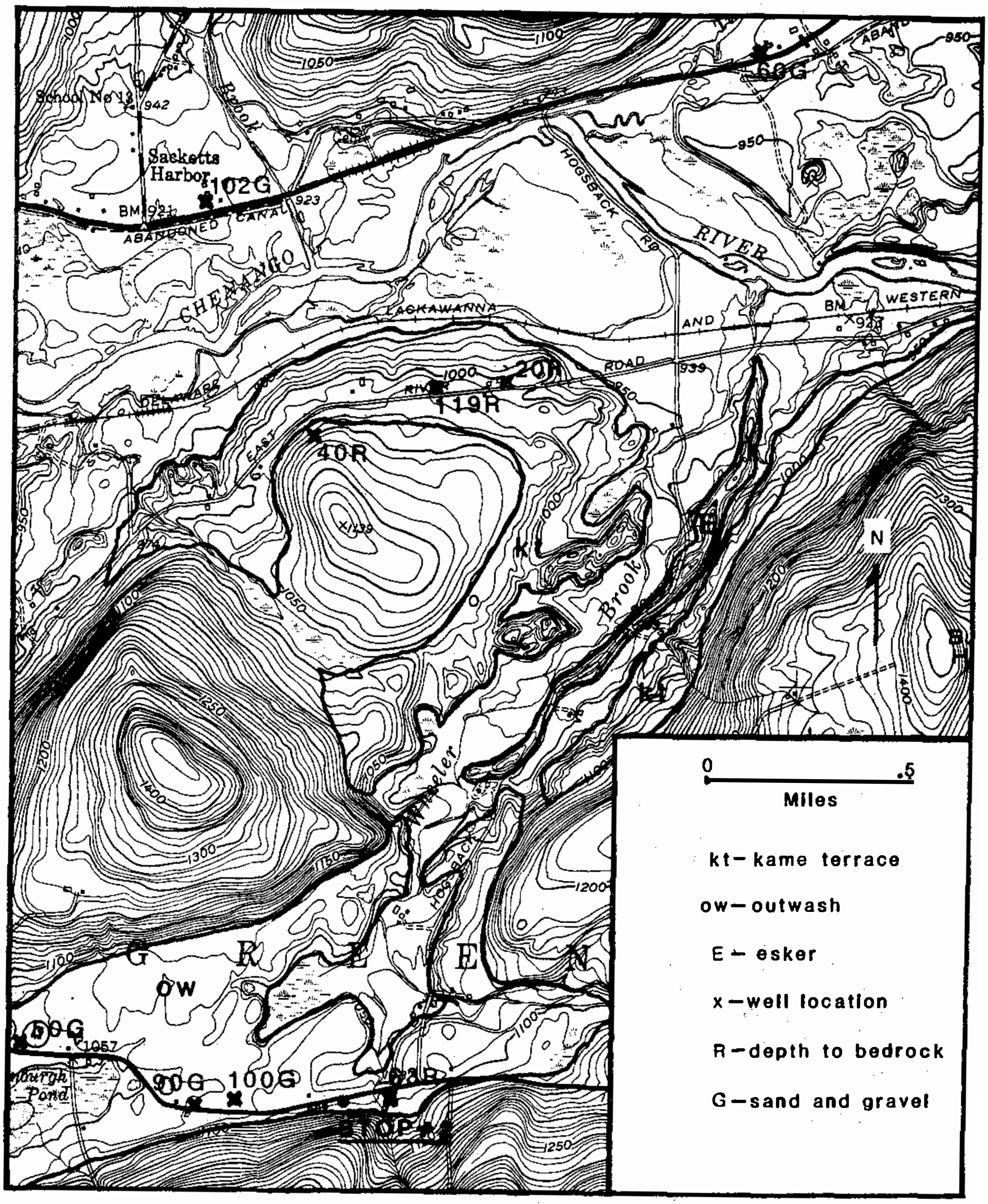
FIELD-TRIP ROAD LOG AND ROUTE DESCRIPTION FOR GLACIAL GEOLOGY OF THE CHENANGO RIVER VALLEY

CUMULATIVE MILEAGE	MILES FROM LAST POINT	ROUTE DESCRIPTION
0.0	0	This road log begins at the entrance to the SUNY Binghamton campus. Proceed west on Route 434.
0.2	0.2	Bear right on Route 201 N.
0.7	0.5	Cross over Susquehanna River.
1.0	0.3	Go around traffic circle and continue north on Riverside Drive West.
1.9	0.9	Junction Route 17. Proceed east on Route 17.
3.7	1.8	View south across Binghamton. Binghamton is built on outwash sands and gravels greater than 50 ft (15 m) thick.
5.1	1.4	Junction Rt. 81. Proceed east on 81 and 17.
5.9	0.8	Exit at N.Y. Route 7 North. Proceed north on Route 7N.
7.9	2.0	Note kame terrace to right.
9.0	1.1	Outwash sand and gravel exposed in the kame terrace.
9.4	0.4	Route 7 rises onto kame terrace at Hillcrest.
10.0	0.6	Route 7 becomes Route 88.
11.9	1.9	Exit Route 88 to Route 369. Turn left (north) on 369. Note gravel pits in sand-and-gravel outwash to the left of Route 369 as you proceed north.
14.2	2.3	Kame is on left.
15.8	1.6	Turn left toward entrance of Chenango Valley State Park.
16.5	0.7	Entrance to Chenango Valley State Park. STOP 1.

STOP 1. CHENANGO VALLEY STATE PARK

All of the glacial features preserved in this park were formed at the retreating Woodfordian ice margin. Massive amounts of sand and gravel was deposited around disintegrating blocks of ice (Refer to the topographic map of the state park and vicinity). These horizontally stratified sands and gravels are at least 200 feet (61 m) thick. An ice-channel filling is preserved between Lily and Chenango Lakes. The pebble lithologies are 10 percent limestone; 75 percent local siltstone, sandstone and shale; and 15 percent exotics (from outside the Chenango River drainage basin).

		Return to entrance of the park, then proceed to the left onto Pigeon Hill Road.
16.6	0.1	Note the sand pit to the left. These sands were deposited in relatively quiet water adjacent to stagnating ice in the Page Brook valley. Meltwater was flowing from the north.
17.5	0.9	Meltwater channel. The sand and gravel here is at least 175 ft (53 m) thick.
18.0	0.5	Note large kettles in outwash gravels.
18.8	0.8	Junction with Route 79. Proceed to the right (north). A radiocarbon age date of 16,650 $\pm$ 1800 yBP was obtained from wood at the base of a kettle to the right, over the hill.
20.2	1.4	Turn left onto unnamed road, toward Greene. This kame terrace has at least 250 ft (76 m) of sand and gravel.
20.4	0.2	Enter Chenango County. This road is on a large kame terrace built when an ice tongue occupied the Chenango Valley. The sand and gravel is at least 196 ft (60 m) thick.
24.7	4.3	A gravel pit to the right (up the road) has outwash gravels. An ice tongue occupied the main valley as meltwater flowed from the ice margin to the east. The lithologies present are 4 percent limestone, 76 percent locals, and 20 percent exotics.
25.9	1.2	Enter Greene.
26.5	0.6	Junction with Routes 41 and 206. Turn right (east) onto Routes 41 and 206. Proceed across the kame terrace.
27.7	1.2	View across meltwater channel and outwash gravels. These sands and gravels are as much as 100 ft (30 m) thick.



STOP 2. Wheeler Creek ice-channel filling (esker).

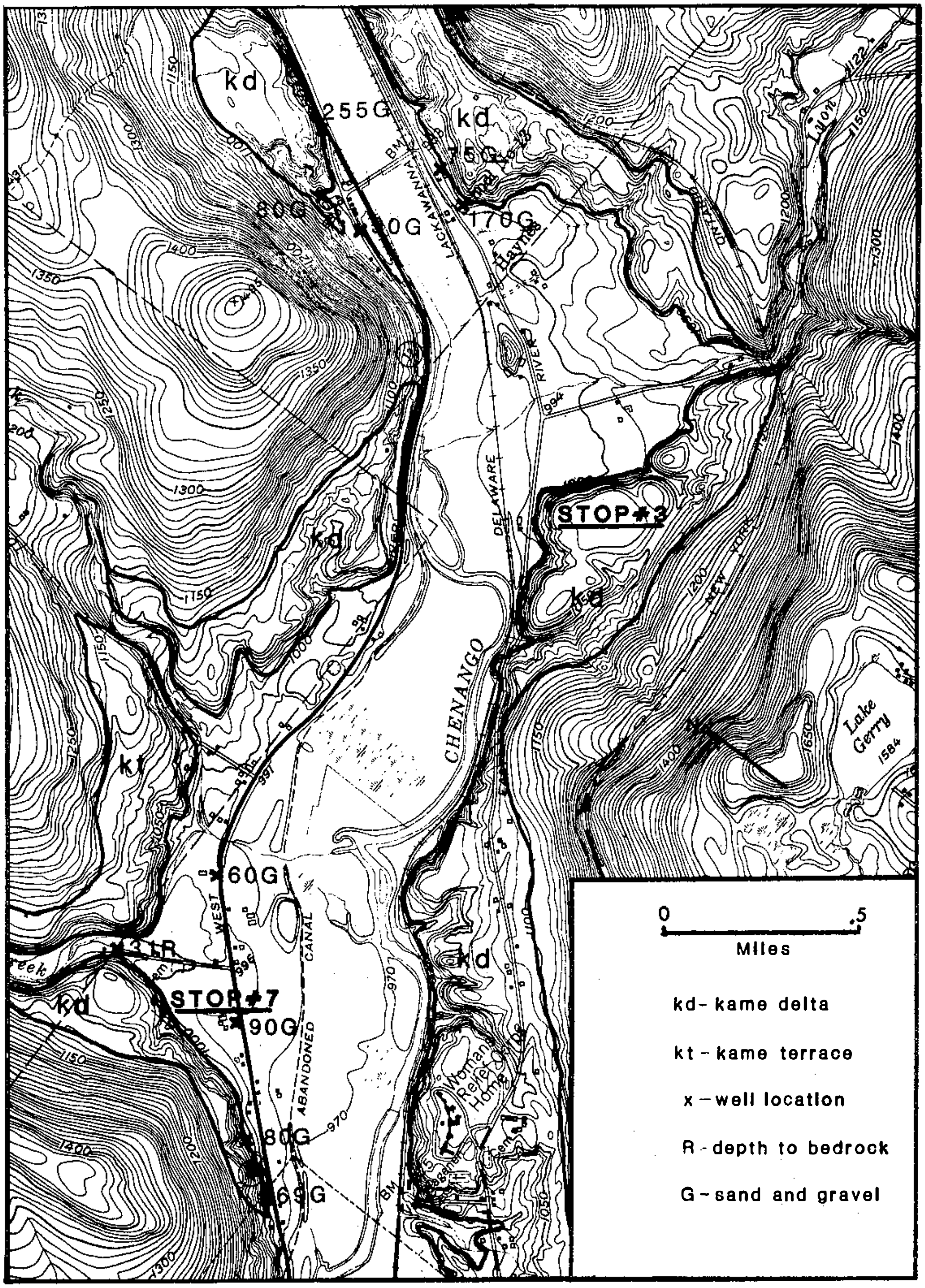
29.9	1.2	STOP 2. WHEELER CREEK ESKER

STOP 2. This will be a brief stop to discuss the formation of the Wheeler Creek esker (Refer to the topographic map). A thin ice margin was located at this point. Meltwater flowed westward from this margin toward the main Chenango Valley. The kame terraces and the esker were deposited adjacent to and beneath the ice, respectively.

		Proceed east on Routes 41 and 206.
30.0	0.1	Turn left onto Hogsback Road.
30.7	0.7	The esker can be seen on the left as you descend from the kame terrace. Meltwater flow was to the south.
31.0	0.3	Road is located on the crest of the esker.
32.0	1.0	Junction with East River Road. Turn right (east).
34.6	2.6	Turn left toward Route 12 (across bridge).
34.8	0.2	Junction Route 12. Turn right (north). Proceed north on Route 12.
40.1	5.3	Note the massive hummocky topography (valley (plug) on the east side of the valley. This sediment and ice served as a base-level control for deposition to the north. The significance of this will be discussed at STOP 3.
42.8	2.7	Traffic light, center of Oxford. Turn right on Route 220. Cross Chenango River.
43.1	0.3	Bear left on 220.
43.6	0.5	Turn left, toward Vets Home, on County Route 32. This road is on top of a large kame terrace.
45.7	2.1	STOP 3. LYNK DELTA

STOP 3. The Lynk Construction Company removes sand and gravel from this large delta/lacustrine complex. Meltwater was flowing from the east along Lyon Brook and sedimentation occurred in a lake adjacent to an ice tongue. Note the planar-surfaced features to the north. The elevation of all these features was controlled by the valley plug noted at mileage 40.1. The lithologies present are: 20 percent limestone, 52 percent locals, and 28 percent exotics.

		Proceed north on county Route 32.
46.6	0.9	Note pit in kame delta to the right.



STOP 3 and STOP 7. Lynk and Emerson kame deltas. These kame deltas were constructed adjacent to an ice tongue in the Chenango River valley.

50.4	3.8	Cross alluvial fan. The thickness of drift in the valley is at least 350 ft (107 m).
52.2	1.8	Intersection with Route 23. Proceed north on county Route 32 (straight through).
53.4	1.2	Junction Route 320. Turn right (east) on Route 320.
58.0	4.6	Turn left onto dirt road. Entrance into the Whapanaka State Forest.
59.5	1.5	STOP 4. WEDGE OR POP-UP?

STOP 4. Walk up into the bedrock quarry. This pit is used by the state for maintenance of the roads. The face of the exposure is oriented N20E along one of the joint planes. The wedges are developed along N75W. One wedge is well developed. Two others are partially developed. The wedges persist for at least 3 ft. to the east.

		Return to Route 320.
61.0	1.5	Turn right onto Route 320 (south).
65.6	4.6	Junction with Route 32. Turn right (north) onto Route 32.
68.3	2.7	STOP 5.

STOP 5. Picture stop of kame terrace. When this kame terrace was deposited an ice tongue occupied the main Chenango valley. Note kame delta on the west side of the valley.

70.1	1.8	Junction Route 12. Turn left onto Route 12 and cross the Chenango River.
70.7	0.6	Bear right toward North Norwich.
70.9	0.2	Bear right toward North Norwich.
71.0	0.1	Turn right on N. Main Street.
71.3	0.3	STOP 6. CHENANGO COUNTY SAND PIT

STOP 6. The exposure in this pit illustrates lake sands, silts,and clays, with current-ripple laminations. Several years ago exposures in this pit had a section beneath the well-stratified lake sands. In this lower section are examples of strata with vertical bedding; horizontal bedding that abruptly ceased into an unstratified zone; and cross-bedded sands with inclinations of up to 85 degrees. The lithologies present include: 18 percent red sandstone, 25 percent limestone, 42 percent locals, and 15 percent exotics.

It is hypothesized that a supraglacial lake developed with deposition of the lower sands. The lake abruptly drained causing frozen blocks of sediment to be rotated. Sedimentation continued around the rotated blocks depositing the ripple-drift sediments.

		Return to N. Main Street, turn left (south), proceed straight through North Norwich.
72.0	0.7	Junction Route 12. Proceed to the right (south) on Route 12.
73.2	1.2	Kame delta on right, with foreset beds dipping to the South.
76.2	3.0	Intersection Route 320. Proceed south on Route 12. Note exposure in kame terrace on right.
77.3	1.1	Junction Route 23. Proceed south on Route 12.
81.7	4.4	Good exposure in kame delta.
84.4	2.7	STOP 7. EMERSON PIT

STOP 7. This kame delta illustrates the variety of sediments and orientations in a delta deposited adjacent to an ice tongue. Refer to the topographic map for STOP 3.

		Proceed south on Route 12.
85.7	1.3	Intersection Route 220 in Oxford. Proceed south on Route 12.
86.7	1.0	Note ice contact kame delta to right. <u>Stop</u> if you have time.
88.0	1.3	View to the left (east) of valley plug that controlled the elevation of the deltas examined at STOPS 3 and 7.
93.6	5.6	Brisben.
99.3	5.7	Light at junction Route 206, Greene. Proceed south on Route 12.
100.7	1.4	View to left (east) to outwash units associated with Wheeler esker (STOP 2).
106.1	5.4	View east to Chenango Valley State Park area.
106.7	0.6	Enter Broome County.
107.3	0.6	Junction Route 79. Turn right (north) on 79.

109.1	1.8	STOP 8. TILL

STOP 8. The drift in this valley includes Olean-type till, Binghamton-type stratified units,and thick lacustrine clays. These units are interspersed between bedrock outcrops. Rotational slump blocks develop by a process of piping in the silt and clay beds.

		Turn around and proceed south on Route 79.
110.9	1.8	Junction Route 12. Turn right, proceed south on Route 12.
113.5	2.6	Good view to the south of Chenango River valley.
117.4	3.9	Junction Route 11. Proceed south on Route 12.
117.5	0.1	Entrance ramp to Route 81 south. Proceed south on Route 81.
120.7	3.2	Junction Routes 81 and 17. Bear right onto Route 17 (west).
123.9	3.2	Exit at Route 201 south (exit 70S). Proceed south on Route 201.
125.6	1.7	Junction Route 434. Follow signs to SUNY campus.
125.8	0.2	Entrance to SUNY-Binghamton.

SUBMARINE DISCONTINUITIES AND SEDIMENTARY CONDENSATION IN THE UPPER HAMILTON GROUP (MIDDLE DEVONIAN): EXAMINATION OF MARINE SHELF AND PALEOSLOPE DEPOSITS IN THE CAYUGA VALLEY

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INTRODUCTION

As defined by Jenkyns (1971) condensed sedimentary sequences are differentially thin deposits recording minimum rates of net sediment accumulation. Such sequences are shown through correlation to be greatly condensed equivalents of thicker deposits; these latter typically display convergence of coeval zones or beds along their lateral periphery. It is possible for stratigraphic sequences to thin from hundreds or even thousands of feet to tens of feet or less while preserving or partly preserving the internal sequence of fossil zones or lithologic subunits. Stable depositional settings such as oceanic-rises and seamounts, regions of the open ocean floor, and broad epeiric shelf areas are commonly characterized by condensed sedimentary sequences. Selective winnowing of fines by currents and bioturbation effects, periods of nondeposition, inorganic degradation and biocorrosion of carbonates, exhumation and reburial of concretions and fossils, and juxtaposition or commingling of fossil zones are all variably characteristic of such condensed sedimentary sequences. Association of these sequences with local and regional discontinuities is also common as will be shown for part of the New York Middle Devonian section in this paper.

The upper Hamilton Group (Ludlowville and Moscow Formations) of the Cayuga Valley includes a thick sequence of marine mudstone characterized by widely spaced, thin limestone units and numerous mappable fossil zones. Several of these are associated with a widespread paraconformity and condensed sedimentary sequence at the base of the Moscow Formation; the differentially thin strata include the Portland Point and Kashong Members which are most compressed stratigraphically in the southeast part of the Cayuga Valley. These units thicken northwestward across the Cayuga and Seneca Valleys, 9.5 feet of strata (Portland Point and Kashong Members) at the Portland Point type section expanding to 100-110 feet in the Canandaigua Valley. The Tichenor-Portland Point Members disconformably overlie the King Ferry Shale in the Cayuga Valley, although evidence of erosion is lacking in most outcrops.

The Canandaigua-western Cayuga Valley area was a region of differential subsidence (trough) during the Middle Devonian. This is evidenced at several levels by differentially thickened units (Jaycox, Tichenor,

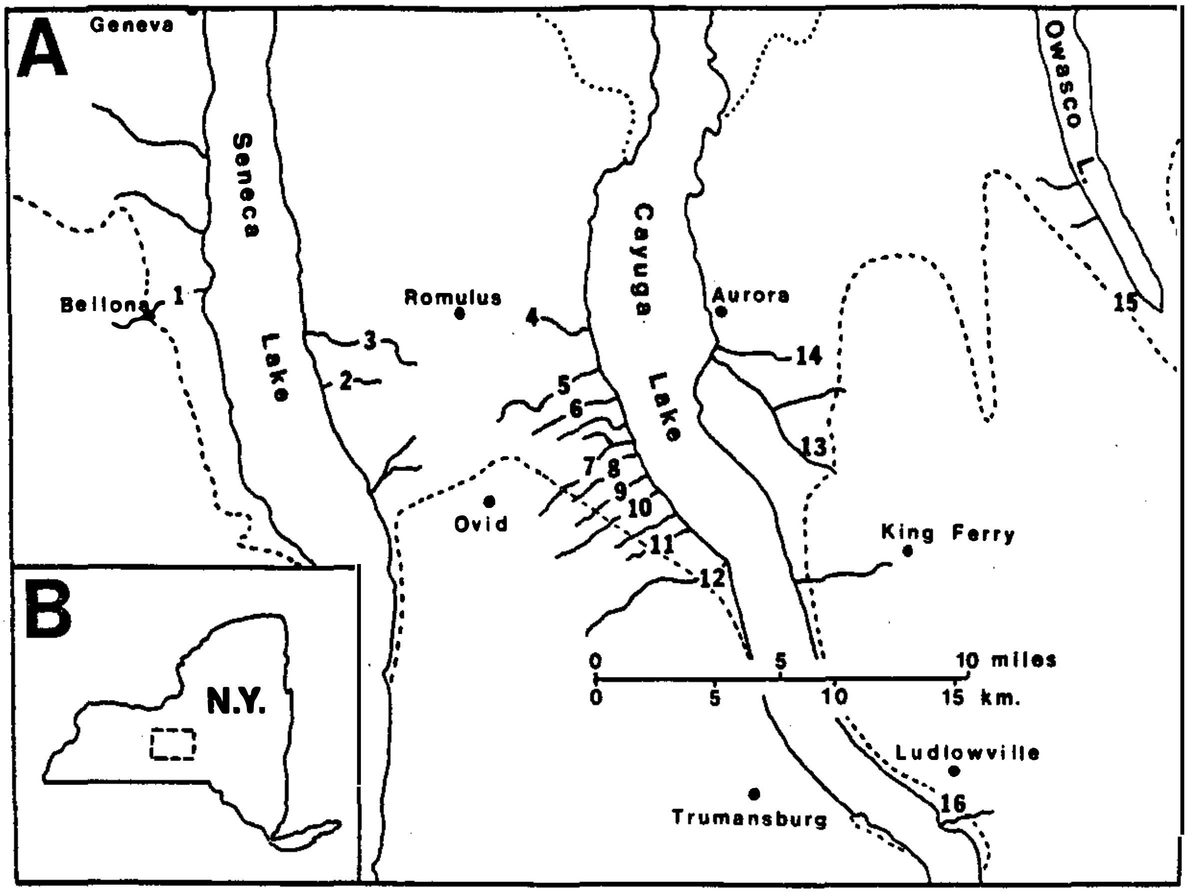


Figure 1. Study area. A is map of Seneca-Owasco Valley region. Figure shows top of Hamilton Group (dashed line) and base of Hamilton (dotted line). Numbered localities are discussed in text. Stops include: 1) Big Hollow Creek (Loc. 5: Stop 1); 2) Bloomer Creek (Loc. 8: Stop 2); 3) Barnum Creek (Loc. 9: Stop 3); 4) Sheldrake Creek (Loc. 12: Stop 4); and 5) Portland Point type section (Loc. 16: optional Stop 5). Inset B shows position of study area in New York State. Modified from Baird (1981).

Deep Run, Kashong Members), transition of calcarenites to argillaceous carbonate facies (Tichenor Member), and local development of black shale facies (Ledyard Member, parts of Wanakah-King Ferry Members 2 levels in Windom Member). Some of this shelf-to-trough facies change will be seen in the 4-mile interval of field trip stops 1-4 west of Cayuga Lake.

The lower (Wanakah-equivalent) part of the King Ferry Shale Member (Ludlowville Formation) displays northwestward facies change from fossil-rich, silty, gray mudstone to dark-gray and black fissile shale across the Cayuga Valley, apparently reflecting lateral paleoenvironmental transition to a poorly oxygenated, deeper water setting. Associated with an area of abrupt, narrow, gray-to-black shale transition west of Cayuga Lake are two diastems; these are characterized by abundant hiatus-concretions which are extensively bored and encrusted by organisms. Submarine erosion which produced the diastems is believed to have been the result of three processes: 1) weak, episodic wave and bottom currents; 2) disturbance and liquifaction of surface muds by infauna; and 3) sediment removal from an inclined sea floor (gentle submarine paleoslope) by current and/or gravity-induced transport. This best explains two closely-spaced, mirror-image discontinuities which display identical regional species-diversity gradients of associated organisms and which cut across a prominent facies change.

The lower Moscow condensed section, the above two diastems, and another discontinuity (phosphatic pebble bed) in the Kashong Shale Member are examined on the field trip. The area of field-trip stops (though small) shows several trends and features associated with the trough margin on the west side of the Cayuga Valley. The Portland Point type section on the east side of Cayuga Lake (Loc. 16), is also included as an optional stop in the road log.

STRATIGRAPHY

The Middle Devonian (Givetian) Hamilton Group is an eastwardly thickening wedge of predominantly terrigenous sediment which is entirely marine except in its easternmost parts. This sequence, ranging from a minimum thickness of 280 feet at Lake Erie to 3,000 feet at the Catskill front, is characterized by detrital sediments which coarsen to the east and southeast. In western New York and through most of the study area (Figure 1) the Hamilton Group is comprised of gray to black marine shale and mudrock with widely spaced thin calcareous intervals. In central and eastern New York shale and limestone are displaced by siltstone and sandstone facies with the appearance of fluvial sandstones and redbed floodplain deposits east of the Schoharie Valley.

The deposits in the study area display marked regional facies change across the Seneca-Owasco Valley region with much of this lateral change occuring in the area of closely spaced numbered localities west of Cayuga Lake (Figure 1). The sedimentary section examined includes the upper one-half of the Ludlowville Formation and the lower 40 feet of the Moscow Formation (Figure 2). These units are moderately to highly fossiliferous and several major units and zones marked by diverse faunas are included. Widespread faunal marker units include: 1) the Tichenor Limestone Member with its diverse and distinct coral, bryozoan, pelmatozoan biota, and 2) the *Rhipidomella*, *Centronella*, *Spinocyrtia* bed (Kashong Member) with its distinctive brachiopod-bryozoan assemblage. Another possibly traceable fossil zone is the Bloomer Creek bed (King Ferry Member) which yields numerous *Pleurodictyum americanum* and *Strophodonta demissa*. This diastem, though a local feature, is at the appropriate stratigraphic level (uppermost Wanakah-equivalent King Ferry Member) to be stratigraphically coeval to the *S. demissa* zone in the upper Wanakah Member of western New York.

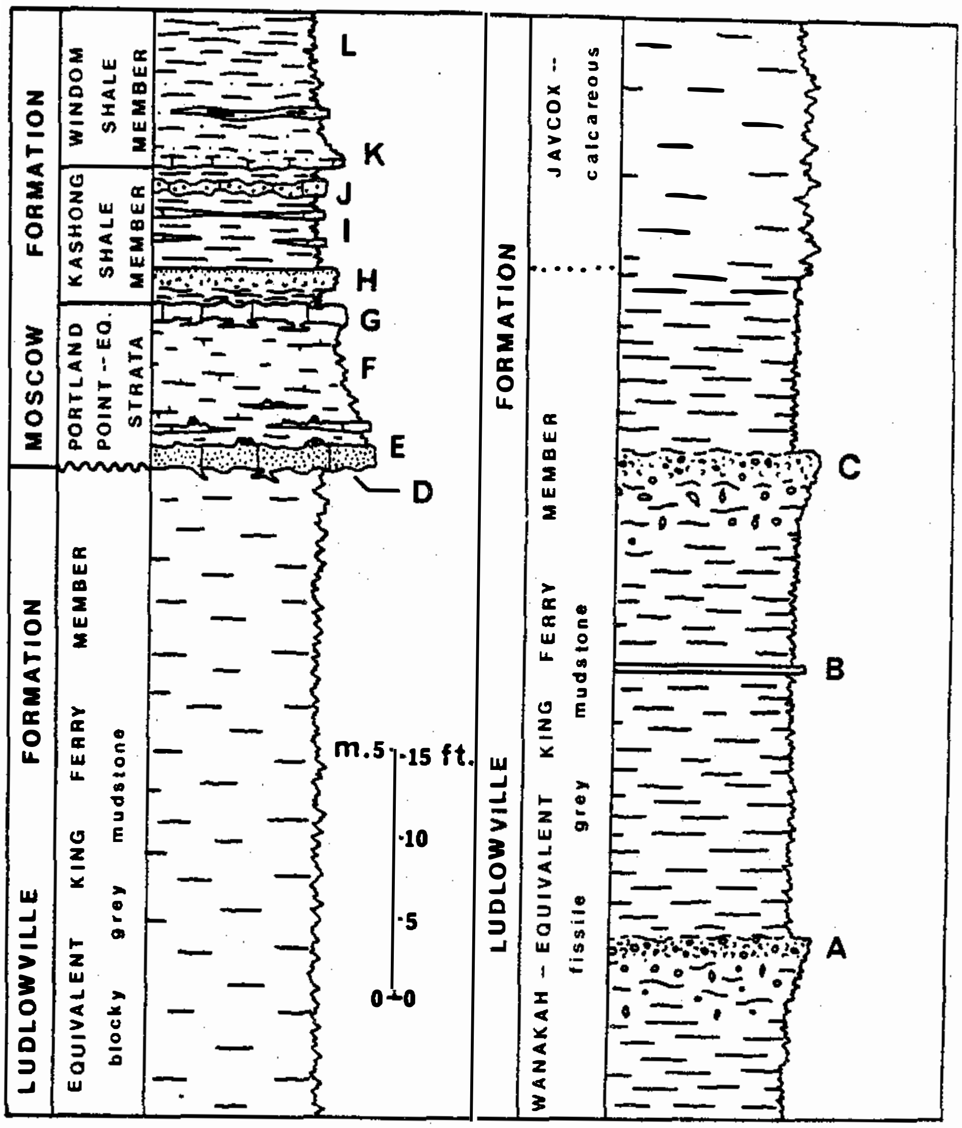


Figure 2. Generalized Upper Ludlowville-Lower Moscow stratigraphic section at Barnum Creek (Loc. 9: Stop 3). Numbered units include: A) Barnum Creek bed; B) Mack Creek turbidite bed; C) Bloomer Creek bed; D) Basal Moscow paraconformity; E) Tichenor Limestone Member; F) Deep Run Shale Member; G) Menteth Limestone Member; H) Kashong Member: Rhipidomella-Centronella bed and basal interval of gray Kashong mudstone; I) Kashong Member: upper Kashong mudstone interval; J) Phosphatic pebble bed (discontinuity); K) Windom Member: basal silty mudstone interval; L) Windom Member: gray, fissile shale of Ambocoelia umbonata-chonetid zone.

DEPOSITIONAL SETTING

Western New York Shelf

The Hamilton sequence of New York includes the northernmost Middle Devonian deposits developed within the Appalachian basin; these sediments accumulated in the northern arm of an inland sea, the deepest part of which was developed south-southwest of the study area in western Pennsylvania, eastern Ohio, and West Virginia (Dennison and Head, 1975; Dennison and Hassan, 1976). The northern and western boundaries of the basin bordered lower relief cratonic shelf regions; these latter areas supplied relatively little detrital sediment to the basin with resultant thin Middle Devonian sections in western New York (Dennison and Head, 1976). A broad, gently-sloping shelf was developed across most of central and western New York during Hamilton deposition (Cooper, 1957; Grasso, 1970, 1973; Heckel, 1973). A shallow sea with a level muddy bottom prevailed over most of this region (Cooper, 1957; McCave, 1967, 1973).

The upper Hamilton formations are composed largely of detrital sediments and include several upward-regressive hemicycles which record basin-filling and general westward migration of the eastern shoreline punctuated by eastward shoreline advance associated with episodic transgressions and/or periods of reduced sediment supply. This westward-thinning detrital sequence is a result of Acadian tectonic events, including uplift and erosion to the east and southeast (Cooper, 1957; Heckel, 1973; Oliver, 1977). It is the initial sedimentary unit of the Catskill delta complex which expanded greatly during the Late Devonian (see Rickard, 1975, for summary of stratigraphy and facies).

Hamilton sediments are characteristically fossiliferous; the rich biotas of the Cayuga Valley section have been described by paleontologists over a period of more than 100 years (Hall, 1843; Cleland, 1903; Cooper, 1929, 1930; Fernow, 1961). The Hamilton sea apparently had near-normal salinity, water temperatures, and circulation as evidenced by presence of diverse stenotopic organisms. However, well-developed faunal endemism in the Hamilton biota suggests partial physical isolation of this sea from "Old World" circum-oceanic waters (Oliver, 1976, 1977).

Inter-Shelf Trough

The western New York shelf was bisected by an actively subsiding trough in the Canandaigua-Cayuga Valley region and particularly at the Seneca Valley meridian. Differential subsidence took place resulting in locally thickened and/or anoxic deposits. This is best illustrated by great local stratigraphic expansion of basal Moscow sediments equivalent to both the Portland Point Member of east-central New York and to the thin Tichenor-Menteth sedimentary sequence in western New York (Figure 3). Nine feet of strata at the Portland Point type section (Figure 4) expand to more than 100 feet of section in the Canandaigua Valley. The Tichenor Limestone Member and lower (Tichenor-equivalent) Portland Point Limestone are thin, widespread calcarenitic units. In central and western New York, respectively, these coeval beds can be

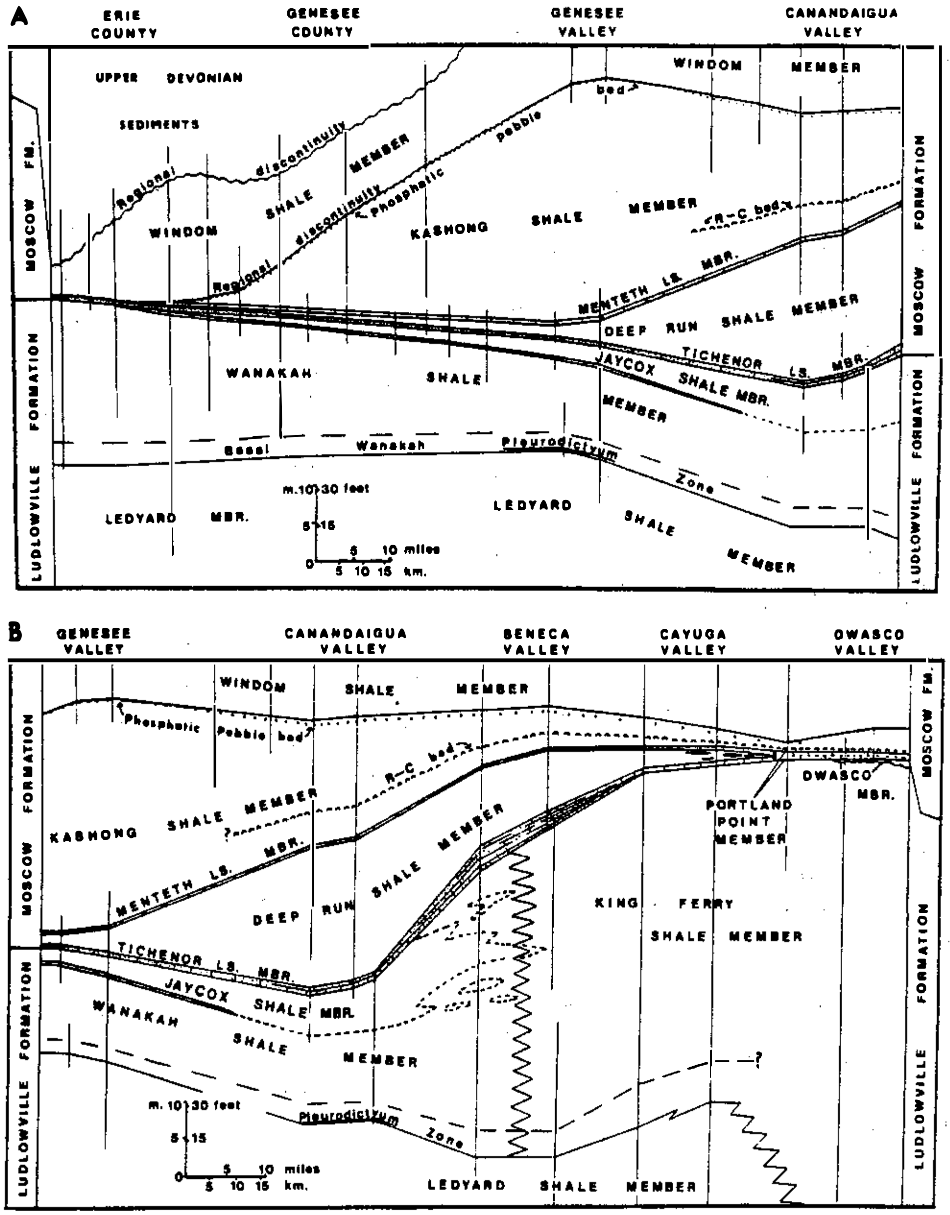


Figure 3. Upper Ludlowville-Lower Moscow stratigraphic relationships observed in present study; A) in western and B) in west-central New York. Vertical lines denote sections. From Baird (1979).

followed from opposing directions into the Seneca Valley where the equivalent sequence is a thicker (10-12 feet) sequence of interbedded calcareous mudstone and fossiliferous argillaceous limestone (Figure 3). The Tichenor and Menteth Members diverge, as the Deep Run Member expands into a thick lentil of calcareous gray mudstone. Moreover, the Jaycox Shale Member thickens eastward from a "feather-edge" carbonate bed in Erie County to 50-60 feet of gray calcareous mudstone in the Seneca-Cayuga Valleys; this sequence apparently thins eastward across the Cayuga Valley (Baird, 1981), but its eastern extremity is not known (Figure 5).

The basal Moscow paraconformity (base of Tichenor Limestone) is visible in all area sections. However, west of Cayuga Lake and in the Seneca Valley, this hiatus is believed to be of lesser magnitude than to the east or west due to differential subsidence and nearly continuous sedimentation in the region. This is suggested by maximum thickness of Jaycox-equivalent strata between Seneca and Cayuga Lakes (Figure 5); this sequence thins both to the east and west with probable erosional loss of beds along the paraconformity.

Black-shale facies displace equivalent gray mudstone and siltstone sequences within parts of the Ludlowville and Moscow Formations mainly in the Seneca Valley. In the Ledyard Member (Ludlowville Fm.), a black-shale facies in the Seneca-Cayuga Valley region grades westward into gray mudstone west of the Canandaigua Valley (McCollum, 1980) and eastward into gray mudstone and siltstone in the Owasco-Skaneateles Valley region (Cooper, 1930; Smith, 1935). In the Wanakah Member and Wanakah-equivalent part of the King Ferry Member, there is a similar lentil of black shale which occupies most of the Wanakah interval at Kashong Glen (Figure 1, Loc.1); this tapers and grades into gray mudstone both to the east and west of the Seneca Valley (Baird, 1979). A lentil of black platy shale (Crurithyris praeumbona Zone) is locally developed in the middle of the Windom Shale Member (Moscow Fm.); this grades westward to gray mudstone between the Seneca and Genesee valleys (Baird and Brett, in prep) and eastward to silty mudstone facies of the Cooperstown Member in central New York. At the top of the Windom, black shale yielding Allanella, "Leiorhynchus," and C. praeumbona reaches its greatest development in the Seneca-Canandaigua Valley region.

Not all units appear to exhibit differential subsidence in the region. The Mententh Limestone Member, Rhipidomella-Centronella bed (Kashong Member), and Tully Formation extend across the region with little or no evidence of change in sediment character or thickness. Thus, the subsidence is believed to have been gentle and episodic; this was a local trough or "saddle" as opposed to a major basin. Deeper water (black-shale) facies in the trough may be a northward extension of more basinal-type sediments (e.g. Millboro Shale) typical of contemporaneous deposits in the central Appalachian basin. Finally, the axis of maximum subsidence was not entirely stationary; a general westward migration of the trough is apparent in the Jaycox-Kashong interval (Figure 3).

Submarine Paleoslope

Associated with facies change across the Cayuga Valley within the medial King Ferry Member is evidence of a gentle but significant submarine slope. This northwestward facies transition from shallow subtidal, fossil-rich, shelf deposits in the Aurora, King Ferry, and

Ludlowville area to deeper water, dark-gray and black shale in the Romulus-Bellona area (see Figures 1, 5) is characterized by presence of a mappable turbidite bed and evidence of local submarine erosion (Baird, 1981). This erosion, expressed by the presence of two unusual diastems in the area of most abrupt facies change, between localities 5 and 11, occurred through combination of physical and biological processes on a sloping sea bed. These diastems, both commencing in shallow-water facies and terminating within dysaerobic-anaerobic sediments, are key correlative horizons allowing time-controlled study of faunal gradients developed across the shelf-to-trough interval.

SEDIMENTARY CONDENSATION-DISCONTINUITIES:

LOWER MOSCOW FORMATION

Basal Moscow Paraconformity

The Moscow Formation is bounded at the base by a discontinuity below the Tichenor Limestone, which is traceable from Lake Erie to eastern New York (Baird, 1979). This hiatus is considered to be a paraconformity. In outcrop there is no observed angularity of Ludlowville beds beneath the break,but regionally there appears to be evidence of erosional overstep of Jaycox beds, particularly to the west.

In Cayuga Valley sections this discontinuity is conspicuous and can be seen in planar view in overhangs associated with waterfalls. At Bloomer Creek (Loc. 8: Stop 2), Barnum Creek (Loc. 9: Stop 3), Sheldrake Creek (Loc. 12: Stop 4), and at the Portland Point type section (Figure 4, Loc. 9: Stop 5) it can be easily observed; the break occurs beneath calcarenitic facies of the Tichenor Member and of the coeval lower Portland Point Member.

The paraconformity is marked by burrow networks (hypichnia) which locally penetrate into the uppermost King Ferry mudstone. These are somewhat analogous to, though smaller than, burrows observed along the base of the Tully Formation (Heckel, 1973). Reworked shale chips and intraclasts are rare in the basal calcarenite sequence in Cayuga Valley sections. However, further east in the Owasco Valley (Figure 1, Loc. 15), intraclasts occur in cross-bedded lower Portland Point calcarenite; erosional truncation of more coherent sediments in the Owasco Valley and further east probably explains the numerous sediment clasts in the Portland Point, this contrasting with submarine erosion of incompetent mud in Cayuga Valley localities.

Portland Point Member and Coeval Beds

East of the Cayuga Valley basal carbonate beds of the Moscow Formation are represented by the Portland Point Member (Cooper, 1930; Baird, 1978); this member is remarkably widespread, being traceable in outcrop from the Portland Point type section (Figure 4, units 2, 3) almost to the Catskill Front (McCave, 1967, 1973). In marked contrast to this uniformity, the Portland Point stratigraphic interval and that of the overlying Kashong Member expand greatly to the west-northwest across the Cayuga and Seneca Valleys (Figure 3). As the Portland Point

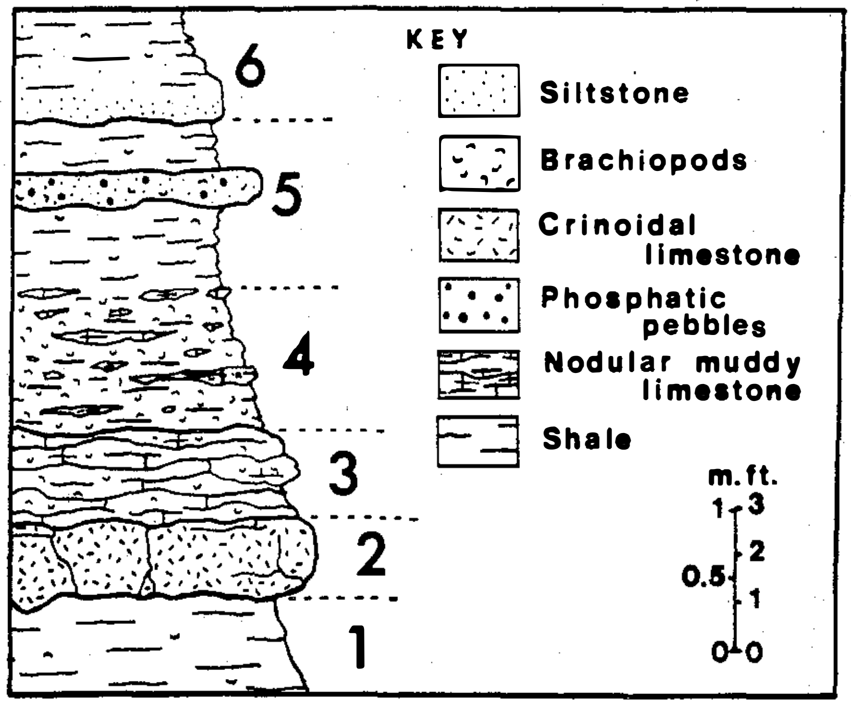


Figure 4. Diagram of Portland Point type section (Loc. 16: optional Stop 5). Units include: 1) Uppermost Ludlowville (King Ferry) shaley mudstone; 2) basal Portland Point (Tichenor-equivalent) calcarenitic bed; 3) Upper Portland Point (Deep Run-Menteth equivalent) muddy limestone; 4) Lower Kashong Rhipidomella-Centronella Bed; 5) Upper Kashong phosphatic pebble bed and synjacent shale; 6) basal siltstone bed and succeeding shales of Windom Member. Units 2 and 3 are Portland Point Member. Units 4 and 5 comprise Kashong Member. From Baird, 1979.

thickens westward from its 4-foot-thick type profile, it becomes differentiable into three distinct members; these are in ascending order: 1) Tichenor Limestone Member; 2) Deep Run Shale Member; and 3) Menteth Limestone Member (see Figure 2, 3). At the Portland Point type section, the Deep Run and Menteth members are undifferentiable, the equivalent interval (Figure 4, unit 3) being expressed as a 2-foot thick, shell-rich argillaceous limestone. At Barnum Creek (Loc. 9, Stop 3), the section shown in Figure 2, Portland Point-equivalent beds include 10 feet of section with the three component members represented. At Localities 7 and 8 (Stop 2), 1.5 miles farther north, this sequence is closer to 15 feet in thickness. The greatest expansion of this interval is between Localities 8 and 1 (see Figures 1, 3), but the basic trend is clearly visible in the field-trip area.

In the Canandaigua Valley the Portland Point-equivalent interval reaches a maximum thickness of 60 feet with the vast bulk of the sequence being composed of Deep Run mudstone. Farther west the Deep Run thins greatly, to a single calcareous mudstone band 1-3-feet-thick, between the Tichenor and Menteth members west of the Genesee Valley (Figure 3A). Thus, a nearly mirror image of sedimentary convergence and condensation is observed with sediments coeval to the Portland Point west of the axis of differential subsidence. The lateral tapering of the Deep Run detrital wedge, both to the east and west, points strongly to the pattern of contemporaneous sedimentary condensation in similar but separate shelf areas.

The Tichenor Member is a conspicuous, though uniformly thin, crinoidal limestone west of the region of differential subsidence. In the Seneca Valley and northwest Cayuga Valley, it is thicker and more argillaceous than in western New York counties. At Kashong Glen (Loc. 1) it consists of interbedded calcareous, gray mudstone and argillaceous biomicritic limestone (Baird, 1979). The fauna is rich and distinctive, being composed of sponges, large rugose corals, fistuliporid and fenestrate bryozoans, diverse brachiopods, and numerous pelmatozoan taxa. Although this diverse fauna is best preserved in the Seneca Lake area, the biota can be followed as a distinct zone both to the east and west. At Bloomer Creek (Loc. 8: Stop 2) the upper part of the Tichenor yields large, characteristic pelmatozoan stems which are massive; these bear pseudocirral protuberances, which apparently acted as attachment structures. These "runners" occur with a rich association of bryozoans, brachiopods, and camerate and inadunate crinoids. Between this locality and Sheldrake Creek (Loc. 12: Stop 4), the Tichenor becomes increasingly calcarenitic and massive, and the distinctive faunal zone at the top of the member largely disappears due to increased condensation and sedimentary disturbance at this level.

The Tichenor calcarenitic band can be followed into central New York as the lower bed of the Portland Point (Baird, 1979). It thins to a discontinuous sheet of calcarenite above the Owasco Sandstone Member east of the Owasco Valley but remains characteristic of the basal Portland Point at least to the Chenango Valley, if not farther east.

The Tichenor-basal Portland Point calcarenite is interpreted as a moderate- to high-energy deposit marking the initial deposit of marine transgression. It overlies the widespread paraconformity which apparently marked even shallower erosive conditions terminating Ludlowville deposition. It is envisioned as a wave-worked, shallow-subtidal sheet-sand facies typical of very shallow conditions on a broad shelf. Pelmatozoan debris comprising the calcarenite is believed to be variably reworked and transported.

The Deep Run Shale Member is a detrital wedge of hard, dark gray, calcareous mudstone, gradational from the distinctive pelmatozoan-coral zone of the upper Tichenor, and at the top, into the Menteth Limestone Member. Although it reaches greatest thickness in the Canandaigua Valley (Figure 3), it is still a prominent unit in the Seneca Valley.

In the Cayuga Valley it can be seen to best advantage at Bloomer Creek (Loc. 8: Stop 2),where it is approximately 10 feet thick. At Sheldrake Creek (Stop 4) it is only 4.5 feet thick and distinctly richer in bioclastic hash and shell debris. The Deep Run lacks any particularly distinctive taxa although it is rich in bryozoans, brachiopods, and bivalves, particularly in the thinner, condensed-edge facies.

The Menteth Limestone Member is alimestone bed 1.5 to 2 feet thick which extends from eastern Erie County eastward into the northern part of the Cayuga Valley and to one exposure (Ensenore Ravine) in the Owasco Valley. It thins southeastward across the Cayuga Valley from 1.5 feet at Mack Creek (Loc. 7) to 0.5 feet at Sheldrake Creek (Loc. 12) and 0.4 feet at Paines Creek (Loc. 13); this may be partly due to sedimentary condensation (thinning) or to submarine truncation. It is not recognizable as a discrete part of the type Portland Point but may be correlative with the uppermost part of the member (Figure 4, unit 3). The Menteth similarly disappears southward along the Owasco Valley as the Portland Point thins, and it is not recognizable at the N.Y. Route 38 roadcut section (Loc. 15) at Cascade.

The Menteth is composed of intensely bioturbated argillaceous limestone which weathers to buff-colored irregular ledges in outcrop. It is commonly gradational with the underlying Deep Run Member; elsewhere prod burrows may extend down from the limestone into the underlying mudstone. Unlike the Tichenor Member, this unit is predominently calcisiltite, composed of comminuted skeletal material, variable amounts of terrigenous silt, and larger brachiopod and trilobite debris, rather than pelmatozoan calcarenite. At Mack and Bloomer Creeks (Loc. 7, 8), the Menteth contains large fragments of the trilobites, Phacops and Dipleura, but these are difficult to extract from the matrix. Farther west, in the Genesee Valley, the Menteth has yielded silicified fossils; delicate juvenile stages of several brachiopod genera have been obtained in siliceous residues from this unit (Clarke and Luther, 1904). Siliceous masses occur locally in the Menteth Limestone of the Cayuga Valley and similar small silicified fossils may be plentiful.

The Menteth is characterized by strikingly uniform thickness regionally, and it extends across the intershelf trough with no significant change in character. Clues to its origin include the general similarity of Deep Run and Menteth fossils and the gradational boundary between the members. The Deep Run apparently represents local rapid detrital infilling of a deeper, lower energy trough on the New York shelf. As sediment input exceeded subsidence, the depression was filled to the ambient level of the surrounding shelf such that the seabed had migrated into a slightly higher energy regime. At a critical depth level, sediment input would have equalled sediment removal through winnowing by currents, this winnowing being selectively aided by intense bioturbation and liquification of surface muds by organisms (see Rhoads, 1970; Rhoads and Young, 1970). Fine clay would have been removed leaving behind a thin lag mantle of intensely burrowed silt and shell debris which would later be expressed as the Menteth. This interpretation holds that the Menteth was a terminal depositional phase of the Deep Run, representing condensation following trough infilling. Such an explanation could account for its widespread thin character and uniformity.

Kashong Shale Member

The Kashong Member is best developed in the Genesee Valley-Livonia area where it is more than 80 feet thick; it thins laterally to the east and west in a pattern very similar to that of the Deep Run Member, i.e. it is a lens-shaped unit with tapering margins (Figure 3). This unit reaches maximum thickness well to the west of the region of maximum thickness of the Deep Run, indicating that subsidence and associated fine mud deposition had shifted westward along the New York shelf.

The Kashong consists of gray, calcareous mudstone with numerous shell beds and lenses of concretionary limestone. It is characterized by diverse benthos (about 60-70 genera) of brachiopods, bryozoans, and pelmatozoans. Characteristic taxa include *Tropidoleptus carinatus*, *Pleurodictyum americanum*, *Orthonota undulata*, and *Dipleura dekayi*. Crinoid colonies occur at several levels with numerous genera represented. In the Cayuga Valley, the Kashong is greatly condensed, and the diverse, delicate fauna typical of sections to the west is somewhat reduced. However, at localities 8 and 12 (Stops 2, 4) many fossil genera including brachiopods, bivalves, and trilobites can be found in this unit. Characteristic *Tropidoleptus*, *Dipleura*, and pentagonal crinoid columnals can be collected on the field trip. The Kashong is only 7-9 feet thick at localities 8, 9, and 12 (Figure 2), and it thins further to 3 feet at the Portland Point type section (Figure 4); however, it is particularly fossil-rich and contains both the distinctive *Rhipidomella-Centronella* bed and the phosphatic pebble (discontinuity) bed (Baird, 1978, 1979; see Figures 2, 3).

The *Rhipidomella-Centronella* ("R-C") bed is a prominent shell-rich unit which is traceable from at least Menteth Glen in the Canandaigua Valley eastward into central New York where it constitutes a basal thin zone of the Cooperstown Member. From the Canandaigua Valley to the Owasco Valley, this unit is of nearly uniform 1.5- to 3-foot thickness. East of the Owasco Valley it gradually thins to a single smear of characteristic brachiopods and pelmatozoan debris and disappears as a recognizable marker east of the Cazenovia-De Ruyter meridian. At Menteth Glen, the R-C bed is separated from the Menteth Member by 10 feet of *Tropidoleptus* and bivalve-rich gray mudstone. This mudstone thins eastward as the R-C bed and Menteth converge; it is only 5-7 feet thick in the Seneca Valley and 0-1.5 feet between localities 7 and 12 in the Cayuga Valley. From Sheldrake Creek (Loc. 12) southeastward, the R-C bed rests directly on Menteth-Portland Point carbonates.

The R-C bed contains a distinctive macrofauna of brachiopods, including the terebratulids, (*Centronella* and *Cryptonella*), the orthid, *Rhipidomella*, and the large spirifer, *Spinocyrtia*, which are all rare to absent in adjacent Kashong beds. The brachiopods are packed in with abundant bivalves and bryozoan debris. Large corals such as *Heliophyllum* and *Favosites* occur rarely in the unit, and unusual taxa such as rostroconchs and a flexible crinoid have also been found. Of particular importance is distinctive mechanical abrasion (faceting) of larger brachiopods which often have the shell worn through and surface detail completely removed. The R-C bed is usually capped by a thin bed,1-2 inches thick, of pelmatozoan-ossicle packstone.

The R-C bed apparently represents a regressive episode within the Kashong; in sections to the west of the Cayuga Valley, shell beds of *Tropidoleptus* and other debris become thicker and more closely spaced immediately below the unit, suggesting a shoaling trend. The presence of large corals and thick-shelled brachiopods indicates further regression and establishment of equitable bottom conditions. Mechanical wear on the brachiopods is strong evidence of wave action and disturbance, the climax phase of this regression is expressed by the thin calcarenitic interval capping the bed.

The phosphatic pebble bed of the uppermost Kashong Member is indicative of a widespread shale-floored discontinuity which is best developed in western New York (Baird, 1978). This discontinuity, marked by a bed 3 to 10 inches thick of phosphatic pebbles and reworked shell debris, is traceable from the Cayuga Valley westward to Lake Erie (Figure 3). West of the Canandaigua Valley the break marks the top of the Kashong Member; it occurs at the boundary of the *Tropidoleptus*-rich fauna of the upper Kashong and *Devonochonetes-Ambocoelia*-rich shale of the lower Windom Member. From the Canandaigua Valley east to the Portland Point type section the discontinuity occurs approximately one foot below a calcareous siltstone bed which marks the base of the Windom (Baird, 1978). The intervening mudstone interval is characterized by a mixture of Kashong and Windom-type fossils. Again, this unit contains unusual faunal elements like those of the R-C bed including rare large rugose corals and faceted valves of *Spinocyrtia*, suggesting a second regressive episode.

From the Canandaigua Valley westward, the bed is rich in phosphatic pebbles and phosphatic fossil steinkerns of Kashong taxa. Some of these are admixed with and encrusted by Windom epizoan taxa (Baird, 1978), indicating that submarine erosion and exposure of the pebbles had continued up to and past the time of initial colonization by Windom organisms. In the Seneca and Cayuga Valleys, the change of faunas is less distinct as the discontinuity becomes less pronounced, but epizoans (*Spirorbis*, bryozoans) are commonly observed on reworked phosphatic material.

Sediment mixing by infauna characterized the erosion process; burrowing and sediment churning by *Zoophycos* organisms caused vertical mixing of shells and nodules such that the discontinuity shows up as a bed rather than as a discrete break. Such a bed strongly resembles a condensed sediment interval, the only difference being evidence of erosional overstep of underlying sediments, which is expressed in westward erosional truncation of the Kashong sequence in Genesee and eastern Erie Counties (Baird, 1978). Both at and east of the Cayuga Valley erosion appears to have been minimal or absent, and the phosphatic pebble bed is more truly a condensed sediment interval.

This bed can be seen at Bloomer and Sheldrake Creeks (Stops 2 and 4) and at the Portland Point type section (optional Stop 5). Phosphatic pebbles are present but not very common in Cayuga Valley sections, and they are hard to extract from the silty, calcareous mudstone matrix. Tropidoleptus is common in the bed; these brachiopods should be examined for phosphoritic interior fillings. Some of these brachiopods are worn and disarticulated, and the black phosphorite is clearly visible showing through the shell or adhering to valves (see Baird, 1978). This indicates that phosphatization occurred within near-surface sediment, often within sediment-infilled shells. Steinkerns of aragonitic shells are commonly observed indicating that shell carbonate was not replaced by phosphorite. Interstitial phosphorite formation appears to have been associated with prolonged intervals of nonsedimentation, sediment mixing, and disturbance prior to submarine erosion.

STRATIGRAPHIC SIGNIFICANCE OF BASAL MOSCOW CONDENSATION AND DISCONTINUITIES

The Portland Point-Kashong complex of beds and zones displays the greatest aggregate convergence of strata in the western New York Middle Devonian. Moreover, the nearly mirror-image thinning, convergence, and truncation of the same strata in western New York clearly point to a major, long-lasting geologic event or closely spaced series of events affecting virtually the entire New York shelf.

The pattern of convergence of many different units and juxtaposition of widespread thin limestone and shell beds indicate that the time interval represented by the basal Moscow sequence and the underlying paraconformity is probably very great. The Portland Point interval in both western and eastern New York appears to represent widespread and greatly prolonged periods of nondeposition of terrigenous sediment in the shelf region, allowing for widespread, uniform stacking of thin carbonate units of significantly different ages. As noted above, this apparent "sediment starvation" may have resulted from selective winnowing of fine-grained sediments in the very shallow shelf area. The thin carbonates grade into relatively thick clastic lentils (i.e. Deep Run and Kashong members) which were deposited rapidly in the slightly deeper, lower energy subsiding trough. It is in the inter-shelf trough that the sequence shows its great time significance by the stratigraphic "ballooning" described earlier.

This sequence of condensed beds should recur in equivalent strata elsewhere around the Appalachian basin, but perhaps not in the basin center where subsidence was greatest. It should be a basin-margin feature associated with one or more discontinuities, coral-pelmatozoan rich beds, or unusual fossil zones.

DIASTEMS IN KING FERRY SHALE MEMBER

Barnum Creek Bed

Two diastems, characterized by abundant hiatus concretions and fossil shell debris, occur in the Wanakah-equivalent part of the King Ferry Member, these are best developed between Powell Creek (Loc. 11) and Mack Creek (Loc. 7). The lower diastem,represented by the Barnum Creek bed (Baird, 1981), is a subtle hiatus lacking any discrete discontinuity surface (Figure 2, 5B). It can be easily overlooked except for the presence of bored and bioencrusted reworked concretions (Figure 6b). From Powell Creek (Loc. 11) and Short Creek (Loc. 14) southeastward, this bed is characterized only by shell debris which is variably biocorroded (Figure 3b). Abundant fossils including the coral, Stereolasma; brachiopods, Athyris spiriferoides, Mediospirifer audaculus; bryozoans, and bivalves characterize this fauna. The mudstone sequence underlying the Barnum Creek bed yields numerous in situ calcareous concretions as well as macrofauna similar to that in the overlying shell bed. Large conulariids, associated with and infilled by concretionary calcareous mudstone, occur within this sequence; these "pre-fossilized" objects appear within the Barnum Creek bed north of locality 11 as the underlying conulariid zone is overstepped.

From Groves Creek (Loc. 10) northward, the Barnum Creek bed is characterized by an interval 5 to 10 inches thick of mixed shells, mud, and hiatus concretions. North of Barnum Creek (Loc. 9, Stop 3) this bed thins, and fossils decrease in abundance and diversity northward to Hicks Gully (Loc. 4) and westward to Sampson State Park (Locs. 2 and 3) in the Seneca Valley. This occurs as synjacent sediments become darker in color and distinctly less fossiliferous, reflecting the regional shelf-to-trough facies transition discussed earlier.

Bioencrusted hiatus concretions and reworked conulariids are characteristic of this unit at Barnum Creek (Stop 3) and Groves Creek (Loc. 10); these objects show a complete spectrum of degradation from minimally or partially exhumed forms to intensely bored and bioencrusted fragments (Figure 5B, 6). Small rugose and auloporid corals (Stereolasma, Cladochonus), bryozoans (Ascodictyon, Hederella, encrusting trepostomes), and pelmatozoans are characteristic encrusters (Figure 6). Borings are predominantly a vasiform morphotype of Trypanites. Numerous parallel and intersecting sets of grooves on concretions may be scratch marks produced by burrowing infauna impinging against disturbed, but not yet exhumed, concretions. These marks are similar to those produced by infauna on Jurassic nodules (Fürsich, 1979). Pre-fossilized conulariids are characteristic of the reworked sediment interval; these are typically three-dimensional and infilled by concretionary cemented mudstone (Figure 6). These are reworked as evidenced by variable degradation of thecae and bioencrustation of exposed concretionary mudstone. North and west of Mack Creek (Loc. 7), epizoans are still observed on hiatus concretions, but epizoan diversity is low overall. At Sampson State Park Locs. 2, 3) only Cladochonus and Ascodictyon are observed on concretions. The hiatus concretions themselves tend to be small, flattened objects reflecting diagenesis in darker, underlying mudstone facies; these are sometimes stacked like shingles along the bed. West of Seneca Lake (Loc. 2) the Barnum Creek bed is no longer recognizable, its position is occupied by dark-gray to black platy shale.

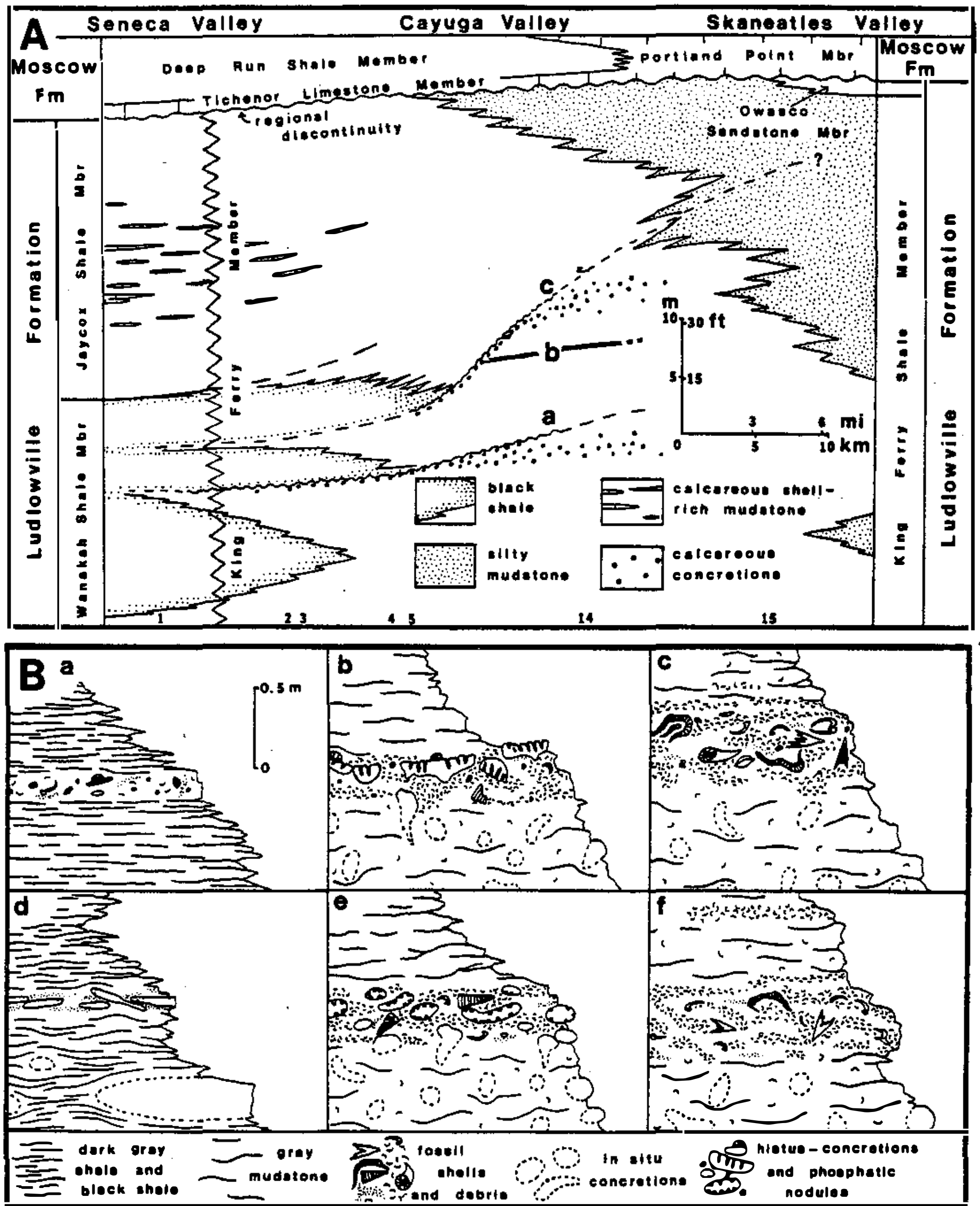


Figure 5. Stratigraphy associated with King Ferry diastems. A shows Barnum Creek bed (a), Mack Creek bed (b), and Bloomer Creek bed (c) relative to regional facies distribution. Note truncation of Mack Creek bed by Bloomer Creek bed. B, schematic vertical profiles of King Ferry diastems. a-c show Bloomer Creek bed at localities 5, 9, and 13 respectively, d-f show Barnum Creek bed at localities 2, 9, and 15. Localities in A correspond to those in Figure 1. From Baird, 1981.

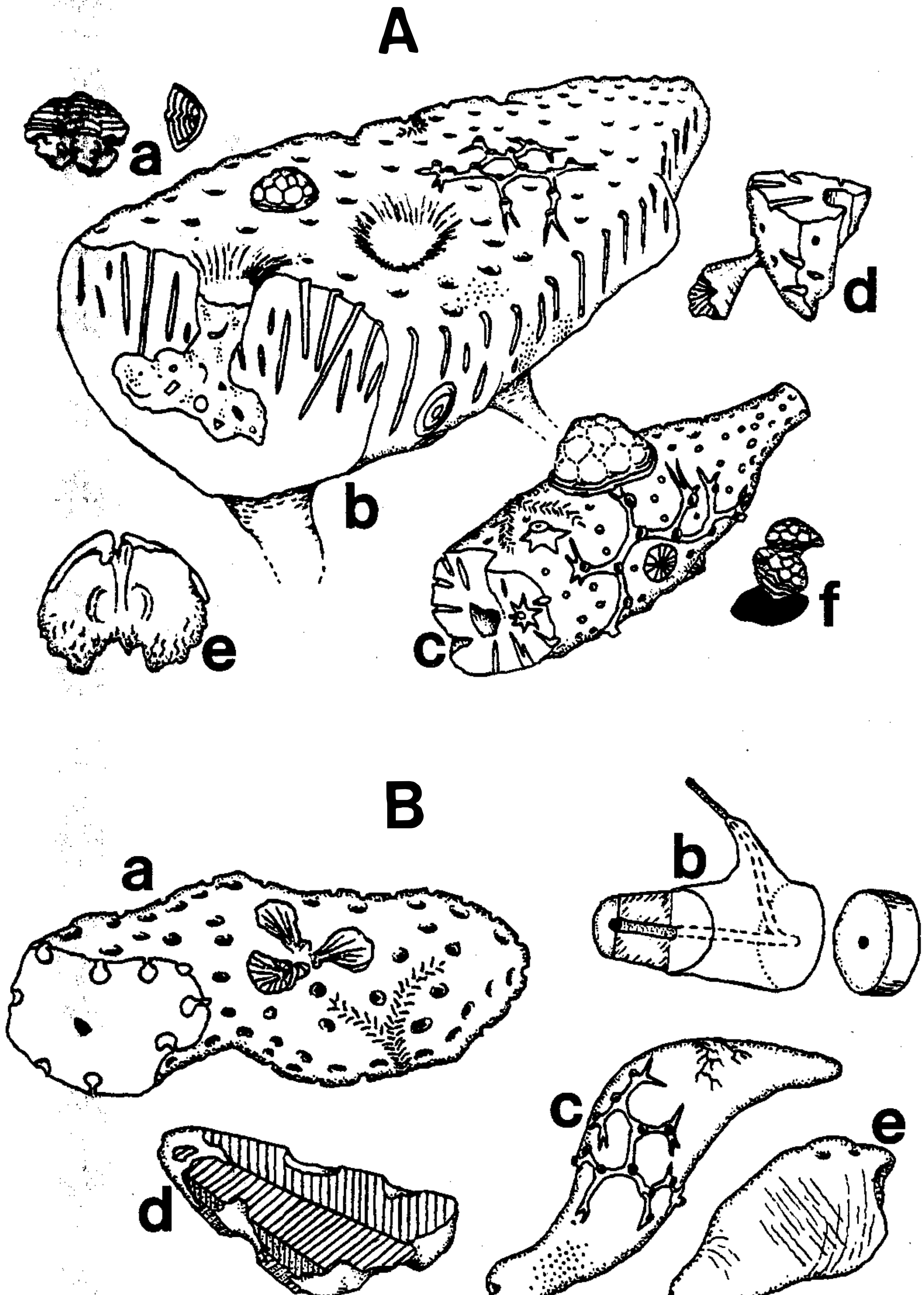
Bloomer Creek Bed

Twelve to thirty feet above the Barnum Creek bed another more pronounced discontinuity is observed; although lacking a discrete erosion surface, the Bloomer Creek bed displays dicontinuity character mainly from Groves Creek (Loc. 10) northward to Big Hollow Creek (Loc. 5; Stop 1). It too, is characterized by vertically-mixed hiatus concretions and shells in the same general area as the Barnum Creek bed. From Powell Creek eastward and southward, the bed generally lacks reworked early-diagenetic material and is characterized by abundant brachiopod and bivalve shells irregularly packed in gray mudstone matrix. In situ concretions are common in the 3-to-6 foot mudstone interval immediately below the bed; these are progressively exhumed north of Powell Creek as the discontinuity oversteps this sequence. North of Mack Creek (Loc. 7) hiatus concretions become less common as the underlying concretion zone is erosionally cut out, and there is a marked drop-off in abundance and diversity of epifauna.

At Bloomer and Barnum Creeks (Locs. 8, 9; Stops 2, 3), the Bloomer Creek bed carries a rich biota, dominated by brachiopods and bivalves. Key taxa include the brachiopods, Athryis spiriferoides, Pseudoatrypa cf. P. devoniana, Strophodonta demissa, Mediospirifer, and numerous smaller forms; bivalves include Modiomorpha, pterioids, and protobranchs. Scattered bryozoan fragments, gastropods (Mourlonia, bellerophontids) and fragmental cephalopods and trilobites are abundant. Small Pleurodictyum americanum characterize the bed, particularly north of Bloomer Creek where they are common. These corals commonly encrust shells, but they are also observed on hiatus concretions and phosphatic pebbles (Figures 6B, 7). Large corals, including Cystiphylloides, Heliophyllum, and Favosites, are scarce but present in the shell-rich phases of this unit.

Hiatus concretions are locally conspicuous features of this bed, particularly at Barnum Creek (Stop 3). Most of these are tube-, spindle-, and turnip-shaped objects; these typically display a central tube or axis which is pyrite-infilled when the concretion is in situ. These tubes appear to be early diagenetic features associated with lebensspuren in the underlying mudstone (Figure 7A). The calcareous concretions represent a later phase of carbonate diagenesis around tubes. The exhumed conretions are variably encrusted by epizoans and typically intensely bored (Figures 6A, 7A); a long, straight variety of Trypanites is typical of this bed. Hiatus concretions are variably degraded by the boring process and multiple episodes of bioencrustation are seen on nodules (Figures 6A, 7c-e), suggesting prolonged exposure of the concretions on the bottom and/or multiple episodes of exhumation and reburial. Small phosphatic pebbles and phosphatic steinkerns of brachiopods, nautiloids, and trilobites are common in the Bloomer Creek bed (Figure 6A). At Big Hollow Creek (Loc. 5; Stop 1) Pleurodictyum and Stereolasma have been found encrusting pebbles and steinkerns only a fraction of their size (Figure 6A); this strongly suggests synchronous coral growth and burial of the undersized-nodule substrate during sedimentation. Presence of both reworked phosphatic pebbles and hiatus concretions indicates that the period of submarine erosion may have been much longer than for the Barnum Creek bed.

The Bloomer Creek bed displays conspicuous local, northwestward erosional overstep of underlying beds; this unit is observed to overstep a conspicuous turbidite which is mappable between the diastems from Mack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Creek (Loc. 7) southeast to Sheldrake and Paines Creeks (Locs. 13, 12). This turbidite, designated the Mack Creek bed, occurs 17 feet below the Bloomer Creek bed and 18 feet above the Barunum Creek bed in southeastern exposures. The Bloomer Creek hiatus concretions are 8 feet above the turbidite bed at Bloomer Creek, and only 4 feet above it on the north branch of Mack Creek. At a small creek (Loc. 6) 1.3 miles north of Mack Creek the diastems are 17 feet apart and the turbidite is missing. At Big Hollow Creek (Loc. 5; Stop 1) 0.75 miles farther north, the diastems are only 12 feet apart. No further convergence is noted to the north and west of this section. Although some of the 20 feet of diastem convergence can be attributed to northward facies change from gray mudstone to dark gray-black shale, the abundant hiatus-concretion material and truncation of the turbidite clearly indicate that much of this change is the result of submarine erosion.

Submarine-Erosion Processes

The King Ferry diastems are termed stratomictic discontinuities (Baird, 1981); these are erosional breaks characterized by beds or zones of mixed and disturbed sediment, fossils, and reworked diagenetic structures rather than discrete erosion surfaces. The sediment mixing is generally the result of interaction of physical sea-floor erosion by currents and disturbance of near-surface muds by infauna. Such beds resemble condensed sedimentary sequences, but the erosional overstep associated with them excludes these units from this category.

Submarine erosion producing the King Ferry diastems occurred through the additive interaction of at least two and probably three processes; these include 1) dissipation-impingement of weak, episodic wave-current

Figure 6. Reworked concretions and fossils from King Ferry diastems. A shows material from Bloomer Creek bed; this includes: a) phosphoritic steinkern of trilobite, Loc. 9 (Stop 3), X 0.75; b) large partially exhumed calcareous concretion showing differential boring-encrustation of its upper surface, Loc. 9, X 0.7; c) tubular hiatus concretion showing hollow central core axis and encrusting Pleurodictyum, Cladochonus, and Philhedra, Loc. 8 (Stop 2), X 0.75; d) biodegraded hiatus-concretion "crumb" with attached Stereolasma, Loc. 8, X 1; e) Athyris valve showing bioattrition, Loc. 7, X 0.8; f) Pleurodictyum encrusting undersized phosphatic pebble, Loc. 5 (Stop 1), X 0.6. B shows material from Barnum Creek bed; this includes: a) bored and bioencrusted hiatus concretion. Note flask-shaped borings and encrusting Stereolasma and Hederella, Loc. 9, X 0.6; b) schematic cutaway view of in situ concretion showing core axis surrounded by diagenetic carbonate, X 0.5; c) bioencrusted tubular concretion with attached Ascodictyon (spots), Cladochonus, and Hederella, Loc. 8, X 0.8; d) reworked prefossilized conulariid, Loc. 9, X 0.8; e) nodule showing scratch marks attributed to infauna, Loc. 9, X 0.4. From Baird (1981).

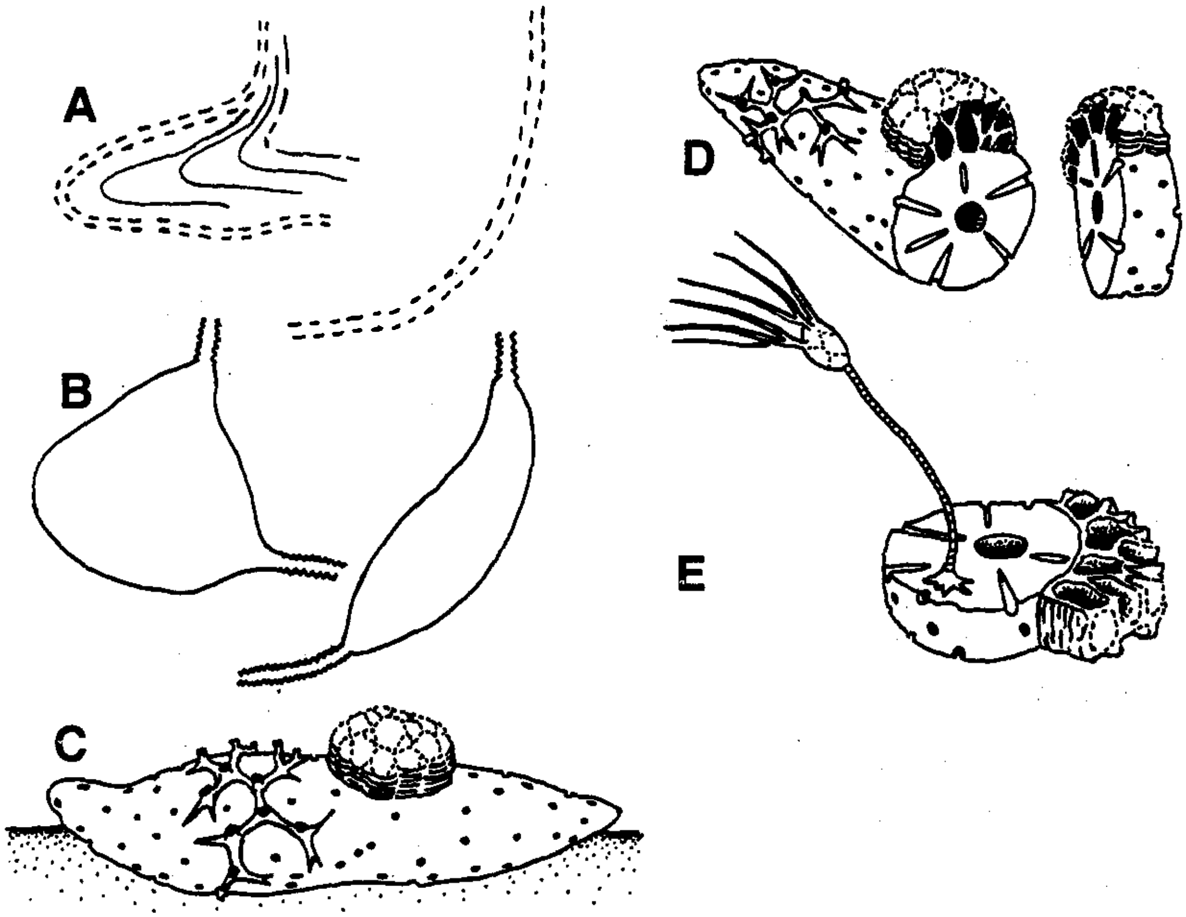


Figure 7. Genesis of "poker chip" hiatus-concretion fragment. A, bioturbation; development of vertical and oblique tubular burrows; B, diagenesic formation of calcareous concretion around burrows and precipitation of pyrite and/or calcite in burrows; C, erosion event; exhumation and bioencrustation-boring of exposed hiatus-concretions; D, hiatus-concretion disintegration; break up of nodule and encrusting coral Pleurodictyum; E, post-breakage bioencrustment; pelmatozoan attachment to transverse break surface.

energy on the sea bottom; 2) disturbance and liquefaction of surface muds by infauna; and 3) downslope transport of fines along a storm wave-induced current-energy gradient and/or through gravity effects. The third is less easy to assess but appears reasonable, given the regional character and distribution of the diastems.

Submarine erosion along the trough margin seems to have started as the result of an outside control such as slight regression, and/or reduction in sediment supply to the region. Such a shift, timed with ongoing bioturbation of shelf-slope muds, would have shifted the sedimentation balance from net accumulation to net loss without radical environmental change. Increased current energy acting on a soft bioturbated substrate would have resulted in resuspension and removal of fines. The activity of bottom organisms is known to contribute greatly to bottom erosion particularly in sloped sea-floor areas (Rowe et al., 1974; Stanley and Freeland, 1978). Animals may cause erosion at current flow velocities much lower than the threshold erosion velocity for a given sediment type by ejecting or scattering clay flocs into suspension, thus causing sediment entrainment in weak currents (Dillon and Zimmerman, 1970; Lonsdale and Southard, 1974). Bioturbation in both the Barnum and Bloomer Creek beds is evidenced by presence of Zoophycos and local abundance of protobranch bivalves (Nuculites, Paleoneilo); although protobranchs left no identifiable lebensspuren, these forms were probably active burrowers and sediment ingesters in the near-surface muds (Bowen, Rhoads and McAlester, 1974; Thayer, 1974). They presumably modified surface sediments, probably pelletizing and liquefying them in a pattern similar to that produced by modern Nucula (Rhoads and Young, 1970; Stanley, 1970).

A thin (2 to 15 inches thick) mantle or flux zone of bioturbated surface mud containing admixed shells and hiatus concretions is believed to have been present during concretion formation (Figure 8). Although reworked concretions were locally abundant within this layer, only a variable proportion of these would have been exposed at the surface at any given time due to vertical mixing of mud by organisms and lateral sediment transport. Because of winnowing of fines by currents, this zone of mixed material would have migrated vertically downward with time, overstepping (cannibalizing) progressively older beds (Figure 8, A-C cycle). Downward probing by infauna would have kept pace with downward advance of the sea bed with continuous incorporation of underlying concretions into the burrowed layer.

A gentle northwestward-dipping submarine paleoslope was associated with diastem formation; this regional slope was probably present through the period of deposition of Wanakah-equivalent King Ferry sediments, the two diastems being in part consequences of it. Important features discussed earlier (Baird, 1981) including: 1) gray-to black-shale facies change in the King Ferry, 2) associated faunal diversity gradients along both diastems, 3) distribution of hiatus concretions, 4) distribution and current-directional indicators of the Mack Creek turbidite, and 5) regional facies patterns in units bordering the King Ferry, constitute evidence for the paleoslope. The facies change from gray to black shale is interpreted as an environmental gradient to deeper water; black shales are usually interpreted as outer-shelf or basinal deposits (Heckel, 1973; Bowen, Rhoads and McAlester, 1974; Rhoads, 1975).

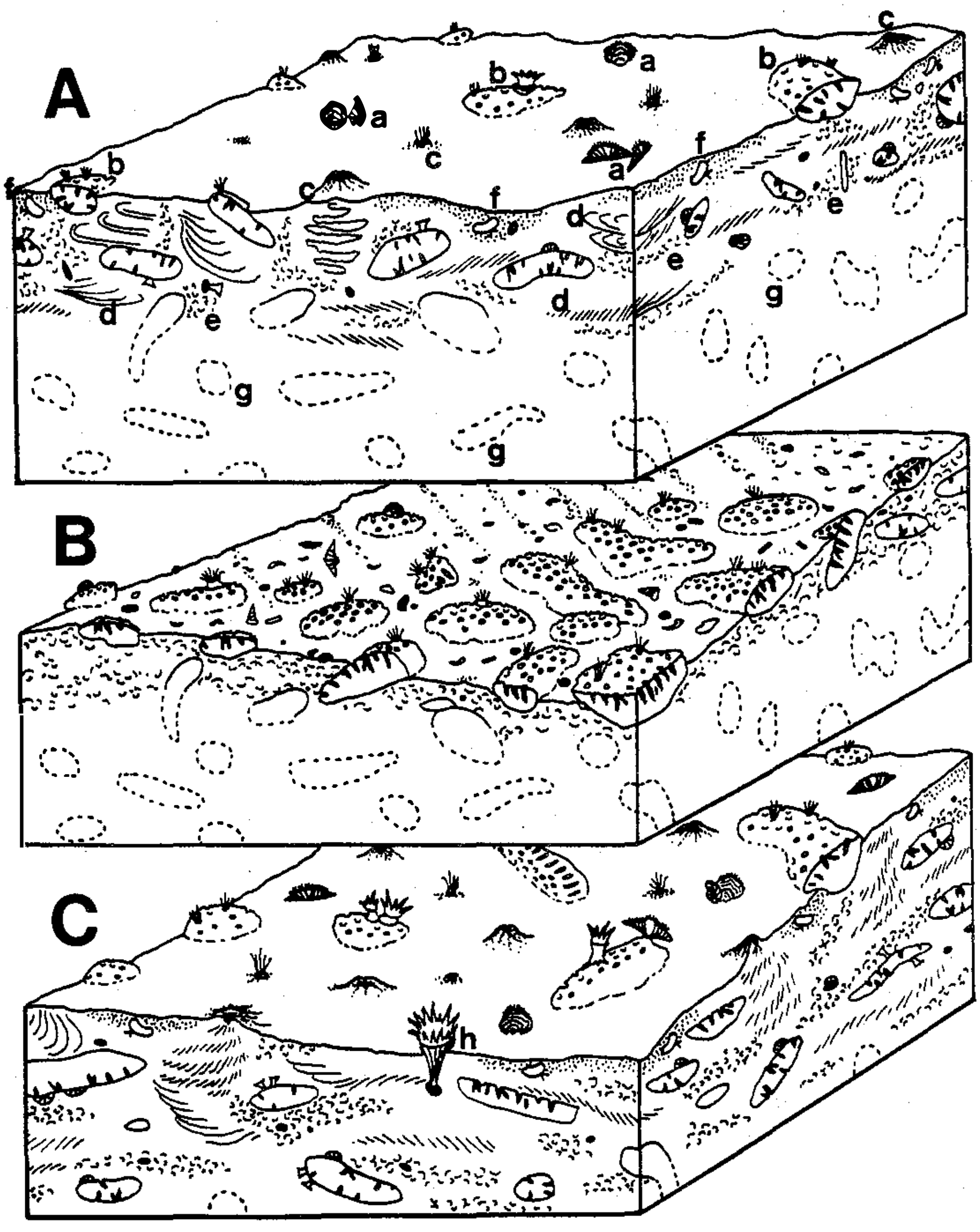
Diastem erosion is believed, in part, to be a result of paleoslope control; both diastems are best developed in the area of maximum facies change and are minimally developed or absent away from the sloped region. Why would these discontinuities be so peculiarly distributed? The answer appears to be related to differential instability of sediments on submarine slopes even including surfaces of less than 1° inclination. The presence of the Mack Creek turbidite with groove casts normal to inferred depositional strike is a strong indication of episodic density-current flow on the paleoslope. Similarly, resuspended fine mud would have been transported downslope (along a gradient of wave-induced current energy); this sediment slowly migrated downslope over time to be deposited as a thin blanket of black mud in deeper, level-bottom trough areas.

ACKNOWLEDGMENTS

The authors are grateful to Drs. Curt Teichert and Robert G. Sutton of the University of Rochester for review and criticism of the paper. Ronald Testa and Robert Eaton prepared the photographs. Elizabeth Moore and Margrit Gardner typed copies of the manuscript during its preparation.

Figure 8. Submarine erosion; schematic reconstruction. A, Biological activity on and within substrate. Development of water-rich burrowed sediment layer. Vertical shell-nodule mixing and loss of fine sediment to currents. B, Increased current scour and accelerated erosion. Concretion exhumation and bioencrustment. Local development of nodule pavements. C, Reduced current scour and erosion. Continued burrowing activity with regeneration of water-rich sediment layer. Downward extension of burrowing activity into undisturbed older muds.

Key: a, brachiopods; b, hiatus-concretion epizoans; c, conjectural surface expression of burrows; d, *Zoophycos* spreiten; e, shell concentrations; f, nuculid bivalves and associated thixotropic (stippled) sediment; g, *in situ* concretions; h, rugose coral displaying growth synchronous with sediment accumulation.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ROAD LOG FOR MIDDLE DEVONIAN DISCONFORMITIES AND SEDIMENTARY CONDENSATION IN THE CAYUGA LAKE AREA

TRIP

CUMULATIVE MILEAGE	MILES FROM LAST POINT	ROUTE DESCRIPTION
0.0.	0.0	Begin trip at Holiday Inn on Vestal Parkway opposite State University of New York, Binghamton campus. Turn east on parkway, but pull off immediately on northbound exit, Johnson City, Rte. 201.
0.4	0.4	Bridge over Susquehanna River, traffic circle at north end of bridge. Continue north on 201 (Riverside Drive).
1.2	0.8	Junction with N.Y. Route 17, proceed west on 17 to Owego exit.
15.4	14.2	Owego exit. Proceed north across Susquehanna River and through Owego on New York Route 38-96.
17.5	2.1	Junction of New York Routes 38 and 96. Proceed northwest on Route 96.
24.8	7.3	Village of Candor. Junction of New York Routes 96 and 96A. Proceed north on Route 96A.
34.8	10.0	Town of Danby, continue north.

39.7	4.9	Town of Ithaca at base of hill. Proceed west on New York Route 96A to junction of Route 13-14-96 with Route 96A.
40.1	0.4	Junction of Routes 13-34-96 with 96A. Turn north on Route 13-34-96.
40.2	0.1	Junction of Routes 13-34 and 89-96. Turn left (west) on 89-96 and cross Cayuga Inlet.
40.4	0.2	Junction of New York Routes 89 and 96. Turn right (north) on Route 89.
41.2	0.8	Leave town of Ithaca, proceed north along west side of Cayuga Lake.
50.2	9.0	Cross mouth of Taughannock Creek at Taughannock State Park. To the left (upstream) from the road bridge a low waterfall is visible; this section includes the Tully Formation (limestone) overlying the uppermost Hamilton Group (Windom Member). Farther upstream is a splendid gorge-waterfall section exposing sediments of the lower Genesee Group (Upper Devonian: Frasnian Stage).
60.4	10.2	Cross Sheldrake Creek.
62.6	2.2	Cross Barnum Creek.
63.6	1.0	Cross Bloomer Creek.
65.6	2.0	Turn left (west) on Swick Road. Proceed up hill.
66.4	0.8	Slight turn in road. Note deep gully of Big Hollow Creek to right.
66.9	0.5	Junction with north-south blacktop road. Turn right and cross Big Hollow Creek, STOP 1.

STOP 1 (1 hour): BIG HOLLOW CREEK SECTION. At this stop we will examine both the Barnum and Bloomer Creek beds; these can be seen in the creek bed below the small farm north of the creek. The Barnum Creek bed is thin, but characterized by abundant flattish to ovoid hiatus concretions which are encrusted by Cladochonus and small bryozoans. The Bloomer Creek bed occurs about 150 feet upstream from the lower discontinuity and 12 feet higher in the section. This 8-inch-thick mudstone bed has numerous small corals (Pleurodictyum, Stereolasma) which are commonly attached to phosphatic nodules and hiatus concretions. Above this unit is 25-30 feet of dark-gray and black shale. In a borrow pit upstream from the road crossing, the black shale grades upward to calcareous mudstone marking the base of Jaycox-equivalent King Ferry Member. Return to vehicle.

67.0	0.1	Turn around on blacktop road and turn left (east) on Swick Road.
68.3	1.3	Junction with New York Route 89. Turn right (south).
70.6	2.3	Turn right (west) by sign for vineyard onto gravel farm road.
70.8	0.2	Park by farmhouse at end of road. STOP 2.

STOP 2 (1 hour and 20 minutes): BLOOMER CREEK SECTION. We will walk north for 1500 feet, crossing the main branch of Bloomer Creek to a more accessible section on its northern tributary. At this section we will observe the Mack and Bloomer Creek beds below the small waterfalls and study basal Moscow beds both at and slightly above the falls. This is one of the northernmost sections at which the Mack Creek bed can be observed. The Bloomer Creek bed occurs 8 feet above this unit; it is rich in brachiopod and molluscan debris and hiatus concretions are common; these display abundant epizoans.

The Tichenor Limestone Member occurs in the recess under the falls; it yields abundant pelmatozoan material including articulated crinoids. The Menteth and overlying Kashong Members occur upstream from the falls. Fossils are plentiful in the Kashong sequence, and they can be easily extracted from the Rhipidomella-Centronella Bed. The phosphatic pebble bed of the upper Kashong is inconspicuous, but it yields phosphatic pebbles and occasional phosphatic brachiopod and trilobite steinkerns. Return to vehicle and proceed back to main road.

71.0	0.2	Junction with New York Route 89. Turn right and proceed south
71.2	0.2	Park vehicles in open area above high falls on Barnum Creek. STOP 3.

STOP 3 (2 hours, lunch first before examining the rocks): BARNUM CREEK, UPPER LUDLOWVILLE SECTION. We will proceed around the waterfall, which is capped by the Menteth Member, and reach the creek by a safer downstream route. Wading up the creek, we will examine, in order, the Barnum Creek bed, Mack Creek bed, and Bloomer Creek bed. This is a classic complete profile shown schematically in Figure 2. Both the Barnum and Bloomer Creek beds yield a rich, shelly biota and numerous hiatus concretions. The disontinuities overlie intervals of mudstone bearing numerous *in situ* concretions which are the erosional source of hiatus concretions. Of particular interest is the occurrence of reworked, prefossilized conulariids in the Barnum Creek bed. Moreover, large differentially bored hiatus concretions occur in the Bloomer Creek bed; these show intense *Trypanites* borings on their upper surface and no boring underneath (Figure 6A). The nodules have different epizoan assemblages on their upper and lower surfaces, indicating development of microhabitats along their exterior. Return to vehicles.

71.4	0.2	Rejunction with New York Route 89. Turn left (south).
73.3	1.9	Turn left (east) on road to Sheldrake.
73.8	0.5	Turn right (south) and proceed to farm above high falls on Sheldrake Creek.
73.9	0.1	Park vehicle and walk to STOP 4.

STOP 4 (45 minutes): SHELDRAKE CREEK, LOWER MOSCOW CONDENSED STRATIGRAPHIC SECTION. The Tichenor-basal Windom interval is represented by only 17-18 feet of section. The Deep Run Member is only 4.5 feet thick as opposed to 10 feet at Stop 2. The Menteth Limestone is thin and irregular in this section, and the *Rhipidomella-Centronella* bed of the Kashong Member is juxtaposed on it. The Tichenor Member is represented by a massive calcarenitic ledge with a sharp basal contact with the Ludlowville. Medial King Ferry beds will not be examined here, but the Mack Creek bed and the Bloomer Creek bed are both visible in the creek bank below the falls. Return to vehicles.

74.0	0.1	Junction with Sheldrake Road. Turn left (west).
74.5	0.5	Junction with New York Route 89. Turn left (south) and return to Binghamton.
134.6	60.1	Return to Holiday Inn. End of trip.
Return from <u>Locs. 1-4</u>		OPTIONAL STOP 5
94.5	0.0	Start at junction of New York Routes 13-34 and 96 in Ithaca, proceed north on Route 13-34.

96.5	0.2	Junction of New York Routes 13 and 34. Turn left (north) and proceed on Routes 34.
101.1	4.6	Junction of New York Routes 34 and 34B in South Lansing. Turn left (west) on Route 34B.
101.8	0.7	Turn left (southwest) on Portland Point Road.
102.1	0.3	Entrance to Penn Dixie Cement Corp. Portland Point quarry on left. The Tully Formation (limestone) is quarried here, and the uppermost beds of the Moscow Formation (Windom Shale) are exposed. For description of this locality see Grasso (1970) N.Y.S.G.A. field trip D and H, Stop 9. An excellent Tully-Windom exposure is developed to the right of the road in Minnegar Brook.
102.4	0.3	Salt mine to right. The Morton Salt Company is mining salt from the Salina Group (Upper Silurian).
102.8	0.4	Arrival at Portland Point at the mouth of Gulf Creek. Stop vehicles and proceed on foot.

STOP 5a (25 minutes): PORTLAND POINT TYPE SECTION (see Figure 4). It occurs in a railroad cut along Cayuga Lake shore 300 feet south of entrance to Gulf Creek. There is noticeable southward dip to the strata so that 10-15 feet of uppermost King Ferry is visible below the Portland Point; the Portland Point-Kashong interval descends to railroad level over a 100-150 foot distance.

Originally, the Portland Point was described as including 9.0-9.5 feet of calcareous strata with a prominent calcarenite bed at the base and shell-rich limestone at the top (Cooper, 1929, 1930). The Portland Point is now redefined to include only the calcarenitic (Tichenor-equivalent) bed and overlying 2-foot muddy, Deep Run-Menteth equivalent limestone (Baird, 1979). 5.5 feet of succeeding beds are assigned to the Kashong Member. A prominent calcareous siltstone bed believed to be equivalent to the phosphatic pebble bed to the west occurs 8.5-9 feet above the base of the Moscow Formation; this may have been the top marker bed of Cooper's Portland Point section.

This bed occurs about one foot below a calcareous siltstone bed marking the base of the Windom. Above the base of the Windom are numerous Windom brachiopods including _Athyris spiriferoides_, _Ambocoelia umbonata_, and chonetids.

STOP 5b (20 minutes): SHURGER FALLS. We now walk back north to the mouth of Gulf Creek. Proceeding east (upstream) for 100-150 feet we come to Shurger Falls. The Portland Point is now at much higher elevation, capping the 40-50 foot waterfall. We will not climb the falls but will collect fossils from loose blocks which have fallen down.

103.8	1.0	Return to vehicles and proceed back to junction with New York Route 34B.
111.1	7.3	Return to starting point at junction of New York Routes 13-34 and 96 in Ithaca.
150.3	39.2	Return to Holiday Inn, Vestal, New York. END OF TRIP.

HYDROLOGY IN RELATION TO GLACIAL GEOLOGY ALONG THE SUSQUEHANNA RIVER VALLEY, BINGHAMTON TO OWEGO, NEW YORK

by

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INTRODUCTION

Over the past 20 years, several regional and localized studies have increased our knowledge of aquifers in the Susquehanna River basin of New York. This article extracts and summarizes some results of these studies that pertain to sites visited on this field trip along the Susquehanna valley from Binghamton downstream to Owego.

DISTRIBUTION OF AQUIFERS WITHIN THE GLACIAL DRIFT

About 85 percent of the Susquehanna River basin of New York is an upland in which shale, siltstone, and fine sandstone of Devonian age is dissected by narrow valleys and mantled by till. Stratified drift is confined largely to broad valleys, 1,000 to 8,000 feet wide, which occupy the remaining 15 percent of the basin.

During the waning stages of glaciation, tongues of ice extended southward along the broad valleys beyond the main ice sheet. Lakes formed near the ends of the shrinking ice tongues, temporarily filling any space vacated by melting of the ice that lay below the level of stratified drift previously deposited downstream. Many of these lakes were extensive, particularly in the deeper valleys, and trapped much clay, silt, and very fine sand. Elsewhere, particularly in relatively shallow valleys such as the Susquehanna River valley from Binghamton to Owego, smaller lakes and stream channels formed atop and against ice tongues that became too thin to flow. As each small lake filled with sediment, others formed nearby, and water velocities were often great enough to keep fine-grained sediment in suspension. The resulting stratified drift is heterogeneous but predominantly coarse grained, like the idealized sketch in Figure 1 and the actual cross sections in Figures 2 and 3.

Only the coarser sands and gravels within the stratified drift yield enough water to be considered aquifers. Several generalizations may be made as to their distribution within the broad valleys of the Susquehanna River basin:

1. As a rule, the greater the depth to bedrock, the greater the thickness of non-water-yielding clay, silt, and very fine sand. In broad valleys northeast of Binghamton, depth to bedrock generally ranges from 250 to 500 feet. Near and west of Binghamton, depths of 70 to 200 feet are typical, although bedrock is deeper than 250 feet locally. Total thickness of gravel and coarse sand rarely exceeds 150 feet in any of these valleys and can be as little as 10 feet.

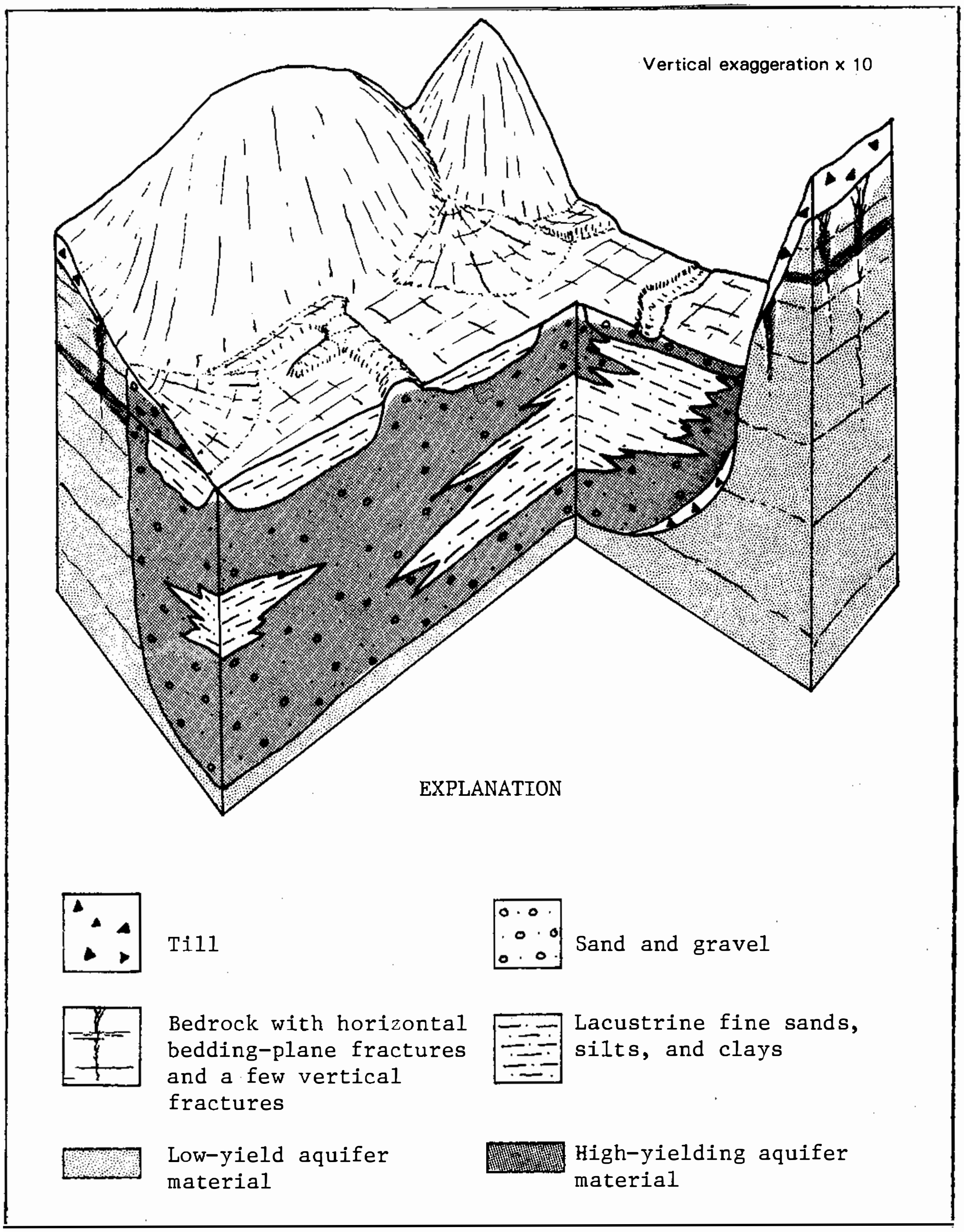

Figure 1. Geology and hydrology typical of broad valleys where ice tongues stagnated. Depth to bedrock is commonly less than 100 feet below stream grade. (From R. D. MacNish, written commun., 1969).

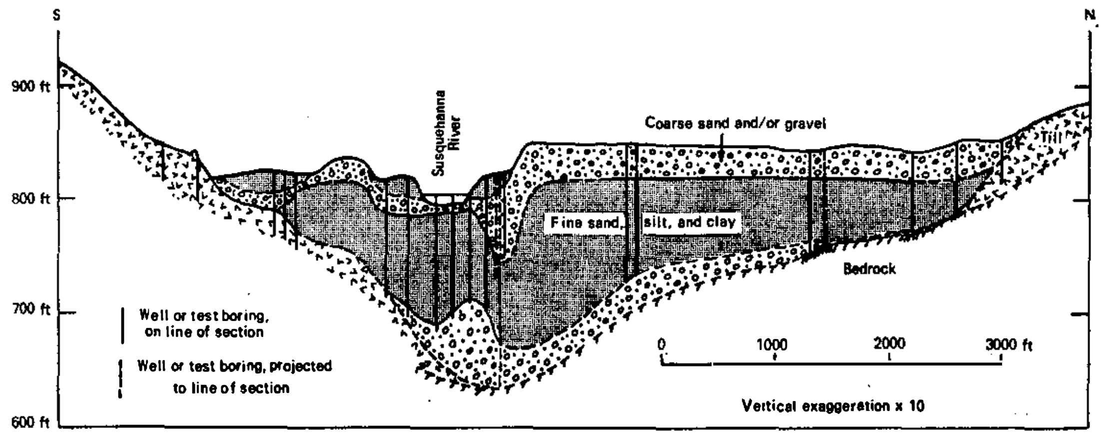

Figure 2. Geologic cross section beneath Route 26 in Vestal and McKinley Avenue in Endicott (Modified from Randall, 1970). Approximate location is shown in figure 7.

2. Sands and gravels are present at or near land surface in all broad valleys. In some places, the near-surface sands and gravels are thin or largely above the water table, but elsewhere they form the most productive aquifers in the basin because they are moderately to highly permeable and generally in hydraulic contact with streams from which water can infiltrate to sustain well yields.

3. Some of the broad valleys also contain basal sand-and-gravel aquifers beneath extensive silts and clays. The basal aquifers tend to be thin or less permeable than the near-surface aquifers and generally yield only moderate quantities of water. The water is commonly of inferior chemical quality, widely characterized by high iron and, in a few places, by high chloride.

4. Locally, the entire thickness of stratified drift is sand and gravel. In the deeper valleys, the condition is generally limited to the sides of the valleys.

LITHOLOGY AS A CLUE TO DEGLACIAL HISTORY AND AQUIFER DISTRIBUTION

MacClintock and Apfel (1944) were the first to call attention to a marked contrast in drift lithology within the Appalachian Plateau of south-central New York. They recognized a "Binghamton" drift whose bright, colorful appearance is caused by numerous pebbles of limestone, chert, quartzite, and other rock types foreign to the plateau, and a contrasting drab "Olean" drift in which limestones generally constitute less than 3 percent of the pebbles, and sandstone or shale of local origin constitute more than 85 percent. They inferred that the two drifts represent successive ice advances from different directions. Later writers

(Merritt and Muller, 1959; Denny and Lyford, 1963; Moss and Ritter, 1962) reinterpreted the drab and bright drifts as facies of a single drift sheet in which the bright facies was largely restricted to some of the major valleys.

In the Susquehanna River valley from Binghamton downstream to the Pennsylvania border, bright outwash forms prominent terraces but is commonly underlain by relatively drab sand and gravel and locally bordered by drab kames or kame terraces at slightly higher elevation (Randall, 1978a). Stratified drift in which 85 to nearly 100 percent of the pebbles are local shale and less than 6 percent are limestone is abundant and may constitute the bulk of the valley fill, even where shallow exposures are much brighter. The bright outwash entered the Susquehanna valley from the broad north-side valleys—Chenango River, Owego Creek, Cayuta Creek—but even these valleys contributed drab sediment, or none at all, when deposition of stratified drift was beginning in the Susquehanna valley.

The stratigraphy described in Table 1 and illustrated in Figure 3 is based on exposures and drilling samples from Binghamton and Johnson City, immediately west of the confluence of the Chenango and Susquehanna Rivers. The bright gravel and sand are part of a valley train extending down the Chenango River valley. Terrace altitudes[1/] descend from 940 feet east of Kattellville, 6 miles north of Binghamton, to 880 feet in Binghamton and to 840 feet near the Broome-Tioga County line, 9 miles west of Binghamton. The terraces do not fall on a single profile, but instead seem to form a set of imbricate profiles dipping downvalley, each steeper than the present stream gradient. This suggests that they may represent successive kame deltas built into a lake or lakes. The bright gravel capping these terraces is highly calcareous and is commonly lime-cemented. In most places, 30 to 50 percent of the pebbles are exotic (that is, unlike the local bedrock). The bright gravel is continuous beneath silt and organic deposits in iceblock depressions and varies widely in thickness, which indicates that it was deposited when stagnant ice was still present in the Susquehanna valley. Drab gravel is exposed locally along the sides of the Susquehanna valley at altitudes that descend from 910 feet in Binghamton to 850 feet near the Broome-Tioga County line. Where penetrated by wells and test holes, the drab gravel ranges in thickness from zero to 75 feet over short distances; much of it is very silty, but clean layers are also common and in places have yielded 250 to 2,500 gallons per minute to wells. These features suggest that the drab gravel originated as ice-contact deposits early in deglaciation. Several holes penetrated sharply contrasting bright and drab gravel separated by many feet of silt, but in holes penetrating only sand and gravel, the content of exotic materials seemed to change gradationally. Till penetrated beneath stratified drift was universally drab (Randall, 1978a).

Distribution of bright and drab sand and gravel near Owego, where Owego Creek joins the Susquehanna River, is similar to that near

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1/ Altitudes here, and in figures 2, 3, and 6, are above the National Geodetic Vertical Datum of 1929, formerly called mean sea level.

Binghamton (Randall, 1978a). Gravel pits in a prominent terrace in the center of the Susquehanna valley reveal at least 40 feet of bright outwash. Near the north end of the terrace, however, a well penetrated sand with drab pebbles beneath the bright gravel. Another well, drilled on the flood plain within the valley of Owego Creek, abruptly entered drab noncalcareous gravel beneath bright gravel (Randall, 1972); although the bright gravel in this well is appreciably lower in altitude than that capping the terrace, it may be a collapsed equivalent. A succession of slightly younger terrace remnants upstream from the village of Owego along Owego Creek are also capped by bright gravel. Drab gravel was not observed along the valley side near Owego, but 4 miles downstream, at Tioga Center and Lounsberry, kames near the sides of the valley are much less bright than younger outwash terraces. Thus, the earliest melt-water streams along both the Susquehanna River and Owego Creek valleys apparently carried very drab sediment (Randall, 1978a).

The evidence summarized above, and additional data described by Randall (1978a), suggest that the last ice sheet to invade this region removed whatever older drift may have been present in the major valleys and that the increase in exotic content of the stratified drift with time is best ascribed to changes in sediment transport during the latest deglaciation rather than to a succession of ice advances. Most exotic materials reached the Susquehanna River valley by transport along broad valleys, probably by one or a combination of the following mechanisms:

1. Preferential flow of ice along the valleys when the continental ice sheet covered the region (Muller, 1965).

2. Tongues of ice continuing to flow south in the valleys beyond active ice in the uplands. Moss and Ritter (1962, p. 104) summarize evidence that such valley tongues were at most a few miles long, however.

3. Englacial or subglacial melt-water streams extending many miles along the valleys, perhaps in active as well as stagnant ice (Randall, 1978a). There is some local evidence that melt-water drainage systems were extensive in or under the retreating glacier. (a) Large volumes of stratified drift were deposited in the broad valleys in proglacial lakes that formed when ice melted below the level of older outwash downstream. By contrast, kames or kame terraces are rare in the uplands, which implies that lakes rarely formed in upland valleys during deglaciation. Some of these valleys drain north for more than 10 miles and descend several hundred feet below the lowest saddles on the divides. The lack of lakes in such valleys seems to require that extensive channels existed in or under the decaying ice sheet to drain melt water into the nearest broad valley. (b) Drab stratified drift between Owego and Waverly, a distance of 14 miles (or 6 miles north-south), is not necessarily all precisely the same age, but all must have been deposited against stagnant ice by south-flowing melt water before the valley of Owego Creek began to carry bright sediment. Thus, at an early stage in deglaciation of the master valley, melt-water drainage on, in, or under stagnant ice must have extended more than 14 miles.

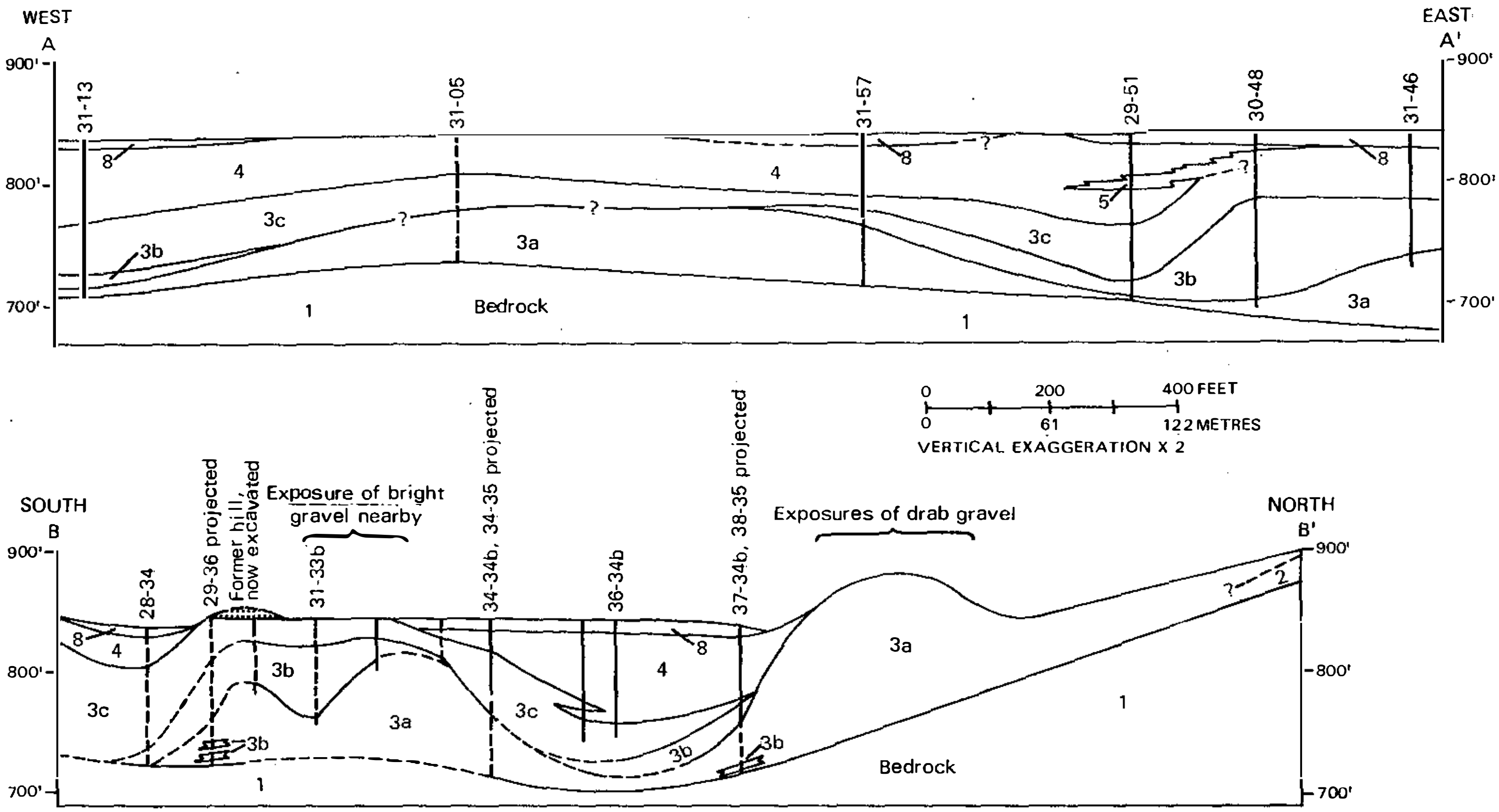


Figure 3. Geologic cross sections in Binghamton (from Randall, 1978a). Numbers in section refer to stratigraphic units in table 1. Wells and test borings are represented by vertical lines, solid where samples were studied, otherwise dashed. Other borings near section B (not shown) were also studied. Vertical numbers are seconds of latitude and longitude ("b" indicates engineering test boring) and indicate that log is published (Randall, 1972). All sites are lat 42°06' N, long 75°55' W (or 75°54' W near east end of section A). Approximate locations of sections are shown in Figure 7.

Table 1. Pleistocene and Holocene stratigraphic units in Binghamton and Johnson City, New York
[From Randall, 1978a]

Stratigraphic unit	Number in figure 3	Lithology, thickness, distribution
Fill	8	Chiefly trash and ashes; some sand, gravel, and other materials. Placed in most natural depressions, 5 to 20 feet thick.
Flood-plain silt	-	Brown silt and very fine sand with roots and little organic matter; typically 5 to 15 feet thick.
Alluvial fan deposits	-	Gravel, moderately sandy and commonly moderately silty; most stones are flat local shale or siltstone. Deposited by small streams entering the Susquehanna River valley.
Older river alluvium	5	Sand and gravel, bright but leached mostly free of limestone; as much as 35 feet thick; interfingers with and overlies unit 4 near Chenango River, where thin silt or silty sand interbeds (unit 4) contain abundant wood and fine organic matter.[a]
Postglacial or late-glacial lake beds	4	Silt and very fine sand with some clay and scattered fine organic fragments, commonly grading up into peat and highly organic silt; as much as 80 feet thick. Fills ice-block depressions in a narrow east-west zone near the deepest part of the bedrock valley.
Stratified drift:		
Bright gravel	3c	Sandy gravel and pebbly sand with variable amounts of silt; highly calcareous. Upper part very bright (35 to 75 percent exotic pebbles), lower part moderately bright (15 to 30 percent exotic pebbles); ranges in thickness from near zero to at least 100 feet.
Lake beds	3b	Silt to fine sand, some clay, no organic matter; lenses may occur anywhere within unit 3, but seem most common between bright and drab gravels.
Drab gravel	3a	Sandy gravel and pebbly sand with variable amounts of silt; slightly calcareous; 10 percent or less exotic pebbles; thickness varies widely.
Glacial till	2	Typically a stony silt; only a foot or so thick in places, but forms low hills in southern part of valley.
Bedrock	1	Shale and siltstone.

[a] Wood from a well at lat 46°06'25" N, long 75°54'50" W (Randall, 1972) at depths of 24 and 45 feet had radiocarbon ages of 2649 ±79 and 3801 ±60 years, respectively (Randall and Coates, 1973).

Knowledge of the drift lithology is helpful in subsurface correlation and in understanding ground-water quality. For example, the downwarp of the upper gravel layer near the east side of Figure 3A was first recognized from pebble lithology; the downwarp was later verified by water-level and water-temperature data (Randall, 1977). Hardness of ground water, which is caused by dissolved calcium and magnesium, is commonly much higher in the bright stratified drift than in the drab (Ku and others, 1975; Randall, 1977).

SOURCES OF RECHARGE TO AQUIFERS

Ground-water-resource appraisals must not only determine the extent and lithology of aquifers but also evaluate the yield obtainable from each aquifer. Many recent appraisals of the water resources of large drainage basins in New York and New England have evaluated potential aquifer yield by determining representative regional rates of ground-water recharge and applying those rates to the dimensions of each aquifer or area of stratified glacial drift (Cohen and others, 1968, p. 24-46; Crain, 1974, p. 45, pl. 3; Kantrowitz, 1970, p. 67; La Sala, 1968, p. 54; Randall and others, 1966, p. 66; Cervione and others, 1972, p. 46-47). Different sources or components of recharge were treated separately in most of these studies. The principal sources of recharge to stratified-drift aquifers in the Susquehanna River basin are described in the following paragraphs.

Precipitation on land surface above the aquifer

Where sand or gravel are present at land surface, nearly all rain and melting snow will infiltrate, and about half will eventually reach the water table as recharge. (The rest is returned to the atmosphere by plants or evaporation.) Thus, the annual volume of recharge from precipitation to a surficial aquifer depends principally on the extent of surficial sand and gravel and on the annual precipitation rate. Randall (1977) calculated recharge from precipitation to a surficial aquifer in Binghamton and Johnson City.

Precipitation on upland hillsides adjacent to the aquifer

Most stratified-drift aquifers in upstate New York are in valleys bordered by till-covered hillsides. Where the till contains a large percentage of silt and clay, as in the Susquehanna River basin, only a small part of the water from rain or snowmelt infiltrates deeper than the top foot or two; the excess moves downslope in rivulets or through shallow openings in the soil. Where upland hillsides slope toward a stratified-drift aquifer (rather than toward an upland stream), runoff that reaches the permeable sand or gravel in the valley infiltrates there. Annual recharge to an aquifer from upland hillsides depends principally on annual precipitation and on the size of upland areas that slope toward that aquifer. Permeability of soils overlying the aquifer is rarely a limitation.

Natural infiltration from streams

Where the water level in a surficial stratified-drift aquifer is lower than the water surface in a stream crossing the aquifer, water will infiltrate from the stream into the aquifer. In the Susquehanna River basin, this occurs naturally wherever a tributary stream leaves its own valley to flow over the sand and gravel fill of a larger valley (Ku and others, 1975). This is true on all scales. For example, as suggested in figure 4, a tiny ephemeral stream descending a steep hillside loses water where it reaches the narrow flood plain of a creek draining perhaps a square mile, and that creek loses water where it crosses the alluvial gravel that borders a large upland stream draining 10 or 20 square miles. Similarly, the upland stream loses water where it enters the valley of a major river and crosses a thick stratified-drift aquifer or its own alluvial fan.

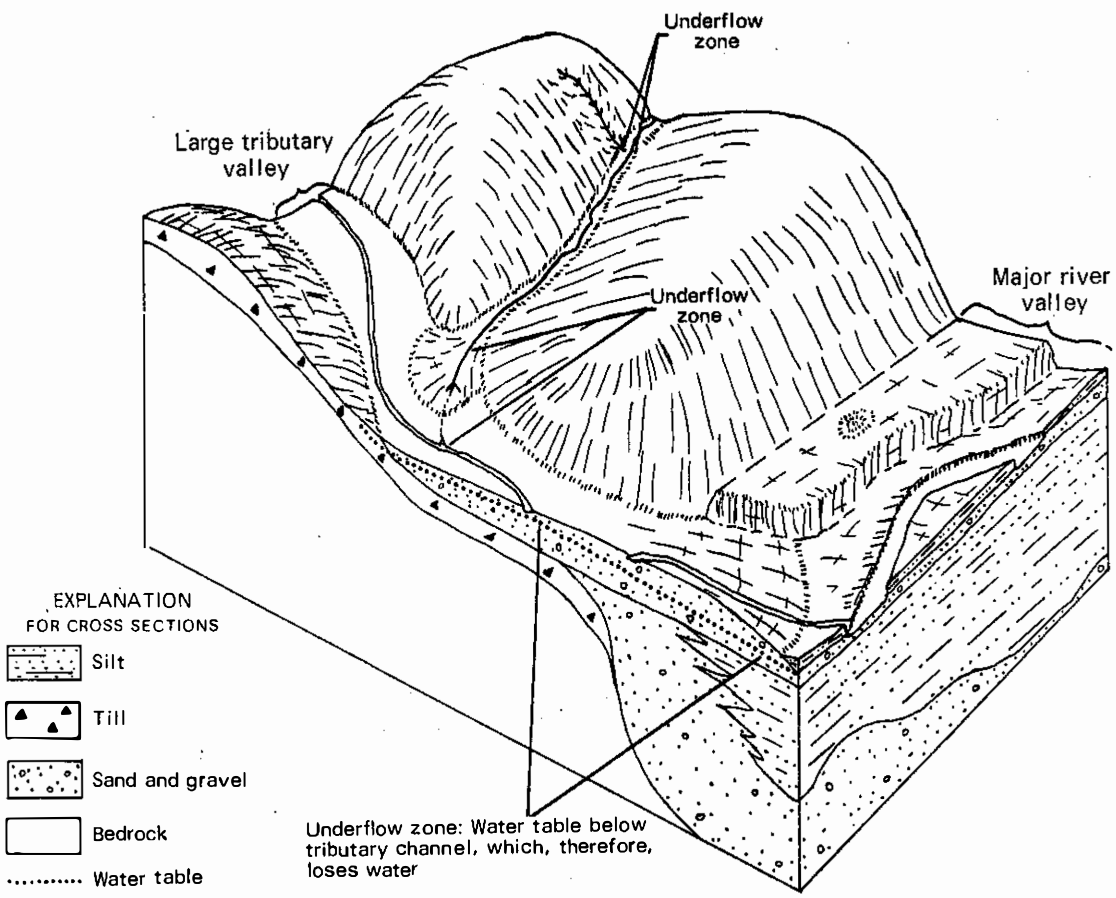


Figure 4. Typical distribution of losing stream reaches. (From Ku and others, 1975).

Infiltration losses from seven tributary streams where they cross stratified-drift aquifers in major valleys were measured and analyzed by Randall (1978b). Flow was measured at the most downstream point at which the channel was known to be cut in till or bedrock, or at least was still clearly within its own upland valley, and at one or more points farther downstream within a major valley, where streamflow losses were expected. Each set of flow measurements was completed within 2 hours. Near the stream reaches studied, the upper 15 to 30 feet of sediment is chiefly compact silty and sandy gravel deposited in alluvial fans by postglacial streams. Upstream from the edges of broad valleys, this alluvium overlies till or bedrock; downstream, it commonly overlies sandy glaciofluvial gravel, and farther downstream a wedge of silt and very fine sand may overlie or replace the glaciofluvial gravel. The glaciofluvial gravel generally has a more varied lithologic composition and a lower silt content than the alluvium and has much greater water-yielding potential. The actual streambed contains loose sandy gravel, generally slightly silty or free of silt.

Each stream studied began to lose water rapidly several hundred feet downstream from where it entered the major valley, or from the lowermost known exposure of till or bedrock in its channel. Measured losses in this zone of rapid loss varied directly with stream length between measurement sites; that is, infiltration per unit length of channel was approximately constant (Fig. 5). By contrast, stream width and depth had little effect on infiltration. Plots of infiltration per unit length against width, and against the product of width and depth, showed no correlation. Along most streams, the uniform maximum rate of infiltration loss per unit length of channel prevails for 300 feet or more when sufficient flow is available. The loss rate for typical streams would be at least 10 liters per second per 100 meters (1 cubic foot per second per 1,000 feet), and average hydraulic conductivity of the alluvium was estimated to be at least 13 meters per day (50×10^{-5} feet per second, or 320 gallons per day per square foot).

Several papers dealing with infiltration from streams state or imply that the rate of infiltration is normally controlled by a thin streambed layer that is less permeable than the underlying sediment (Walton, 1963; Walton and others, 1967; Norris, 1970; Moore and Jenkins, 1966). Nevertheless, evidence cited by Randall (1978b) suggests that infiltration from most tributary streams in the Susquehanna River basin is controlled by permeability distribution within the alluvium or stratified drift rather than at the streambed. Along some tributary streams, rapid infiltration loss begins near where the depth to till and bedrock increases, but along other streams till and bedrock seem to lie far below the channel more than 300 ft upstream from where infiltration increases (fig. 6). Furthermore, the point at which rapid infiltration begins shifts somewhat from one date to another. These changes may be caused by changing water-table configuration within the alluvium—a function of prior infiltration and rainfall as well as of permeability distribution.

EXPLANATION

▲ Point at which stream went dry (infiltration equaled inflow)

● Point of flow measurement downstream from inflow measurement

●——▲ Measurements made on same date

—— Trend line, estimated average infiltration rate

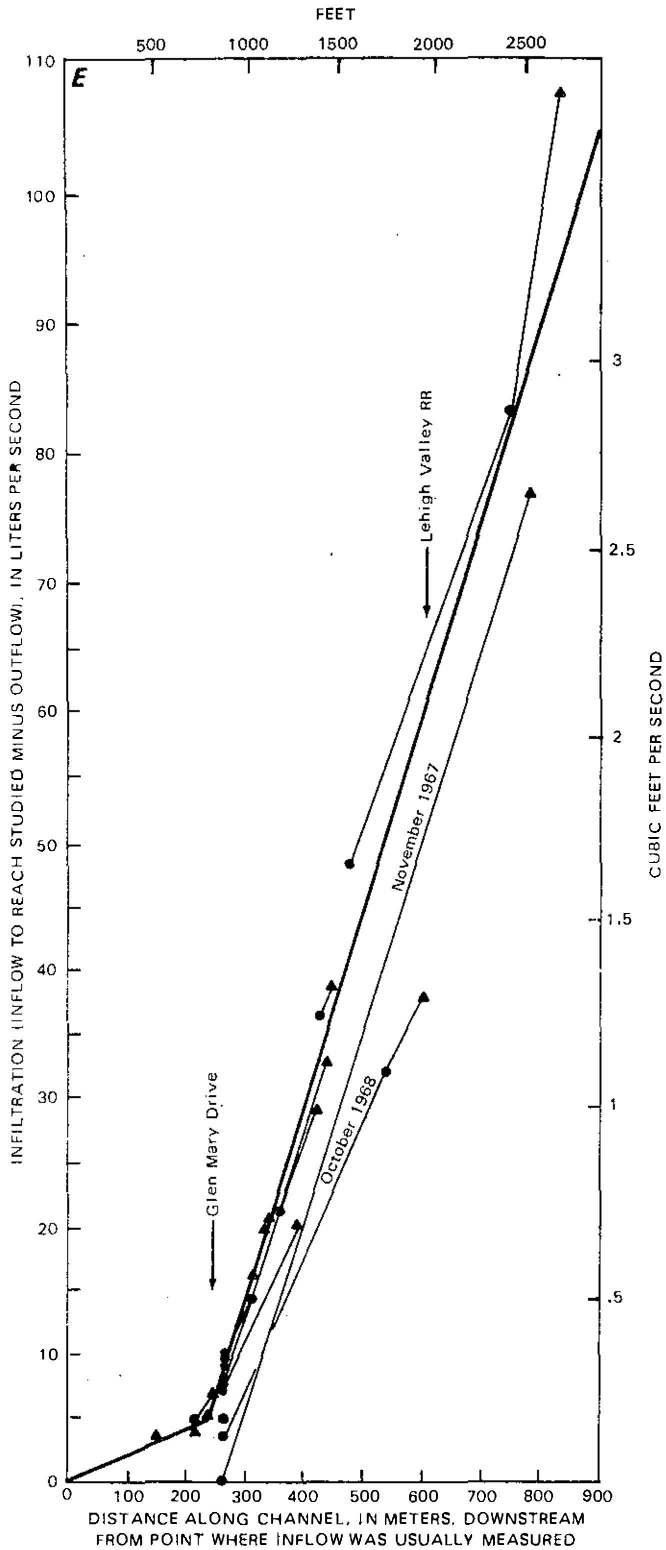


Figure 5. Loss of water from Thorn Hollow Creek in relation to stream length. (From Randall, 1978b.)

EXPLANATION
SYMBOLS IN CROSS SECTIONS
Line of section is along creek

Top of bedrock

Well or test boring, projected perpendicular to stream except as noted. Solid line represents casing, dashed where casing has been removed (test borings) or not used (bedrock). Log available (Randall, 1972) unless otherwise indicated

Streambed cut in dense silt
Stream and point of dryness
Water level in well
} Only on date indicated

Water table

Contact between lithologic units

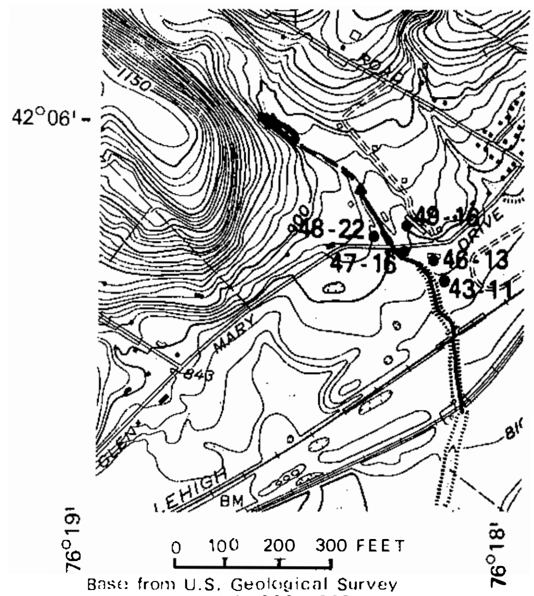
Silty sandy gravel, yields little water to drilled wells; tributary creek alluvium

Sandy gravel or sand, variably silty, in part water yielding; glaciofluvial valley-train deposits

Silt to fine sand; lake-bottom deposits

Glacial till

Uncertain



Base from U.S. Geological Survey
Owego, N.Y., 1:24,000, 1969

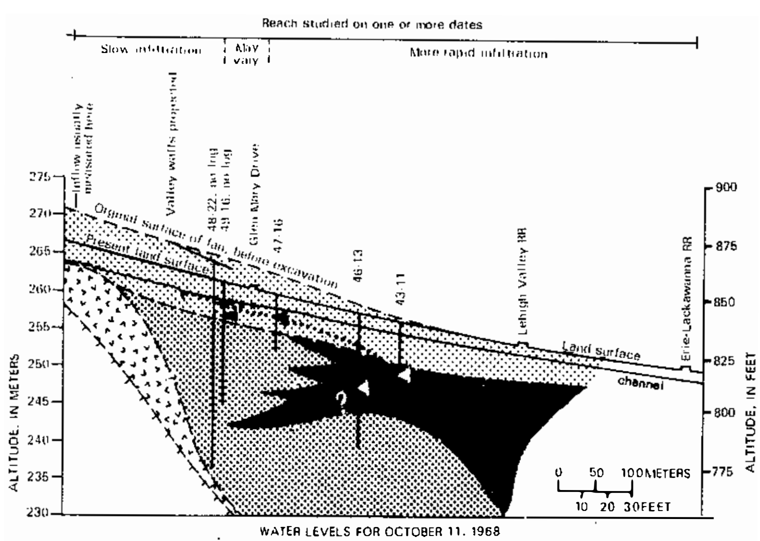


Figure 6. Geohydrologic features along Thorn Hollow Creek. (Modified from Randall, 1978b).

Induced infiltration from large streams

Whenever water levels in surficial aquifers are drawn below stream stage by pumping, infiltration is induced from stream reaches that do not lose water ordinarily. The rate of induced infiltration depends on many factors, including the distribution of wells, pumping rates, hydraulic conductivity of the streambed and nearby parts of the aquifer (which together may be termed "effective streambed permeability"), and changes in stage, bottom area, and water temperature of the stream. Potential induced infiltration from the Chenango River to the Clinton Street-Ballpark aquifer in Binghamton was estimated by Randall (1977, Appendix F), and induced infiltration from the Susquehanna River to the South Street well field in Endicott was estimated by Ground Water Associates (1978). In 1981, water-level distribution around several municipal well fields on the banks of the Susquehanna River will be measured under steady-state pumping conditions and under the transient conditions caused by starting or stopping of pumping. Results are expected to be useful in calibrating a digital model of the stratified-drift aquifers, which should lead to a better understanding of potential induced infiltration and aquifer yield.

GROUND-WATER PROBLEMS AND TRADEOFFS

In most urban areas, numerous competing demands are placed on the local earth resources and landscape. The use or modification of the land to meet human needs generally has some impact, negative or positive, on the quantity or quality of ground water. Several examples from the Susquehanna River valley are mentioned briefly below.

Excavation of river channels

A village of Endicott municipal well 180 feet from the north bank of the Susquehanna River continuously delivered more than 1 million gallons per day of bacteria-free water for 19 years until December 1964, when coliform bacteria were detected in routine weekly water samples. Since then, coliforms have never been absent for more than a few weeks, despite greatly reduced pumping. Randall (1970) demonstrated that the bacteria came from the river and argued that the most likely explanation was repeated excavation of the river bed within 200 feet upstream and downstream of the well for pipeline and bridge construction in the early 1960's. The backfill that replaced the naturally stratified streambed sediments may have been more permeable, permitting greater induced infiltration but also greater movement of bacteria.

Riverbeds have been excavated for many reasons. The Chenango River between the Erie Railroad and DeForest Street in Binghamton has been relocated and deepened at least twice to accomodate dike and highway construction (Randall, 1977). According to a local contractor and dealer in earth materials (R. Murphy, oral commun., 1981) large volumes of gravel were removed before 1970 from four reaches of the Susquehanna River channel between Johnson City and Endicott by draglines that may have dug as deep as 40 feet locally. Each reach was at least 1,000 feet long.

Presumably silt settled in the resulting abnormally deep pools, but its thickness and the net effect of excavation and siltation on infiltration have not been studied.

Lining of stream channels

Several reaches of tributary streams crossing the Susquehanna River valley within the Triple Cities (Endicott, Johnson City, and Binghamton) have been encased in culvert pipe or routed in open channels having a concrete floor and sides. This was done to eliminate the meandering and bank erosion that are characteristic of natural channels but costly in areas of urban development. Randall (1977, p. 31) noted that continuation of this process would eliminate recharge from such streams and suggested that recharge could be increased if needed by digging multiple parallel channels or by replacing the oily and silty streambed sediment in other reaches by clean gravel.

Paving of recharge areas

Pavement and buildings covered a substantial part of the Clinton Street-Ballpark Aquifer in 1967, and Randall (1977, appendix E) estimated that, as a result, about 2 inches of recharge that would have occurred annually under natural conditions was lost as storm runoff. However, he also estimated that evapotranspiration had been reduced by at least 4 inches because of lowered water tables due to intensive ground-water development and through the replacement of soil and plants by buildings and paved surfaces. If so, the amount of water available for pumping under the degree of development prevailing in 1967 exceeded that which would have discharged naturally to streams had the city not been there.

Landfills

Innumerable studies have shown that leaching of municipal refuse by infiltrating precipitation or rising ground water produces a strong chemical solution characterized by a large oxygen demand, an offensive odor, and several thousand milligrams per liter of dissolved solids (Zanoni, 1971). Traces of leachate have been recognized at distances of several thousand feet from landfills in permeable glacial outwash (Kimmel and Braids, 1974).

In the 1950's and earlier, some industrial wastes reportedly were dumped east of Charles Street in Binghamton, above part of the Clinton Street-Ballpark aquifer. In 1958, traces of chemicals ascribed to the wastes reportedly were detected in nearby GAF wells 2 and 4 and led to the abandonment of these wells, although no problems were reported at other nearby wells (Randall, 1972; A. Schmidt and others, GAF Corp., oral commun. 1965).

In the late 1960's, the Broome County Board of Supervisors considered establishing a regional municipal landfill in a large gravel pit on Prentice Road in Vestal, which offered advantages in location and capacity.

However, the edge of the pit was only a few hundred feet east of Vestal municipal wells 4-2, 4-3, and 4-4, and the likelihood of adverse impact on ground-water quality was a major factor leading to a decision to look for a site elsewhere (R. J. Martin, consulting engineer, written commun. 1969).

In the 1970's the Village of Endicott operated a large landfill immediately west of their sewage-treatment plant near the north bank of the Susquehanna River. No problems due to ground-water contamination have been reported (R. Austin, Broome County Health Dept, oral commun. 1981). Note, however, that during construction of the Endicott sewage-treatment plant, dewatering pumps withdrew about 5,000 gallons per minute to lower the water table 25 feet (Randall, 1972), which demonstrates presence of a highly productive aquifer; the dedication of this area for a sewage-treatment plant and landfill may have precluded for the time being any use of that aquifer.

Organic fluids

In urban areas such as the Triple Cities, hydrocarbon fuels (including gasoline, kerosene, and fuel oil) and liquid organic chemicals (including solvents and cleaning fluids) are widely used, transported, and stored. Although most of these liquids have slight solubility in water, the minute quantities that can dissolve cause objectionable tastes and odors in drinking water and (or) have been shown to be toxic. Some are volatile to the extent that vapors evolved from films floating on the water table can be explosive or toxic in basements. Over the years, several instances of leaks, spills, or disposal of these liquids have been reported, and the geologic setting of each has influenced the outcome.

In 1965, a petroleum pipeline crossing the State University of New York campus in Vestal was ruptured by a backhoe excavating for new buildings; 29,000 gallons of gasoline erupted and flowed downslope toward the Susquehanna River. About half of it was recovered within hours by pumping from the initial excavation and from ditches and sumps dug downslope (Binghamton Press, Sept. 1965; C. J. Yablonski, Sun Pipeline Co., oral commun., 1981). No ground-water contamination was reported, probably because the rupture and runoff of gasoline was confined to an area underlain by impermeable till in which most buildings were served by public water systems.

In 1979, a leak was discovered in a large tank used to store solvents at a factory in Endicott. Investigation disclosed more than 12,000 gallons of chlorinated hydrocarbons (chiefly methyl chloroform and trichloroethylene) in the subsurface (Dames & Moore, 1980). More than 100 observation wells were drilled to define the extent of the contamination and help devise a method of recovery (Dames & Moore, 1980). Efforts to recover the solvents have been simplified by the stratigraphy in the immediate locality, which consisted of a surficial gravel generally 15 to 35 feet thick having a saturated thickness of 3 to 15 feet, resting upon a few tens of feet of silt and clay (Dames & Moore, 1980; Randall, 1970, fig. 5). The silt and clay must have greatly retarded downward migration of the solvents, which are heavier than water and sank to the bottom of the surficial aquifer (Dames & Moore, 1980). The surficial aquifer is not tapped as a source of water locally but seems to be continuous laterally with thicker aquifers elsewhere along the valley.

In 1980, water samples collected from public-supply wells and numerous points in the water-distribution systems of the Triple Cities were analyzed for several organic chemicals (New York State Department of Health, written commun. 1981). Samples from Vestal municipal well 4-2 on Prentice Road consistently contained 1,1,1-trichloroethane, trichloroethylene, and tetrachloroethylene, in concentrations as high as 217, 92, and 14.8 micrograms per liter, respectively (New York State Department of Health, written commun, 1981). The New York State Department of Environmental Conservation has set a limit of 10 micrograms per liter for discharges of trichloroethylene to potable ground water; no standards have been set for the others (New York State Department of Environmental Conservation, 1980). The New York State Department of Environmental Conservation (1980) and Parratt-Wolff, Inc. (1980) have investigated the contamination, and both report that one possible source is a chemical plant 200 feet west of well 4-2 that repackages chemicals including 1,1,1-trichloroethane, trichloroethylene, and other chlorinated solvents, and has discharged water from rinsing empty containers to a leach pit. In this locality, the glacial drift consists almost entirely of sand and gravel of variable silt content (Parratt-Wolff, 1980; Randall, 1972), which may have favored migration of the organic solvents to the depth of the well screen.

A LIMITED ALTERNATIVE: THE BEDROCK AQUIFER

Most studies of ground water in the Susquehanna River basin in New York have emphasized the stratified-drift aquifers because of their potential for high-yield wells, even though these aquifers underlie no more than 15 percent of the basin. Bedrock, chiefly siltstone and shale of Devonian age, underlies 100 percent of the basin and has been tapped by many wells to meet domestic and other small yields. A well drilled at any point in the basin stands a very good chance of obtaining enough water for a single-family home, although "dry holes" yielding almost no water are occasionally reported (Wetterhall, 1959; Soren, 1963; Randall, 1972). Wetterhall (1959) reports the average yield of wells tapping bedrock in Chemung County to be 8 gallons per minute. This average may be somewhat misleading, however, for two reasons: First, it is based on mostly domestic wells in which drilling was stopped as soon as the owner's needs were met, without attempting to determine the maximum yield of fresh water obtainable from the bedrock. Second, no one has studied the yield of the bedrock aquifer to clusters of wells, as distinguished from the yields of individual wells. One might ask: if 200 homes were built on contiguous half-acre lots and were supplied by individual wells, would the yield of all these wells average close to 8 gallons per minute? It seems reasonable to expect the cone of depression to be deep enough that wells near the center would have smaller yields than wells near the perimeter, but how much smaller cannot be easily predicted at present.

Near the Triple Cities, residents of some planned or unplanned concentrations of homes that were initially supplied by private wells later voted to form water districts and import public water, in part because of fear of water shortages. Study of several such clusters should lead to at least a semiquantitative relationship between maximum drawdown and cluster density, total demand, and topographic setting. In principle, it should be possible to specify limits of size and location within which a cluster of individual wells could be expected to function indefinitely, thus avoiding the cost of providing duplicate water systems and making maximum use of local resources before importing water.

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ROAD LOG
HYDROLOGY IN RELATION TO GLACIAL GEOLOGY, SUSQUEHANNA VALLEY, BINGHAMTON TO OWEGO

[Route and location of stops are shown in Figure 7]

CUMULATIVE MILEAGE	MILES FROM LAST POINT	ROUTE DESCRIPTION
0.0	0.0	Intersection of Route 434 (Vestal Parkway) and main entrance of State University of New York campus in town of Vestal, New York. Proceed west on Route 434 to Owego.
11.7	11.7	Marshland Road on right; continue on Route 434; note level terrace surface on right.
12.7	1.0	Pause on road shoulder.

HESITATION STOP. Pit on left, in the highest (oldest) stratified drift in this part of the valley: cobble-boulder gravel to coarse sand, variably silty (piles of coarse gravel rejected by operator are visible); pebbles are relatively drab (few limestones); topography is irregular, maximum altitude 880 feet; wet areas on pit floor suggest till at shallow depth. Barn on right marked "Tioga Manor"; trip will return past this barn.

17.7	5.0	Turn right (north) on Route 96, cross steel bridge over Susquehanna River, enter Village of Owego.
18.0	0.3	Turn left (west) on Route 17C (Main Street).
18.8	0.8	Cross bridge over Owego Creek.
19.3	0.5	Deep Well Motel is on right; Route 17C is on nearly level surface of valley train.
19.5	0.2	Entrance (on left) to pit owned by Concrete Materials, Inc.
19.6	0.1	Entrance (on left) to pit owned by C & C Ready Mix Corp.

STOP 1. KAME DELTA, BRIGHT GRAVEL. Stop 1 will be in whichever of these two pits offers the better exposures on the date of the trip. Topset and foreset beds of bright gravel have been observed beneath the valley-train surface; lower terraces are capped by bright fluvial gravel overlying fine-grained lake-bottom sediments.

--	-?-	Leave pit, turn left (west) on Route 17C.
23.8	4.2	Tioga Center; school on right at intersection.
24.0	0.2	Bear slightly right on dirt road.
24.1	0.1	Turn right at second driveway; proceed to gate at entrance to pasture and gravel pit owned by Kenneth Pipher. Be sure to close gate after entering.

STOP 2. KAMES AND ICE-CHANNEL FILLING, DRAB GRAVEL. Several pits in the kamic features in this locality revealed drab gravel and sand with very few limestone or other exotic rock types; west-dipping foresets have been observed. Maximum altitude is 860 feet, well above the bright valley-train surface to the east.

--	-?-	Return to Route 17C, turn left (north).

26.9	2.8	Turn left on Glen Mary Drive, just before Route 17C rises to cross RR.
28.1	1.2	Cross Thorn Hollow Creek, park along road.

STOP 3. GROUND-WATER RECHARGE FROM THORN HOLLOW CREEK. In 1967-68, streamflow measurements on 15 dates showed a loss of about 0.02 ft^3/s (10 gal/min) per 100 feet of channel upstream from Glen Mary Drive and 0.16 ft^3/s per 100 feet downstream. During most of the year, this stream goes dry somewhere on its alluvial fan.

Continue northeast on Glen Mary Drive.

29.1	1.0	Intersection; turn right and immediately cross RR.
29.6	0.5	Intersection; turn left.
29.7	0.1	Intersection with Route 17C; continue east (straight ahead).
30.1	0.4	Cross bridge over Owego Creek.
30.9	0.8	Turn right, follow Route 17C past county courthouse.
31.0	0.1	Drive straight ahead across bridge (Route 17C turns left).
31.3	0.3	Turn left (east) on Route 434.
34.9	3.6	Turn left on Marshland Road, go under expressway.
35.7	0.8	Note road bordered and arched over by large maples.
36.2	0.5	Tioga Manor barn on right. Behind barn is a hummoky surface 820-830 feet in altitude; a small pit revealed 1.5 feet of highly calcareous,bright, fine-pebble gravel with abundant limestone atop several feet of drab gravel.
36.5	0.3	Pause beside road.

HESITATION STOP. View to right of oval hill in midvalley, composed of (or perhaps heavily mantled with) till. Terraces at about 830 feet altitude between the road and the hill, now largely excavated, were capped by bright fine-pebble gravel and coarse sand.

38.1	1.6	End Marshland Road; turn left (east) on Route 434.

42.2	4.1	Tracy Creek; pause on road shoulder near bridge.

HESITATION STOP. The reach of Tracy Creek beneath the bridge visible to the right was dry on all eight occasions when visited by the U.S Geological Survey in the summers of 1962-65 as part of a study of low streamflow in the Susquehanna River basin. Like most tributary streams, Tracy Creek suffers severe seepage loss when crossing the stratified drift in the Susquehanna River Valley. Upstream 2,000 feet, where the creek has cut a gorge through till and bedrock, a small flow continued throughout the 1962-65 drought.

44.6	2.4	Cross Choconut Creek.
45.1	0.5	Turn left at traffic light on Bridge Street.
45.3	0.2	Pass under expressway, turn left on narrow road.
45.4	0.1	Turn left on dirt road; Vestal municipal well is in well house on right.
45.6	0.2	Cross dike, bear right; park.

STOP 4. VESTAL WATER DISTRICT 1 WELL FIELD. Three municipal wells tap sand and gravel between 70 and 150 feet in depth. Induced river infiltration is potentially an important source of recharge; temperature profiles in April 1981 indicate river water infiltrated past well 1-3 at a time when only well 1-2 was in use. The sand and gravel seems to be relatively drab, perhaps totally drab near the base.

46.1	0.5	Return to Bridge Street, turn left.
46.2	0.1	Turn right (east) at traffic light on Old Vestal Road.
49.2	3.0	Turn left on Prentice Road.
49.5	0.3	Park beside road.

STOP 5. VESTAL WATER DISTRICT 4 WELL FIELD. To the west is municipal well 4-2, water from which was found to contain organic solvents in 1980 (see text). To the north are two more municipal wells. To the south are numerous oil tank farms. To the east, the land was once level with Prentice Road, but has been excavated for gravel; the owner plans to mine gravel below the water table by dragline both east and west of the road. The pit to the east was once considered as a site for a municipal landfill

--	-?-	Turn around, return to Old Vestal Road.
49.8	0.3	Turn left (east) on Old Vestal Road.

51.5	1.7	Turn left on ramp up to Route 201, toward Johnson City.
52.2	0.7	Traffic circle on north side of bridge, continue on 201 (2/3 of the way around the circle).
52.9	0.7	Bear right on ramp, join Route 17 east.
53.9	1.0	Take exit for Stella-Ireland Road.
54.1	0.2	Turn left (north) on Stella-Ireland Road.
55.0	0.9	Flood-control reservoir is on right, on Little Choconut Creek.
55.5	0.5	Turn left at traffic light (Lewis Road); then immediately turn right on Rhodes Road.
56.0	0.5	Top of steep grade, stop beside road.

STOP 6. VIEW OF HOUSING CLUSTER. All homes in this area have public sewers and on-lot wells tapping bedrock. In 1981, water levels were as deep as 100 to 130 feet seasonally in some wells on Rhodes Road, and yields of a few wells were not as large as desired by occupants. Further building has been proposed downslope.

--	(?)	Turn around; retrace route following Stella-Ireland Road, Route 17 west, Route 201 south; cross Susquehanna River on Route 201, go past Old Vestal Road.
60.7	4.7	Bear right on ramp to Route 434 east.
61.0	0.3	Entrance to State University of New York campus.

END OF TRIP

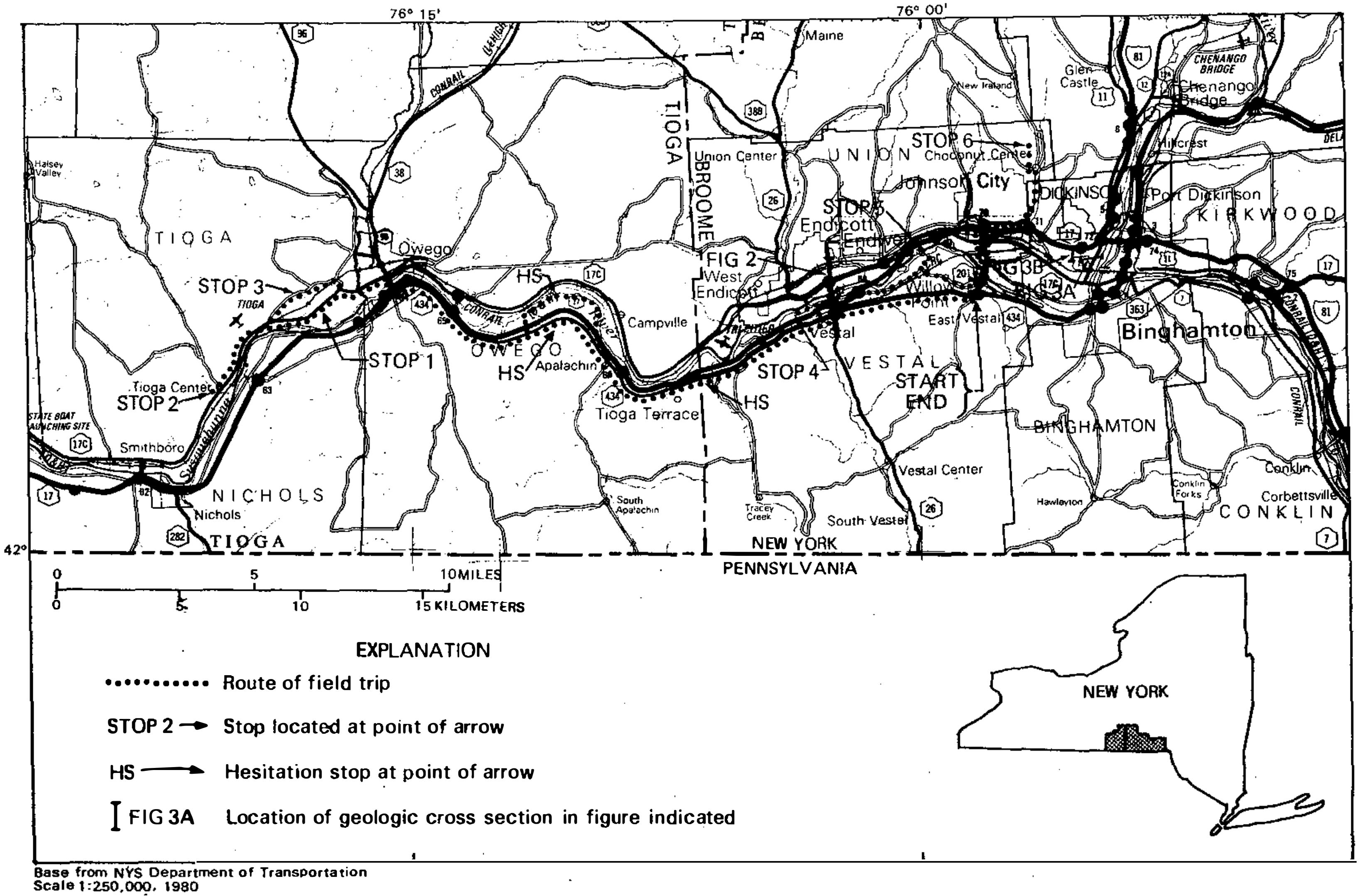

Figure 7. Location of field-trip stops and geologic cross sections.

GEOMORPHOLOGY OF SOUTH-CENTRAL NEW YORK

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PHYSICAL SETTING

South-central New York is part of the Glaciated Appalachian Plateau (Fig. 1). The topography is diverse, however, as indicated by the fact that parts of three different geomorphic sections comprise the region. Nearby adjacent physiographic provinces further lend great variety to this New York-Pennsylvania twin-tiers area.

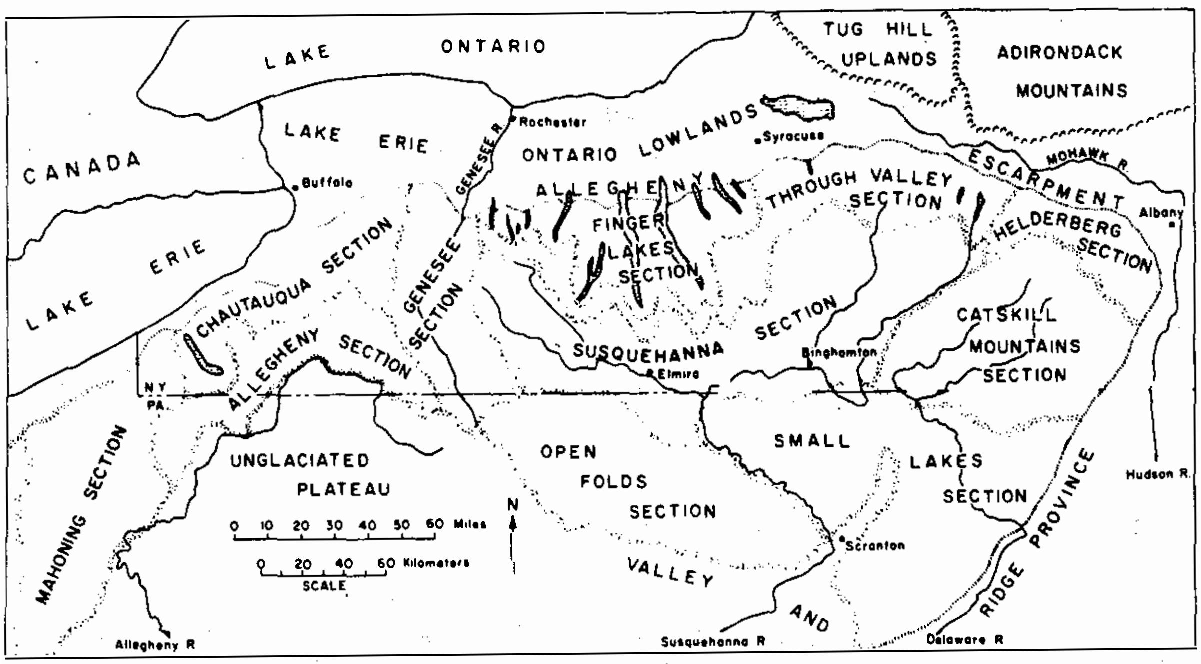


Figure 1. Map of geomorphic sections of the Glaciated Appalachian Plateau with adjacent provinces.

The Binghamton metropolitan area (of 200,000 people) is nestled in the Susquehanna section. As can be seen in Figure 2, three rivers dominate in the scene -----the Susquehanna, the Chenango and the Tioughnioga Rivers.

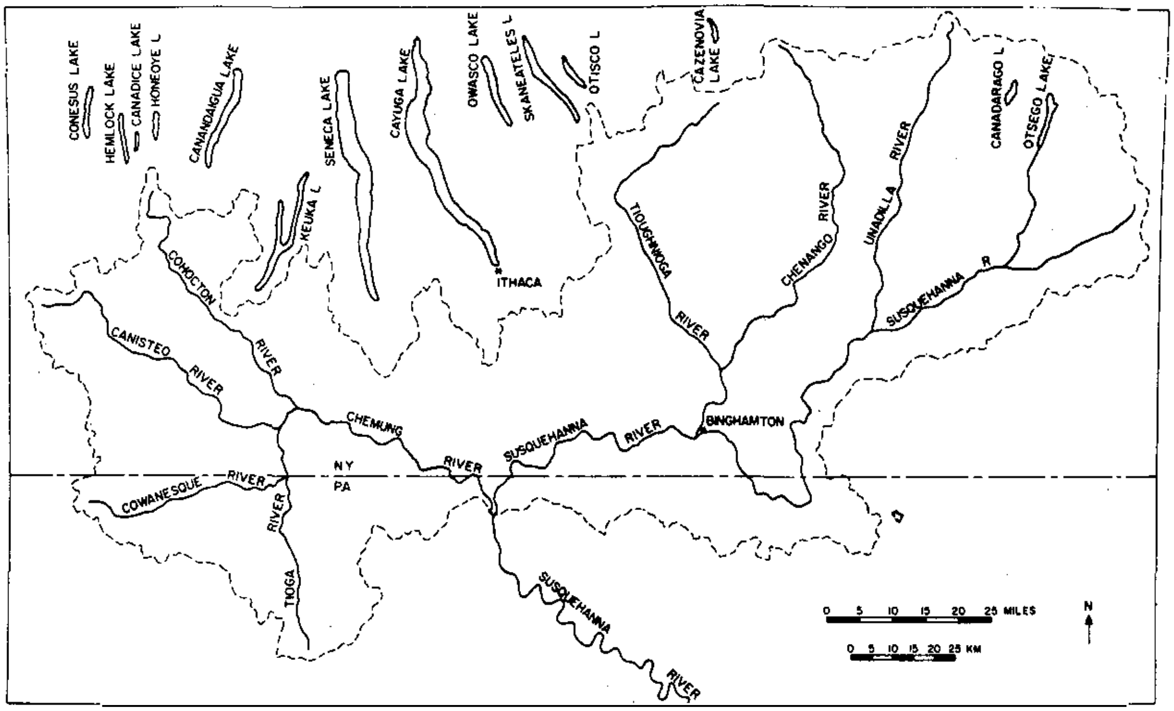


Figure 2. Map of Susquehanna River Basin hydrographic features.

In the east, the more rugged Catskill Mountain Section rises to elevations in excess of 4,000 ft. In the south, the Small Lakes Section is dotted with several hundred lakes and wetland areas. The Binghamton area lies at the heart of the region and is particularly interesting because it: (1) forms the transition zone between the higher and more rugged Catskills on the east with the more typical Appalachian Plateau to the west, (2) contains the unique configuration of the Great Bend area, (3) illustrates the problem of the evolution of two major river systems, the Susquehanna and the Delaware, and (4) contains a wealth of unusual landforms perhaps unmatched in most other parts of the plateau.

Relief in the Binghamton area exceeds 1,000 ft. with river elevations about 800 ft. and higher hills rising above 2,000 ft. A particularly unusual feature of the region is the highly erratic course of the Susquehanna River which changes directions several times before finally leaving the region at Waverly and eventually reaching sea level at

Chesapeake Bay. For millennia the landscape has been sculptured by fluvial, gravity, and finally glacial processes. Without glaciation, the region would more closely resemble parts of West Virginia with its steep ridge-and-ravine topography. Instead the New York hilltops have been reduced in elevation, and the valleys partially filled with deposits giving an overall effect of reduction in relief and hillslope steepness. Without the advantage of glaciation, the population would be much smaller because floodplains would be narrow, groundwater supply would be less plentiful, and there would be an absence of abundant sand and gravel deposits which provide an important mineral base to the economy.

PREVIOUS STUDIES

Most of the early surface geology work in the region was focused on the problem of explaining the evolution of drainage (Coates, 1963a). Three general views prevailed concerning the evolution of streams and the topography: (1) They are mostly derived from rather continuous denudation processes dating from post-Permian time. (2) They are largely post-Cretaceous and inherited from either a peneplain condition or a cover of Cretaceous sediments long since removed. (3) The features have developed from drainages that have become adjusted to the structural fabric.

The work of Tarr et al. (1909) provides one notable exception to the rule that very little glacial investigation of the region had been done prior to 1960. Rich (1935) provided the most detailed mapping of the Catskill region. Although MacClintock and Apfel (1944) and Fairchild (1925) briefly looked at the south-central New York localities, many of their conclusions have proved erroneous. Even soil surveys (as in Tioga County) have been shown as inaccurate on such matters as character and thickness of till. The following list indicates several of the early misconceptions about the region:

1. The region was often characterized as containing dendritic drainage. However, on the broad-scale drainage is arcuate with individual segments displaying pseudo-parallel patterns that also contain trellis drainage as in the upper reaches of the Chemung, Cohocton, and Canisteo Rivers.

2. Effects of glacial erosion were considered minimal. Although ice erosion was not as severe as in the Finger Lakes to the north, when combined with glaciofluvial erosion,the landscape produced by glacial processes has been greatly modified from its original fluvial character.

3. The thickness of till is minimal, ranging from 3-10 ft. Coates (1966) has shown that the average thickness of till exceeds 60 ft.

Starting in the 1960's, the SUNY-Binghamton group began to develop a series of publications and reports dealing with the surface geology of the region. Glacial features were discussed in the following works:

Coates (1963a,b, 1966); Harrison (1966); Coates, Landry, and Lipe (1971); Cadwell (1972); Coates (1973), King and Coates (1973); Kirkland (1973); Coates (1974); Coates and Kirkland (1974); Fessenden (1974); Coates (1976b); Aber (1976, 1979); Caprio (1979); Gillespie (1980); Stone (1981); and Phelan (1981). Hydrology characteristics were emphasized by Coates (1972); Flint (1968); and Rideg (1970). Topographic relations were stressed in Coates (1963a, 1963b); Conners (1969); and Coates (1976a).

BRIEF GEOLOGY HISTORY

Bedrock in the region is predominantly composed of clastic sedimentary strata of Upper Devonian age. West of Windsor, except near hilltops, rocks are mostly shale and silitstone of marine origin. East of Windsor and south of Great Bend rocks are largely sandstone of terrigenous origin. All rocks were part of the Catskill delta-alluvial-plain environment. In this region conglomerates, red beds, and limestone are exceptionally rare.

Although the rocks appear to be horizontal at any given outcrop, when viewed on a larger basis the architectural flavor of the region is a homocline with a regional dip of 10 to 40 ft/mi in a southerly direction. The strata have also been gently flexed into a series of east-and northeast-trending folds, but with dips less than 1^{o}. The Catskill Mountains are generally conceived as a gentle synclinorium with axes that plunge southwestward from the upturned rim on the east.

The topographic character of south-central New York has been sculptured by many millennia of fluvial and gravity processes after the region was uplifted. Then, finally, during the past few million years glacial processes have placed in indelible overprint on the terrain. The key to this evolution has been the development of the Susquehanna River and its bizarre geometry....probably the most erratic of all American rivers (Fig. 2).

A variety of hypotheses have been presented in attempts to explain the anomalous course (for a fuller discussion of these ideas see Coates, 1963a, p. 22-24). For any hypothesis to be valid, it obviously must be in accordance with facts. Among these should be the consideration that: (1) The flow of several rivers is contrary to the regional structural fabric, such as the direction of flow of the Tioughnioga, Cohocton, Canisteo, and Chemung. (2) The arcuate character of the major part of the Chemung and Susquehanna Rivers in New York (Fig. 2), is an anomaly. (3) The magnitude of post-Cretaceous denudation could exceed 1 mi, and denudation since post-Permian time could greatly exceed 2 mi. (4) The present-day structures and lithologies where streams are currently eroding may not correspond with conditions when streams were eroding at formerly higher elevations. Figure 3 provides a list of the many different ideas that have attempted to explain drainage throughout the region, and Figure 4 illustrates some of these.

Figure 3. Listing of different hypotheses as explanations for the origin Appalachian drainage.

1. The original drainage was northward, but, through a reversal, the rivers now flow south.

2. The original drainage was west and southwest due to the development of consequent streams flowing first off the Adirondacks and later off the Catskills.

3. Drainage evolution through capture processes by headward-eroding streams that became adjusted to structural trends along joints, faults, and folds.

4. A variation of No. 3 with successive stripping of various thrust belts and sheets. First proposed by Meyerhoff and Olmstead (1936) but can now be resurrected as byproduct of the new plate-tectonics and overthrust belt ideas.

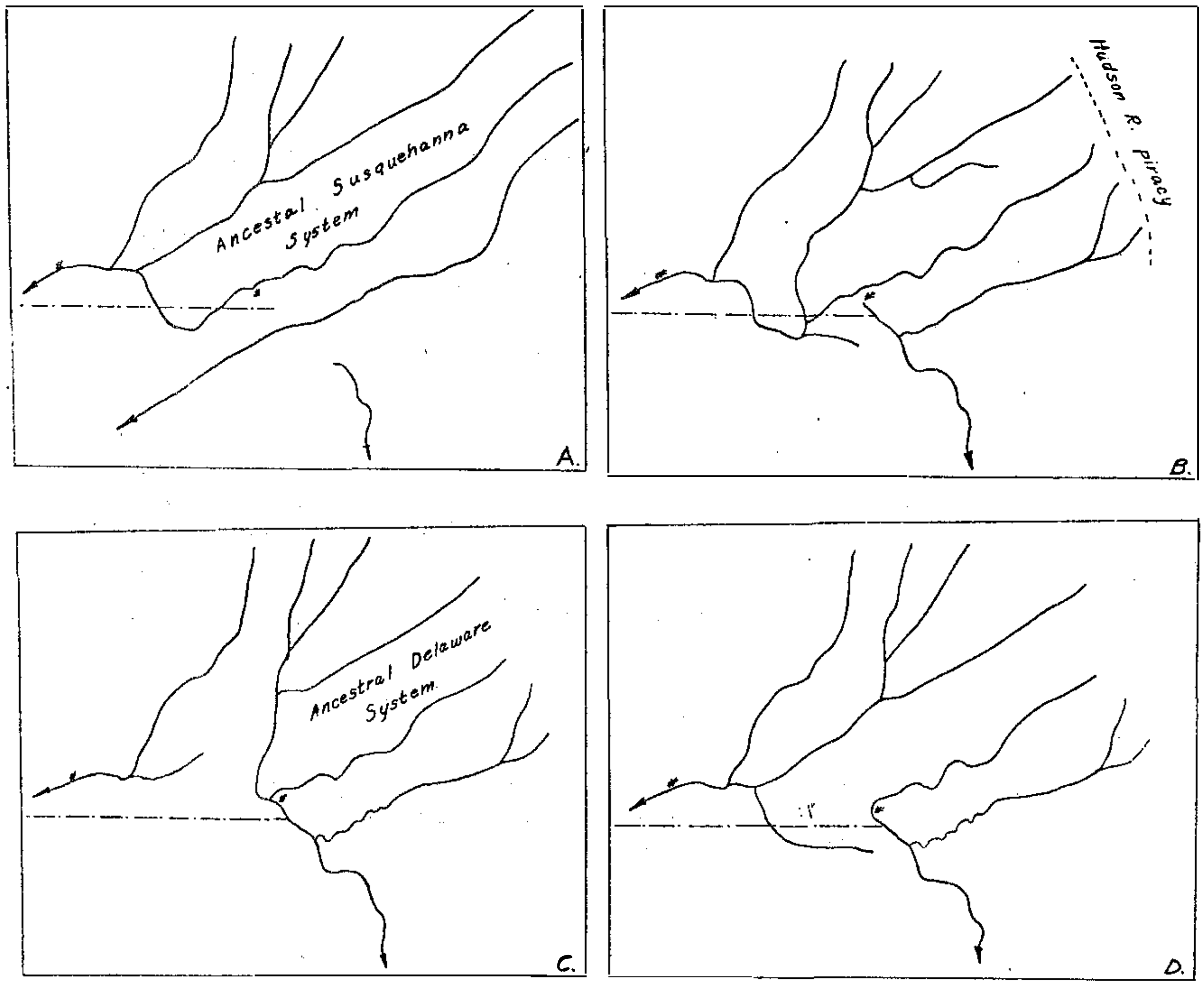
5. Development of a Tertiary peneplain with drainage superposed through the underlying sedimentary rocks.

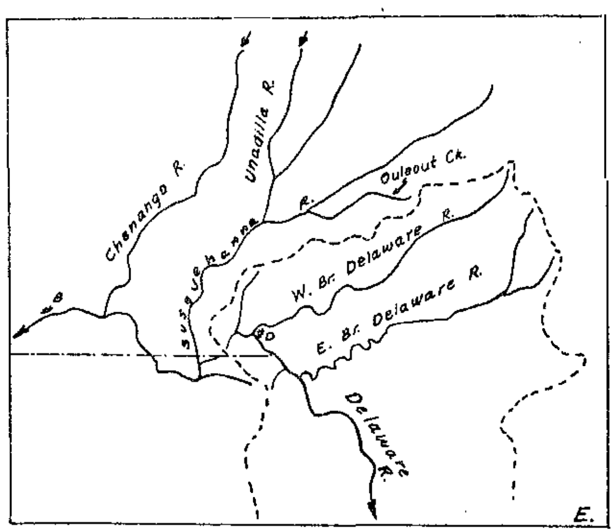
6. A regional covering by Cretaceous sediments with subsequent superposition through them onto underlying Paleozoic rocks.

7. Capture of the southwest- and west-flowing Susquehanna by the Delaware system because of the Delaware's advantage of steeper gradient and shorter distance to sea level.

8. Capture of the Delaware system by the Susquehanna system.

9. Large-scale drainage derangements caused by glacial diversions over the nearly 2 million years of Quaternary time.





A-B Ruedemann Theory. Ancestral Susquehanna is master system of primitive drainage. The Delaware as a subsequent stream captures west-flowing drainages.
C-D Mackin Theory. Ancestral Delaware is original stream of primitive system. The Susquehanna by headward growth captures northern headwaters of the Delaware.
E. Present relation of the Susquehanna and Delaware drainages in the central and western Catskills.

Figure 4. Theories of drainage development in the Catskill Mountains.

Whatever the final verdict, it is obvious the Susquehanna has had a long and checkered history. The river has probably involved a combination of consequent flow, piracy of other stream segments, and glacial diversions. For example, the turnaround of the river in the Great Bend area is a particularly vexing problem, especially when the unusual differences in valley-fill sediments are considered. Bedrock depths are generally 200 ft. or so northeast and west of Great Bend whereas in the curvature segment bedrock is less than 100 ft. deep.

The final chapter for landscape development of the region was written during the ice ages of Quaternary time. Thus about 2 million years of glacial and interglacial episodes have been available to cause a re-fashioning of what formerly had been a predominantly fluvial terrane. The Laurentide ice sheets buried the Binghamton area with ice more than 3,000 ft. thick and the glacial margin extended south into Pennsylvania for 60 additional miles. The heritage of these times is reflected not only in landform changes caused by ice and meltwater erosion and deposition, but also by periglacial activities and more recent geomorphic events.

A particularly intriguing series of landforms occur along the valley wall on the south side of the Susquehanna River between Binghamton and Waverly. Instead of a somewhat continuous series of parallel hill-slopes, there are many scallop-shaped reentrants or arcuate hollows. The SUNY-Binghamton campus is situated in one, and other good examples occur 2 mi west and again 5 mi west. It may be of importance to note that the location of the reentrants is usually in line with a south-trending drainage that joins the Susquehanna on the opposite side of the valley. Could their origin be linked to ice and meltwater erosion of valley-type ice tongues, or to a much-displaced thalweg of the Susquehanna River?

In summary, present-day landforms are a blend of erosional and depositional features that have been imprinted by many different processes acting under different environmental conditions.

GLACIAL EROSION

The movement of ice throughout the region was related to the amount of relief and the glacial source area. The main ice sheet at first advanced southward thrusting into the embayment now formed by the Finger Lakes to the north, where considerable erosion occurred. For example, as much as 1,200 ft of deepening was created in the Seneca and Cayuga troughs. South of the Finger Lakes, major valley erosion was accomplished in suitably oriented valleys, those with orientation mostly parallel with the outflow from the Ontario basin. This radial belt suggests divergence of the ice as it moved away from the confining topography of the Allegheny escarpment. However, this radiating pattern is interrupted by the arcuate geometry of the Chemung and Susquehanna Rivers and major tributaries. Coates (1974) has suggested the positions of such valleys may have been influenced by ice-margin drainage, in much the same fashion that parts of the Missouri and Ohio Rivers reflect ice-margin positions. East of Binghamton the direction of ice flow was influenced by the higher topography of the Catskill Mountains which caused a more southwesterly flow as indicated by striations in the area. Such ice flow was also enhanced during glacial maxima when in the north the ice sheet overtopped the Adirondacks producing southwesterly flow. This chain of events is described by Coates and Kirkland (1974) in which they trace the history of growth and decay of the ice sheet.

Two other aspects of the processes associated with glacial and interglacial times have served to flavor the landscape; the sequential series of landforms that have developed and the accentuation of forms with each succeeding glacial and interglacial stage. The following evolution of landforms was described as a continuous hierarchy of features throughout the region (Coates and Kirkland, 1974): (1) notch stage, (2) col stage, (3) overflow-channel stage, (4) sluiceway stage, (5) through-valley stage, (6) finger-lake stage, and (7) composite-valley stage. The notch stage forms when proglacial meltwaters create a V-shaped entrenchment at a drainage divide. Continued erosion, aided by advance of the ice,serves to hollow out the notch to produce a gentler and flatter floor during the col stage. Depending on the vigor of the processes and character of preglacial terrane, an overflow-channel stage or spillway may be formed. This occurs when a flat-floored channel extends immediately down-gradient from the drainage divide. The sluiceway stage evolves when the meltwaters have moved some distance from the divide and are deeply entrenching a gorge at lower elevations. Such valleys are larger, longer, and lower than the overflow channels. These long, narrow, steep-walled chasms at first only contain minor tributaries which provide evidence for their glacial rather than fluvial heritage and origin. For example, the anomalous morphology of such valleys proves they did not form through a normal fluvial cycle with headward erosion and the development of a nicely adjusted and articulated drainage network. Instead,the glacial meltwaters incised new terrain that is dissimilar to adjacent fluvial systems. The sluiceway north of Waverly and the sluiceway south of New Milford, Pa., are excellent examples of this development (Fig. 5). The most complex of these forms is the one associated with the Tioughnioga River. This sluiceway has a "beaded" valley configuration whereby amphitheater-type valley segments alternate with severely narrowed and constricted portions. Such aberrant features,when added to the unusual southeast valley gradient (which goes against the topographic grain of the region), indicate the Tioughnioga is a multicyclic sluiceway that required more than a single glacial episode for its formation. Thus it was repeatedly occupied by meltwaters during various glaciations.

Another landform created by glacial meltwater erosion is the umlaufberg (Fig. 6). There are many of these features with various sizes and shapes through the region. In this region, umlaufbergs are bedrock outliers within a valley surrounded by glacial drift that is

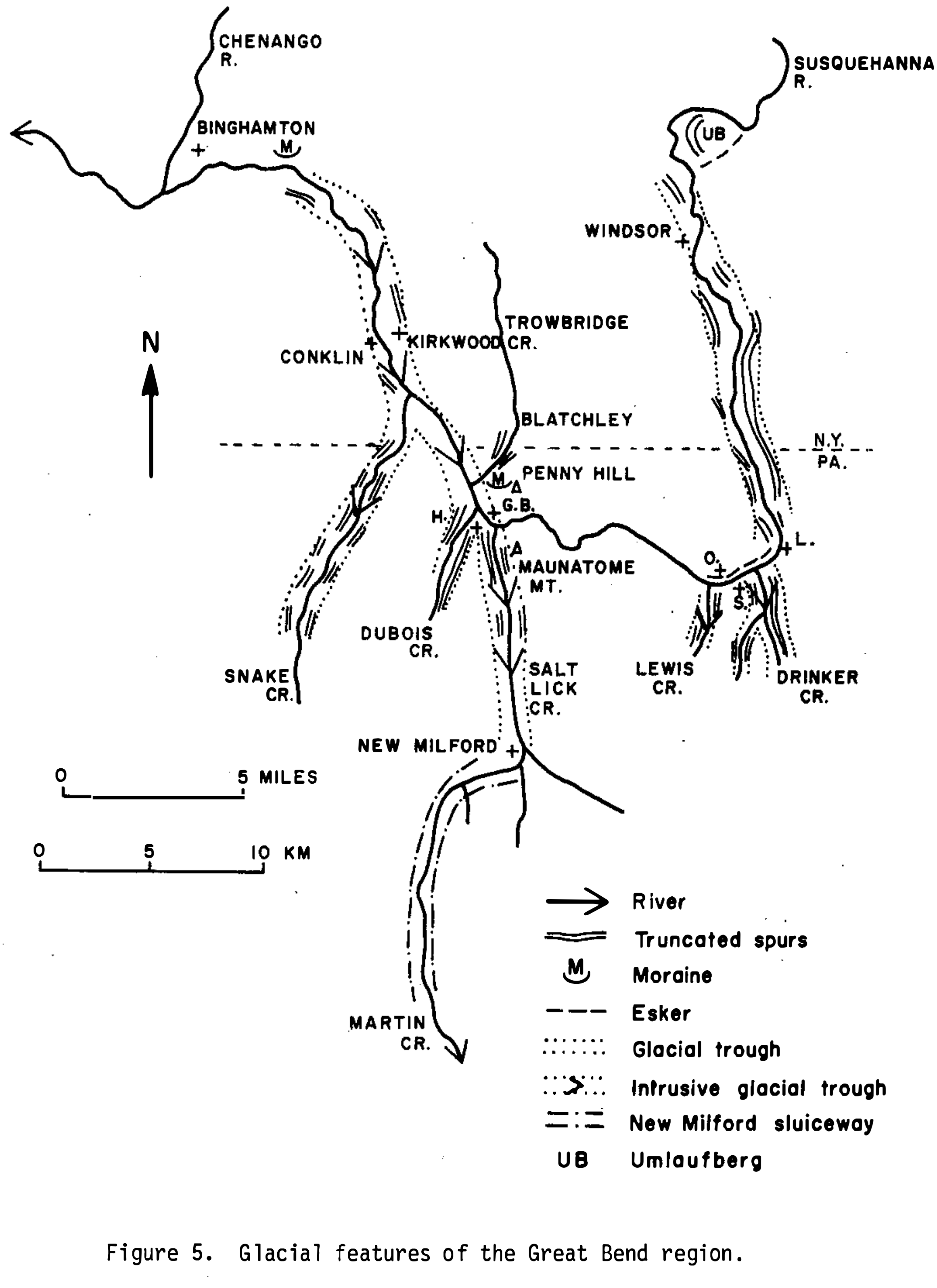


Figure 5. Glacial features of the Great Bend region.

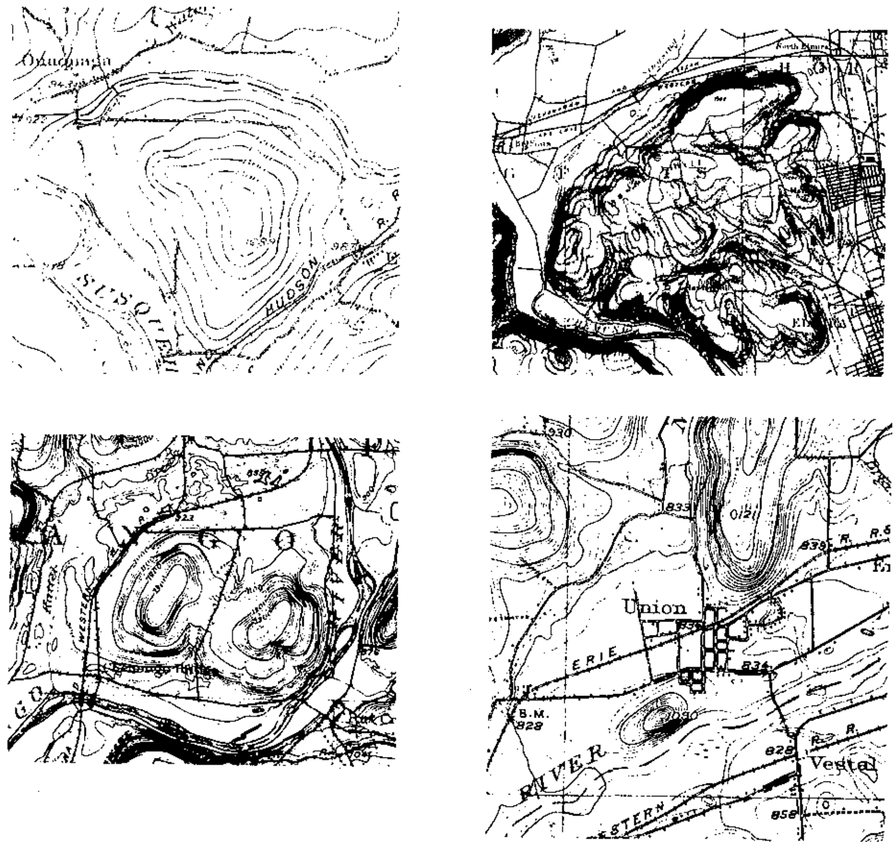


Figure 6. Maps showing variations in umlaufberg size and perfection in development. Copied from U.S.G.S. topographic maps, scale 1:62,500.
A. Single cycle. Susquehanna River near Windsor (Ninevah Quad.). Umlaufberg is 1.6 mi long.
B. Double cycle, Chemung River near Elmira (Elmira Quad.) Umlaufberg is 5 mi long.
C. Multicycle. Chenango River near Binghamton (Binghamton Quad.) Umlaufberg is 2 mi long. Note that it is double crested.
D. Multicycle end member. Susquehanna River at Union. (Apalachin Quad.) For scale the "Union" lettering is 0.4 mi long.

typically stratified. Their common bond is their origin. They form when an ice margin forced meltwater drainage to flow over what formerly was part of a bedrock spur or divide. Such drainage diversion produced a new entrenched channel, and upon retreat of the ice a bedrock knob remained with valleys surrounding it. Umlaufbergs can be important in deciphering glacial history because they range in type from those formed by a single glacial episode to those that are multicyclic (Coates, 1974). The size and shape of the umlaufberg, the character of the surrounding valleys, and the degree of tributary development in the contiguous terrain aid in determining the relative age of the features. The best examples in the Binghamton area occur at Roundtop Hill, Vestal; Roundtop, Endicott; and at Chenango Bridge. There is even a completely buried bedrock hill at the confluence of the Susquehanna and Chenango Rivers which means during former times the Chenango River flowed southwest instead of south at that position.

GLACIAL DEPOSITION

There is a large variety of glacial deposits and their landforms in south-central New York. The materials include different types of till, as well as both ice-contact and outwash stratified sediments of meltwater origin. Furthermore each of these sediment types comprise special landforms that are distinctive in their morphology.

Till Composition

Two distinct types of till occur in the region. The Olean or "drab" till was named from its type locality at Olean, N. Y., and the Binghamton or "bright" till was described for our area of the same name. MacClintock and Apfel (1944) did the first study of these materials and argued they were formed by separate ice sheets with the Olean being older. Olean till is clay-rich with a high percentage of local rocks in the gravel-size range. The erratic count (or "exotic" rocks) is generally less than 8 percent of far-travelled materials. Such composition imparts a dull or "drab" appearance. In contrast, the Binghamton till is more sandy, contains clasts that are more rounded, and may contain erratic gravel that comprises 20 percent or more of this size range. These exotics (rocks from outside the drainage area) are generally limestone, chert, red sandstone, and igneous-metamorphic crystallines. The Valley Heads drift, which occurs north of our study region, has many similarities with the Binghamton composition. Denny (1956) was the first to challenge the age concept of MacClintock and Apfel because he found no evidence of Binghamton drift in the Elmira area (also see Denny and Lyford, 1963). Instead he proposed that the Binghamton was contemporaneous with the Olean and had formed as the result of incorporation of erratic-rich valley gravels. Moss and Ritter (1962) also support this view and used pebble lithologies, heavy-mineral coatings, and texture ratios as documentation. Coates (1963a) described the Binghamton drift as simply the valley facies of the Olean which was usually an upland till. Thus these units were regarded as end-members of a lithologic continuum of Woodfordian age (Late Wisconsinan). In a study of valley-fill sediments and well borings, Randall and Coates (1973) show that ice-margin oscillations occurred, and the presence of transitional drift types indicate only a single major glaciation period.

More recent studies of upland tills in the region have revealed some new insights on their characteristics. Aber (1976) studied the lithology of upland drift by examining cuttings from deep water wells. He discovered gravel units interbedded between till units and also found interstadial conglomerates in the material. He ascribed these units to different time zones, interpreting the section as representing an older Wisconsinan event that was subsequently covered with a newer Woodfordian event.

In attempts to obtain greater resolution and understanding of upland drifts, Caprio (1979) and Gillespie (1980) undertook quantitative mineral and chemical analyses of tills to discover if there were significant differences, and, if so, whether such variations could be attributed to time lines. These studies showed that the amount of exotic rock-mineral matter in the drift is a function of textural size. In general, there are higher ratios of exotics for smaller grained materials. Even the sand-size material in Olean drift may contain appreciable amounts of exotics. The use of a ratio between purple and red garnets was also found to be instructive. For example, high ratios (more purple than red) indicate a primary provenance from the Montreal, Canada, region via Lake Ontario and then southward transport of materials. The lower ratios (with increased red garnets) indicate provenance from either the north and central Adirondacks or the Canadian shield west of Toronto, Canada. These detailed studies are supportive of the glacial-flow model developed by Coates and Kirkland (1974) for movement directions of the ice sheet during advancal and retreatal phases.

The Laurentide ice sheet invaded the region, moving through the St. Lawrence Valley and Lake Ontario, then spread out, radiating southward from the Ontario Basin. The Adirondacks acted as a barrier to flow until ice thickness reached sufficient depths to overtop the mountains, which then become an outflow center. During deglaciation and ice thinning, glacial movement once again was restricted to flow through the St. Lawrence and Ontario areas. This chain of events would be largely repeated during each glacial stage of Quaternary time. During interglacial stages, the valley-fill sediments become reworked by fluvial processes and lay in wait for the next ice inundation to incorporate the gravels into the newest drift.

A problem still remains concerning what to do with the name "Olean". As Coates (1976b) indicated, it should only be used as a lithologic description for one type of drift. It cannot be used in a time-stratigraphic sense because the age of such materials range from perhaps 50,000 ybp (years before present) in some parts of the state to 13,000 ybp in other parts. Clearly it represents several different time episodes.

Till Distribution and Landforms

Till is widespread throughout the region and nearly ubiquitous on all hillslopes and uplands. However, it is not of uniform thickness. Although average till thickness is more than 60 ft, south-facing slopes contain six- to ten-times-thicker drift than north-facing slopes. The steep slope south of the SUNY-Binghamton campus is representative of those sites with thin till. Till thickness is also small in the high-level cols across which ice passed. The upland topography away from major valleys contain hills with pronounced asymmetry, with south-facing slopes only half as steep as those on the north. Such hills in the Binghamton area have an average relief of 340 ft. The asymmetry is a function of the amount of underlying drift. Thus the till thickness is usually more than 90 ft. on south slopes but less than 20 ft. on north slopes. At sites such as near Hawleyton (south of Binghamton) and at West Windsor (east of Binghamton) till thickness reaches 250 ft. These asymmetric topographic forms have been called "till-shadow hills" (Fig. 7) by

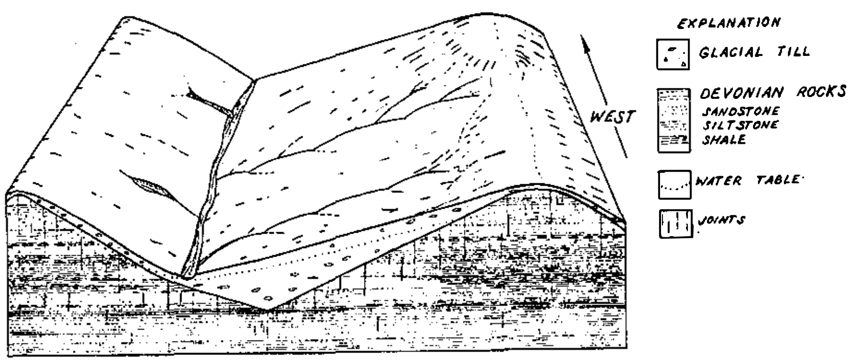


Figure 7. Diagrammatic sketch of a typical till-shadow hill.

Coates (1966) because the thick drift on the south is clearly a function of glacial deposition on the down-current side of an obstacle (in the same manner that water and wind deposits form on the lee side of a rock).

Till is also the primary constituent of several other landform types. Hubbard (1906) was the first to describe mounds of till in valleys which he compared to drumlins. He called them "drumlinoids", although they do not possess elements of streamlining. These small hills are rarely more than 60 ft high and contain slopes of 12-15 percent. They are restricted to valleys with roughly north-south orientation and occur as isolated hillocks within the valley. Castle Creek and Choconut Creek are typical sites where seven to eight of the forms occur along the valley trend. A detailed study has not been done of these forms, but it is known their composition is largely till with thicknesses to 100 ft.

Several other landforms comprised of till occur in the region. There are more than 20 examples of arcuate hollows near drainage divides of south-sloping valleys where lakes or wetlands have till mounds or ridges that help form the depression. St. Johns Pond (Fig. 8), Beaver Lake, Ansco Lake, Sky Lake, and Deer Lake are typical sites. Between Binghamton and Elmira, several north-draining valleys contain blockages of till hills. During deglaciation imponded waters behind the till plugs were forced to incise new channels, and in some cases, as at Tracy Creek, the new channel missed the old valley and was instead cut into bedrock. The

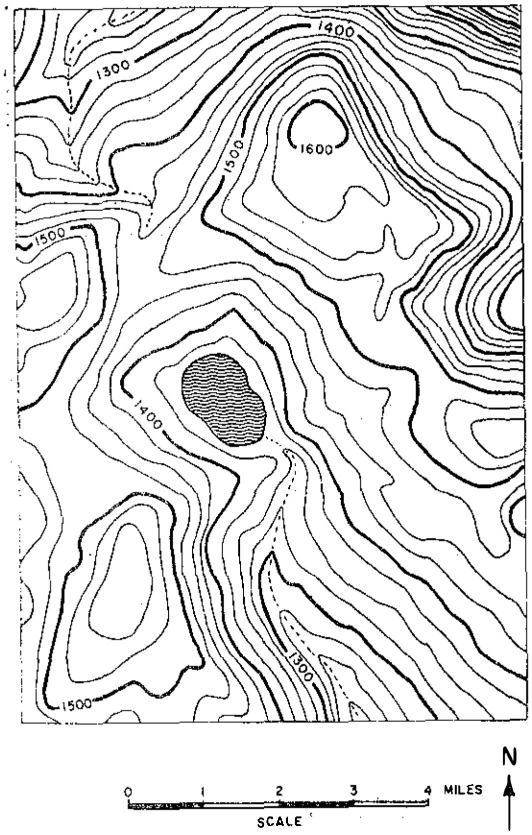


Figure 8. Simplified topographic map of St. Johns Pond area, north of Binghamton, N. Y. Drawn from U.S.G.S. Castle Creek, Quadrangle, 1:24,000 scale.

SUNY-Binghamton campus is built on still another variation of till landform...a ribbed moraine. This feature is a transverse ridge constructed of glacial deposits as the ice was ascending the valley sideslopes during its southward movement. Other examples of these moraines also occur in the region.

GLACIAL MELTWATER DEPOSITS

Glaciofluvial and glaciolacustrine deposits are largely confined to the valleys, but occur locally on valley walls. Valley fill in the major rivers, as the Susquehanna and Chenango, is rarely more than 200 ft, and for smaller streams is rarely more than 100 ft. The economic significance of such deposits should be stressed because the glacially-deposited sands and gravels are not only important mineral resources, but, where below the water table, comprise the only significant aquifers in the region. The texture of these deposits can be dependent on valley orientation as Coates (1972) has pointed out, and can even influence stream regimes (Fig. 9). For example, valley fill in south-draining

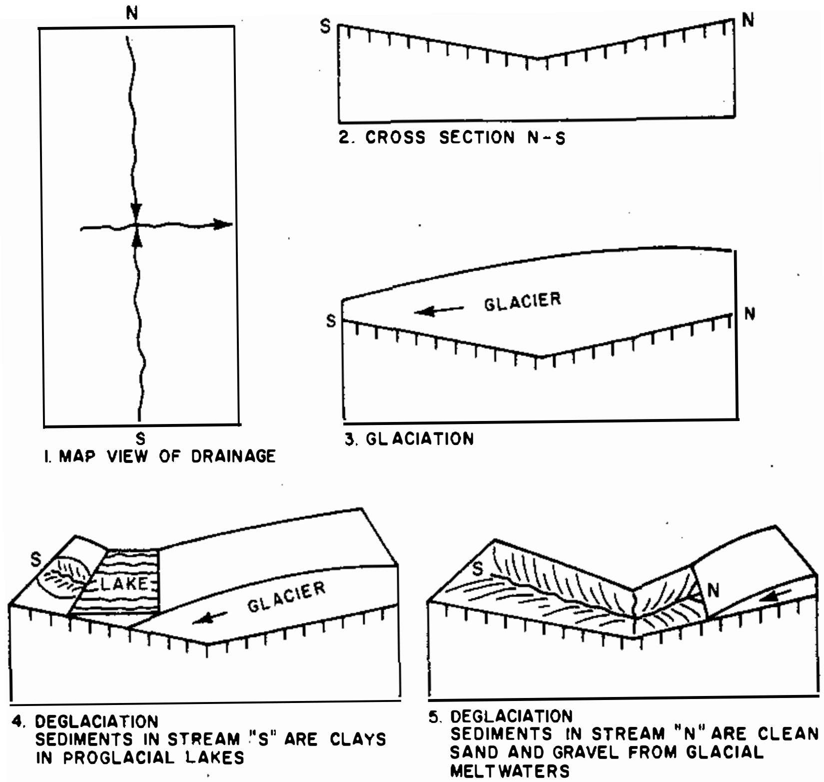


Figure 9. Diagrammatic sketch illustrating differences in meltwater sedimentation in north- and south-flowing stream channels.

valleys is coarser because the fines could be removed by free-draining meltwaters. However, in north-draining valleys, valley fill is finer because of increased blockages and ponding of water which prevented the flushing out of the fines.

South-central New York has a rather complete array of meltwater sediments and landforms they create. Good examples are abundant for both ice-contact deposits and outwash.

Eskers

The two largest eskers are those at Center Lisle (west of Whitney Point) and Oakland, Pennsylvania,(20 mi. southeast of Binghamton). The eskers are several miles long and reach heights of 100 ft. Several smaller eskers are present at some of the stagnant-ice sites such as north of Windsor and at Apalachin.

Kames

These hills composed of ice-contact sediments occur at numerous localities including Apalachin, Kattelville, Whitney Point, Windsor, Gulf Summit, Chenango Forks, and Tioga Terrace. They contain a wide assortment of sediment sizes and bedding characteristics. Many are composed of drab drift with appreciable contents of silt and clay. The best kame terrace and valley train in the region is immediately north of Binghamton at Hillcrest. It can be followed for several miles and forms a rather level surface upon which the community has been built.

At several localities, there is a mixture of both ice-contact features and outwash materials. The best example of this is the topography that forms Chenango Valley State Park and adjacent areas. Here 'dead-ice terrane' is prevalent throughout a several square-mile area. Much of the park contains many tens of kettle holes that formed when ice blocks melted within outwash materials creating a pitted outwash plain. Kame and kame-delta deposits occur both north and south of the park. Other occurrence of these features are in the dry valley north of the Chenango Bridge umlaufberg - the site of the former thalweg of the Chenango River.

Outwash Deposits

Outwash forms the great majority of meltwater deposits in terms of area covered and volume of sediments. These materials constitute the most productive sand and gravel operations (Stone, 1981), and are important aquifers for the villages of Endicott, Johnson City, and Vestal. These deposits are restricted to valley environments so their landform equivalent is referred to as a "valley train". Sediments in the Barney & Dickenson gravel pit typify such deposits. Here the materials are commonly well sorted and show cross bedding and other features indicative of the changing character of the meltwater streams that formed them.

Because there is an absence of well-defined recessional moraines in the region (as compared with the excellent development of such landforms in the Midwest) early workers argued whether the decay of the ice sheet occurred by a process of downwasting or backwasting. Cadwell (1972) and Kirkland (1973) tested the morphologic sequence model, first used successfully in New England, to determine if the same deglaciation characteristics were appropriate for the New York region. Indeed the model works. For example, in the Chenango Valley, Cadwell (1972) showed there were six primary zones of backwasting, with several smaller episodes within each major set (see Cadwell, this guidebook). Thus each zone is marked by a distinctive meltwater sequence that includes kames, kame deltas, and outwash materials all graded to the margin of an ice tongue in the valley.

PERIGLACIAL FEATURES AND HOLOCENE TIME

Periglacial processes can be important terrain sculptors in regions contiguous to ice margins or at higher elevations above the glacial ice. There is a variety of features that developed from periglacial activity in the Binghamton region. These include modest-size block fields, tors, patterned ground, and solifluction lobes. These landforms are especially well expressed in the Windsor quadrangle (USGS 1:24,000). King and Coates (1973) have described some of the forms, and especially call attention to the concavo-convex slope profiles which are so common throughout the area (Fig. 10). For example, southwest of

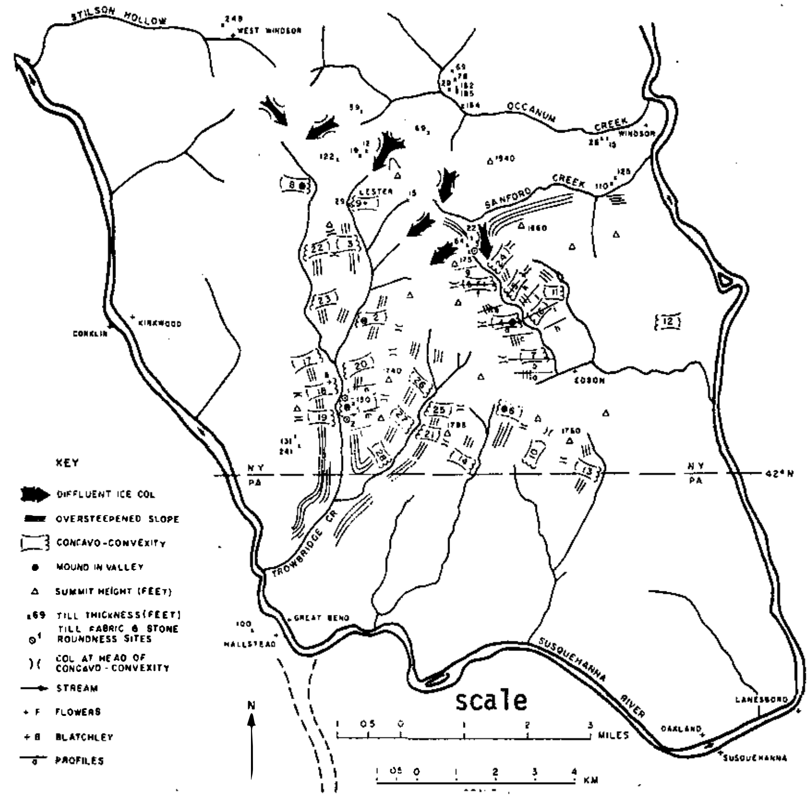


Figure 10. Distribution of periglacial-type concavo-convex landforms in the Windsor-Great Bend area of the Susquehanna. (From King and Coates, 1973).

Windsor, thick till deposits mantled the slopes along the tributaries of south-draining valleys. During the vigorous periglacial period, and probably prior to complete forest development, the hillslope materials underwent extensive creep and spread out as lobes into the valley. These solifluction lobes caused streams to take winding courses between the east and west competing lobes.

Another indication of the absence of slope-stabilizing vegetation in the region is the occurrence of many alluvial fans. These developed in postglacial times when the hillslopes were mostly barren. Fluvial processes combined with sheet flow and mudflows to entrain and mobilize the surface materials and move them downslope into the larger tributaries. These deposits clogged the termini of the channels and built the alluvial fans which also diverted the channel of the master stream at the base.

Tors and small-scale rock cities can be found at a number of localities in the Binghamton area. These include the bedrock features at Ingraham Hill; Table Rock Ridge; Progy Hill; Penny Hill and Roskelly, Pennsylvania; and interfluves of Martin Creek, Pennsylvania. It is possible that other sites in the region may have undergone nivation processes with the creation of features that resemble nivation hollows and altiplanation surfaces.

The glens and waterfalls of the region also attest to a direct lineage from glacial times. Although these are fluvial features, they did not form from the usual fluvial landscape process. Instead they originated when a former fluvial valley was buried with till. During the re-establishment of drainage in postglacial times, the consequent stream failed to discover its former channel. Instead it started its new life at a position above bedrock which it thereby incised as a superposed stream. Doubleday Glen was thus carved more than 100 ft. into bedrock, and of course Watkins Glen and Enfield Glen in the Finger Lakes are prime examples of this type of heritage.

The major rivers of the region, such as the Susquehanna and Chenango, are now busily at work excavating the thick glacial valley-fill sediments. It is of interest to note that the Susquehanna does not contain the free-swinging type of meanders such as the Mississippi River. Instead there are many positions where the Susquehanna has occupied its same channel for more than 4,000 years. Other streams in the area also have unusual qualities and many are misfit. Some, such as the Tioughnioga River, are overfit, a condition where the stream should have carved a larger valley. Many other are underfit, where the valley is much too big to have formed from the small amount of water the stream now carries.

The valley-fill sediments are not exclusively glacial in age. Holocene deposits have been dated at several sites where they overlie glaciofluvial materials. At the junction of Little Choconut Creek with the Susquehanna River, floodplain silt was dated by wood fragments 6-8 ft below the ground surface as being 1690 ± 100 ybp. At another site on the west side of the Bevier Street Bridge, Chenango Valley sediments with wood and other organic materials were encountered in test well borings. Sediments at a depth of 24 ft were 2648 ± ybp, and those at 45 ft were 3801 ± 60 ybp. These deposits occurred in an environment interpreted to be kettle holes that had been infilled with more recent materials (Randall and Coates, 1973).

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ROAD LOG FOR WESTERN CATSKILL - GREAT BEND AREA

The purpose of this trip is to examine geomorphology in the region to the east and south of Binghamton. The evolution of the Chemung, Delaware, Susquehanna, and Tioughnioga Rivers will be discussed. A variety of erosional features will be seen and much emphasis will be placed on glacial deposits and the landforms they create. The starting point of the trip will be the upper part of the SUNY Binghamton campus.

CUMULATIVE MILEAGE	MILES FROM LAST POINT	ROUTE DESCRIPTION
0.0	0.0	STOP 1.

STOP 1. CAMPUS OVERVIEW. This site is the southernmost of three east-trending ridges which comprise a ribbed moraine sequence. The SUNY campus is situated on two of the ridges. Furthermore, the campus is nested in a reentrant in which the bedrock hillslopes have a scalloped form. Roundtop Hill, an umlaufberg, can be seen to the north. The drainage divide to the south contains a col, or saddle, which was the position for override by the glacier. The moraine is composed of bright till (Binghamton-type) with a fabric transverse to the ridge (or parallel with ice transport). Bedrock is more than 120 ft below land surface. All glacial deposits on this trip are Late Wisconsinan. Embark. Travel east on ribbed moraine with SUNY nature preserve and wetlands to south.

1.0	1.0	East entrance of SUNY campus. Continue east to Murray Hill Road and turn right (south).
1.4	0.4	STOP 2.

STOP 2. STAIR PARK. Walk into park going south along Fuller Hollow Creek. Stream flows on bedrock, yet the floodplain appears to be in alluvial materials. Why has stream chosen a more difficult channel to occupy? Walk to 30-ft exposure of bright till (Binghamton lithology). What is the fabric of the till? Within Fuller Hollow there are good examples of fabric wherein some show extending flow and other compressing flow characteristics of the glacier. Such bright drift shows the history of materials. The ice moved south over the valley-fill sediments of the Susquehanna River, incorporating them into the basal load and then depositing them south of the river along thrust planes as the ice was forced to move up and over the hillsides. South from this stop the drift changes composition, passes into a transitional facies and finally becomes true "drab" till (Olean lithology) higher up on the hillsides. Embark. Travel north on Murray Hill Road.

2.1	0.7	Route 434. Turn left (west). In distance note the Roundtop Hill umlaufberg. The highway uses the meltwater channel that separates the hill from the former upland parent spur to the south.
2.6	0.5	Turn right on Johnson City bridge over the Susquehanna River and travel to Oakdale Mall area. The river flows west on valley-fill sediments about 200 ft thick comprised largely of glaciofluvial and glaciolacustrine materials. Near Route 17, wood at about an 8-ft depth in alluvial materials was dated at 1690 ± 100 ybp.
4.7	2.1	STOP 3.

STOP 3. This 25-ft exposure east of Oakdale Mall consists of drab till (Olean lithology). This is typical of lodgment till plastered on the south sides of hills that produce asymmetry of hills, or the "till-shadow effect". What type of fabric would you anticipate from such a depositional environment? Embark and travel to Route 17 going east at Exit 70.

6.4	1.7	Bedrock exposure of Upper Devonian Rhinestreet Formation consists of marine shale with some siltstone.
8.7	2.3	Chenango River. Note floodwalls, and revetment-type levees (armored with riprap) built by U.S. Army Corps of Engineers. Near this site, floodplain sediments in a former kettle were dated at 3801 ± 60 ybp at a 45 ft-depth. Thickness of valley fill is about 200 ft.

9.1	0.4	Turn right toward I-88 and Route 7 going north, exit 72N.
9.6	0.5	Binghamton brickyard and hillside quarrying of Rhinestreet Formation. The highway is adjacent to old Chenango Canal that connected Binghamton with the Erie Canal. The trend of the canal along here is called the "tow path" by residents.
12.3	2.7	Exposure of kame-terrace valley-train sediments that comprise Hillcrest. Note highly cemented character of materials.
13.2	0.9	Turn left over Chenango River at Chenango Bridge. Here valley fill is less than 100 ft thick. Note the umlaufberg, which is double crested. This part of the valley is more youthful because river was diverted from its earlier course on north side of umlaufberg. Keep right on north side of bridge, through stoplight and travel on Kattelville Road.
14.8	1.6	Turn left on Prentice Road. Note the route has been along west side of umlaufberg into the old valley of the Chenango River. This is now a "dry valley" filled with a congestion of glacial deposits and sand and gravel pits.
15.1	0.3	STOP 4.

STOP 4. ROSE GRAVEL PIT (now owned by Barett Paving, Inc.). Travel into east part of pit, 1.2 mi. These materials occur in a very complex sedimentary environment. This can be considered a pitted outwash plain (sandur), that also includes valley train, kame, kame terraces, and extensive progradational deltaic sequences. Although the deposits were formed during a single glacial stade and are comprised by materials of the same general provenance, there is great heterogeniety in the sediments from outcrop to outcrop. Thus the pit contains both pure outwash-type sediment as well as ice-contact deposits. There is a full range of textures from clay to silt, sand, gravel, cobbles, and boulders. Bodies of till also occur within some sequences. The extent, both laterally and vertically, of the cemented units pose serious problems for the miner. Geologically the problem is to determine the origin and time of cementation. For example, did it form in the vadose or groundwater zone? Did it form soon after deposition, or later in the Holocene? Aber (1979) has attempted to answer some of these points. Although much of the material is bright, there are also occurrences of drab drift. Why? Embark and continue on Prentice Road to Route 12.

16.7	1.6	Route 12 turn right (east).
17.7	1.0	Highway ascends a wind gap. This is one of three high-level notches along a north-trending ridge formed as overflow channels when ice occupied the Chenango Valley and meltwater streams discovered high outlets.
20.6	2.9	STOP 5.

STOP 5. CHENANGO FORKS. This is junction of the Chenango River flowing southwest and the Tioughnioga River flowing southeast. The exposure typifies the many different sedimentary features at junctions of two different ice lobes, with contrasting histories and meltwater regimes. This is a kame environment with deposits that range through glaciofluvial, glaciolacustrine, and unsorted materials. What is the origin of the large bouldery unit? Possibilities include flowtill, mudflow - debris flow, ablation till, etc. Are there others? Embark and travel north on Route 79.

21.9	1.3	STOP 6.

STOP 6. LANDSLIDE. This landslide consists of a series of slumps that cover a 7-acre area. It was caused when the road was cut into the lower hillslope. The increased slope and removal of material destabilized the materials. The hydraulic head increased and drainage from a road above the slide funneled water onto the mass. The hillside materials are composed of sediments susceptible to piping which enhance lubrication and movement of materials in the mass. Some pipes have been measured to be about 2 ft in diameter and extend many feet into the hill. The south side of the area is composed of drab till. Most of the slide occurs in glaciolacustrine silt and fine sand with clay more than 60 ft thick. Glaciofluvial beds occur immediately north of the slide area. What was the environmental setting when the deposits formed? Embark and travel north.

22.2	0.3	Bedrock.
22.5	0.3	Drab (Olean lithology) till. These sequences are typical of the Tioughnioga River, a multicycle sluiceway or beaded river. The river is anomalous because it flows southeast across the structural grain of the region. The valley alternates between narrow walls which are younger parts of the system, and wider valleys which represent positions where the river discovered its previous valley. Return south to Chenango Forks.

22.4	1.9	Turn left at Chenango Forks and then take first right over the Chenango River to Pigeon Hill Road and continue south. The road passes between many kettles which comprise this pitted valley train. Cadwell (1972) dated peaty materials in a kettle as 16,650 ± 1,800 ybp.
27.2	2.8	Old gravel pits on both sides of road. Sedimentation is outwash type.
27.8	0.6	STOP 7.

STOP 7. CHENANGO VALLEY STATE PARK. Drive to parking lot near twin lakes. Separating Lily Lake to the north and Chenango Lake to the south is a crevasse filling. The lakes occupy kettle holes. The park contains tens of kettles of various sizes. These were stranded "land icebergs" that were overwhelmed by meltwater deposits when active ice was to the north. Embark from park going east to Route 369.

29.4	1.5	Route 369 south, turn right.
30.0	0.6	Turn right at Virginia City. STOP 8.

STOP 8. HALE GRAVEL PIT. Good example of outwash deposits. Bedding shows characteristics of the shifting meltwater channels that provided avenues for deposition of the materials. Although these are outwash deposits, they contain deleterious ingredients and so fail to pass state (magnesium sulfate soundness test) requirements for use as highway concrete (>18 percent loss on 10-cycle test). Why? What is the lithology of the sediment?

31.5	1.5	Return to Route 369 and turn right going south.
33.9	2.4	Bedrock. The Chenango River enters a more youthful part of its system, with the umlaufberg separating it from the former thalweg. Bedrock in or near a stream invariably means the channel is not located at the major thalweg.
34.9	1.0	I-88 going east.
44.1	9.2	Belden Hill. One of highest elevations in this area. Drainage divide between the Susquehanna and Chenango Rivers.
46.8	2.7	Take Route 79 south through Harpursville.
48.3	1.5	On east side of Susquehanna is "The Plains" one of largest terraces in the region. Along the road there are many stagnant ice features.
49.2	0.9	Start of 0.6-mi - long esker that parallels the highway on east side.

51.2	2.0	Turn east over bridge at Ouaquaga and bear left on other side of bridge. The road arcs around the Windsor umlaufberg.
54.2	3.0	STOP 9.

STOP 9. TOWN OF WINDSOR GRAVEL PIT. Turn right onto dirt road and proceed to working area. This is a beautiful "textbook" example of a kame, a hill composed of ice contact deposits. Meltwater erosion was responsible for dissecting the umlaufberg from the upland connection to the east. The deposits fall within the range of drab lithology and contain glaciofluvial, glaciolacustrine, and poorly to unstratified drift. What is the origin of the latter? Embark, travel east to East Windsor Road and south to old "17".

55.4	1.2	Note kame-and-kettle topography and the character of the umlaufberg,which can be considered "youthful" because of the lack of tributaries on it and the adjacent highlands.
57.6	2.2	Turn east (left) on old "17".
59.2	1.6	Damascus. Continue east on old "17" which now rises on a steep grade up Tuscarora Mountain.
61.4	2.2	Sky Lake, one of three natural lakes in Fly Creek. This is drainage divide between the Delaware and Susquehanna Rivers. We are now in the Catskill Mountains, both geologically and geographically. The bedrock now consists mostly of terrigenous sandstone. Relief is 1500 ft and more, and the Delaware system drains the west and southwest part of the mountains.
62.0	0.6	Deer Lake, second in the lake chain in Fly Creek.
64.2	2.2	STOP 10.

STOP 10. FLY POND. Picture stop. A till barrier impounds this and the others have similar dams. What is the origin of these barriers? Depth to rock is reported to be 160 ft in the valley. There are also two smaller till mounds in Fly Creek. Embark and continue east on old "17".

65.8	1.6	STOP 11.

STOP 11. KAME DELTA. At 1450 ft elevation this is an unusually high level deltaic deposit. However, it forms part of the large picture with a major lake that occupied the Gulf Summit area.

66.2	0.4	Turn left (south) toward Gulf Summit. Note this area appears to be a major topographic low. Within it are numerous kames and kettles.
70.3	4.1	Travel down road and observe character of the drainage divide at Gulf Summit. Return to railroad and view the deep bedrock notch. This was the spillway for drainage of the proglacial lake that occupied this topographic low. Although Wisconsinan-age meltwaters helped to incise the lowermost 70 ft, the higher features may have been initiated during earlier glacial stages.
70.5	0.2	Turn left on dirt road going southeast.
71.1	0.6	STOP 12.

STOP 12. VIEW OF SPILLWAY AND OVERFLOW CHANNEL. This is headwater area of Cascade Creek, a Susquehanna River tributary. The Catskill sandstone shows vestiges of stream erosion. North of the road the hillslopes contain varying degrees of blockfields, a reminder of periglacial climatic conditions. Embark and continue southeast on dirt road.

76.6	5.5	Turn south (left) on blacktop road toward triple cities of Pennsylvania, Lanesboro, Susquehanna, and Oakland. This is the Great Bend area where Susquehanna River arcs from a south flow, to an east flow, then a north flow. The unusual stone masonry bridge was built in 1848 by a Scotsman, Kirkwood, after several earlier contractors had gone broke attempting to build a span across the valley. Route 171 going north.
78.9	2.3	Cross bridge and Susquehanna River and turn right.
79.5	0.6	STOP 13.

STOP 13. OAKLAND ESKER, largest in the region, with total length about 4 mi. Here it is on north side of the river but it can be followed northward where the river has incised it and the esker continues again on east side of river. The esker is not a simple ridge, but is double in several parts, such as near the outlet on the west side of Oakland. Side slopes are steep and esker exceeds a height of 100 ft. At the top there are depressions and undulating topography. Exposures here and at other outcrops show bedding ranging from horizontal to some units whose dips exceed 45°. Till and flow till are interpreted to be part of some sequences. The sparse number of exotics indicate the drift is drab type. Was this a supraglacial (open to the sky) or subglacial esker? Return to Route 171 and continue west.

81.4	1.9	Western limit of esker.
88.3	6.9	Penny Hill on north side of road. There is a "rock city" at top of the hill, another indication of periglacial conditions.
88.8	0.5	Take I-81 going south out of Great Bend, Pa.
91.2	2.4	Hanging delta on east side of highway. This is the valley of the north-flowing Salt Lick Creek. The straightness of the valley, truncation of spurs, and anomalous southern extension indicate valley sculpture by ice and meltwater erosion.
96.1	4.9	Exit on Route 492 to New Milford and to Route 11.
96.8	0.7	Turn south on Route 11 at New Milford and enter the New Milford sluiceway.
98.3	1.5	The most recent meltwater erosion starts here.
98.9	0.6	STOP 14.

STOP 14. PICTURE STOP FOR THE NEW MILFORD SLUICEWAY. This is the south-trending Martin Creek valley, with bedrock walls of Catskill cross-bedded sandstone. This is a "youthful" type or first-cycle sluiceway because of the imperfect development of tributaries and absence of till. The feature was developed when meltwaters from a proglacial lake to the north drained south through spillways formed during earlier glacial times. The Great Bend region had a long and involved history throughout Quaternary times. The unusual geometry of the Susquehanna River implies it is an ice-marginal valley, such as parts of the Missouri and Ohio Rivers.

100.9	2.0	Return to New Milford and proceed north via Route 492 and I-81.
106.4	5.5	STOP 15.

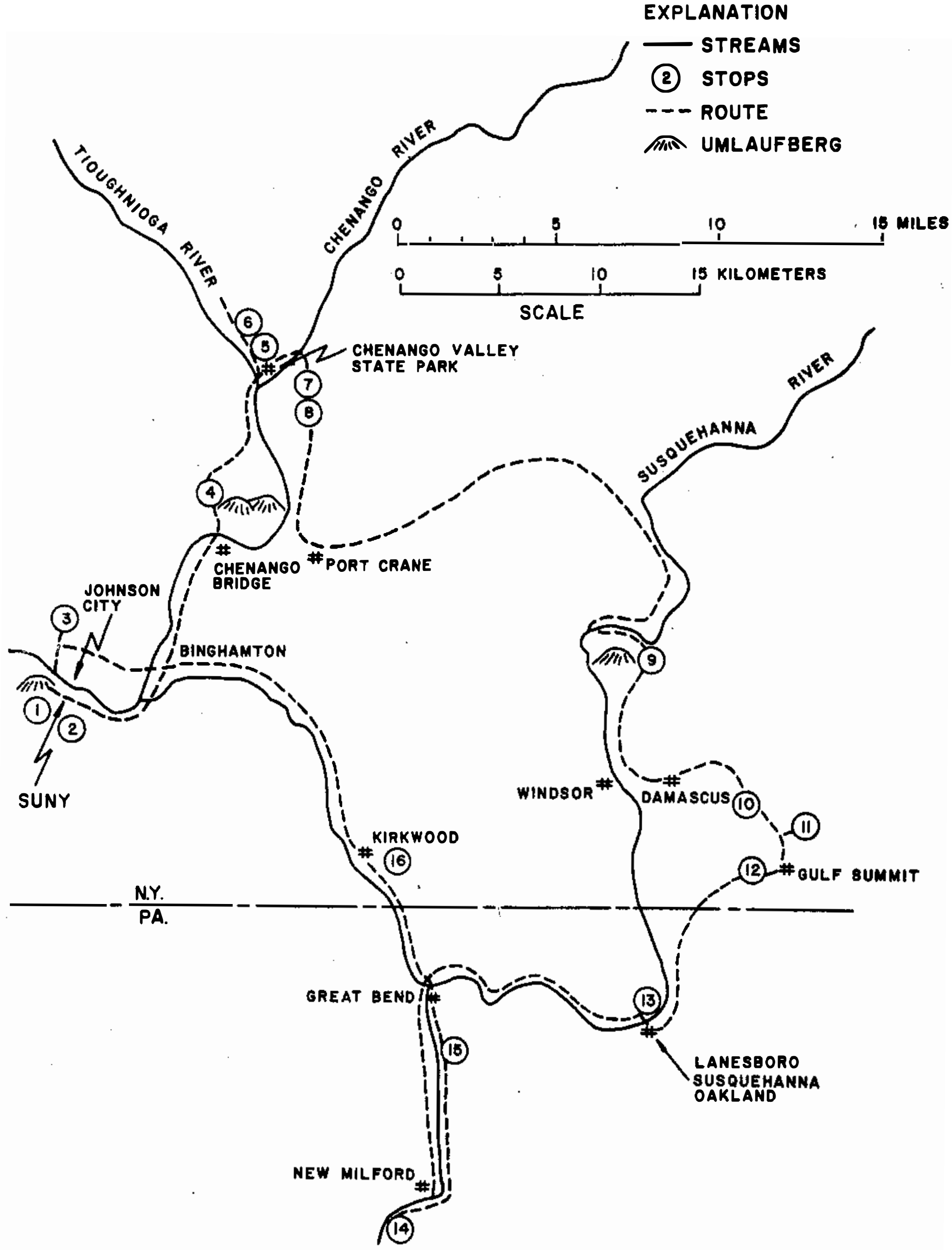
STOP 15. HANGING DELTA. This delta owes its development to events in Little Egypt Creek, the north-flowing stream on the east side of Maunatome Mtn. Along this north-south ridge there are a series of high-level notches occupied by ice and meltwaters during different glacial and proglacial times. Drainage in Little Egypt was blocked by ice to the north causing the eventual overflow of the proglacial lake into Salt Lick Valley, where another lake received the sediments. Typical deltaic structures occur: topsets, foresets, and a general lobate pattern of strata that dip away from the central core. The materials are drab and currently used by the Pennsylvania Highway Department. The deposits are more than 150 ft thick. How did the massive sandstone block become emplaced within the sediments? Continue north on I-81 to New York State line.

115.0	8.6	Take Kirkwood exit and bear right.
115.3	0.3	STOP 16.

STOP 16. DOUBLEDAY GLEN. The Rhinestreet Formation forms bedrock of the area. Note the fabric of unconsolidated materials on top of rock. This is colluvium. Although composed of till it is not in place, having moved downslope by solifluction processes which produced the preferred orientation of the slabby clasts. Walk into the glen. Evolution of this feature is somewhat similar to the famous Finger Lake glens. They developed during postglacial time when a high-level consequent stream starts its new trajectory. Because till covered the preglacial valley and diverted streams to the south, many of the newly developing consequent streams were forced to carve new bedrock chasms. Thus when streams miss their former valleys, they incise rock and form narrow, steep-walled valleys. Where former valleys were encountered, the walls are gentler, not in rock, and wider. Return to I-81 toward Binghamton.

122.2	6.9	New York State Hospital built on unusual till knob. What is its origin?
131.6	9.4	Return to SUNY campus via Route 17 and Exit 70S.

FIELD GUIDE MAP



BIOCLASTIC CARBONATE UNITS IN THE CATSKILL CLASTIC WEDGE

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INTRODUCTION

This trip we will visit two quarries in northern Pennsylvania (Fig. 1) which expose thick carbonate units nearly unique in the "Catskill delta". Our purpose is to examine these units and to discuss their origin, which in many respects remains enigmatic despite considerable study.

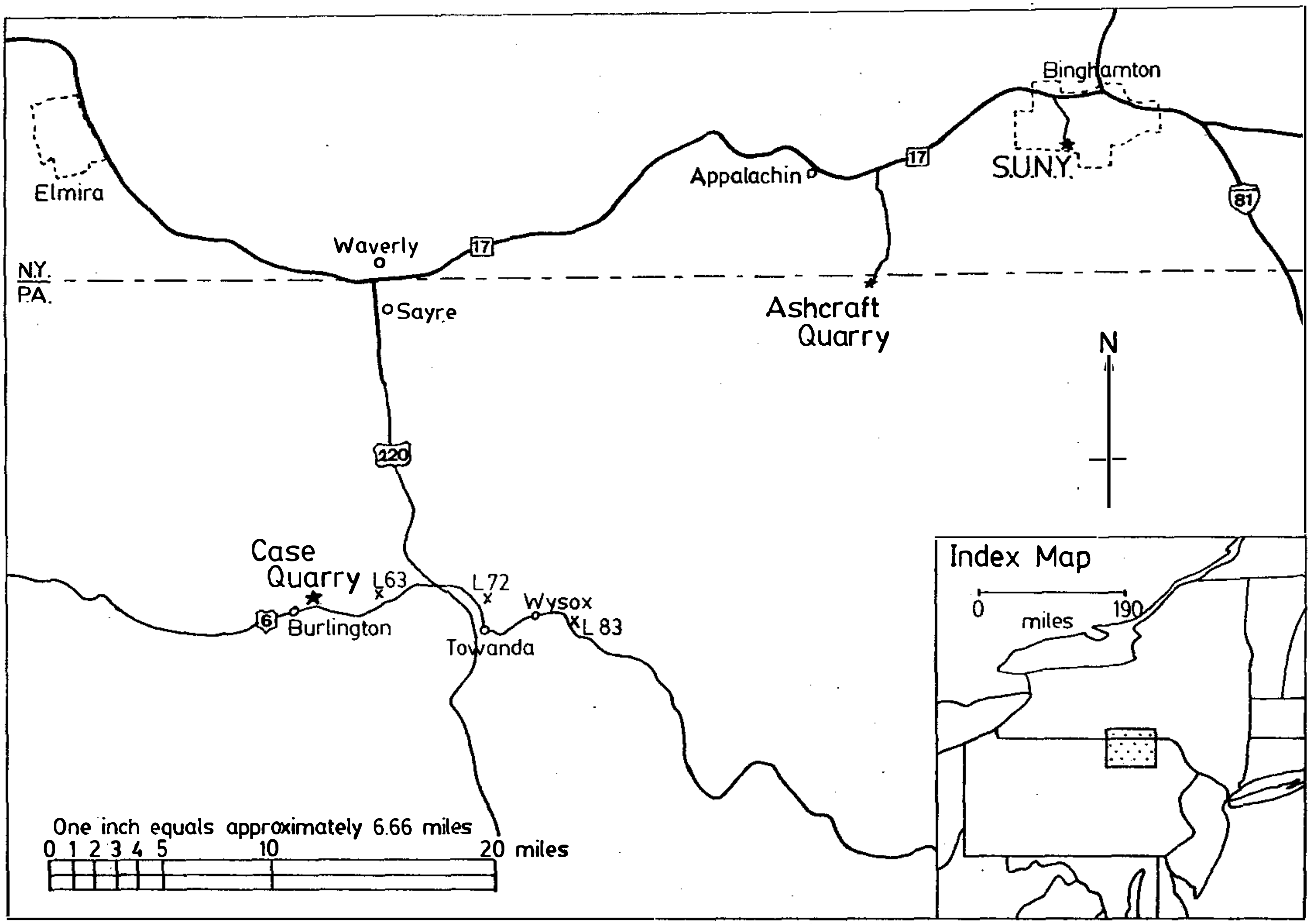
Each of us has contributed to this guidebook in a different manner. All of us have examined both quarries in varying levels of detail. Woodrow and Bottjer measured sections in the Case quarry and considered the trace fossils and petrography there. Minero measured the Ashcraft quarry sections. Enos and Woodrow have supervised the work of several students in both areas.

STRATIGRAPHIC CONTEXT OF THE CARBONATE ROCKS

Rocks seen in the two quarries are part of the Catskill clastic wedge (Woodrow, 1968). Carbonate-rich rocks are unusual in this sequence which is typified by sandstones, siltstones, and shales. Both units were worked initially for agricultural lime,and quarries were later opened because the carbonates provide the best material for riprap available for many miles.

Stratigraphic relationships within the Catskill clastic wedge have intrigued geologists since the middle of the nineteenth century. The major outlines of the stratigraphy have long been known and great strides have been made recently toward an understanding of the sedimentology. Uncertainties exist in both areas, however, even after 100 years of concerted effort by several generations of geologists. That this is so is understandable, especially in northern Pennsylvania, considering that:

a. The rocks are not well exposed, a limitation only partly offset by a few exploratory gas wells drilled in the region.
b. The rocks are gently folded and are broken by small faults.
c. The rock sequence is subject to major facies change over short distances.
d. Trends of facies change, local structure, and regional dip all differ within this area (Fig. 2).


Binghamton
S.U.N.Y.
Elmira
Appalachin
17
81
Waverly
N.Y.
PA.
Sayre
Ashcraft
Quarry
N
120
Case
Quarry
L63
L72
Wysox
L 83
6
Burlington
Towanda
Index Map
0
miles
190
One inch equals approximately 6.66 miles
0 1 2 3 4 5 10 20 miles

Stratigraphic disclaimers notwithstanding, it is possible to arrive at some reasonable judgments about the stratigraphic position of these carbonate units based on physical stratigraphic techniques (Woodrow, 1968; Woodrow and Nugent, 1963). The biostratigraphy of the carbonate rocks has not been studied in detail.

The limestone unit at the Ashcraft quarry apparently occurs at the horizon of or just above the dark Corning Shale (Rickard, 1975). The carbonate at the Louis Case quarry is part of the Luthers Mills Coquinite of Willard (1936), a member of the Towanda Formation (Woodrow, 1968); the quarry is cut into rock approximately 30 to 50 m above the Dunkirk Shale (Rickard, 1975; Woodrow, 1968). It is, thus, considerably younger than the limestone at Ashcraft quarry (Fig. 3). Based on these rock-stratigraphic assignments, the Ashcraft limestone is of mid-Frasnian age and the Luthers Mills Coquinite is of earliest Famennian age.

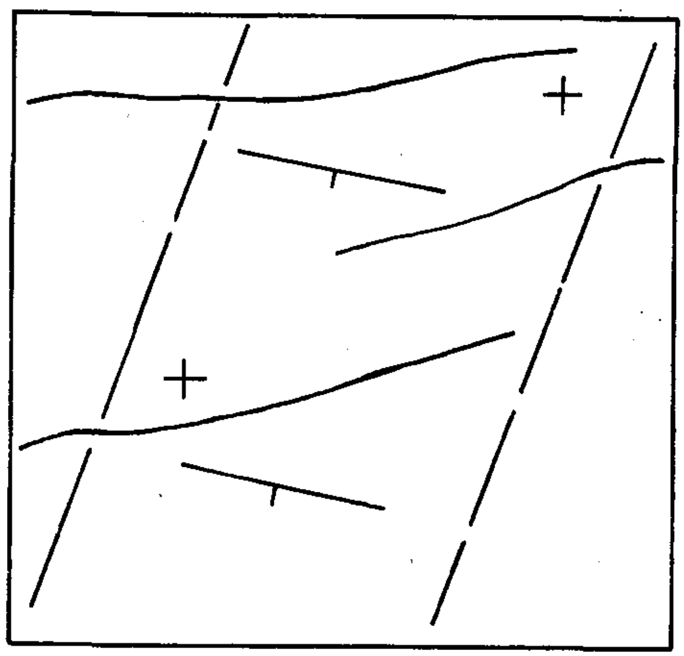
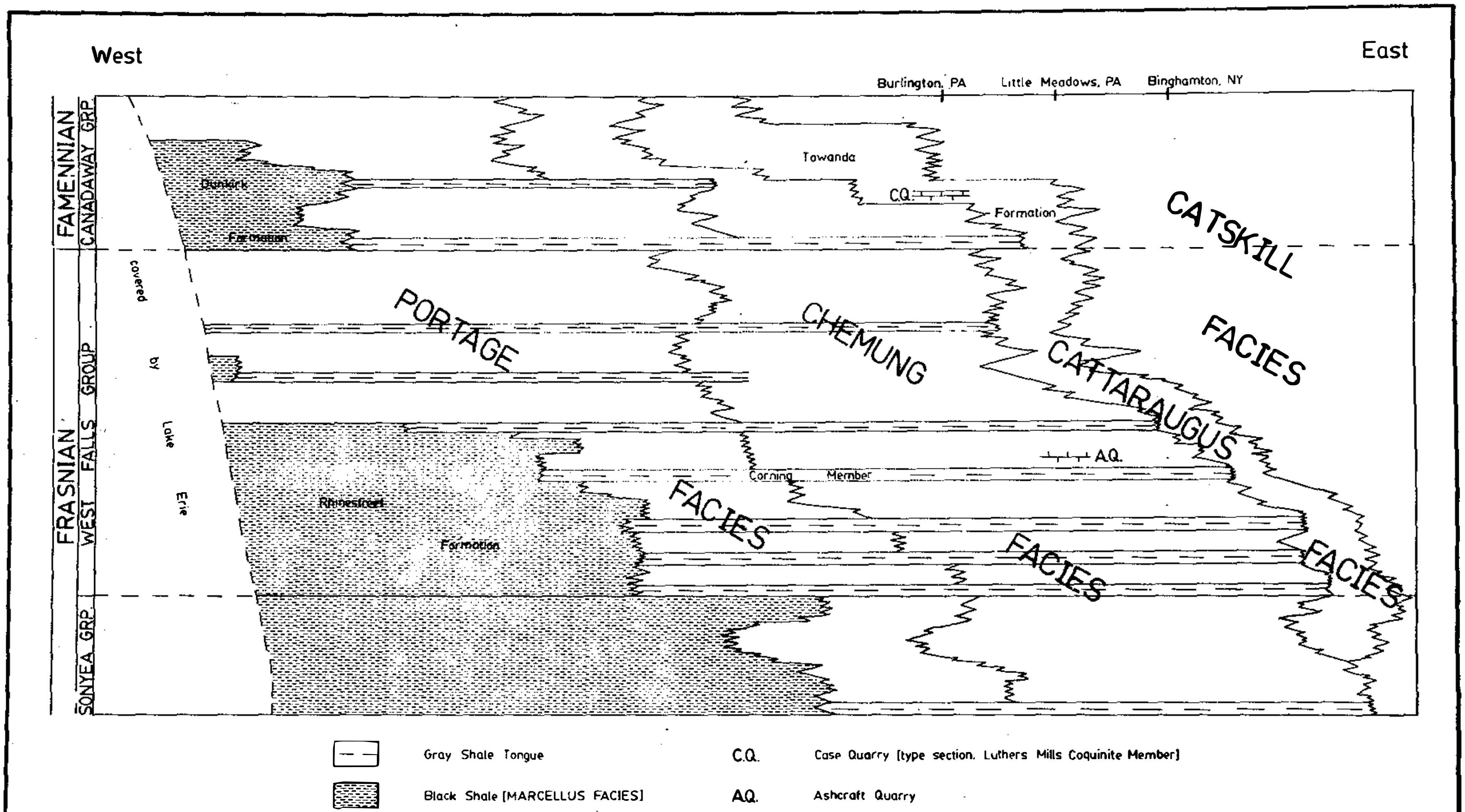


Figure 2. Structural, facies, and regional basin trends near the carbonate quarries (shown with crosses). Facies trends: dashed lines; anticlinal axes: solid lines; basin trend: strike and dip symbols.

◄ Figure 1. Location map showing Ashcraft and Case quarries. Localities L63, L72, and L83 are outcrops of Luthers Mills Coquinite from Woodrow (1968).



Upper Devonian Facies Relationships

Southern New York and Northern Pennsylvania (modified from Rickard, 1975)

ROCKS AT ASHCRAFT QUARRY

The quarry is located on the northwest-facing slope of a hill east of the village of Little Meadows, Pennsylvania (Fig. 1). Rocks exposed in the quarry are part of the West Falls Group and they probably lie less than 15 m above the Corning Shale Member within the Gardeau Formation (Fisher and others, 1970; Rickard, 1975). Local dip is generally southwest at about 1:50 (100 feet per mile). The trend of the quarry face is 65 degrees. The quarry wall is approximately 15 meters high and 200 meters long.

Three subdivisions of the rock sequence are readily apparent (Figs. 4 and 5). A basal unit made up of sandstone and shale is exposed in the sump pit at the northeast end and patchily on the quarry floor and walls. This is overlain with an erosional contact by the limestone unit that is exposed throughout the quarry. The upper unit is in sharp contact with the limestone and extends to the top of the quarry wall for a thickness of 10-12 meters. It is interbedded shale, siltstone, and sandstone.

Basal Unit

As seen in the sump pit (Fig. 5), this unit is greenish-gray shale and lenticular siltstone overlain by gray, medium-bedded sandstone with trough cross bedding. Horizontal burrows and small plant fragments are common in the silt and shale. Crinoids, brachiopods, large carbonized wood fragments, and intraclasts form coarse-grained lenses in the sandstone. Shallow trough cross-beds indicate paleocurrents toward 175°. Asymmetric straight-crested ripples (paleocurrent azimuth 120°) exposed on the quarry floor were locally eroded and reformed into linguoid ripples directed 290°, showing current reversal.

Limestone Unit

The quartzose carbonate unit is the quarryman's objective and the unit of major geologic interest. It varies in thickness from about 5.5 m in the center and eastern part of the quarry to 3.5 m toward the west. Core drilling on the hill south of the quarry demonstrates considerable thickness variation (some possibly the result of glacial or preglacial erosion) with perhaps a general thinning trend toward the southwest (Figs. 6 and 7). With the limited number of core holes and the till cover, the geometry of the body is conjectural, particularly toward the southeast. Previous workers (Krajewski and Cuffey, 1976; Bowen, 1978) have concluded that the unit is a N-S elongated mound confined within the limits of the hill. This cannot be verified from the existing data and must therefore be regarded as only one possible configuration. A tabular body—with internal wedge-shaped units—truncated by erosion and covered by till is another possibility.

< Figure 3. Upper Devonian facies relationships (from Rickard, 1975).

←0.2"

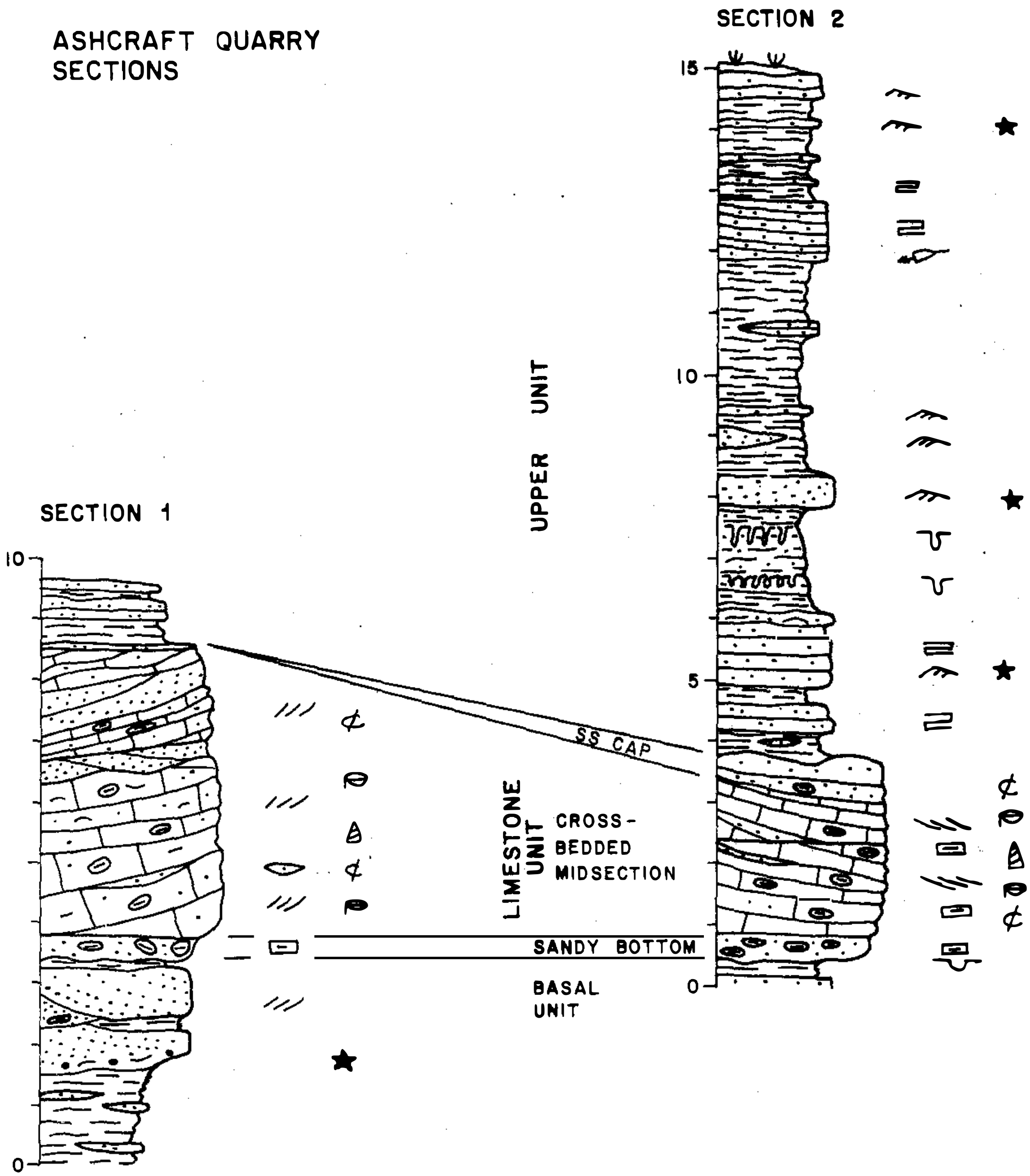


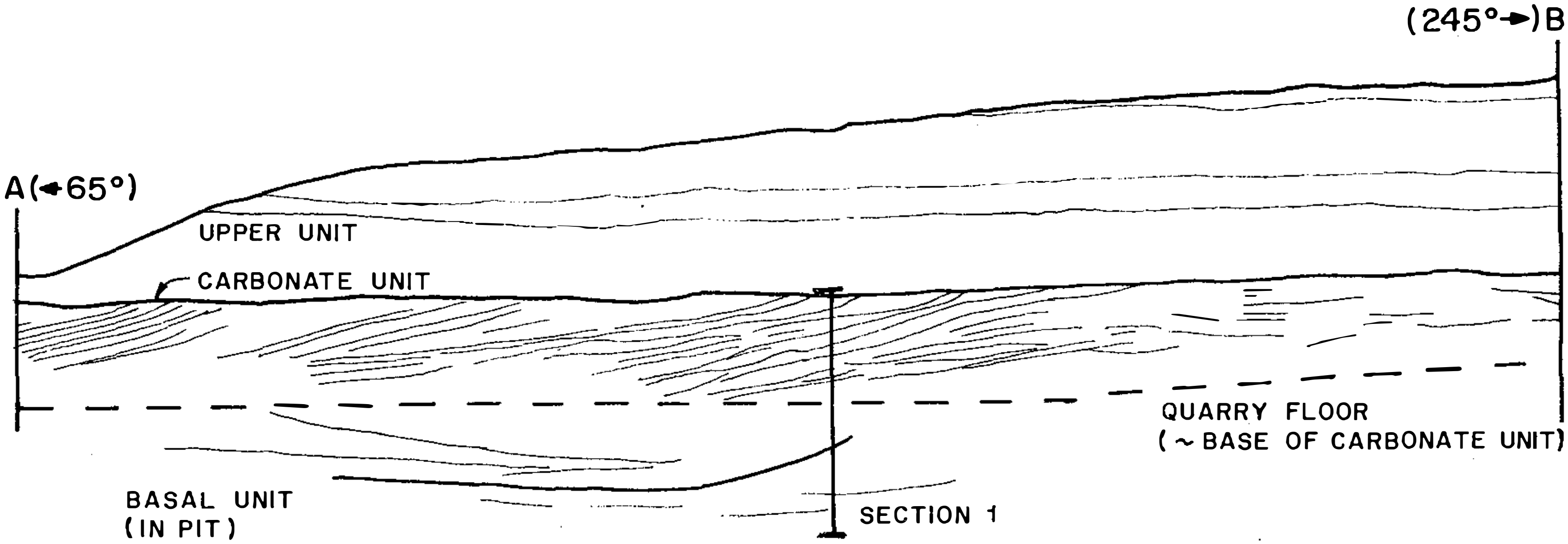
Figure 4. Sections at Ashcraft quarry, measured by Minero. Locations are shown in Figure 5. For symbols see Figure 10.

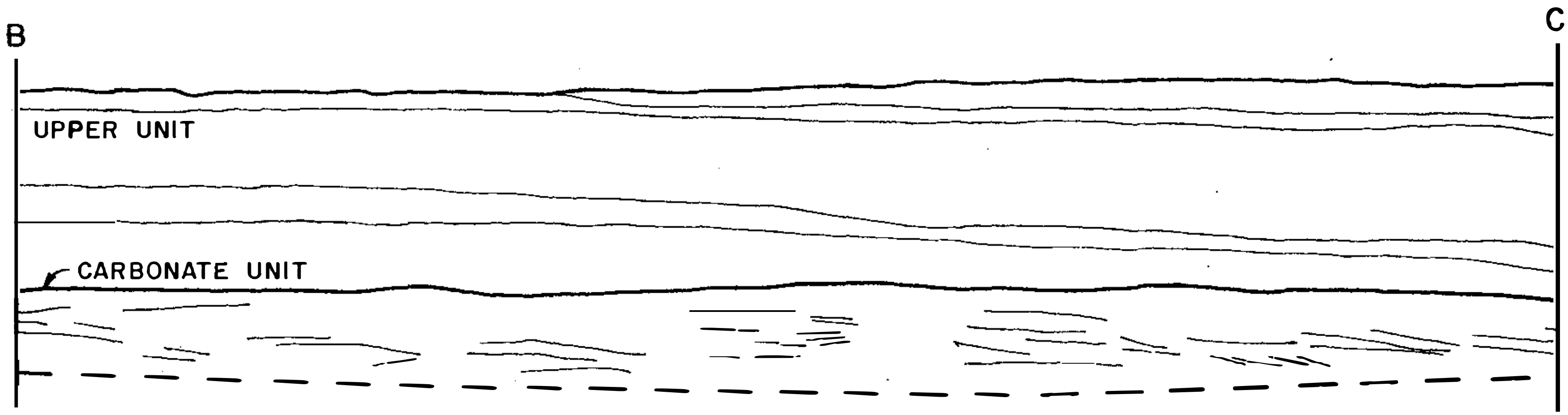
In general, the limestone unit consists of quartzose, skeletal-fragment lime grainstone and skeletal quartz sandstone. Grain size fines upward in the unit. Top and bottom contacts are sharp. Three subdivisions can be recognized, a sandstone bottom, a cross-bedded midsection, and a thin sandstone cap.

Sandstone Bottom. Medium- to coarse-grained, lithoclastic calcareous sandstone, generally less than 60 cm thick, overlies the basal unit with a sharp, apparently erosional, contact. Locally the base of the sandstone is load deformed and small, truncated shale diapirs occur at the top of the underlying unit. In addition to quartz sand and some shell fragments, the sandstone contains numerous large, gray shale clasts and reworked ankerite $Ca(Mg, Fe)(CO_3)_2$ nodules (Minero, unpub. report). The nodules weather to an orange-brown oxide (?) which stains the rocks in this subdivision. Plant fragments, locally permineralized by pyrite, are also common. Low-angle cross bedding is locally visible.

Cross-Bedded Midsection. The bulk of the limestone unit is cross-bedded, quartzose, skeletal-fragment lime grainstone. Trough and wedge-shaped cross-bed sets are up to several meters thick, although most are tens of centimeters thick. Reworked ankerite nodules, shale clasts, and carbonized plant fragments are all common. The ankerite weathers to an orange-brown oxide (?) to produce a mottled appearance against the light-gray background. Rusty yellow streaks on the weathered surface mark the position of large plant fragments, most of which are partially pyritized, some with exquisite preservation of detail. Highly abraded shell fragments, quartz, and some feldspar comprise most of the sand-sized grains of the rock. The quartz is fine- to medium-grained and appears angular, but quartz overgrowths and etched margins filled by calcite suggest that original grain shape has been highly altered. Most skeletons are abraded fragments and are also diagenetically altered to calcite-filled molds or neomorphic calcite, so that few can be identified except by general shape. Recognizable fragments include brachiopods, gastropods, bivalves, crinoids, and fish plates. Shelly portions of the carbonate unit are cemented by medium- to coarse-grained calcite spar; quartzose portions have some clay matrix and fine-grained carbonate. Modal analyses of Bowen (1978) indicate an average of about 80 percent carbonate, including cement, sparse matrix, local replacement dolomite, and more than 30 percent fossil fragments.

Calcareous quartz sandstone is interspersed with the limestone on scales ranging from small lenses, representing isolated ripples or dunes, to the largest cross-bed sets. Inter-leaving of large wedge-shaped units is best seen at the east end of the quarry and in the northwest corner. Quartz-sandstone wedges appear to laterally replace limestone ones at both locations, suggesting that the limestone-dominant rocks are indeed limited to a central "mound" (Krajewski and Cuffey, 1976; Bowen, 1978), but more lateral control is needed. Lateral-accretion bedding of the wedges dips toward the east at the east end of the quarry and a northwest-directed mode is found toward the west end (Fig. 8A). Throughout the quarry, however, smaller scale troughs open toward the west (Fig. 8B).


A(←65°)
(245°→)B
UPPER UNIT
CARBONATE UNIT
QUARRY FLOOR
(~ BASE OF CARBONATE UNIT)
BASAL UNIT
(IN PIT)
SECTION 1


B
C
UPPER UNIT
CARBONATE UNIT

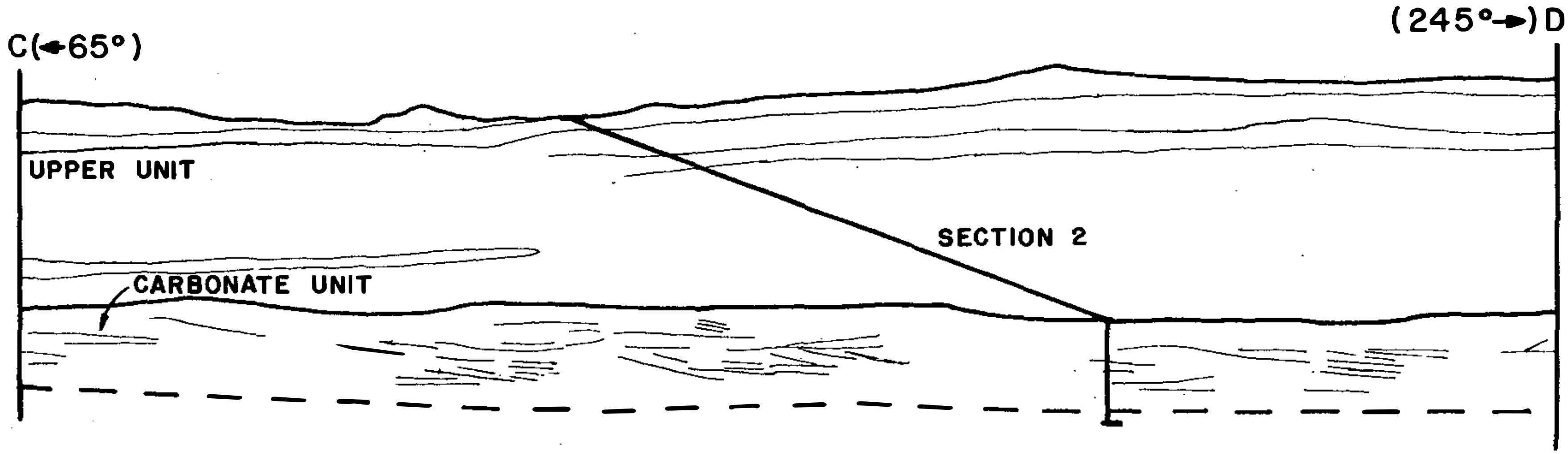


Figure 5. Sketch section of the high wall (SE) of Ashcraft quarry from a photo mosaic by Minero. Sequence of panels from top (A-B) to bottom (C-D) join left to right as indicated by matching letters. Scale is inexact because of photographic distortion; each panel is about 65 m wide; maximum height is about 15 m. Location of sections in Fig. 4 are indicated.

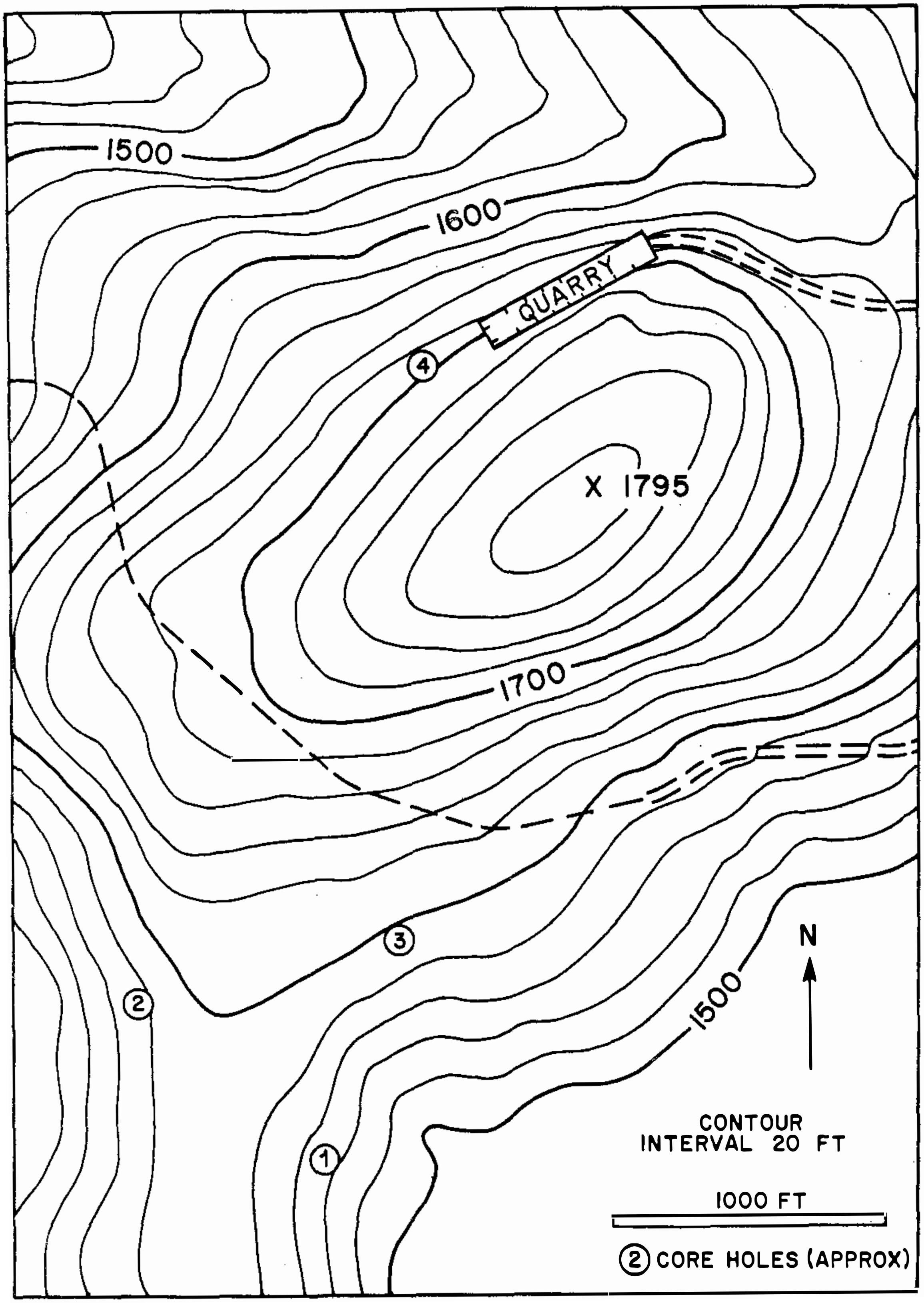


Figure 6. Ashcraft quarry area, showing approximate location of core holes (from Krajewski, unpub., and Ashcraft Excavating Company files). Contour base from Friendsville 7-1/2 min. quadrangle.

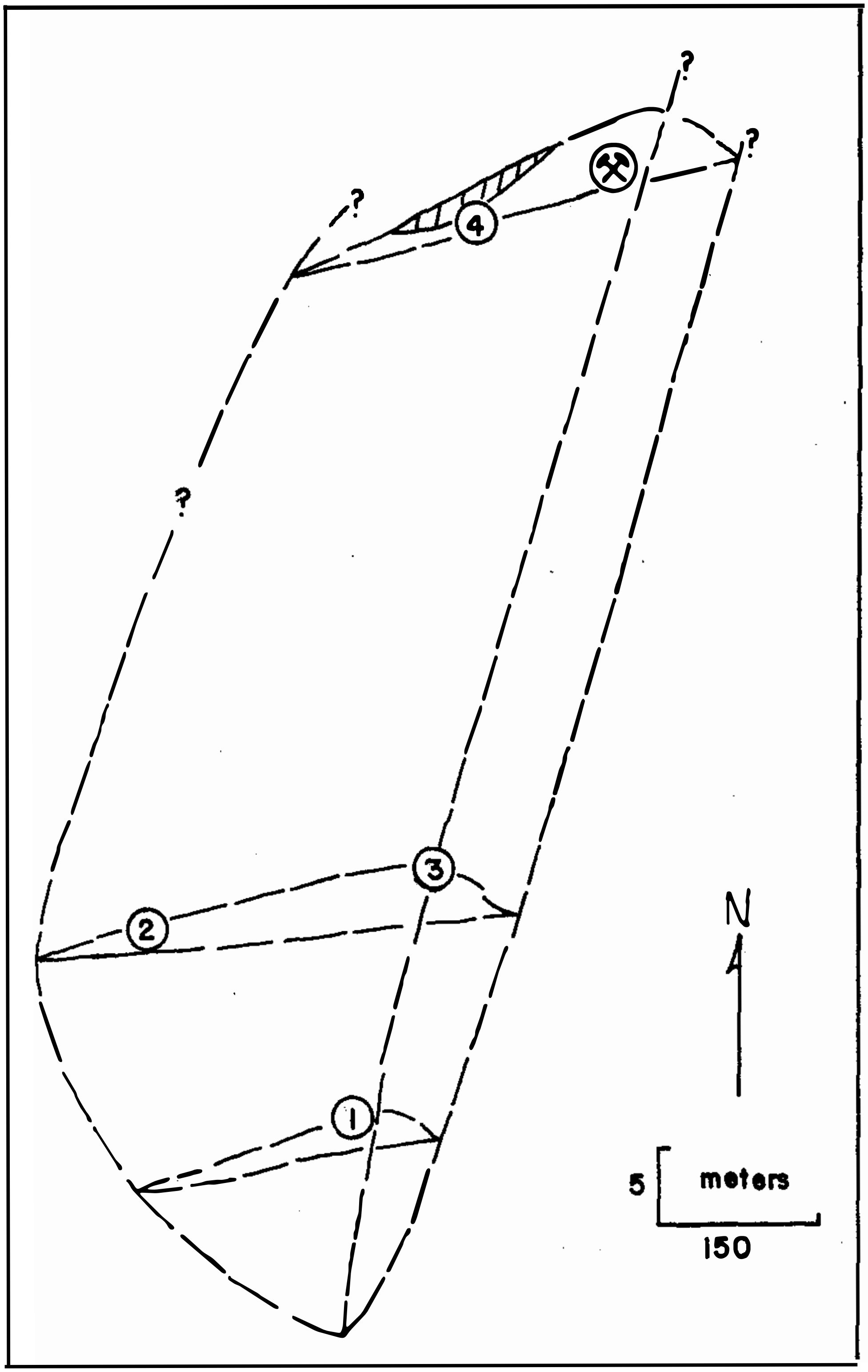


Figure 7. One possible reconstruction of the limestone body: The Wind and a Prayer Model. Location of core holes numbered. Till rests on carbonate at no. 4; erosion is assumed. Northward extent of the body is obscured by erosion; southward limit is also conjectural. Core data from Ashcraft Excavating Co. files.

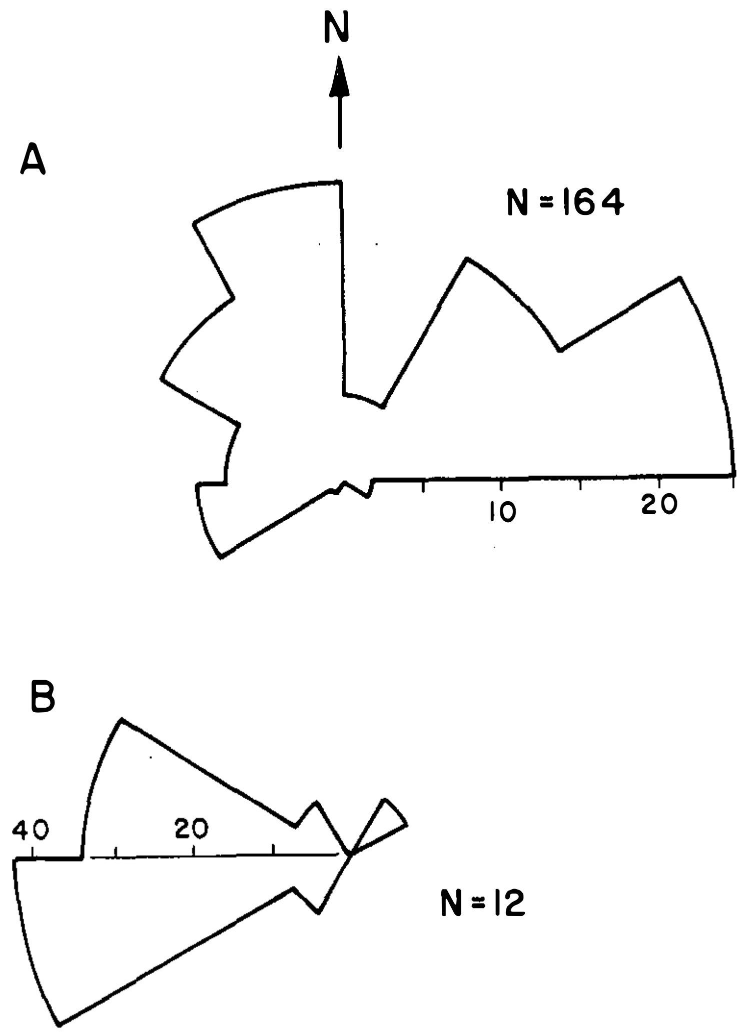


Figure 8. Paleocurrent measurements from carbonate unit in Ashcraft quarry. Scales are percentage of total measurements (N) in 30° intervals. A. Dip directions of 164 cross-beds, measurements by S. A. Krajewski (unpub. data). These are primarily large-scale lateral-accretion bedding, although some trough limbs are included. The east-directed mode is primarily from the east end of the quarry, the NNW mode is from the northwest corner, and the weak WSW mode is from the southwest corner (Krajewski, written comm.).
B. Axial trends of 12 trough cross-bed sets.

<u>Sandstone Cap</u>. A light-gray quartz sandstone that weathers dull brown discontinuously veneers the limestone unit. Although the contact with the cross-bedded midsection locally appears gradational, it also bevels inclined accretionary wedges (Fig. 5). Cross-bedding is generally not apparent within the sandstone, and there are hints of bioturbation that may explain its absence and the gradational lower contact. However, the top of the unit is wavy in many places, and a large bench cut on this unit at the southwest end of the quarry displays numerous nearly straight-crested dunes formed by paleocurrents, mainly toward 110°. The dunes are ornamented by wave ripples, and locally by current ripples, whose variable orientations show that the dune forms channelled subsequent flow. Trace fossils are also common on this surface, including very large horizontal tubes (to 5 cm in diameter) and clusters of long curving tubes.

Identifiable fossils are rare in the sandstone cap; they include a few brachiopods and gastropods. Some large fish-bone and plant fragments have been recovered.

In summary, the limestone unit consists of wedge-shaped cross-bed sets of skeletal lime grainstone and quartz sandstone sandwiched between two thin layers of sandstone. Detrital ankerite nodules, plant fragments, and fish bones are present throughout, but decrease in abundance upward as grain size generally fines. The ankerite nodules range from 0.5 to 7 cm in diameter and consist of micritic to mm-sized carbonate crystals that generally coarsen toward the margins of the nodule in a crudely concentric pattern. One nodule analyzed contained 8.8 percent iron oxide. Most nodules are rounded and they are locally size sorted. Quartz grains within the nodules are more angular, much finer, and less numerous than in the surrounding rock. Thus, the nodules are clearly rewored (Minero unpub. report). Plant fragments are mainly woody trunk material. Some fish bones have been tentatively identified with the arthrodire, <u>Dinicthys</u> (Fig. 9; R. Caprio, unpub. report). The preserved invertebrate fauna consists mainly of brachiopod fragments, gastropod molds, and rare bivalves and crinoids. Burrows are not visible throughout most of the unit, especially in the cross-bedded portion, but are abundant on the top surface. Cross-bedding in the large-scale sets indicates lateral accretion toward the east and northwest (Fig. 8A), away from the center of the quarry. Depositional relief is clearly indicated with some truncation probable at the top of the unit (Fig. 5). Trough cross-bedding is directed westward throughout (Fig. 8B). Preserved dunes were formed by southeasterly currents and ripples are quite variable in orientation.

<u>Upper Unit</u>

Dark-gray and olive-brown shales and mudstones and light blue-gray, fine- to medium-grained sandstones make up the remainder of the quarry high wall (Figs. 4 and 5). Individual sandstones are several centimeters to a meter thick and vary in thickness across the quarry. Many sandstones are lens-shaped over tens of meters suggesting current bedforms of low amplitude. Most sandstones exhibit either ripple cross-stratification or horizontal lamination. Shallow troughs or channels are visible in a few beds. Large load casts or flow rolls deform some channels. The shales

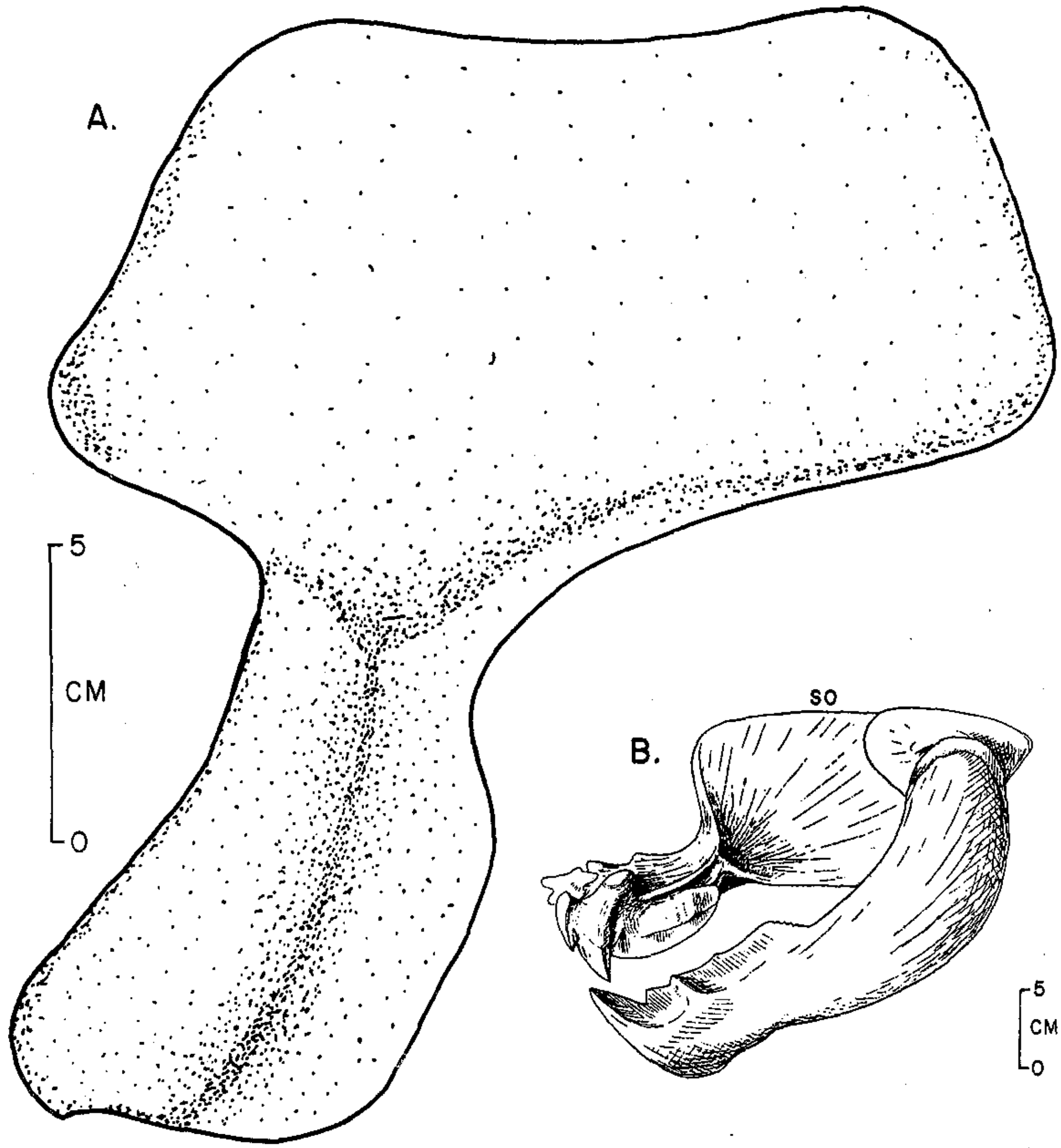


Figure 9. Fish plate from Ashcraft quarry. A. Sketch of the plate, probably the suborbital bone of the arthrodiran placoderm, Dinichthys. B. Diagram of the jaw of Dinichthys showing an internal view of the suborbital (SO), from Heintz (1931, p. 229). Sketch and identification are by Richard Caprio (unpub. report, 1978).

and mudstones are bioturbated, locally to the degree that stratification is totally disrupted and sand-shale layers are homogenized (Fig. 4). Trace fossils include copious trails on bedding surfaces and Skolithus-like vertical tubes. Few body fossils are found except in thin brachiopod-crinoid coquinites at the base of some sandstones. A few lingulid brachiopods have been found in float that probably came from this interval. Plant and fish fossils are relatively rare. Grooves and faint flutes on the base of one sandstone indicate average paleocurrent directions of 298° (5 readings). Ripple cross-lamination is generally west directed and several sandstone beds appear to thin westward.

ROCKS AT LOUIS CASE QUARRY

The quarry is on a south-facing slope located at the north side of U.S. Route 6, one mile east of the village of Burlington, Pennsylvania (Fig. 1). It was opened recently to exploit the Luthers Mills Coquinite as riprap. An older, smaller quarry, about 100 m to the east of the Case quarry, was opened in the 19th century to exploit the coquinite for agricultural lime and is the type locality of the Luthers Mills Coquinite (Willard, 1936).

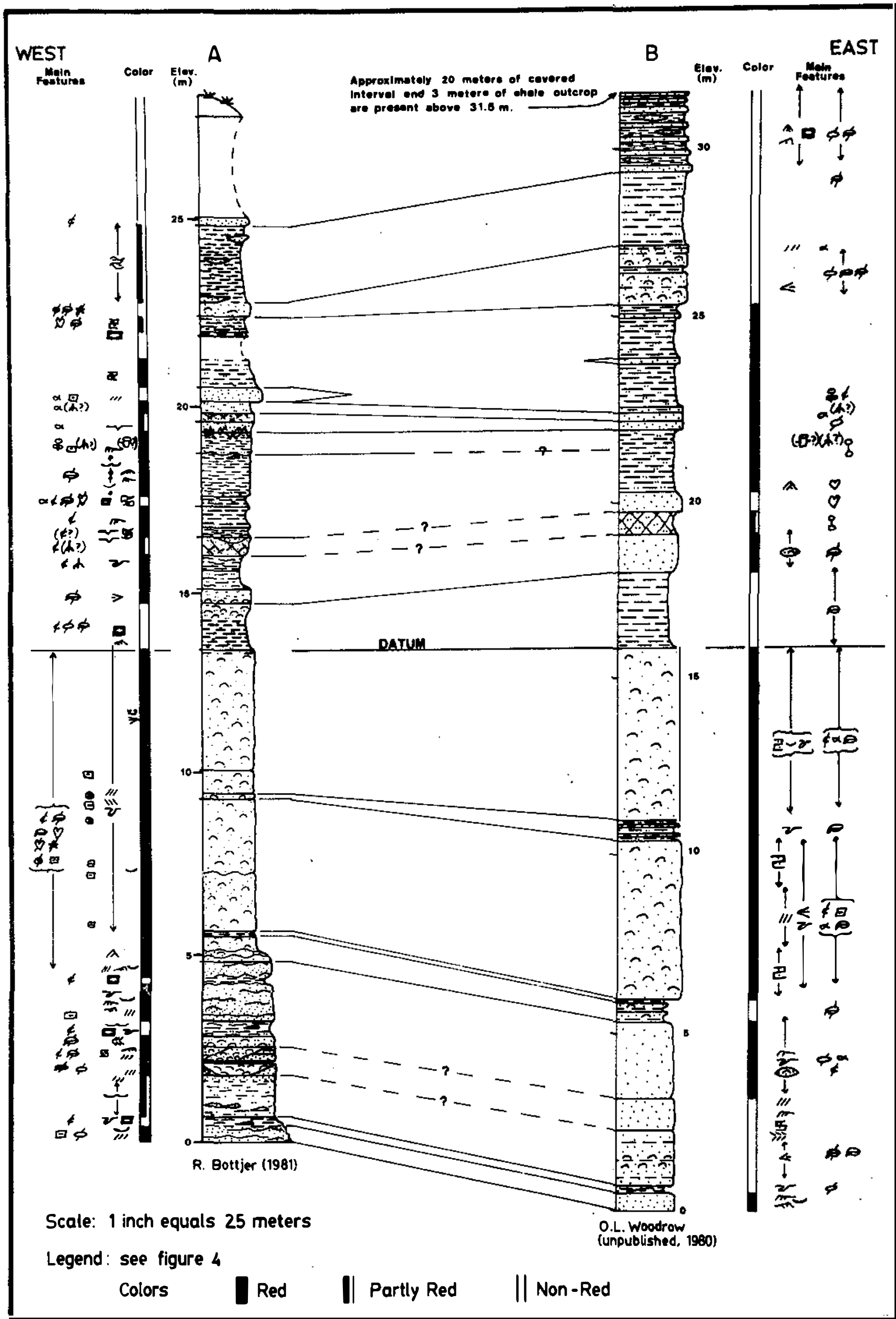
Stratigraphic placement of the quarry sequence (Fig. 3) has it in the Upper Devonian Towanda Formation (Woodrow, 1968) or the Lock Haven Formation of Pennsylvania useage (T. M. Berg, Pennsylvania Geological Survey, pers. comm., 1981). The thick coquinites were referred to by Willard (1936) as the Luthers Mills Coquinite, a distinctive, discontinuous rock unit which extends from Burlington at least as far east as the hills east of the Susquehanna River at Towanda (Fig. 1).

Within the quarry, the rocks are divisible into the Luthers Mills Coquinite at the base and the overlying shales and sandstones (Fig. 10). The base of the Luthers Mills is beneath the quarry level, but its sharp upper contact is clearly visible.

Geologic structure in the region is more pronounced than at the Ashcraft quarry. Locally the dip is north at 7° on the north limb of the Towanda anticline.

Luthers Mills Coquinite. This striking rock unit is at least 16 meters thick (the basal contact is not exposed) and of that thickness more than 80 percent is shell-rich. Strata are red or greenish gray with broken, abraded, or, rarely, whole shells arrayed in layers tens of centimeters thick. Most beds persist for tens of meters with sharp contacts. Cross-stratification and clay-draped ripples (flaser bedding) are common. Sandstone and mudstone beds are more common in the lowest 5 m of the exposure.

Sedimentary structures include plane bedding; wave, interference, and current ripples; cross-stratification; and load casts. Troughs of cross-strata open toward the west (260° to 310°) and ripple asymmetry is bimodal (Fig. 11). In some troughs, ripples were developed by paleo-currents perpendicular or opposed to those which formed the troughs.


WEST
Main Features
Color
Elev. (m)
A
B
Elev. (m)
Color
Main Features
EAST
Approximately 20 meters of covered interval and 3 meters of shale outcrop are present above 31.5 m.
DATUM
R. Bottjer (1981)
O.L. Woodrow (unpublished, 1980)
Scale: 1 inch equals 25 meters
Legend: see figure 4
Colors
Red
Partly Red
Non-Red

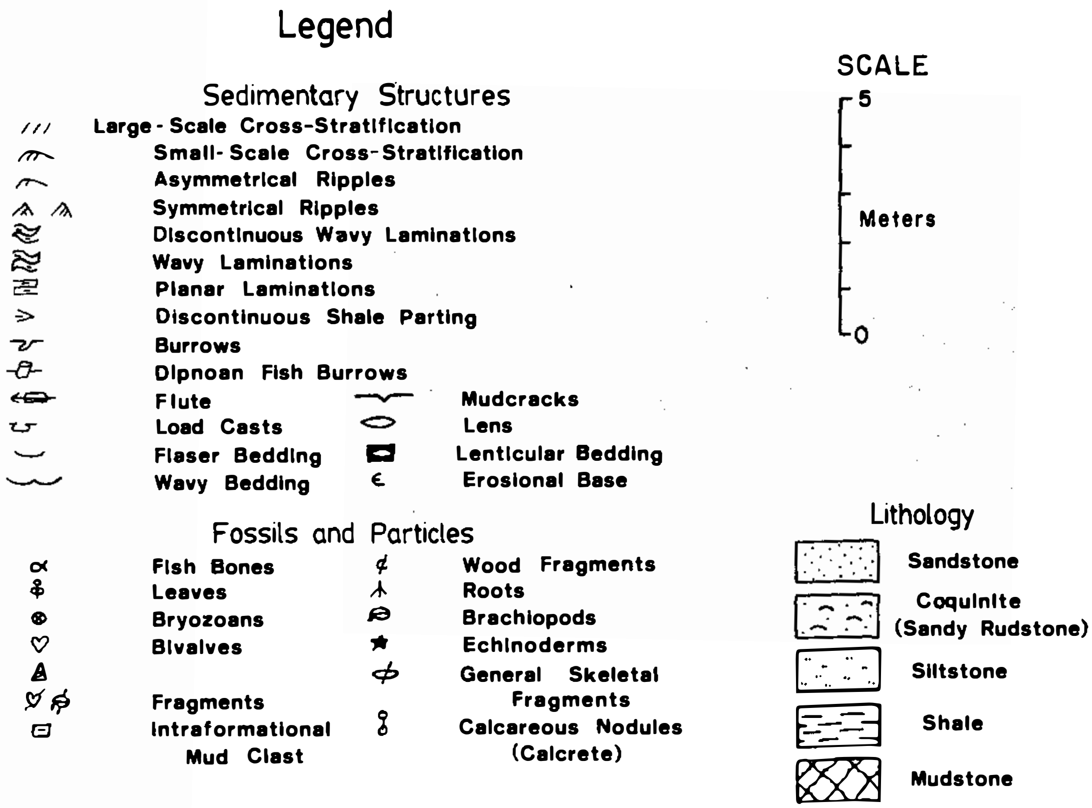


Figure 10. Measured sections from Case quarry. A is from the west center of the quarry; B is at the east end. Symbols for measured sections (including Figure 4) are above.

Quartz, rock fragments, and shells are the framework elements in these rocks (Table 1). Matrix consists of red hematitic mud and fine carbonate. Cement is predominantly calcite. Dolomite rhombs in the matrix and in internal skeletal cavities filled with red mud constitute several percent of some shelly samples.

Table 1. Major Components in Coquinite at the Louis Case Quarry, in Volume Percent.

	Shell-rich Rock	Shell-poor Rock
Quartz and chert fragments	14.7	21.5
Rock fragments	21.3	37.5
Fossils	32.9	9.7
Matrix and cement (carbonate)	17.4	18.5
Red mud	13.6	12.9

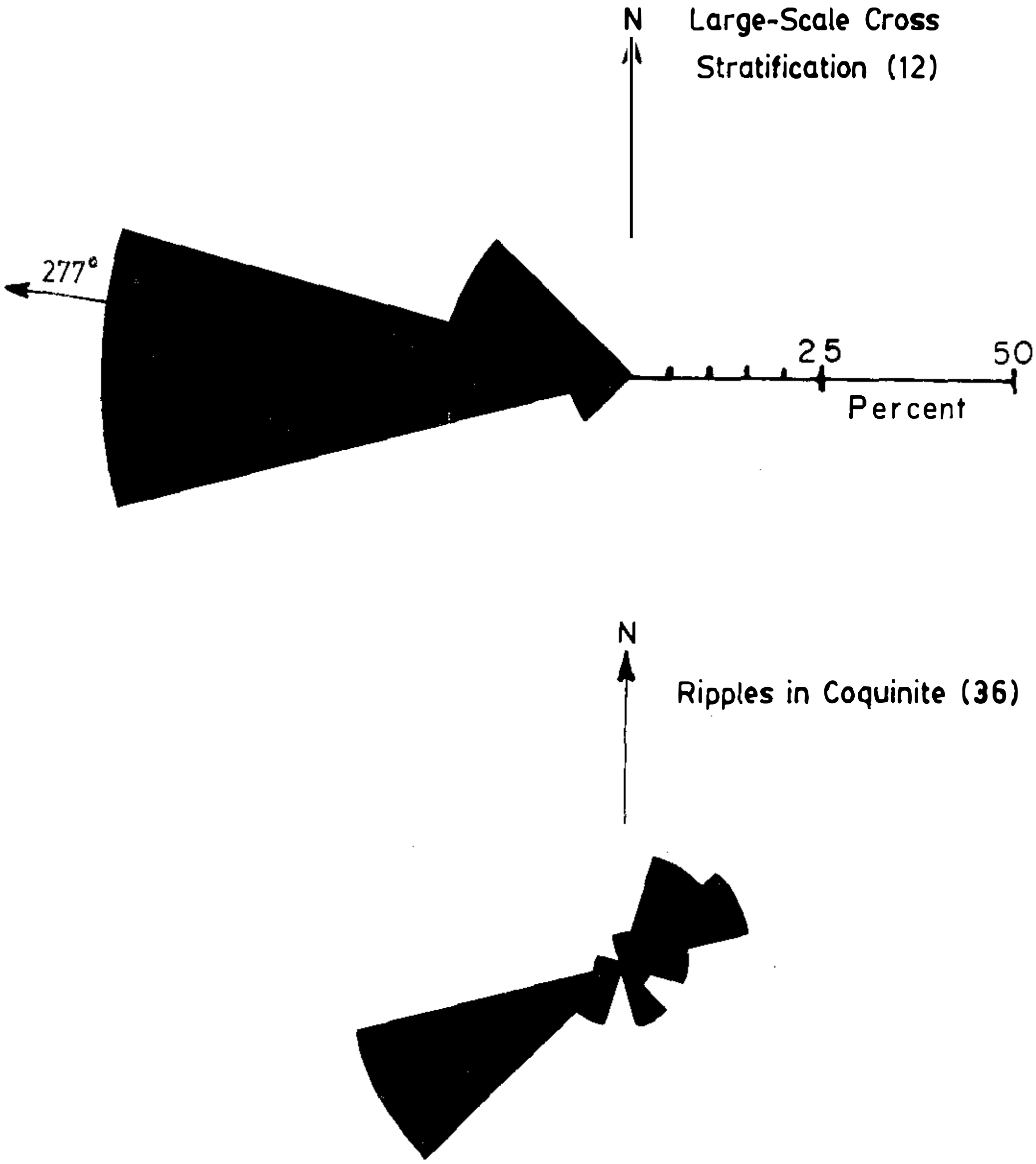


Figure 11. Paleocurrents from Case quarry. Scales are in percentage of total measurements in 30° intervals. A. Large-scale cross stratification, 12 readings, from the Luthers Mills Coquinite. Most readings are axial trends of troughs; the remainder are dips of cross-beds. B. Transport directions indicated by 36 asymmetrical ripple marks.

The hematite occurs as thick rims on well-rounded grains and locally partially replaces some grains. It is the pervasive coloring agent in the matrix. Two matrix samples contained 7 and 11.8 percent iron oxide but an average of only 4.4 percent Fe_2O_3. Rock fragments are mostly red and green shale, gray and green mudstones, and a variety of low-grade metamorphics. Many are rimmed with hematite. Fossils found in the coquinite are commonly broken and abraded. However, articulated brachiopods are not uncommon, and both colonies and individual branching bryozoans were found in two layers (about 9 m in Fig. 10A). Bivalves, crinoids, and bone fragments are less common. Many of the zooceia in bryozoa have been filled with hematite-rich mud which contains tiny rhombs of dolomite. Similarly, some bone and crinoid fragments have been replaced or infiltrated by hematitic mud in which dolomite rhombs are found.

The fauna represented in the coquinite is composed mainly of shelly invertebrates. The most common brachiopods are Cyrtospirifer sp. and Athyris sp.; they far outnumber the other fossils. Among the others are brachiopods, Camarotoechia sp. and Cryptonella sp.; bivalves, Grammysia circularis, Nuculoidea corbuliformis, Eoschizodus chemungensis, (Bottjer, 1981); and bryozoans, Leioclema cf. subramosa, Eridotrypella parvulipora and a bifoliate cryptosome (R. J. Cuffey, written comm., 1981). The bryozoans occur as single specimens, as colonies up to 70 cm across and 10 cm high, and as binders or encrusters enclosing other fossils. Trace fossils are a distinctive feature of the coquinite (D. B. Hutchinson, unpub. report, 1981). Two distinctive trace fossils are shown in Fig. 12; one (12B) is certainly a feeding trace. Other trace fossils found include: straight vertical burrows, U-shaped vertical burrows (rare), resting traces, Cruziana (?), tracks and trails. In many examples, activities of the trace-forming organisms have apparently moved shells enough to give beds a chaotic fabric.

Overlying Shales and Sandstones. The rocks extending about 15 m from the coquinite to the top of the quarry wall are a complexly interbedded sequence of gray-green silty shales; red silty shales; and red or green, lenticular, muddy sandstones and siltstones. Several thin shell-rich beds are found above the thick coquinite (Fig. 10). Small in situ carbonate nodules are found in red mudstones and shales at 19.5 and 21.5 m (Fig. 10B) along with some poorly preserved mud cracks. Vertical and inclined tube-casts up to 5 cm in diameter were found at 19 m (Fig. 10A) and 21.5 m (Fig. 10B). Similar casts have been variously interpreted as root impressions, fish burrows, or bivalve burrows. Large numbers of small, disarticulated fish plates and bones and delicately preserved plant imprints were found at 22 m (Fig. 10B). One of the plants has been identified as the fern, Archeopteris hallensis (J. D. Grierson, pers. comm., 1981). A bone and shell lag caps the overlying unit at 26 m. The uppermost unit exposed in the quarry high wall is a complex of thin, rippled sandstones and fissile silty shales with brachiopods throughout. Some flat mud clasts in these sandstones may be rip-ups of desiccation-cracked muds. The brachiopods Cyrtospirifer sp. and Hamburgia vera were collected from a gray-green shale directly above the coquinite. Bivalves have also been noted, among them Glossites depressus, (Bottjer, 1981).

ORIGIN OF THE CARBONATE UNITS: THE REALM OF HYPOTHESIS

Both carbonate units are unique in this part of the Catskill clastic wedge; to our knowledge, units as thick as these have been described from nowhere else in the sequence. Both are well exposed in the quarries where it is possible to observe many features. Both are reasonably well fixed in the stratigraphic sense and their facies context is known to the

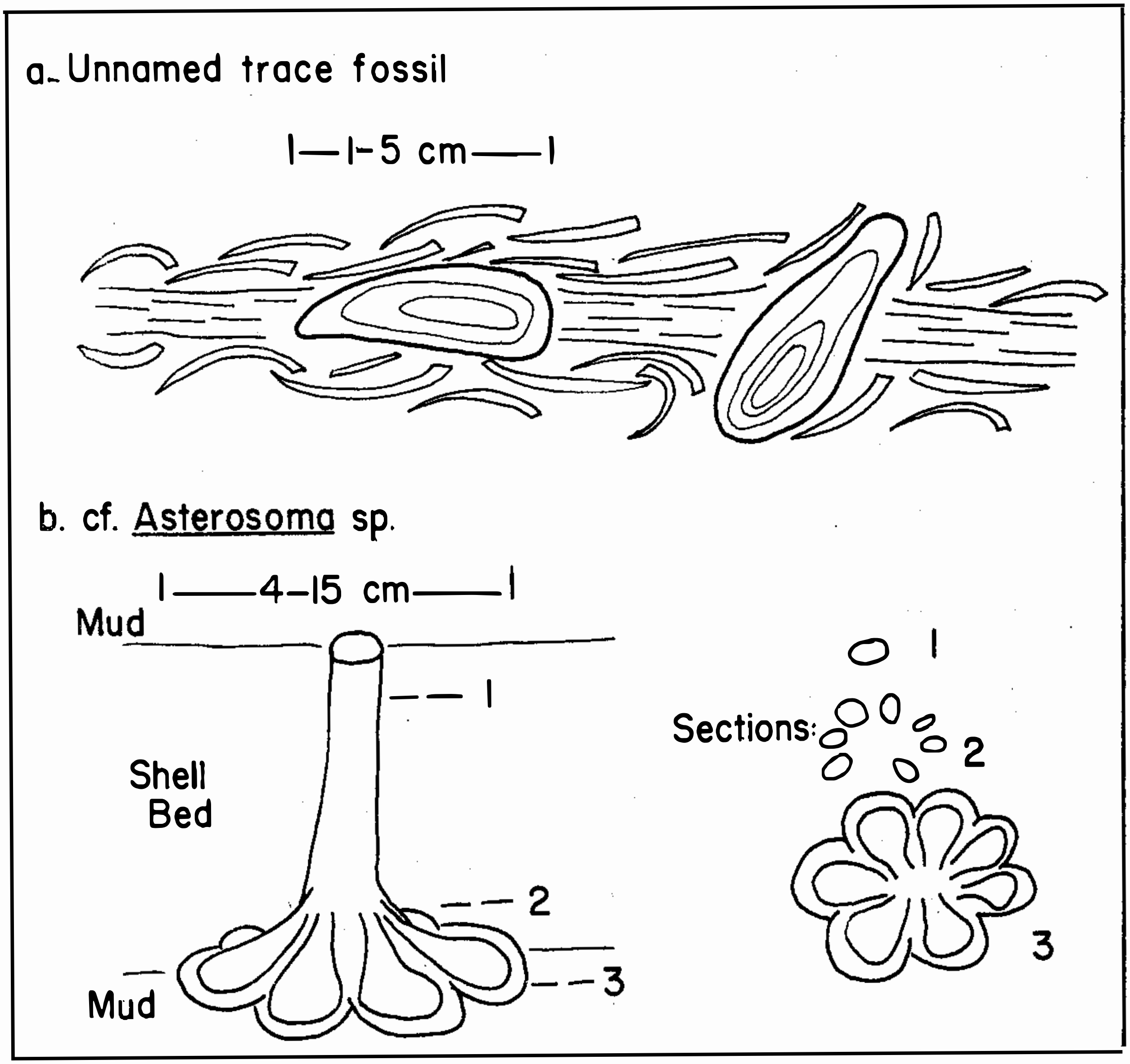


Figure 12. Trace fossils from Case quarry. A. Horizontal burrows with spreiten, shown in vertical section. B. Asterosoma-like feeding burrows. Penetration is 5 to 10 cm. After Bottjer (1981).

reconnaisance level. Salient features of the two units, including those which show similarities and those which set them apart are summarized in Table 2. Even with this information available, a fully satisfying interpretation of the origin of these rocks remains elusive. We offer here working hypotheses in order to stimulate discussion.

Table 2. Comparison of the Carbonate Sequences at Ashcraft and Case Quarries.

A. Features common to both:

Skeletal carbonate content is high in contrast to the enclosing terrigenous clastic rocks.

Bodies are discontinuous or isolated stratigraphically.

Location is toward landward edge of transgressive sequence.

Brachiopod shells dominate with some mollusc and crinoid debris.

Fish-plate fragments and large plant fragments abound.

Shale clasts are common.

Cross-stratification is abundant.

Trough cross-bedding opens westward (seaward).

Current directions are bimodal or polymodal.

B. Contrasting features:

	Ashcraft Quarry	Case Quarry
Geometry	Elongate mound ?	Biostrome, prism, or blanket
	Isolated?	Lateral equivalents
Bedding	Large-scale accretion bedding in inclined wedges	Planar beds enclose cross-bed sets
Color	Light gray, brown mottles	Red
Lithology	Skeletal grainstone (limestone)	Shelly quartz wacke (sandstone)
	Mud-free	Mud common as matrix and clay drapes
	Ankerite nodule clasts	Replacement dolomite
Skeletal alteration	Severely abraded	Generally disarticulated and fragmented but relatively unabraded

Table 2 (continued).

	Ashcraft Quarry	Case Quarry
	Diagenetically altered	Well preserved
Biota	Crinoids rare, bryozoans not observed	Crinoids and bryozoans abundant
Trace fossils	Confined to top surface	Numerous and diverse throughout
	Little bioturbation, especially in shelly beds	Bioturbation common in shelly beds
C. Contrasts in overlying rocks		
Lithology	Sharply bounded sandstone and shale	Intergradational silty shale, siltstone, and sandstone
Bedding	Distinctly layered	Complexly interbedded, lenticular
Color	Medium to dark gray	Gray-green and red
	Fish plates and plant fragments rare	Fish plates and plant fragments locally abundant
Trace fossils	Small, vertical "skolithus" tubes abundant at two levels	A few large ovoid tubes
	Common and varied traces on sandstone soles (hypichnia)	Hypichnia not observed

Ashcraft Quarry

The features of the limestone unit and of the strata surrounding it are explainable in terms of sedimentary process operating in shallow-marine environments where both tidal currents and waves were effective. There are few features of the limestone which suggest subaerial exposure, while the presence of brachiopods and crinoids and the development of a variety of sedimentary structures suggests deposition in shallow marine water. Reversals in current direction indicated by the paleocurrents, the textural maturity of the sand lenses within the limestone, and the lack of mud within the carbonate suggest a shallow-marine setting in which strong tidal currents winnowed out the fines and abraded the shell materials.

Shells, quartz sand, shale clasts, reworked nodules, plant fragments, and bones were all transported to this locality. The nodules and shale clasts must have been transported by strong currents from sources not too far distant, judging from the large clast size and relative lack of abrasion. The shells must have been transported from the nearby sea floor with extensive concentration and reworking to account for their numbers and abraded condition.

The internal plan of the carbonate mass seems to require three kinds of sedimentary processes: a) erosion and transport of sand, shells, nodules, etc. to the site; b) vertical accretion to form positive depositional relief; and, c) lateral accretion to produce the ultimate accumulation as either an isolated mound or a tabular body.

Thus, the carbonate mass appears to have been an accumulation built up above its surroundings by currents bringing to the site materials from the nearby sea floor or lagoon and from nearby terrestrial sources. Five hypotheses of accumulation have been suggested, no one of which is completely satisfactory or can be tested adequately with the available data. They are offered here as a basis for discussion.

Deposition as a subaqueous "shell mound" seaward of a beach/lagoon complex and/or a tidal flat was suggested by Krajewski and Cuffey (1976). Bowen (1978) postulated a "break-point bar" offshore from tidal flats. Both Krajewski and Cuffey (1976) and Bowen (1978) suggested significant contributions of skeletons produced in situ once initial relief above the sea floor was established. Another possibility is deposition on the subaqueous part of a land-tied accretion lobe or spit (Enos and Minero). Accumulation in the form of a tidal delta with both ebb and flood deposits represented has been suggested by John Bridge (pers. comm., 1981). Finally, some of the features in the unit are compatible with development as a small, mostly submerged, barrier bar as hypothesized by Woodrow.

In any event, nodules and other clastics were brought to the site of accumulation as the result of fluvial or shore-zone erosion and transport. A swamp seems the most likely source of ankerite nodules and abundant plant material. Shale clasts could have come from either the landward side or the seaward side of the carbonate accumulation. The source must have been one where mud normally accumulated and was periodically eroded. Wave and/or tidal scour of the nearby shelf provided the shells. Tides and waves reworked these particles into the limestone we see now.

Case Quarry

The points of similarity between the two carbonate bodies (Table 2) are such that we conclude that both were formed in tide-influenced marine environments. However, contrasts between them are sufficient to make it clear that the local geography of the Catskill shoreline at the two sites must have been different. The abundance of brachiopods and the presence of crinoids and in situ bryozoa demonstrate that the environment of deposition of the coquinite was subtidal marine. Tidal effects are indicated by material from both marine sources (shells) and terrestrial sources (plant fragments, red muds), by the intimate intermixture of sandstone and shale (including flaser bedding), and by some examples of reversals in current direction. Most of the large-scale cross-strata record flow toward the west or northwest suggesting ebb-dominated deposition.

If the question of process is tentatively answered, we are left with the question posed by the mixture of materials in the coquinite. The source of the shell material was the local sea floor, despite the apparent ebb dominance, but the source of the red mud is less obvious. A source of highly oxygenated muds is required, perhaps from the higher, subaerial parts of a tidal flat or nearby alluvial plains. Plant debris could derive from the same sources. Shells in great quantity require a large community of epifauna either close to or as part of a tidal shore. The bryozoan clusters, at least, point to some production _in situ_.

The coquinites of the Case quarry have been interpreted as a part of a tidal delta (Woodrow) and as a subtidal zone of a tide-dominated fluvial delta (Bottjer). Strata above the coquinite indicate alternations of shallow marine and non-marine or supratidal environments. They contain mudcracks, _in situ_ pedogenic carbonate nodules in red mudstones, and the enigmatic burrows or root cases; interlayered are beds containing marine fossils.

The local Upper Devonian shoreline was dominated by tidal processes, not by large streams and fluvial-dominated deltas. Material delivered to the shore appear to have been carried by small streams draining an alluvial plain or in channels of a tidal coast. Evidence of a large tidal range is lacking; a range of less than a couple of meters would probably account for the tidal features. Flow velocities were high enough in some areas, however, to move coarse material (shells and sand) as well as fines and to deposit them as migrating dunes and ripples. The localization of thick carbonate units may result from local amplifications in the tidal range or tidal velocity perhaps caused by indentations in the shoreline (bays).

In summary, both sequences appear to result from the concentration of shell debris, quartz, rock fragments, nodules, and wood by high-energy sedimentary processes in ebb-dominated tidal environments. Differences in lithology and geometry between the two units probably reflect local differences in the shoreline.

ACKNOWLEDGEMENTS

We are indebted to Louis Case and to Kirk Ashcraft for permission to study the quarries and for permission to visit the quarries on this field trip as well as on many previous occasions. Stephen A. Krajewski and Roger J. Cuffey generously shared unpublished data on Ashcraft quarry and Cuffey identified the bryozoans from Case quarry.

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ROAD LOG FOR BIOCLASTIC CARBONATE UNITS IN THE CATSKILL CLASTIC WEDGE

CUMULATIVE MILEAGE	MILES FROM LAST POINT	ROUTE DESCRIPTION
0.0	0.0	Begin trip at main entrance to SUNY Binghamton (Harpur College) campus, jct. of Glenn G. Bartle Drive with N.Y. 434. Go west (left) on N.Y. 434, but hold right to exit on N.Y. 201.
0.15	0.15	Exit right (N) onto N.Y. 201 toward Johnson City.
0.7	0.55	Bridge over Susquehanna River.
1.0	0.3	Enter Johnson City traffic circle (with caution!); prepare to exit right on Riverside Drive west (3rd exit).
1.1	0.1	Exit right (NNW) into Riverside Drive (N.Y. 201). Follow signs to N.Y. 17 west.
2.0	0.9	Exit right onto cloverleaf entrance to N.Y. 17 west.
4.6	2.6	Susquehanna River.
10.8	6.2	Apalachin, N. Y., exit. Continue west on N.Y. 17.
14.0	3.2	Esker on north (right) side of N.Y. 17. Much of esker and associated kamic deposits have been removed for aggregate.
19.7	5.7	Rest area, westbound lane.
20.3	0.6	Steep wooded hill with small exposures of Chemung (shallow marine) facies.
30.7	10.4	Road cut, Chemung facies.
32.7	2.0	Susquehanna River.
36.9	4.2	Exit right for U.S. 220. Turn left (S) on U.S. 220 at stop sign.
40.1	3.2	Chemung River.
40.9	0.8	Exposure of Chemung facies on right (W).
41.8	0.9	Jct. Pa. 199; continue south on U.S. 220.
47.0	5.2	Road cut in Wiscoy Fm. (?), Chemung facies, (shallow marine).

48.0	1.0	Blinker light at Bridge St., Ulster, Pa.
51.3	3.3	Road cut in lower Towanda Fm. Cattaraugus facies; red beds, some coquinite.
52.6	1.3	Turn right (W) on U.S. 6.
53.6	1.0	Road cut in Towanda Fm. with abundant flow rolls (pillow structures).
57.6	4.0	Community of Luthers Mills.
59.3	1.7	Enter Louis Case quarry on right (N) side of U.S. 6. Obtain permission from Louis Case in houst at west edge of quarry. STOP 1.

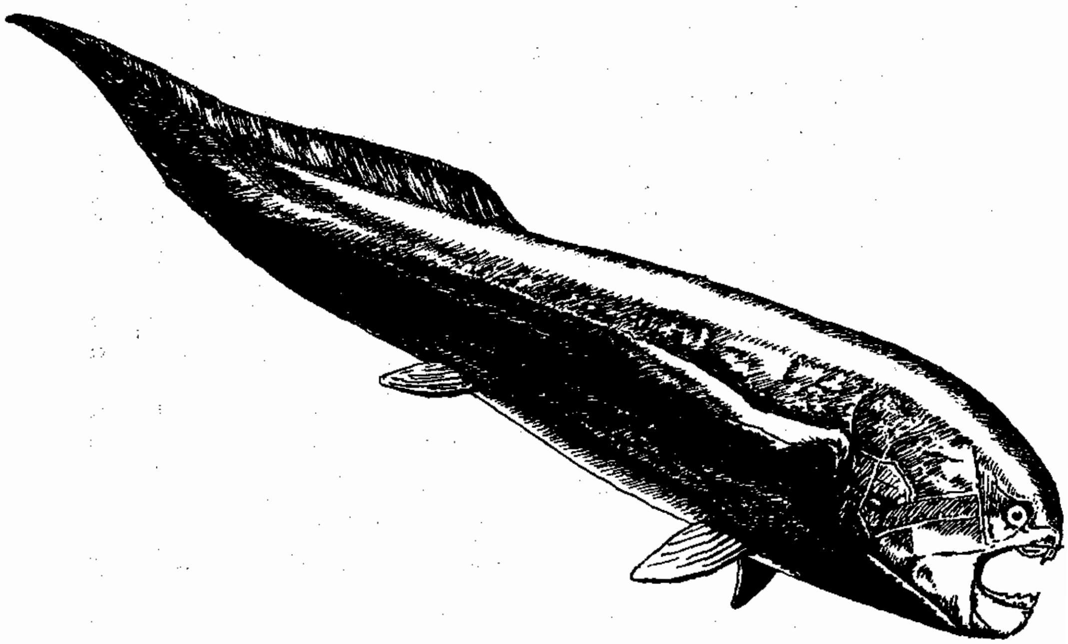
STOP 1. LOUIS CASE QUARRY, Burlington, Pa. Start in small face below main bench of quarry. Work upward toward east end of quarry.

59.4	0.1	Enter U.S. 6 eastbound by turning left at quarry entrance. Retrace route to Apalachin, N. Y.
59.6	0.2	Kiln on left (N) side of road where coquinite was exploited for agricultural lime and mortar. This is the type locality of the Luthers Mills Coquinite.
63.6	4.0	Jct. township road. Large blocks of coquinite exposed in bed of Sugar Creek at sharp curve to west approximately 1/2 mi. north of U.S. 6. Long road cut in Towanda Fm. beginning at curve on township road.
65.5	1.9	Overview of Susquehanna Valley to north (left).
66.1	0.6	Turn left (N) on U.S. 220.
66.8	0.7	Overview up Susquehanna Valley to north.
67.4	0.6	Road cut in Towanda Fm.
71.7	4.3	Road cut in Wiscoy Fm. (?)
75.0	3.3	Milan, Pa. Continue north on U.S. 220.
81.8	6.8	Turn right (E) onto N.Y. 17.
89.4	7.6	Rest area at Tioga Park race track (a monument to human folly). Potential lunch stop.

100.8	11.4	Owego, N. Y. Continue east on U.S. 17.
108.1	7.3	Exit to right for Apalachin (Exit 66). Turn left (E) immediately on N.Y. 434 at stop sign.
110.2	2.1	Broome-Tioga County line.
110.6	0.4	Turn right (S) into Tracy Creek Road. Follow it through community of Ross Corners.
111.3	0.7	Outcrops of Rhinestreet Fm., Chemung facies, to right in bed of Tracy Creek.
115.4	4.1	Jct. with O'Connell Rd. Tracy Creek Rd. becomes Collins Rd. Continue ahead and bear right up hill.
116.3	0.9	Pa.-N.Y. state line.
116.35	0.05	Turn right on unmarked road at locked gate. Entrance to Ashcraft quarry. Permission to visit quarry is obtained from Ashcraft Excavating Co. on Pennsylvania Ave., 2.5 mi south of N.Y. 434 in Apalachin, N. Y.
117.3	0.95	Follow road into Ashcraft Quarry. STOP 2.

STOP 2. ASHCRAFT QUARRY, Little Meadows, Pa. Start in pit near east end of quarry. Work along main face to west end. Climb onto bench at west end.

118.25	0.95	Return to Collins Rd. at gate. Turn left (W) and follow Collins Rd.-Tracy Creek Rd. back to N. Y. 434.
124.0	5.75	Tracy Creek Rd. ends at N.Y. 434. Turn right (E) on N.Y. 434 and continue through Vestal.
131.3	7.3	Main entrance to SUNY Campus, Bartle Drive. END TRIP.



Dinichthys, the largest Devonian placoderm.
From B. Kummel, 1970, History of the Earth,
p. 238. Redrawn by Kevin Enos.

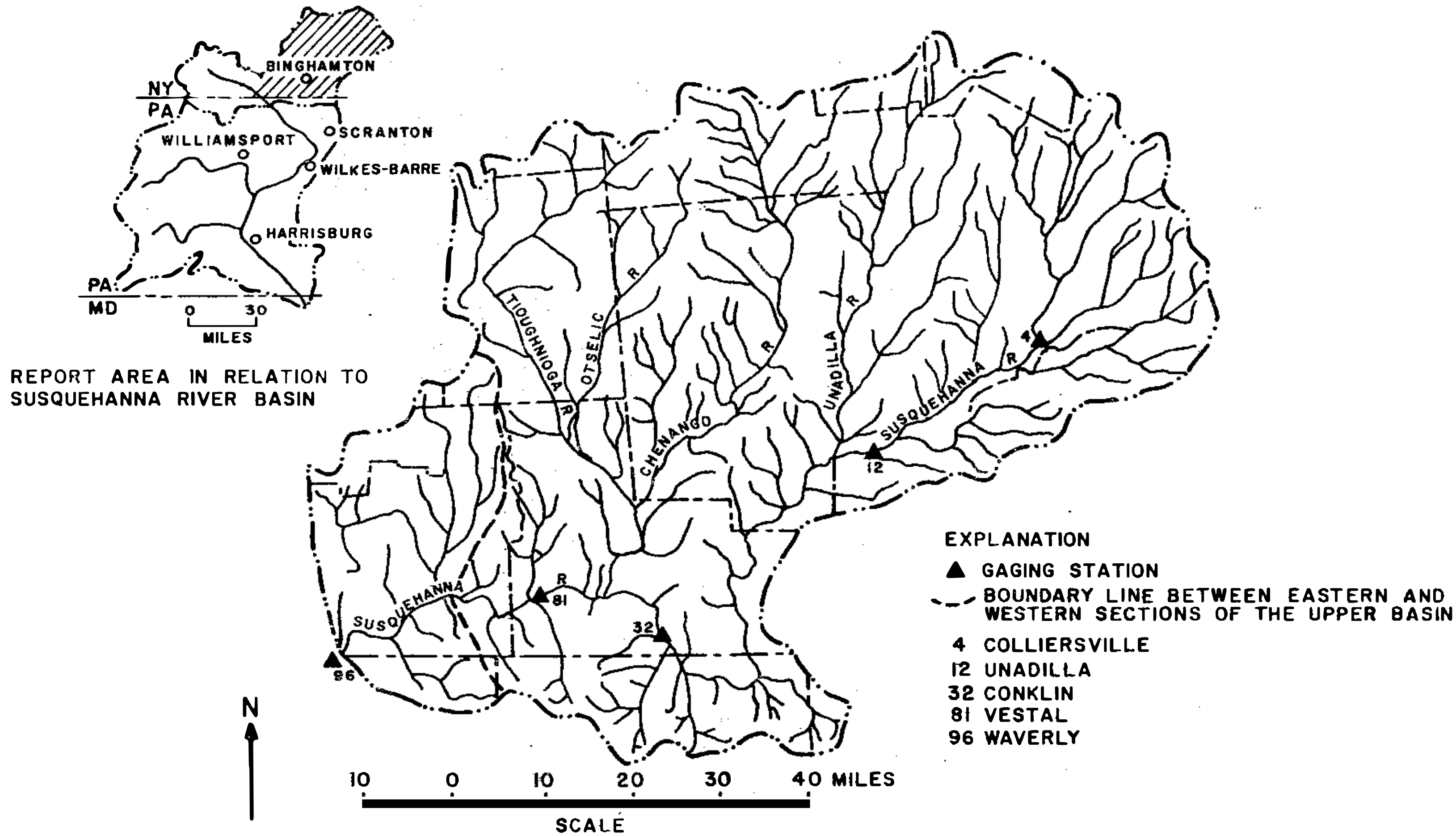


Figure 1. Index map of the eastern Susquehanna River basin.

SOME ENVIRONMENTAL PROBLEMS OF THE BINGHAMTON AREA

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THE SUSQUEHANNA RIVER BASIN

The Susquehanna River system is the largest drainage network in the northeastern United States draining to the Atlantic Ocean. Starting in central New York, the river flows southward through Pennsylvania and Maryland, draining 27,510 square miles (Fig. 1). We are concerned only with the eastern Susquehanna basin of New York in this report. The river here drains an area of 4,780 square miles in New York and Pennsylvania above and through the Binghamton area to Waverly, where it turns south into Pennsylvania and leaves New York. The major tributaries of the eastern Susquehanna River are the Otselic, Unadilla, Tioghnioga and and Chenango Rivers (Fig. 1). The Chenango joins the Susquehanna River at Binghamton. Indeed, this confluence determined the location of Binghamton.

The eastern Susquehanna River basin lies in the Appalachian geomorphic province. The bedrock is sedimentary sandstone, siltstone, and shale of Devonian age. The strata are essentially horizontal, but are slightly arched up into broad, gentle folds with axes oriented northeast-southwest. The folding generally has not markedly affected the basic dendritic drainage pattern of the section.

The region has been glaciated, resulting in a somewhat subdued topography. Hills have been smoothed and rounded and are commonly asymmetrical with steeper slopes on the north. Elevations range from 2500+ feet on the uplands to 750-850 feet along the river bottoms. The major valleys were broadened and deepened by glaciation and filled with thick deposits of glacio-fluvial sands, gravels, silts, and, in some cases, lake clays. Many of the small postglacial streams have cut steep, narrow gorges through bedrock. The combination of stream types and broad, open uplands gives a pleasing esthetic quality to the region.

Glaciation had a significant effect on drainage, not only in ways already mentioned, but also by disrupting and blocking pre-glacial drainageways. The extraordinary path of the Susquehanna as it loops down to Pennsylvania and back into New York east of Binghamton is a reflection of events during deglaciation. Many tributaries flow in "misfit" valleys which are too large for them. Drainage divides occur in "through valleys", i.e., a valley which is occupied by streams one of which flows north and the other south. Many obvious drainage diversions can be seen throughout the region.

Besides such changes, the glaciers exerted their influence on the Susquehanna drainage through the deposits they left. The uplands and valley side slopes of the watershed are covered with glacial till. This results in soils which are generally impermeable and poorly drained.

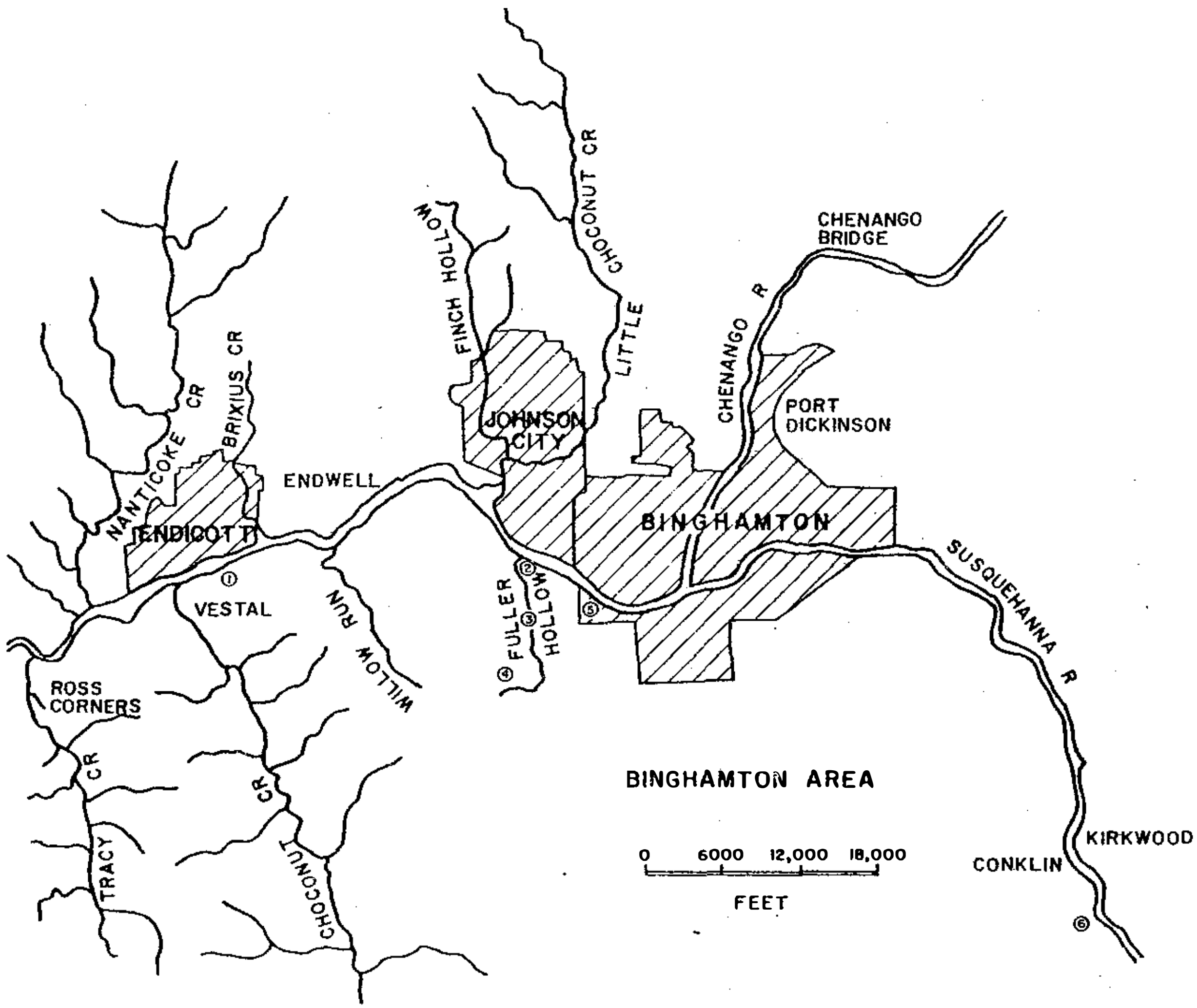

Figure 2. Map of the Binghamton area with stops indicated.

Hence, runoff is rapid and many tributaries are "flashy", i.e., have a quick rise and fall of discharge. The thick glacio-fluvial fills in the major valleys are good ground-water reservoirs which sustain flow of the larger streams throughout the dry summer months (Ku, Randall and MacNish, 1975).

The soils of Broome County were formed in glacial till, glacial outwash, glacial-lake deposits and more recent alluvial deposits. Soils in the low-lying areas, along the floodplains, are mostly of the Tioga-Chenango-Howard association. These are soils that are deep, well-drained, and gently sloping and are, therefore, very suitable for development. The main problem associated with these soils is that of occasional flooding.

The terraces bordering the floodplains are primarily Chenango, Howard, and Unadilla soils. Like those found in the floodplains, these soils are deep and well-drained (S.C.S., 1971).

In most of the county, particularly in the uplands, soils of the Volusia-Mardin association are formed on deep, gently sloping to very steep glacial till. These soils are not suitable for most types of development, because they exhibit a slowly permeable fragipan. A fragipan is a dense subsurface layer of soil; it is indurated, hard and slowly permeable. The Volusia fragipan is composed of grayish-brown silt-loam at a depth of 15-22 inches. This is not to say, however, that development has not occurred in areas with these soils; there has been little choice because these soils cover about 90 percent of the county.

The glacial modification of the topography has largely determined the human geography of the region. Population is mostly concentrated on the broad flood plains and terraces which are locally as much as two miles wide. Broome County has the higheset population density in the eastern basin, with development concentrated in the Triple Cities (Binghamton, Johnson City, Endicott) section along the Susquehanna (Fig. 2). The other counties in this watershed are primarily rural. Land use shows the effect of soil type. Upland and valley slopes in till are generally forested or in pasture. Much of the agricultural land is on the broad flood plain composed of glacio-fluvial deposits.

A conflict in use arises since the flood plains are also the places most easily and economically developed. The aquifers in the valley fill and the permeability of the sands and gravels for septic systems make the valleys more desirable for housing. During the post-World War II development boom, extensive urbanization occurred in the valleys, along the Susquehanna River itself and up larger tributaries. At present, 66 percent of the population resides in the strip of flood plain along the Susquehanna River. The steep slopes of the uplands tended to act as natural development barriers. It is only recently, with continued growth and some realization of the dangers of building on flood plains, that urbanization has spread to the flat upland summits and the valley side slopes. Urbanization of these seemingly innocuous areas also brings on drainage and river problems, as will be seen.

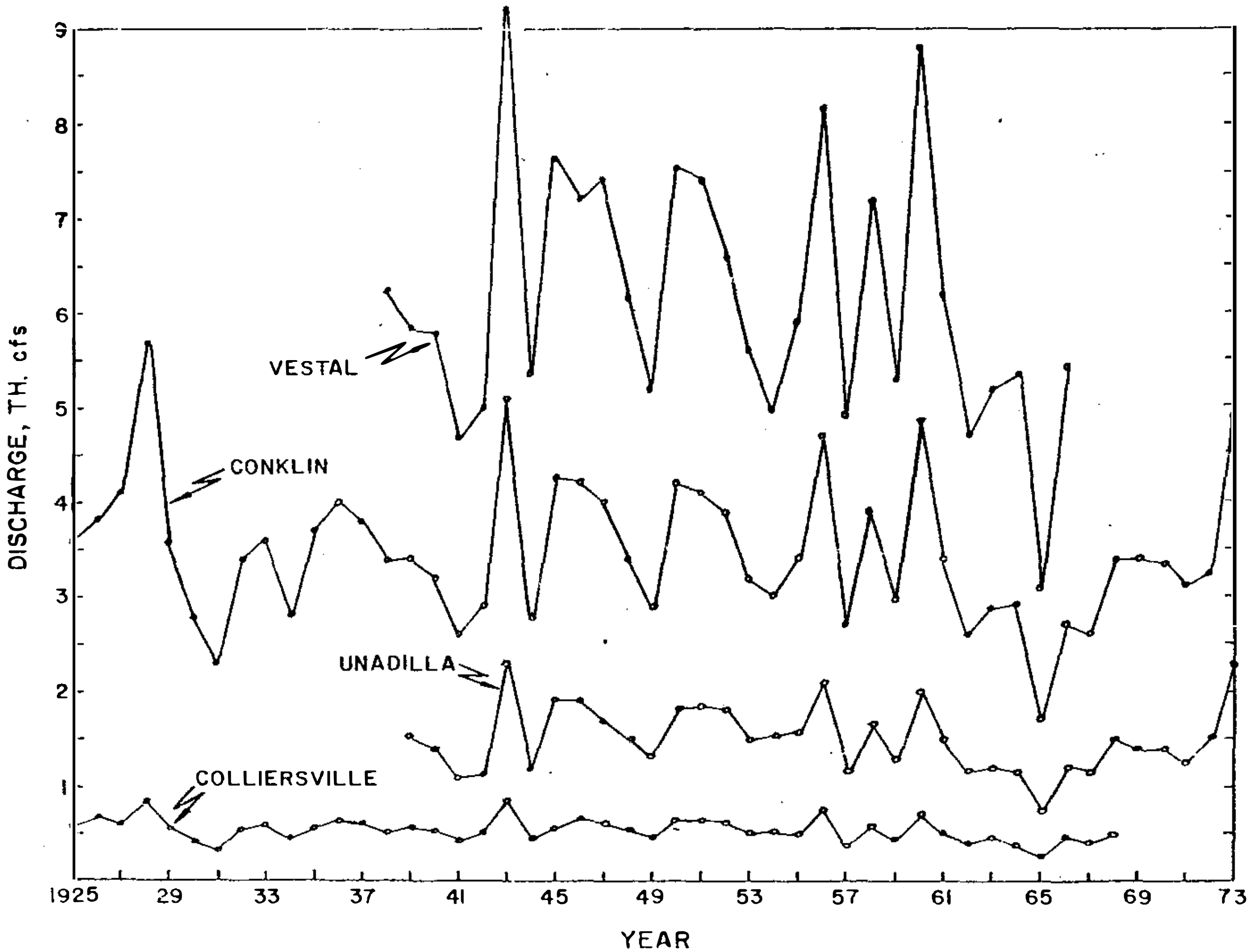

Figure 3. Annual discharge on stations at or upstream of the Binghamton area. From Morisawa and Vemuri (1975).

Hydrology of the Eastern Susquehanna Basin

The area has a humid, continental climate with an average precipitation of 36-40 inches per year. Precipitation is generally of the frontal type, where polar air masses meet the more humid warm air masses moving northeastward. The record flood of 1936 was produced by such frontal precipitation combined with a spring thaw (Susquehanna River Basin Study, 1970). Although the summer is dry, intense local thunderstorms may occur. The region also lies in the path of tropical hurricanes. These storms, originating in the Atlantic or Caribbean, sometimes swing inland bringing intense and excessive rainfall. Severe damage has been caused in the past by these tropical storms. More recently, Agnes (1972) and Eloise (1975) caused considerable damage on smaller tributaries, but did not cause damaging floods on the main stem.

Table 1

Highest Floods of Record, Binghamton Area

	Date	Stage ft.	Estimated Discharge, cfs
Susquehanna River			
Conklin	Mar. 1936	20.14	61,600
Vestal	Mar. 1936	30.5	107,000
Waverly	Dec. 1952	19.7	112,000
Chenango River			
Chenango Forks	July 1935	20.3	96,000
Broad Acres	July 1935	20.6	96,000

Data from Susquehanna River Basin Study, 1970.

The water budget reflects the difference between precipitation over the watershed and discharge flowing out of the basin. The runoff (20.8 inches) reflects 54 percent of the mean annual precipitation. Forty-six percent of the rainfall is lost by evapo-transpiration because the area is well forested and 87 percent of the watershed is agricultural or vacant.

There are four gaging stations on the eastern Susquehanna River main stem. Annual flow for the periods of record and flow-duration curves are shown in Figures 3 and 4. The four stations below Colliersville (Fig. 4) reflect the contribution of the thick valley fill which act as acquifers contributing to stable base flow. This is denoted by the levelling off of the curves at approximately 98.99 percent of the time with a good discharge. Note the difference between the tails of these curves and that of the Colliersville station.

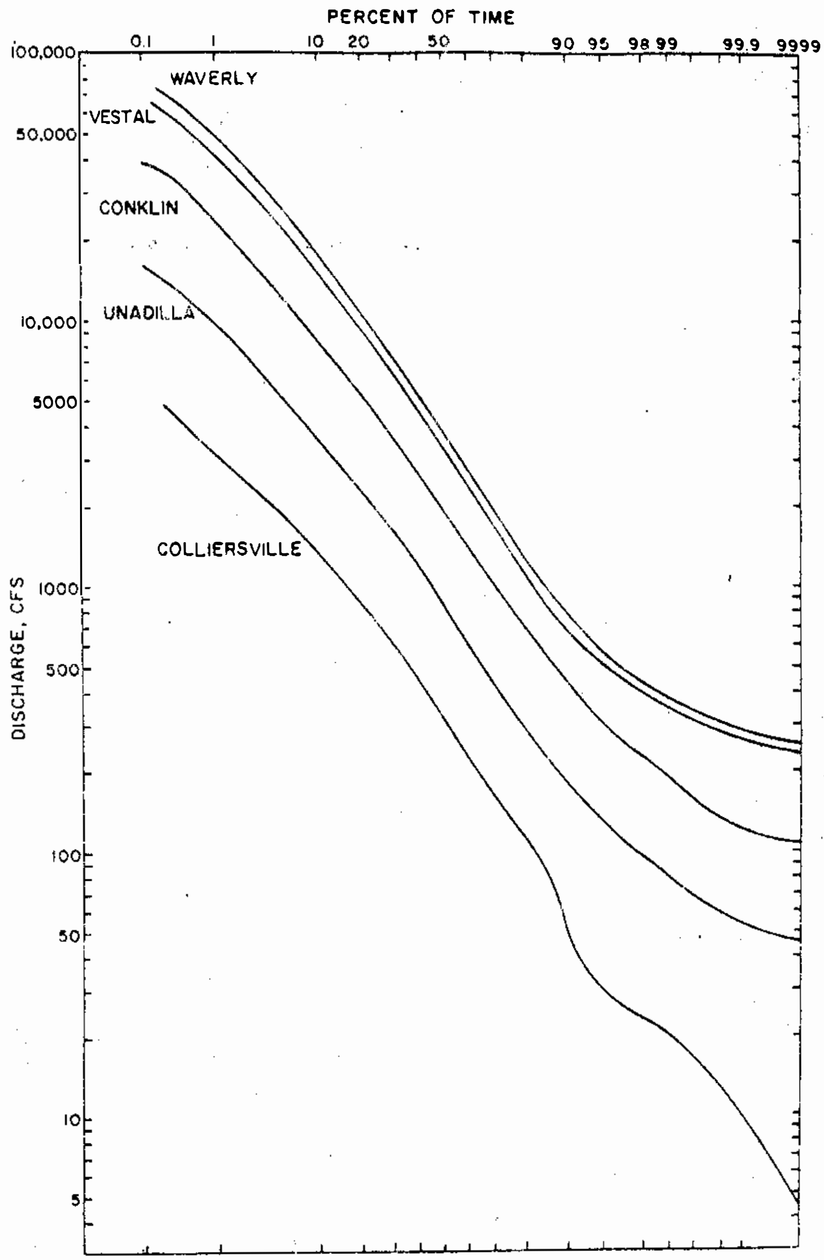

Figure 4. Flow-duration curves, Susquehanna River at Waverly and upstream. From Morisawa and Vemuri (1975).

The recurrence-interval curves (Fig. 5) indicate the average time interval at which a given discharge recurs. These can be used either to make predictions of peak flow or to determine the frequency of a given storm. For example, at Vestal a high discharge of 70,000 can be expected every 5 years. On the other hand, the peak flow of the storm Eloise in September of 1975, which was 61,500 cfs in Vestal could be expected about every 3 years or so, i.e. a 3-year recurrence interval.

The graph (Fig. 6) and the regressions which relate drainage area on the Susquehanna to mean annual discharge and to peak flow allow a prediction of these discharges, if one knows the area of basin above any point on the main stem. Cortland and Sherburne are on the Tioughnioga and Chenango Rivers, respectively, so their peak annual discharges lie somewhat off the regression line for the main stem Susquehanna River. Regressions of mean annual discharge ($\overline{Q}$) and mean annual peak flow (Qp) to watershed area (A) are:

$$\overline{Q} = 1.5\ A \qquad r = .99$$

$$Q_p = 25.7\ A^{0.93} \qquad r = .96$$

Note that the scales on the graph of Figure 6 are logarithmic. That is, the regression equations can be written:

$$\log \overline{Q} = \log 1.5 + \log A$$

$$\log Q_p = \log 25.7 + 0.93 \log A$$

Of particular importance in understanding the hydrology of the Binghamton area are the gages at Conklin above Binghamton and at Vestal below it. Table 2 gives the mean annual discharge per square mile of drainage area of the Conklin and Vestal stations. Since the area between these two gaging stations represents much of the urbanized stretch of the region, an attempt was made to evaluate the change in discharge which might be attributed to urban growth. In order to discount the amount of water carried into the Susquehanna by the upper Chenango, the Vestal flow minus the discharge at Chenango Forks was used (column 3). This was then recalculated to account for the increase in area of the Susquehanna Basin to Vestal over the area of the Chenango. To minimize precipitation variability, a ratio was calculated (column 5). This ratio represents the proportionate contribution of the basin over the Binghamton reach to the flow of the Susquehanna. Several points should be kept in mind. First, the Chenango River contributes a great deal of flow to the Susquehanna River. This is important at times of peak flow, because the city of Binghamton lies at their confluence and backwater effects at the junction can be disasterous. Also, at time of drought the low flow of the Susquehanna is augmented by discharge from the Chenango valley outwash deposits. Finally, the table shows that through 1956 there was a fairly constant ratio of discharge above Binghamton to that contributed by the urbanized stretch. However, a spurt of development in the late fifties resulted in a jump in this ratio after 1956.

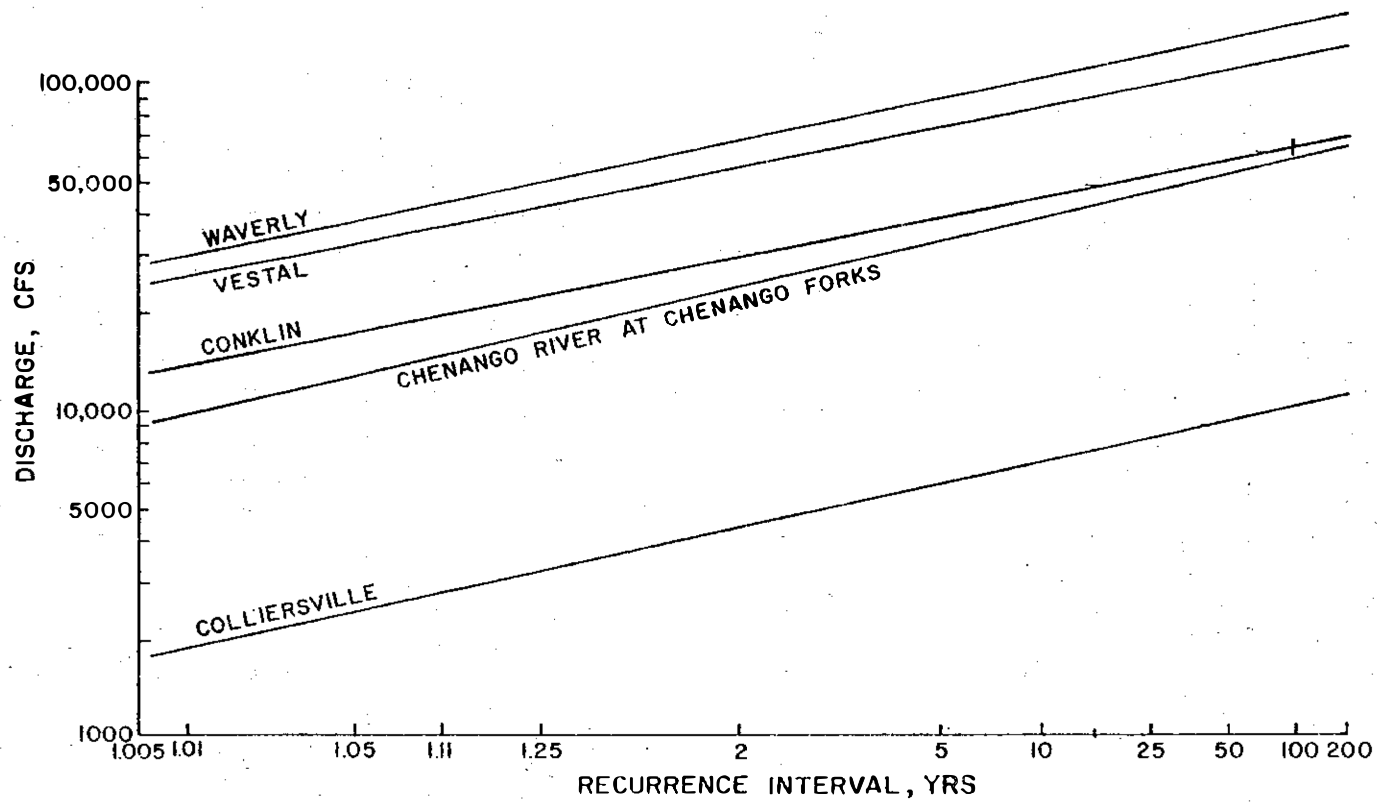

Figure 5. Flood-recurrence-interval curves, eastern Susquehanna Basin. From Morisawa and Vemuri (1975).

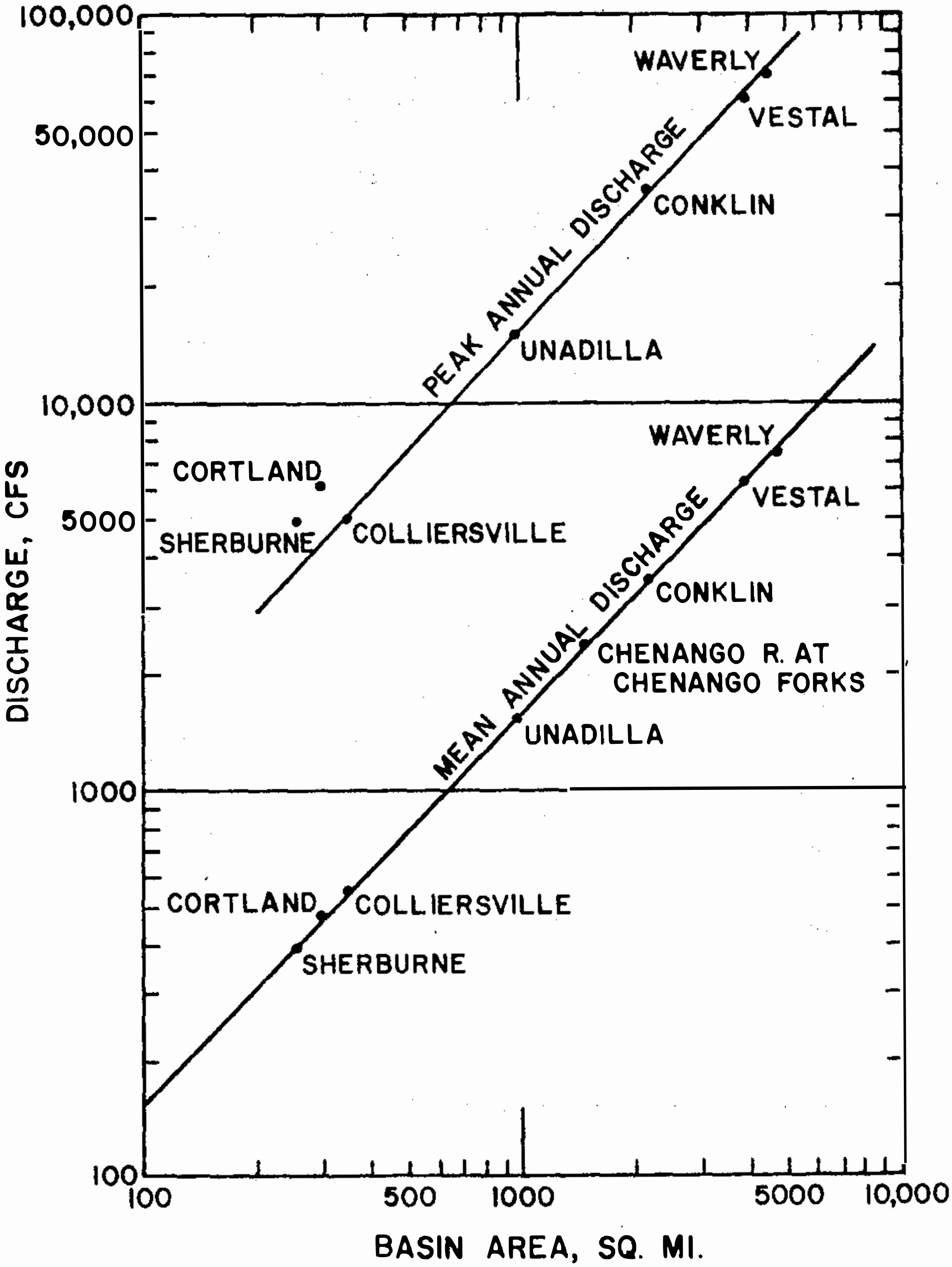

Figure 6. Relation of mean annual discharge and peak flow to basin area, eastern Susquehanna watershed. From Morisawa and Vemuri (1975).

Table 2

Ratios of Discharge per Square Mile, Susquehanna River at Binghamton

(1) Year	(2) Vestal cfs/mi^2	(3) Vestal minus Chenango Forks cfs/mi^2	(4) Conklin cfs/mi^2	(5)* V-CF/C
1941	1.18	1.14	1.16	.98
	1.27	1.28	1.30	.98
	2.33	2.26	2.31	.98
	1.35	1.25	1.26	.98
1945	1.92	1.87	1.90	.98
	1.83	1.87	1.90	.98
	1.88	1.76	1.78	.99
	1.55	1.52	1.53	.99
	1.32	1.30	1.31	.99
1950	1.90	1.86	1.87	.99
	1.86	1.83	1.85	.99
	1.68	1.70	1.74	.98
	1.42	1.42	1.45	.98
	1.25	1.34	1.36	.98
1955	1.50	1.50	1.53	.98
	2.07	2.09	2.10	.99
	1.25	1.23	1.20	1.02
	1.81	1.82	1.76	1.03
	1.35	1.35	1.32	1.02
1960	2.22	2.27	2.17	1.05
	1.58	1.56	1.50	1.04
	1.19	1.16	1.16	1.00
	1.30	1.32	1.28	1.03
	1.34	1.30	1.30	1.00
1965	.78	.72	.75	.96

*Column 3/column 4

Flow data from U.S. Geol. Survey computer printouts.

Fig. 7 shows that 1945 was a year of high rainfall, yet the ratio (V-CF/C) remained the same as in 1952, a low-rainfall year. This probably reflects the effect of ground-water storage. Even though 1957 was a dry year, culminating a downward trend in precipitation, the ratio increased and remained high until the excessively dry 1965, after 5 years of drought. Unfortunately, the Vestal gaging station was discontinued after 1965 so data beyond that year is not available. It should also be noted that Binghamton gets its water from the Susquehanna River below the Conklin gaging station and this may account for some loss of water in the urbanized area. These ratios indicate that there has been an increase in the mean annual discharge per square mile in the urbanized Binghamton region, a result of growth and development.

EFFECTS OF URBANIZATION

One of the major problems in the metropolitan areas of the eastern Susquehanna River basin (as in many other watersheds) is urban growth and the settlement pattern. Early settlers established the city along the river in the broad flood plain at the junction of the Susquehanna and Chenango Rivers. Since this was the easiest, most economical, and most accessible place; the town grew by spreading along the river channel.

The settlers did not understand the fact that a river develops its network pattern and channel morphology in adjustment to the prevailing environmental conditions of the geology, topography, and hydrology of the watershed. The flood plain is an integral part of the river's drainage system, especially during times of peak flow. At such times the river overflows its normal channel and flows out over its extra-channel right-of-way, the flood plain. The flood plain is thus a normal escape valve for exceedingly high discharges and acts to increase flow capacity. It also serves to decrease velocity, acts as temporary storage, and promotes infiltration into the flood-plain sediments. Floods also serve to replenish the fertility of the flood-plain soil. Disruption of the natural way in which the stream discharges excessive flow is dangerous.

Urbanization disturbs the natural system of land drainage. Denudation of the surface and covering the land with buildings, streets, and parking lots changes the run-off and, thereby, the hydrologic balance. Rain water, no longer able to infiltrate the permeable sand and gravels of the flood-plain, runs off immediately into the rivers. In fact, development generally aids this run-off by supplying ditches, sewers, and storm drains to move rainfall quickly to the local streams. Such a practice increases peak flows and shortens the time lag to peak discharge. It also reduces ground-water recharge and thus reduces low-flow rates.

The filling of channels and flood plains to reclaim more land for development or for highways reduces channel capacity and, again, increases the potential for flooding. As urbanization spreads flood hazards grow, since runoff increases with a given rainfall. Also, as time goes by the probability of more extreme rainfall events increases.

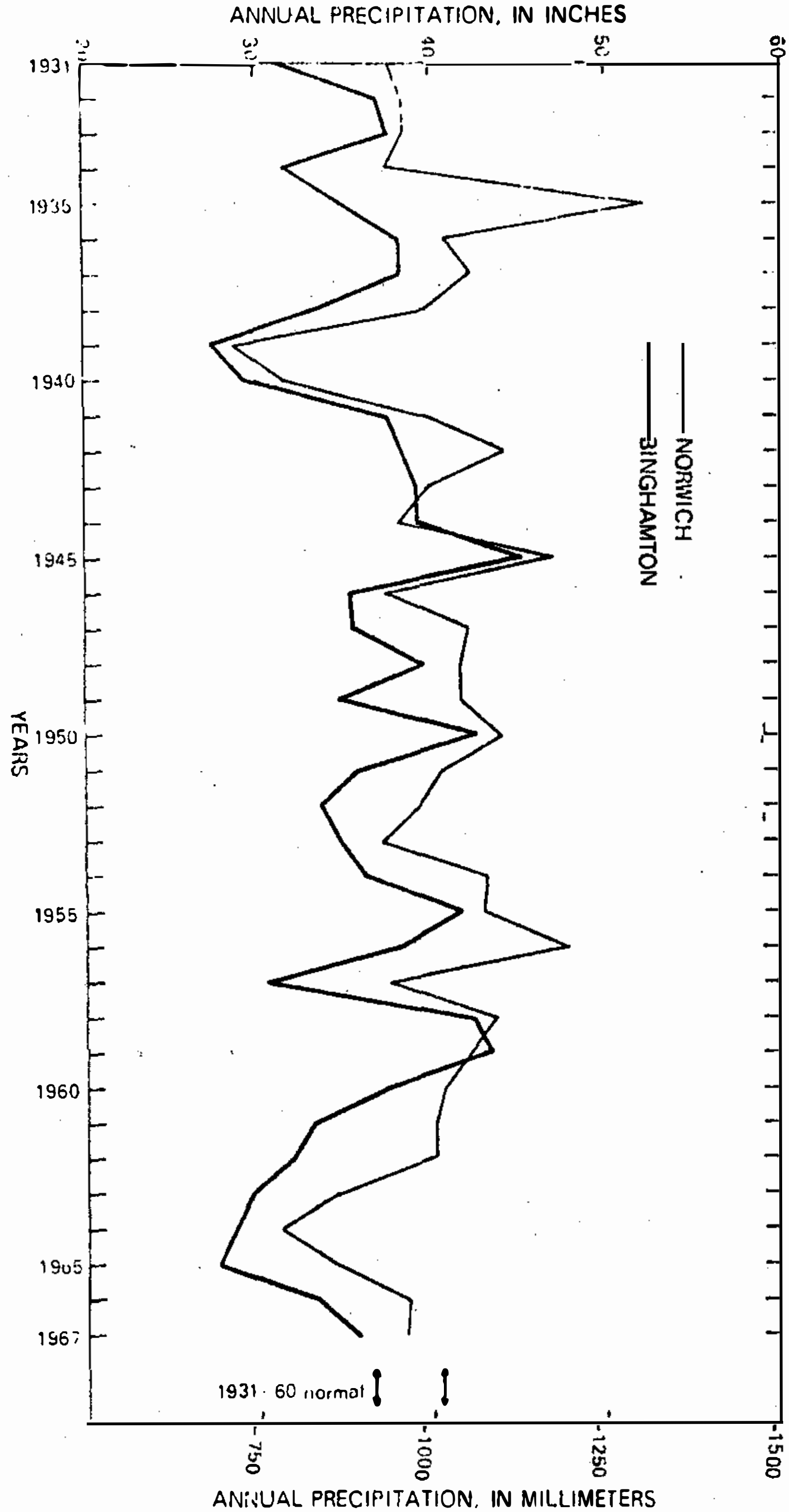

Figure 7. Annual precipitation at Binghamton and Norwich. (From Ku, Randall and MacNish, 1975).

Thus, by encroaching on the stream right-of-way, the flood plain, and by converting land to impervious surfaces, man has intensified the flood hazard. Floods occur and so man reacts, and his reactions have traditionally been in terms of structural measures to "control" the river. Instead of treating the illness, he treats the symptoms. He scratches the itch instead of controlling the allergy.

Flooding In The Binghamton Area

The flood history and its solution in the Binghamton area is similar to that of other watersheds. Following heavy flooding in 1935-36, the City of Binghamton promoted the sale of $200,000 in flood-control bonds. Money from this fund was used to construct flood walls on both sides of the Chenango River and along the north bank of the Susquehanna in Binghamton and Johnson City. This was complemented by the Corps of Engineers' construction of a major flood-control project in 1943, building levees, flood walls, and various channel improvements (especially near Conklin and Kirkwood) in the immediate area of Binghamton on the Susquehanna and Chenango Rivers. Work was later extended to Vestal, Westover, Endicott, and West Endicott. Total federal costs of these projects exceeded 13 million dollars (Table 3). The local costs amounted to over 1 million dollars.

Table 3

Costs of Flood Protection, Study Areas (from Tkach, 1975)

River	Structural Cost
Brixius Creek[a]	$ 322,000+
Choconut Creek[b]	194,000*
	250,000**
Fuller Hollow[ab]	60,000*
Little Choconut[a]	678,730+
	84,000*
Willow Run[b]	144,000*
Susquehanna-Chenango[c]	11,381,228[c]

[a] plus unknown additional amount for channelization

[b] plus Corps of Engineers' diking near the mouth

[c] protection of Binghamton, Endicott, Johnson City by the Corps of Engineers. Includes flood walls, dikes, levees.

* cost of channelization

** diking

\+ dams and flood-retarding structures

Upstream controls by the Corps of Engineers consists of two major reservoirs; Whitney Point Dam on the Otselic River (upper Chenango basin) and East Sidney Dam on Ouleout Creek (upper Susquehanna watershed). The Whitney Point Dam, completed in 1942 at a cost of 5.5 million dollars, controls 255 mi^2 of drainage, and the East Sidney Dam, controlling 102 mi^2 and completed in 1950, cost over 6 million dollars (Susquehanna River Basin Study 1970). These dams reduce flood heights on the Chenango River and Susquehanna River through the Binghamton area.

Since these projects, urbanization has continued to increase in the Triple-Cities area causing or aggravating drainage problems in major and minor tributaries. After the floods of 1960 Broome County received government approval for the largest single flood-control project in the United States (PL 566). The project is a comprehensive plan for nine watersheds and includes dams, channelization, and other channel "improvements" at a federal cost of 6 million dollars and a local cost of over $750,000.

Table 4

Average Annual Flood Damages, Susquehanna River, Binghamton
($1000 at March, 1974, Price Level)

Locality	Normal Existing Conditions	Normal Growth Increment*	Economic Growth Increment**
Conklin-Kirkwood	136.80	10.79	85.12
Chenango River-above Binghamton	87.70	26.30	40.13
Binghamton-Vestal	287.73	6.69	115.50

Increases should be added to existing damage for totals.

*Damages which will occur if future flood-plain development is controlled.

**Damages associated with improvements and contents within the flood plain.

Data from Table III-5, Eastern Susquehanna River Basin Board, 1975.

The desperate need for an overall solution to the growing drainage problems of the Triple Cities region was shown by the effects of two recent storms, Agnes in 1972 and Eloise in 1975. Although the upper Susquehanna River basin was treated lightly by Agnes, damage in local watersheds amounted to 1.25 billion dollars. Damage by Eloise amounted to 1.5 million dollars (Vincent Vaccaro, personal communication). Therefore, despite the fact that a great deal of money has already been spent in protecting the Binghamton area from flood damages, the hazard grows (Table 4). Moreover, the likelihood exists that an extremely rare storm might overtop or break through the flood walls and levees in the Binghamton area, as happened at Wilkes Barre during Agnes. Damage and loss of life could be staggering, since the flood-protection structures have provided a false sense of security for increased flood-plain development. The normal growth increment is damage over and above existing damage which would occur under controlled development of the flood plain. Economic growth increment is the increased amount of damage with improvements and expansion of present flood-plain development.

FULLER HOLLOW CREEK

Fuller Hollow Creek is located on the south side of the Susquehanna River in the Town of Vestal, west of Binghamton (Fig. 2). The creek has its head on the north-facing slopes of Ingraham and Bunn Hills. Below Fuller Hollow Road the stream flows through a broad, wooded flat area with steep sides and into a city park where the channel is on bedrock. The flat above the park is an effective storage area for excessive runoff from above. However, the bedrock is not far below the surface as evidenced by the outcropping in the stream bed at the park. Once the water reaches the bedrock section where impermeable shale underlies the flood plain as well as the channel, water drains out and into the stream, increasing the discharge. Below the park the stream has been straightened, shortening its length by 200 feet. The creek has been channelized where it flows through the S.U.N.Y. campus athletic field and below to its mouth. Total drainage area is 3.8 square miles.

The State University lies within this watershed and is a cause of minor development of the nearby lower part of the basin, below the park. Urbanization has crept up the valley, and since 1970 the area at the head, above Fuller Hollow Road, began to be developed. Now almost the entire upper hillside, once densely forested, has been devegetated, bulldozed, and covered with a 300-home subdivision. The surface has been modified and tributaries and streets sewered to drain storm waters directly and quickly into Fuller Hollow Creek.

A typical hydrograph of stream flow below the subdivision is shown in Figure 8. Urbanization has not only increased storm runoff, but the augmented flow also rushes down the straightened section below the park with great vigor, eroding backyards and deepening the channel. The debris is carried off and deposited downstream. Two major sites of deposition of the debris eroded from upstream are a cemented channel below the Route 434 bridge and the mouth where Fuller Hollow Creek enters the Susquehanna River.

The delta deposited in the Susquehanna by Fuller Hollow Creek was mapped during the summer of 1975 when the flow was low. Much sediment had been carried down since Agnes, creating a sizeable mass of deposits. It is assumed that much of this debris was a result of housing construction in the subdivision and erosion of upstream bed and banks during high-runoff periods. Peak discharge during Agnes swept away the numerous deposits at the mouths of tributaries in the area. The delta was mapped again in October 1975, after Eloise. Peak discharge of Eloise on Fuller Hollow Creek, calculated from floodmarks after the storm and

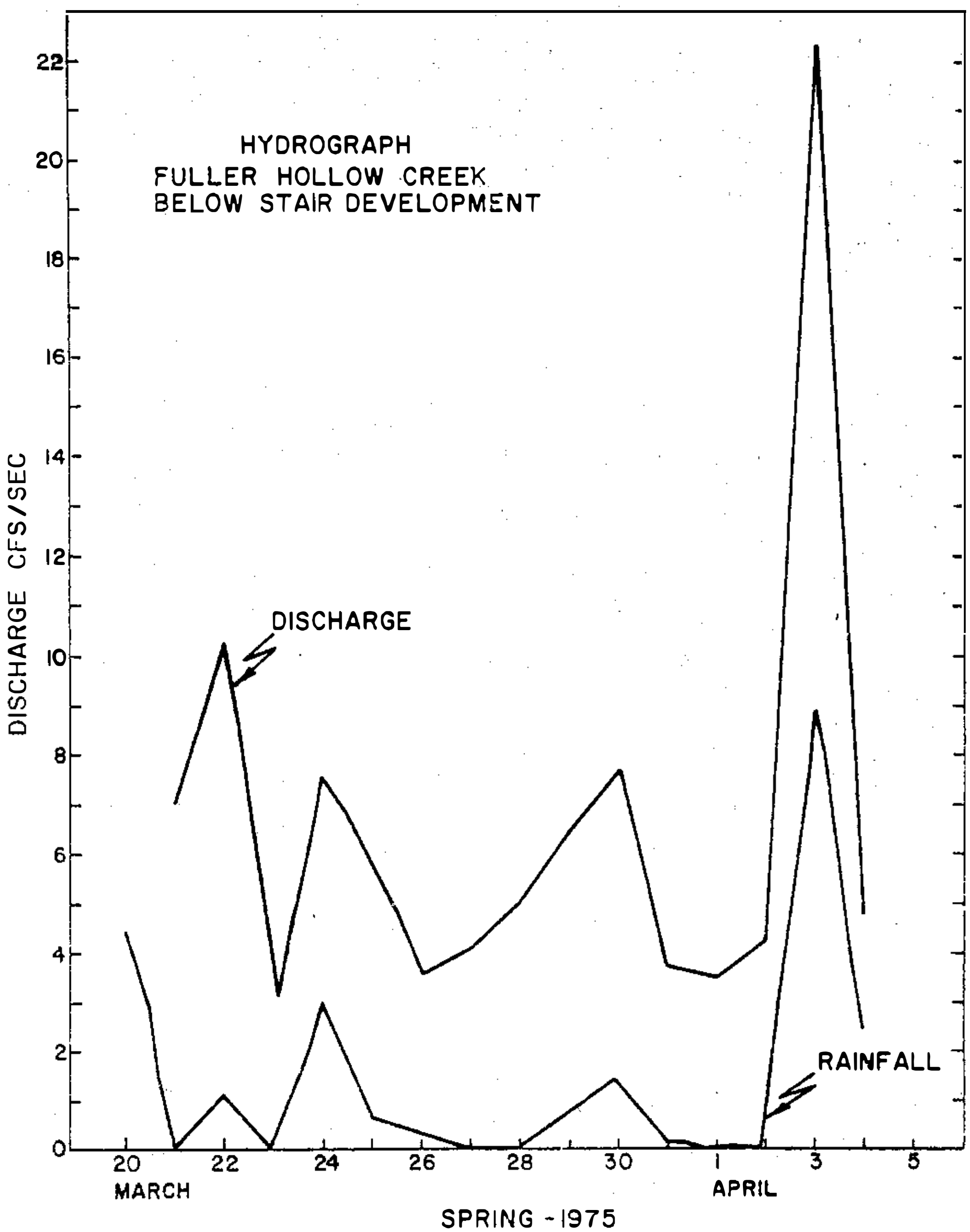

Figure 8. Hydrograph of discharge in Fuller Hollow Creek and rainfall for the same dates. From Morisawa and Vemuri (1975).

observations during it, was 550 cfs. Despite the high water on the Susquehanna during Eloise, the delta grew in size from 1570 sq. ft. before Eloise to 2051 sq. ft. after it. This is a large delta for such a small creek. The channel-bar deposits and the delta indicate that excessive erosion is taking place in Fuller Hollow Creek as a result of urbanization.

On such small creeks, developers should be required to provide storage for runoff during storms rather than sewer the rainfall excess directly into nearby stream channels. There is a large natural basin at the head of Fuller Hollow Creek in the valley south of Fuller Hollow Road where storm water could be detained. An alternative would be to drain the excess rainfall underground. Straightening the channel below the park to provide for development there was a mistake which should be avoided in the future. Such straightening increases the velocity of the water, adding to the energy which is used for erosion. Development should not be allowed on the west bank of the creek north of Fuller Hollow Road and should be barred from the flat between Fuller Hollow Road and Stair Park.

FLOODING IN CONKLIN, NEW YORK

The low-lying areas of Conklin, along the Susquehanna River (Fig. 2), have been flooded frequently, as a result of both spring rains combined with snowmelt and of winter ice jams on the river.

Table 5 shows the most severe floods that Conklin has experienced. Although the 1936 flood did not constitute a 100-year flood, it was

Table 5

Past Floods in Conklin

Date	Discharge, cfs	Flood Elevation, ft.
3/18/36	61,600	861.09
3/22/48	60,500	860.78
4/1/40	51,800	860.08
3/28/13	51,400	859.25
3/10/64	50,200	859.21
3/7/79		858.21

Gage Height = 840.95 ft.

100-year flood = 64,000 cfs.

Flood stage = 11 feet (elev. 851.95)

Sources: Dunn (1970), John May (pers. comm., Jan., 1980), U.S. Army Corps of Engineers (1971).

devastating. Specifically, the flood level was reached in twenty-four hours, and the river was out of its banks for five days (U.S. Army Corps of Engineers, 1971).

This area is also characterized by more minor, localized flooding. As an indicator of this, the Susquehanna River overtopped its banks sixty-five times in Conklin during the 30-year period between 1935 and 1964. The most recent flood occurred on February 11, 1981, as a result of ice jams. This flood was 6 feet above flood stage.

Development in the floodplain in Conklin is relatively recent (mostly within the last 25 years), and therefore the history of flooding is well known to local residents. There are currently no structural measures in effect to protect Conklin, although a channel improvement project, consisting of seven miles on the Susquehanna River, was undertaken to provide relief in the event of smaller floods (U.S. Army Corps of Engineers, 1969). However, directly following the February, 1981, flood; attempts began (and still continue) to persuade the Corps of Engineers to construct a floodwall in Conklin. To date, the Corps has not agreed because of a low benefit-cost ratio, and because of probable adverse effects on downstream communities.

MASS MOVEMENTS

In a study of landslides in the Binghamton region, Ott (1979) identified 83 known slides and inferred an additional 55 using air photos and field checks (Table 6). From a frequency of occurrence, he rated soils as to susceptibility of sliding. Volusia, Mardin, Canaseraga, and Unadilla soils (S.C.S., 1971) were most susceptible to mass movements. He also found that north-facing slopes were more susceptible to failure. Soil characteristics contributing to slope instability were seasonally high water table coupled with slow permeability and a dense fragipan.

Two areas we will examine have failed primarily because they are slopes cut to a steep angle. Both are underlain by Canaseraga soils. Canaseraga soils are slowly permeable, have a seasonal high water table, often have local seeps, have a high available moisture capacity, and are susceptible to differential frost heave. Cut slopes are unstable and the soil surface is easily erodible.

The north face of Pierce Hill was cut into for road material and oversteepened during construction of Route 434 in the late 1960s. It has since been cut back even more for development. Since that time the slope has failed in a number of places. The Red Lobster and Howard's Florist have both gone to great expense in attempts at stabilization.

The slope on the east side of the Vestal Plaza was cut into to provide as eastern access to the Plaza. Failures occurred very shortly on the north-facing slope. Mass movements on the south-facing side have taken place over the last 3 years. The town of Vestal has to bulldoze the material from the road regularly, especially in the spring.

TABLE 6

Number of Landslides Per Quadrangle

Quadrangle	No. of Known Landslides	No. of Inferred or Questionable Landslides	Total No. of Landslides
Endicott	29	37	66
Binghamton West	33	10	43
Binghamton East	21	8	29
Totals	83	55	138

From Ott, 1975

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ROAD LOG FOR ENVIRONMENTAL PROBLEMS OF THE BINGHAMTON AREA

CUMULATIVE MILEAGE	MILES FROM LAST POINT	ROUTE DESCRIPTION
0.0		Bartle Drive Main Entrance to SUNY. Turn left (west) on Route 434.
3.5	3.5	Make a U-turn just past Red Lobster onto Route 434.
3.75	0.25	STOP 1 in parking lot of Gertrude Hawk Candies (Route 434).

STOP 1. PIERCE HILL CUT. This is a slope with active movement of material downslope. Originally a borrow pit, the slope has been cut back even more for the commercial development you see. Debris slides, slumps, rilling and rock fall are modes of downslope movement of the glacial materials. Both Howard's Florist and the Red Lobster have gone to considerable expense to stabilize the slope. One debris flow behind the Red Lobster reached the back door, covering several cars in the way. Subsequently, the wall and drainage pipes were installed.

7.4	3.65	Turn right (South) onto Murray Hill Rd. just east of SUNY campus.
7.5	0.1	STOP 2 along Murray Hill Rd. opposite East Gym.

STOP 2. LOWER FULLER HOLLOW CREEK. Here the creek has been riprapped to prevent erosion of the bed and further down the banks are riprapped. The riprap has progressively deteriorated, large blocks have removed and side-walls have slumped.

8.6	1.1	STOP 3 along Murray Hill Rd. at Stair Park.

STOP 3. MIDDLE FULLER HOLLOW CREEK. Evidence of destructive erosion can be seen here. The foot bridge was washed out in the spring of 1981. Note widening of the channel. Excessive runoff from storm drains have caused much erosion here. Along the downstream reach many landowners are losing their back yards.

8.9	0.3	Turn right (west) onto Fuller Hollow Rd.
10.4	1.5	Turn around in driveway on right. Proceed east on Fuller Hollow Rd.
10.8	0.4	STOP 4. Martin House on Fuller Hollow Road.

STOP 4. UPPER FULLER HOLLOW CREEK. From this viewpoint, one can see the 300 home Stair development. The whole hillside was wooded until approximately 1974-75. Tributaries and storm runoff are piped underground directly to the creek, greatly augmenting flow during storms. Directly below is a meadow through which the main creek flows. This would have been an ideal spot for a detention pond to which runoff could have been piped.

11.8	1.0	Turn left (north) onto Murray Hill Rd. to end.
13.35	1.55	Turn right (east) onto Route 434.
13.8	0.45	Turn right into Vestal Plaza and proceed to southeast corner behind the Grand Union.
14.2	0.4	STOP 4 in southeast corner of Vestal Plaza behind Grand Union.

STOP 5. VESTAL PLAZA SLOPE. This cut has failed in many places since it was made. The north-facing slope has moved much more and did so more quickly than the south-facing slope. The cut is in glacial material. Much of the fine sediment has been removed by mud flows which cover the road whenever it rains. Buildings and parking lots on the surface above the slopes have contributed to mass movement.

		Go out southeast entrance of Vestal Plaza and turn left at Club House Rd. (top of hill).
14.4	0.2	Turn right (east) onto Route 434.
16.5	2.1	Conklin Avenue east off Route 434. Turn left onto Tremont and then right onto Conklin Ave.
16.9	0.4	STOP 5 in Crowley's Parking Lot.

STOP 6. ROCKBOTTOM DAM. This dam is currently being rebuilt after years of deterioration on the older one. The dam is designed to retard the flow of water and to produce ponding so that the water intake for the City of Binghamton's water supply is below the surface even in dry years.

17.8	0.9	Cross Pierce Creek on Conklin Ave. Channelization is evident.
21.5	3.7	STOP 6. Tier gasoline station.

STOP 6. SUSQUEHANNA RIVER. The bend in the river at this point led to flooding of the area between the river and the gas station during the ice jams in February.

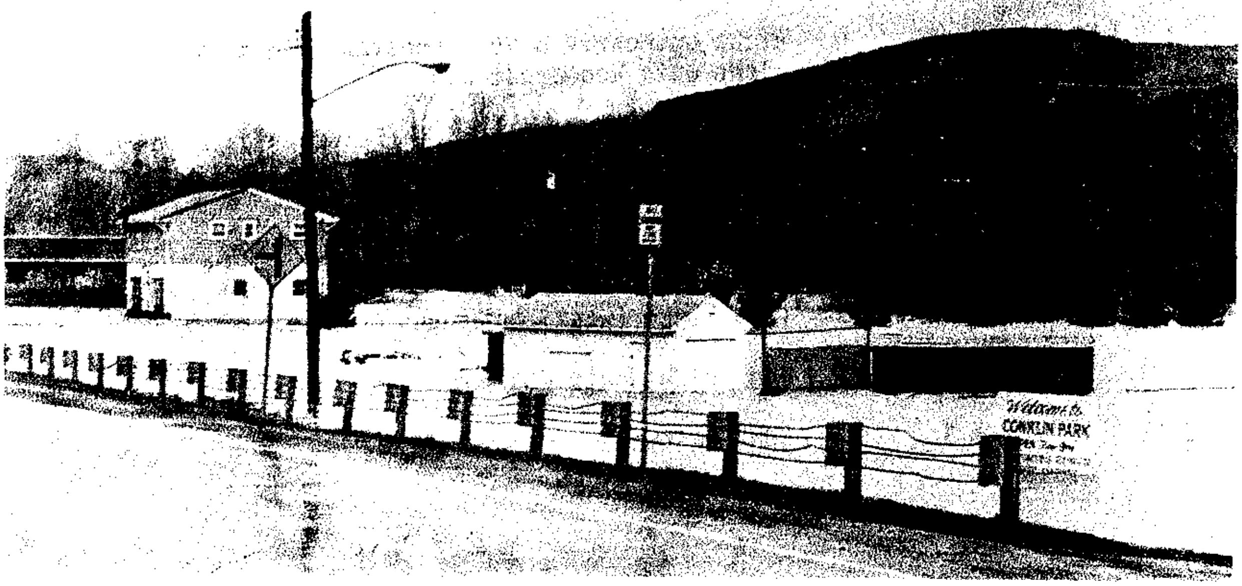
23.3	1.8	STOP 7 Conklin Park then return west on Conklin Avenue.

STOP 7. CONKLIN PARK. This park was flooded entirely during the February 1981 flooding. Water levels reached up to the park sign. Although there are some buildings in the park, they are for storage, primarily. This park is a good example of how flood plain areas should be developed.

24.7	1.4	Right on Morris Blvd. Continue and curve to right onto Wooderest Way.

Note the houses in this area (which was also flooded in February). Each has a nice view of the river and gets flooded almost annually. These are obvious examples of uneconomic floodplain development.

25.5	0.8	Turn right on Inamure.
25.7	0.2	Turn right (west) onto Conklin Ave.
31.7	6	Bear right onto 434 west.
34.7	3	SUNY entrance. Bartle Drive.



Flood at Conklin Park (Stop 7) caused by an ice jam, February, 1981.

Bridge at Stair Park (Stop 3) before it was washed away, Spring, 1981.

DEPOSITIONAL ENVIRONMENT OF THE ONEONTA FORMATION (CATSKILL FACIES), NEAR UNADILLA, NEW YORK

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INTRODUCTION

A large roadside exposure on Interstate 88 near Unadilla (Fig. 1) offers a unique opportunity for detailed study of both vertical and lateral lithofacies variations in the Upper Devonian Catskill facies. The rocks are assignable to the Oneonta Formation (Fletcher, 1963; Rickard, 1975), and the postulated early-middle Frasnian age (*Polygnathus asymmetricus* conodont zone) of this formation is supported independently by the miospore studies of J. B. Richardson (pers. comm.). This paper presents a detailed description and interpretation of the lithofacies, and a brief comparison with other published studies of Catskill-facies sedimentology. For convenience, the rocks were separated into two main facies associations which can usually be clearly distinguished in the field.

THE SEDIMENTARY LITHOFACIES

Facies Association 1

This association comprises mainly fine- to very fine-grained, moderately to well-sorted sandstones, medium-dark to medium-light gray (N4-N6) in color (Goddard *et al.*, 1970), and muddy sublitharenite (McBride, 1963) in composition. Gravel-sized intraformational sandstone, siltstone, and mudstone fragments are common, locally concentrated as breccias above erosion surfaces. These breccias also contain plant remains, fish fragments, and reworked calcareous concretions. In color, they range from dark yellow orange (10YR6/6) to medium-dark gray (N3), depending on the type of fragments present. Discontinuous layers of medium-dark to olive gray (5Y4/1) siltstone and mudstone also occur locally.

The dominant internal structures are small- and large-scale trough cross-stratification, and horizontal stratification. Small-scale trough cross-stratification is usually the climbing type (Type A of Jopling and Walker, 1968); set thickness is about 1 to 3 cm, and trough width, 3 to 6 cm. Thickness of the large-scale cross-bed sets lies in the range 5 to 30 cm and trough widths are 10 cm to 3m. Horizontal stratification (which may actually have a slight inclination) is comprised of mm-scale laminae and is typically associated with parting lineation. Commonly interbedded with horizontal stratification are large-scale nearly planar cross-bed sets.

The spatial organization of texture and internal structure within the four exposed sandstone bodies of facies association 1 is complex, but systematic. Major erosion surfaces with lateral extent on the order of hundreds of meters underlie each sandstone body, but may also occur within a sandstone body, imparting a 'multistory' character (Fig. 2). Immediately overlying beds are usually intraformational breccias, with clasts up to decimeters across set in a fine to very fine sandstone matrix.

Within each sandstone body it is usually possible to recognize bedsets (sedimentation units) up to 1 meter thick, which can be traced laterally for many tens of meters (Fig. 3). Commonly, the base of a bedset is a relatively minor erosion surface with intraformational breccia, variably developed along the bedset. For instance, the breccias lower in a sandstone body are typically better developed and have larger clasts (up to 40cm). The top of each bedset may fine locally to siltstone or mudstone; sandstone bedsets overlying the thicker occurrences of fines (Fig. 3) may display load-casted as well as eroded bases. Sedimentary structures may also vary vertically within a bedset, for example horizontal stratification may be overlain by small-scale cross-stratification and finally capped by ripple marks (Fig. 3). Figure 3 further shows that the thickness, texture, and internal structure of any bedset may vary laterally as well as vertically.

Bedsets may be inclined up to about 10^{o} with respect to the base of a sandstone body (i.e. epsilon cross-stratification, Allen, 1963) or occur as concave upwards channel-fills, evident in sections transverse to paleocurrents (Fig. 2, body 1 and 2). Where paleocurrents essentially parallel the outcrop, bedsets are broadly parallel to the basal erosion surface (e.g. body 3). As explained below the orientation of the bedsets may vary within <u>and</u> between the different stories of a sandstone body.

Within each story of a sandstone body is large-scale vertical and lateral variation in bedset orientation, thickness, texture, and internal structure (Figs. 2 and 3). In some instances, there is an overall fining-upwards tendency associated with a systematic variation in internal structure; however, this is not ubiquitous. The uppermost story of sandstone body 1 (Figs. 2 and 3) shows epsilon cross-stratification becoming steeper to the northeast, and finally changing to a channel fill; there is an associated change of vertical facies sequence. The immediately adjacent channel fill is also complex, with more fines in the base of the channel than higher up.

Paleocurrents are consistently unidirectional for a given story, and the range of values is less than 30^{o} (Figs. 2 and 3). Mean directions <u>between</u> different stories may be distinctly different, thereby assisting in the recognition of separate stories in each sandstone body (c.f. Puigdefabregas and Van Vliet, 1978).

In the top meter of each sandstone body (and in the overlying facies association 2) are abundant siltstone casts of <u>in situ</u> and transported plant roots and stems, about 1 to 3 cm across, and up to 30 cm long (Fig. 5 E, F).

<u>Facies Association 2</u>

Facies association 2 comprises complex interbedding of very fine grained, moderately to poorly sorted sandstones, siltstones, and mudstones.

The color of sandstones varies from medium gray (N5) to grayish red (10R4/2), with siltstones and mudstones ranging from medium-dark gray (N4) to grayish red (10R4/2).

The beds are arranged in sets, centimeters to decimeters thick; each fines upwards. Bases of the bedsets are erosional with relief of up to 5 cm and common intraformational rock fragments up to 10 cm across. The fragments commonly match immediately underlying beds. Flow-parallel furrows (about 5 cm deep and 10 cm across) and load casts are locally associated with erosional bases.

Typical vertical variation in texture and sedimentary structure of facies association 2 is summarized in Figure 4. Large-scale cross-stratification is most commonly of the trough-type, set thicknesses 5 to 20 cm thick, and troughs 10 to 100 cm across. Small-scale cross-stratification (sets 0.5 to 3 cm thick) may be trough (widths 1 to 5 cm) or planar type; troughs are most common. Climbing types correspond to type A of Jopling and Walker (1968). Asymmetrical and symmetrical ripple marks up to a centimeter high and a few centimeters long are common on bedding surfaces (Fig. 5G). Desiccation cracks are ubiquitous, and rare raindrop imprints can be seen (Fig. 5 C,D,G).

Large vertical (and partly horizontal) burrows with roughly circular sections up to 10 cm across occur mainly in siltstones and mudstones. They are commonly filled with relatively coarse sediment and mud chips. Two varieties of surface trails (as yet unidentified) have also been found (Fig. 5 A,B).

Drifted and in situ plant remains, siltstone casts of stems and rootlets, are common in this facies association. The casts and immediately surrounding rock may display a local color change to greenish gray (5G6/1). Concretions, moderate to dark yellowish brown in color (5YR3/4 to 10YR4/2), occur in very fine sandstones and siltstones and are commonly associated with large branching rootlet traces (Fig. 5F). They are irregularly globular, up to 3 cm across, and are similar to type A of Allen (1974).

The bedsets may occur as laterally extensive sheets (tens or hundreds of meters) or filling broad channels (Fig. 2). The channels are typically asymmetrical in section, 3 to 30m across and 0.3 to 1m in maximum depth. Thickness, texture and internal structures of the bedsets change laterally in both sheets and channels (Fig. 3). It appears that the coarser grained representatives of this association occur directly on top of facies association 1 (Fig. 3), and in sandstone body 2 there is lateral transition into association 1. Also there are two examples of small, isolated coarse-grained channel fills, one of which has an asymmetrical section and adjacent epsilon cross-stratified sandstone (Figs. 2 and 3).

Paleocurrents from large-scale cross-stratification, parting lineation, and channels indicate unidirectional flow, subparallel to those from facies association 1 (Fig. 3). Directional variation between bedsets of up to 30° is common, but small-scale cross-stratification and asymmetrical ripple marks are more variable. Crestlines of symmetrical ripples marks show no systematic orientation (Fig. 3).

INTERPRETATION OF LITHOFACIES

Facies Association 1

The sedimentary characteristics of facies association 1 strongly suggest deposition in river channels migrating laterally across alluvial plains. The major laterally extensive erosion surfaces are ascribed to lateral migration of erosion-dominated areas deep in river channels. It is well known that slumped fragments of alluvial-plain sediment from adjacent retreating cut-banks accumulate as lag gravels in these channel deeps. Accordingly, the overlying sandstones and breccias represent the deposits of laterally migrating channel bars or coarse-grained channel fills.

The large-scale vertical sequence of texture and internal structure in each sandstone story has parallels in some modern single-channel streams with sinuous talwegs (e.g. Harms et al., 1963; Davies, 1966; Sarkar and Basumallick, 1968; Bernard et al., 1970; Shelton and Noble, 1974). Large- and small-scale trough cross-stratification record the downstream movement of three-dimensional dunes and ripples, whereas horizontal stratification was deposited on upper-stage plane beds. Climbing ripples and upper-stage plane beds generally indicate the presence of significant suspended load as well as bedload transport within the channels. The vertical juxtaposition of bed configurations and textures reflects their original spatial distribution on the laterally migrating inclined bar surface, in response to locally variable velocity, depth, and slope. It is worth remembering that the facies sequence predicted by the well-known fining-upwards model (e.g. Allen, 1970) will only be present in the downstream part of channel bars (Jackson, 1976; Bridge, 1978).

Perhaps the strongest support for lateral deposition on a channel bar comes from epsilon cross-stratification, where the low-angle stratification surfaces dipping approximately normal to local paleocurrent direction represent ancient bar surfaces. Small-scale vertical facies variation within the individual bedsets that defines the epsilon cross-stratification can also be seen in modern channel-bar deposits (see references above). The upward decrease in grain size and change in internal structure in each bedset probably records deposition during falling flow stages. Desicatted low-flow deposits may be incorporated as intraformational fragments in the deposits of an ensuing flood. The restriction of the in-situ plants to topographically high areas of the bars, and the relative lack of finest sediment grades within the lower parts, suggests that the rivers were perennial.

The channel fills with asymmetrical cross-sections have well-exposed cut banks, supporting an interpretation as sinuous, laterally migrating channels. The sediment in the fills is relatively coarse, and the spatial facies variations within the fills are complex (see later discussion).

The multistory character of the sandstone bodies is expected when lateral bar migration is combined with net floodplain aggradation (Bluck, 1971; Bridge, 1975; Bridge and Leeder, 1979). Each story is a single bar deposit which has been overriden and eroded by a different channel segment and associated bar. The unidirectional paleocurrents in a single story are consistent with channel deposition; however, if the migrating channels had any curvature, it is reasonable to expect different mean paleocurrents from superimposed stories (Figs. 2 and 3).

Sandstone body 1 (Figs. 2 and 3) records evidence of particularly complex channel behaviour. Channel fill 1A is probably associated with the immediately adjacent bar deposit, by virtue of a common basal erosion surface and similar paleocurrent direction; however there is evidence of subsequent erosion of the bar deposits low in the channel. These bar deposits form the cut bank of channel 1B, thereby predating it. Both channels may have existed simultaneously for some time, however, being separated by the earlier bar deposits of channel 1A. The steepening of the epsilon cross-bed sets and change in facies as channel 1B migrated laterally is a result of change in hydraulics and channel geometry. This could result from either (a) observation of the same bar at different longitudinal positions in a given river bend (e.g. Bridge, 1978), or (b) changes in plan form and hydraulics during channel migration, such as increase in sinuosity and decrease in mean slope and velocity. The laterally extensive erosion surface (overlain by sandstone) that truncates the top of the epsilon sets appears to be associated with the last of the coarse-grained fill of channel 1B. Finer grained beds overlying this fill are laterally equivalent to beds only a few meters from the base of channel fill 1A, implying that channel 1A was at least partly open once 1B had filled up. In fact, Figure 2 shows evidence that a substantially shallower channel 1A was actively migrating in the final stages of filling.

A possible sequence of events to explain these complicated facies patterns is:

(a) lateral migration of channel 1A,
(b) chute cut-off and formation of channel 1B,
(c) lateral migration of channel 1B, while channel 1A was filling and partially eroding its previous bar deposits,
(d) gradual filling of channels 1A and 1B; the relatively coarse fills imply that the channels still carried an appreciable discharge, and
(e) diversion of more discharge back into channel 1A, with some renewed bank erosion and lateral deposition during the final filling stage.

Sandstone body 4 (Fig. 2) also shows lateral and vertical transition to a major channel fill, the exposed top parts of which are filled with bedsets of facies association 2.

Facies Association 2

Each bedset is interpreted as the deposit of a single flood on a floodplain. Lower bounding surfaces record some erosion of previously desiccated deposits prior to deposition. Large- and small-scale cross-stratification and horizontal stratification were produced by bed-load deposition of sand moving as dunes, ripples, or an upper-stage plane bed. However, textures and internal structures indicate substantial transfer of suspended sediment to the bed during deposition. The vertical sequence in a bedset records waning overbank flood velocities, with the siltstone and mudstones representing purely suspended-load deposition of the final flood stages. Wind action on ponded areas is recorded by wave ripples, and the raindrop impressions and abundant desiccation cracks indicate subsequent exposure of the floodplain surface. Concretions represent the calcareous soil horizons common beneath floodplains in arid and semi-arid climates (see Allen, 1974, for summary).

Thicknesses of these postulated flood deposits are consistent with observations from modern rivers (e.g. Bridge and Leeder, 1979). The coarser and thicker channel-filling facies are similar to modern crevasse-channel and splay deposits (e.g. Kruit, 1955; Coleman, 1969; Singh, 1972). The two isolated coarse-grained channel fills immediately below sandstone body 2 are specifically interpreted as crevasse channels that were probably open during the deposition of sandstone body 1. The more sheet-like facies closely resemble modern levee deposits (e.g. Fisk, 1947; Singh, 1972; Ray, 1976), although some of these bedsets may have occurred during the final stages of filling of major channels (Figs. 2 and 3). The locally increased dip of bedsets to the west of sandstone body 4, and their sheet-like geometry, reflects the topographic dip of a levee away from a major channel. The finest bedsets have parallels in the flood-basin deposits of modern floodplains (Jahns, 1947; Allen, 1965).

The most abundant *in situ* and drifted plant remains occur in facies that are interpreted here as upper channel bar, levee, and crevasse deposits; that is, topographically high areas immediately adjacent to major channels. It appears that during floods much plant debris was buried rapidly before the fragments could be transported into the flood-basin (see also Bridge *et al*, 1980). The restriction of the major flora to areas near channels may be due to the presence of a locally high water table near perennial streams in a semi-arid climate. The close association of many concretions with *in situ* plant roots suggests that concentration of carbonate in the groundwater is associated with loss of water to the atmosphere through plants. The color difference between red overbank and gray channel deposits is most probably due to differences in the oxygenation of the groundwater during early diagenesis.

The origin of the trace fossils cannot be ascertained conclusively due to lack of preservation of organisms responsible. The larger burrows are similar to those interpreted elsewhere as dipnoan aestivation burrows (Woodrow, 1968) or maybe even arthropod burrows (Rolfe, 1980). Arthropods were probably responsible for surface traces shown in Figure 5A and B.

Sandstone body 2 (Fig. 2) shows a lateral transition from a postulated crevasse-splay deposit to a single major channel with evidence of lateral accretion; the channel then became progressively shallower. This behaviour is consistent with the known tendency of streams to change course during major floods by diversion of discharge through crevasse channels (i.e. avulsion). In this case, the diverted channel did not develop a major channel belt, as represented by the other sandstone bodies. The spatial organization of facies associations 1 and 2 can, however, be explained by avulsion of major channels whilst net floodplain aggradation proceeds (Allen, 1965, 1974; Bridge and Leeder, 1979). Many authors suggest that the presence of appreciable fine floodplain sediment is a result of a degree of channel belt stability present only in meandering streams. Bridge and Leeder (1979) show theoretically that the relative proportion of channel and overbank deposits in an alluvial succession is unlikely to be diagnostic of channel planform. It is the detailed nature of the channel sandstone bodies that supports the single curved-channel interpretation here.

COMPARISON WITH OTHER PUBLISHED INTERPRETATIONS

The foregoing discussion agrees with the generally accepted view that the Catskill facies is alluvial in origin (e.g. Shepps, 1963; Rickard, 1975). The Oneonta Formation in this region has been more specifically interpreted as meandering-river deposits of a lowland alluvial plain (Woodrow and Fletcher, 1967; Johnson and Friedman, 1969). Although many of the sedimentary features described here have been recognized in previous detailed studies of Catskill facies (e.g. Allen and Friend, 1968; Johnson and Friedman, 1969; Allen, 1970) the observations at the Unadilla outcrop have allowed a much more refined and unambiguous interpretation than has been possible to date. Paleoclimatic implications of this study concur with Woodrow *et al.*'s (1973) reconstruction of the Upper Devonian paleogeography of this area.

ACKNOWLEDGEMENTS

We are most grateful to Paul Enos, Jim Sorauf, and Dick Beerbower for reviewing the manuscript and to Olga Kurty for drafting the diagrams.

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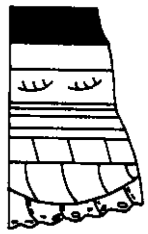
ROAD LOG

CUMULATIVE MILEAGE	MILES FROM LAST POINT	ROUTE DESCRIPTION
0.0	0.0	Trip starts at Bartle-Drive (main) exit of SUNY-Binghamton campus. Turn east on to NY 434 (Vestal Parkway) towards Binghamton.
3.0	3.0	Bear to the right off NY 434 immediately after crossing the State Street Bridge in Binghamton. At this point, follow the road signs to NY 17 and I 81, but subsequently stay in the lane for NY 7N and I 88 towards Oneonta.
9.0	6.0	Chenango Bridge; NY 7N becomes I 88. Continue on I 88.
45.0	36.0	Stop at roadside exposure on right-hand (south) side of I 88 just past Unadilla exit.

STOP. ONEONTA FORMATION, CATSKILL FACIES.

Key to Figures 2-4



Siltstone or mudstone
Indistinct bedding
Small-scale cross-stratified sandstone to silty sandstone
'Horizontal' stratified sandstone or silty sandstone
Large-scale planar cross-stratified sandstone
Large-scale trough cross-stratified sandstone
Scoured surface overlain by intra-formational breccia

Epsilon cross-stratification
Asymmetrical ripple marks
Symmetrical ripple marks
a
b
c
Mean paleocurrent azimuths (N to top of page) derived from:
a) Channels
b) Large-scale cross-stratification and parting lineation
c) Small-scale cross-stratification or asymmetrical ripple marks

Load casts and flame structures
Raindrop imprints
Desiccation cracks
Concretions
Burrows
Tracks and trails
Drifted and insitu plant remains (various sizes)
Fish bone fragments

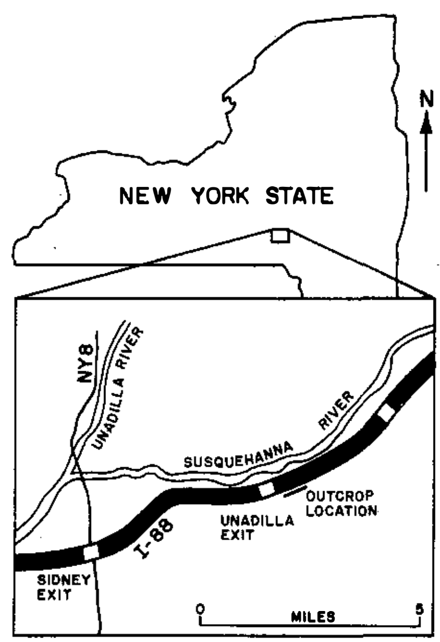
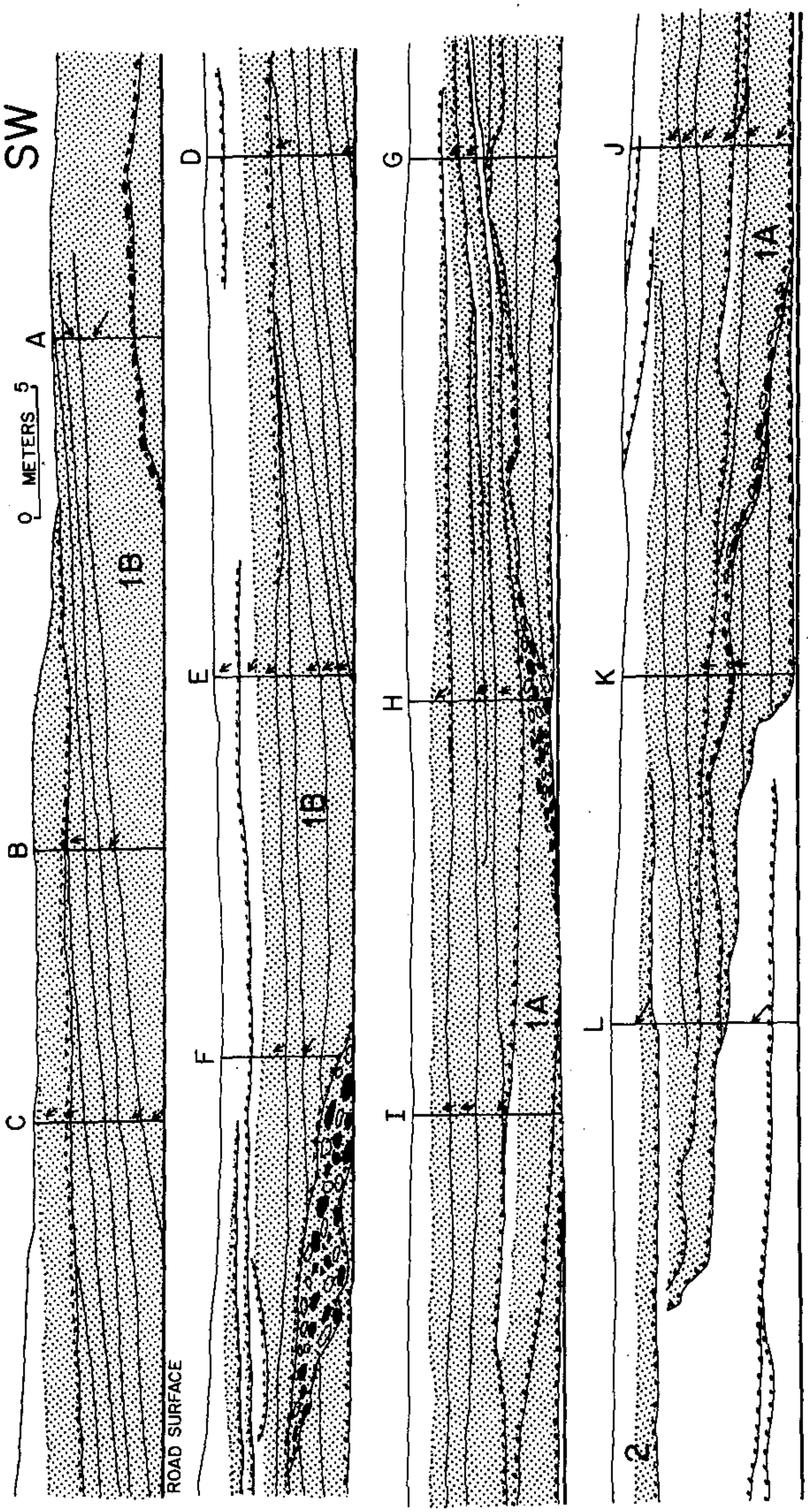


FIG. 1 - Location map of outcrop.

FIG. 2 - Scale diagram of outcrop studied, showing major bedding features and facies associations. The diagram represents a continuous section, starting at the southwestern edge of the outcrop (nearest Binghamton) and finishing at the northeastern edge. The vertical scale is not strictly correct for the upper part of the outcrop because of some distortion in the photographs from which the diagram was drawn; however true vertical thicknesses can be obtained from detailed measured sections (Fig. 3), the positions of which are marked by letters. Stippled areas are sandstone bodies of facies association 1, with some important interbedded shales unstippled. Facies association 2 is unstippled, except for parts of sand body 2 that are transitional with facies association 1. Only representative bedding surfaces are marked, the size of the overlying ornament reflecting degree of erosion and development of intraformational breccia. Paleocurrent azimuths (north to top of page) are shown; the azimuth of the outcrop is approximately N50^{o}E. Individual sandstone bodies are numbered for reference from text.


SW
0 METERS 5
A
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C
1B
ROAD SURFACE
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F
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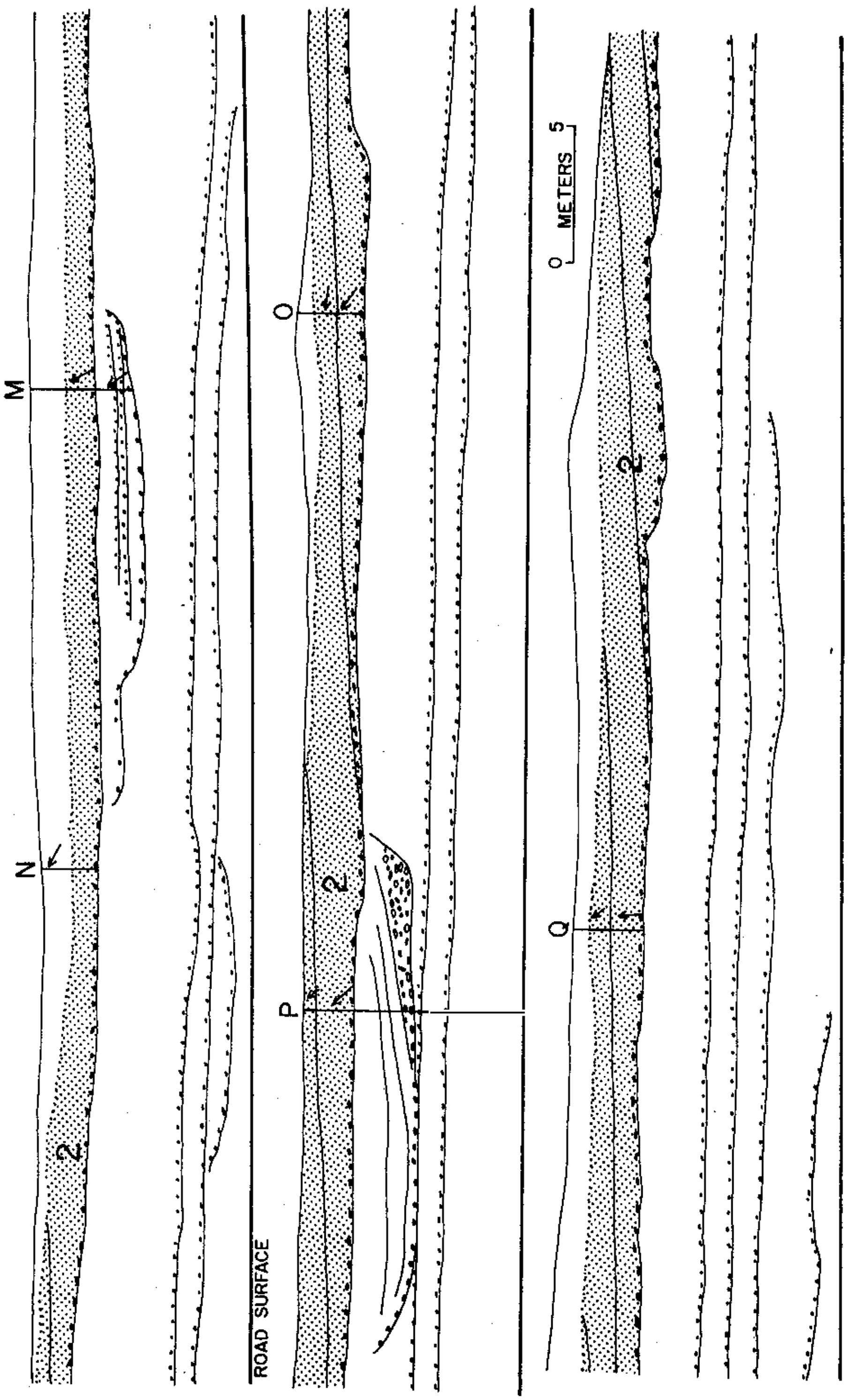
G
H
I
1A
J
K
L
1A
2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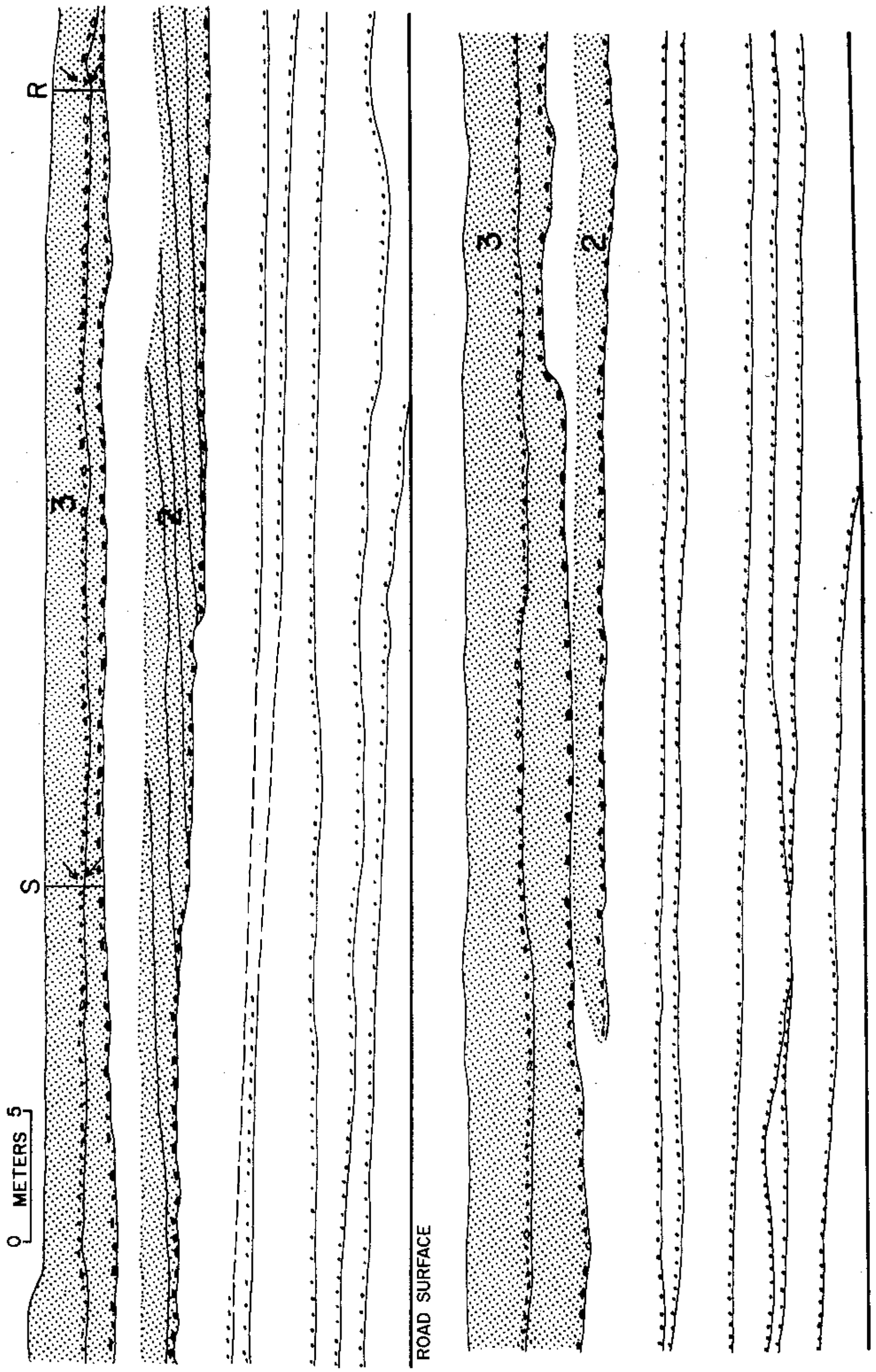
N
2
ROAD SURFACE
O
2
P
0 METERS 5
Q
2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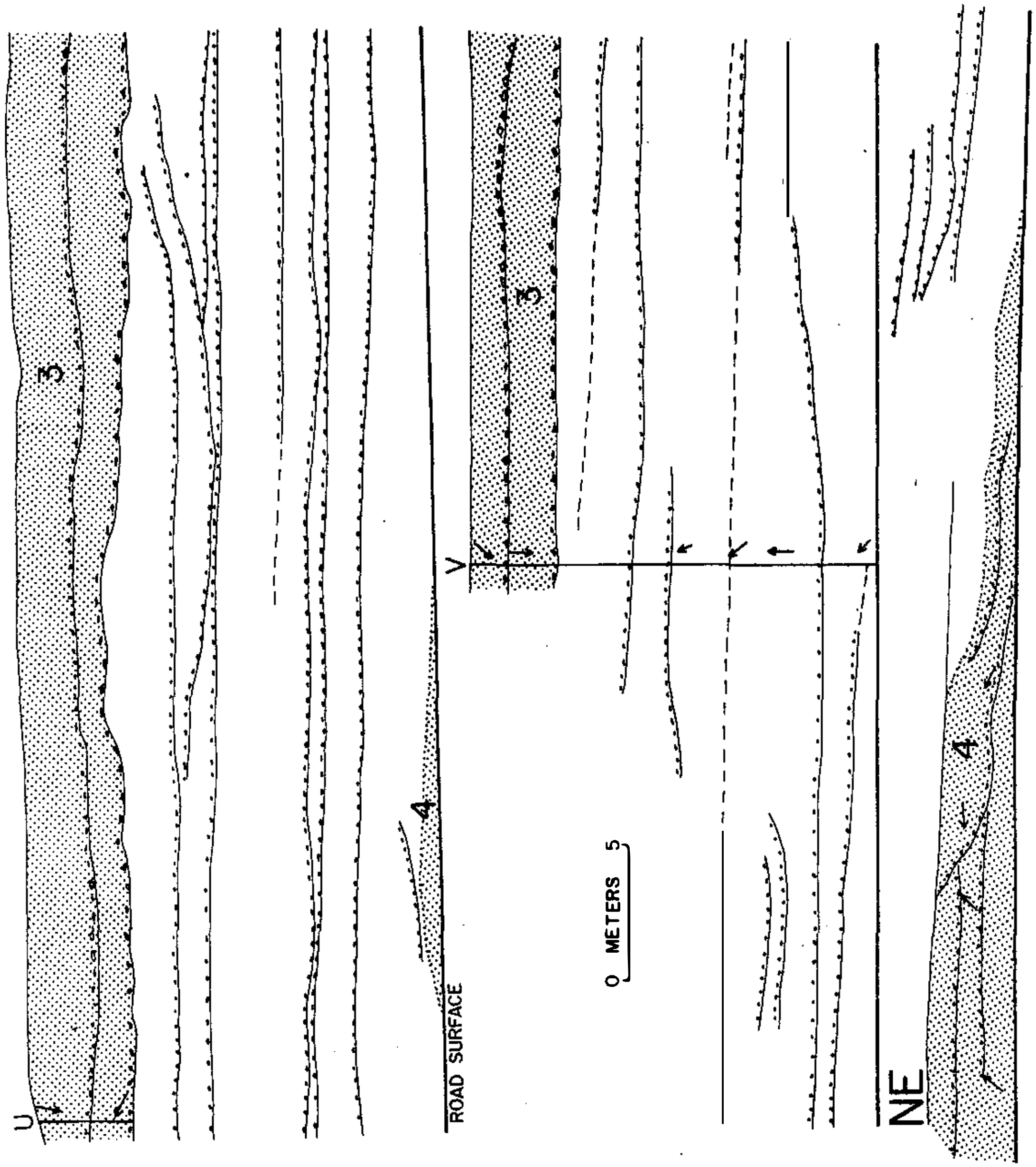
S
3
2
0 METERS 5
ROAD SURFACE
3
2


U
V
3
3
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ROAD SURFACE
0 METERS 5
NE

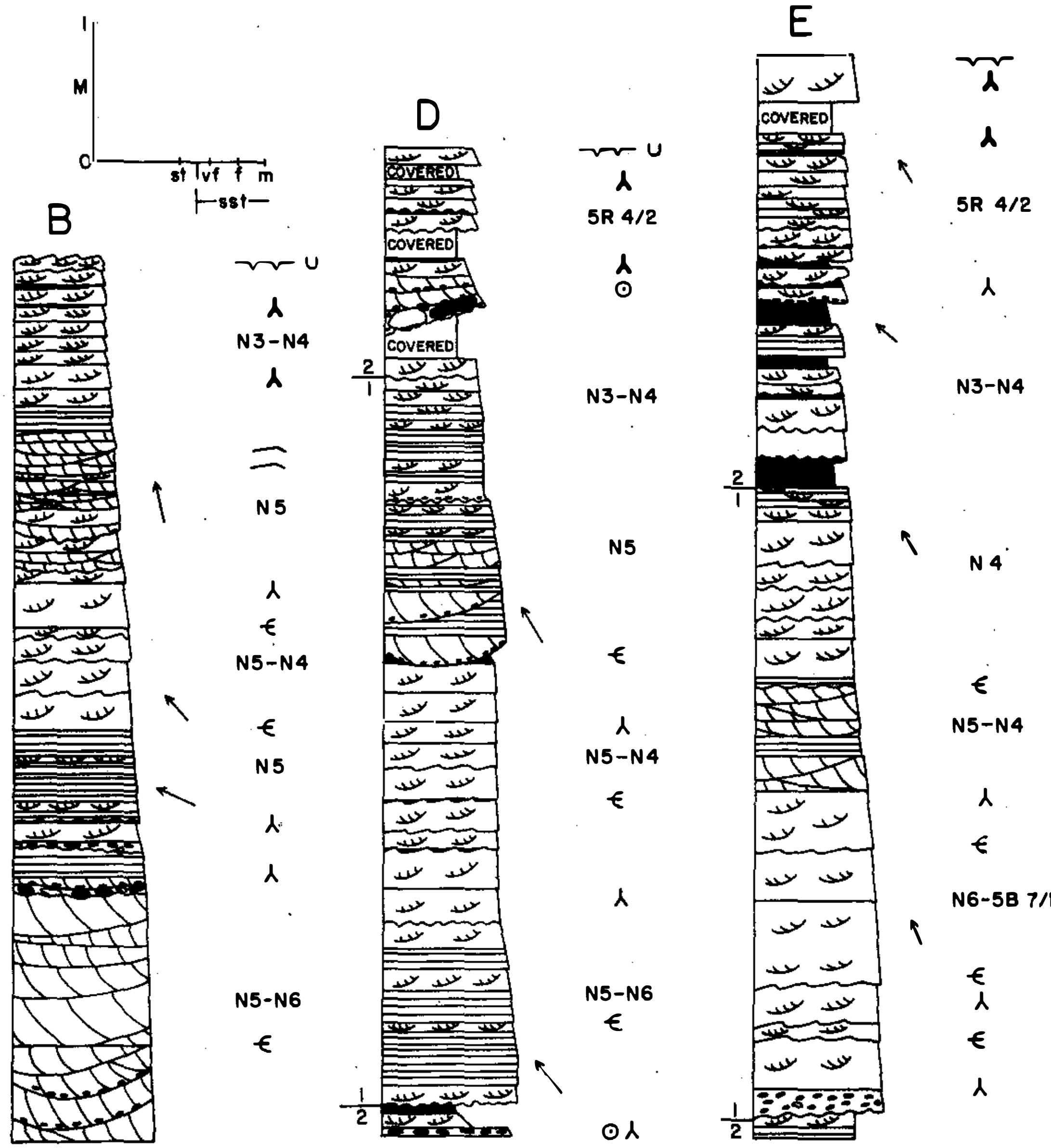
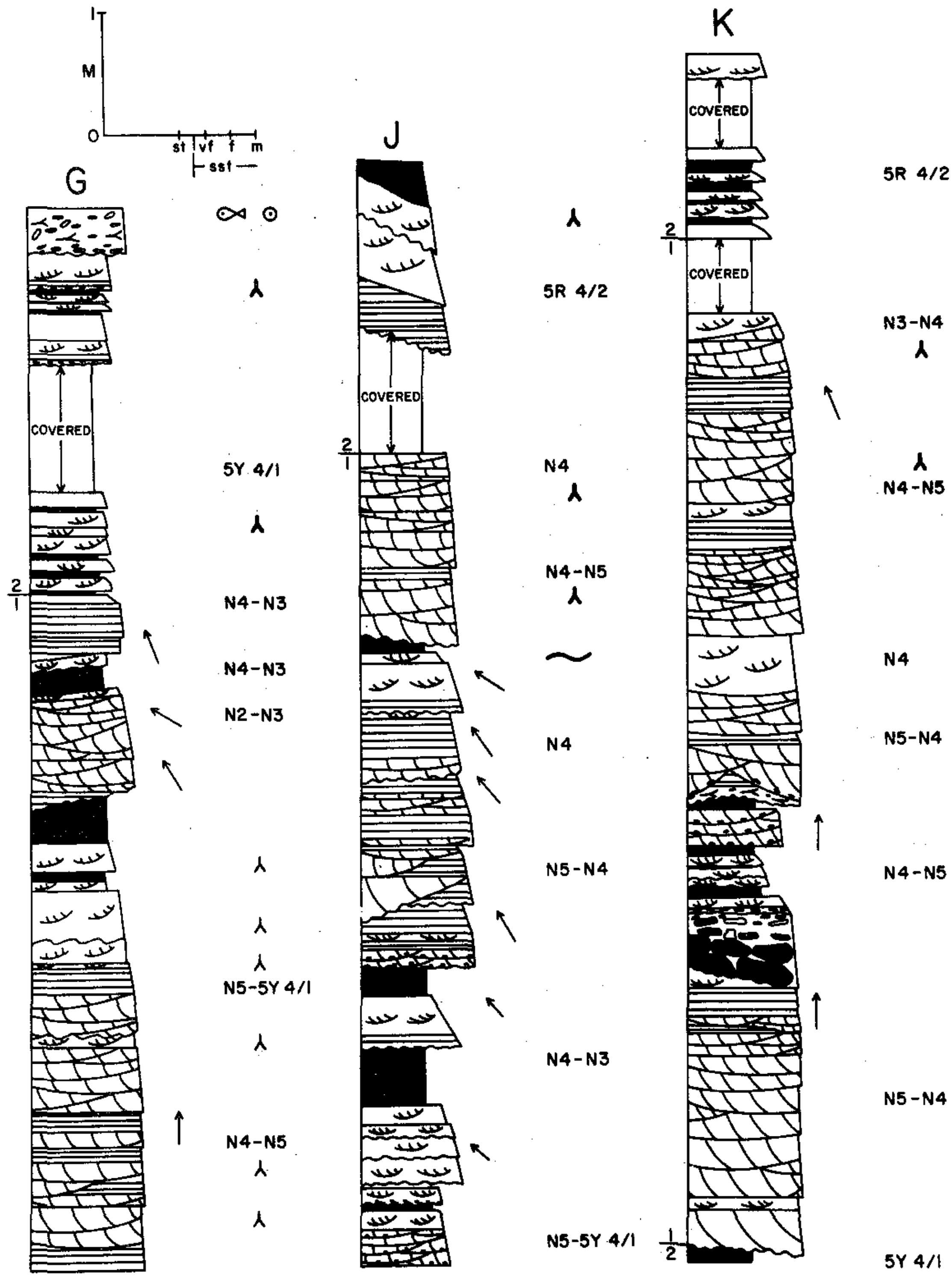


FIG. 3 - Selected detailed vertical sections, as shown on Fig. 2. See legend. Paleocurrents (north to top of page) and additional sedimentary properties are shown to right of each graphic log. Facies association given to left of each log.



M
O
st
vf
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m
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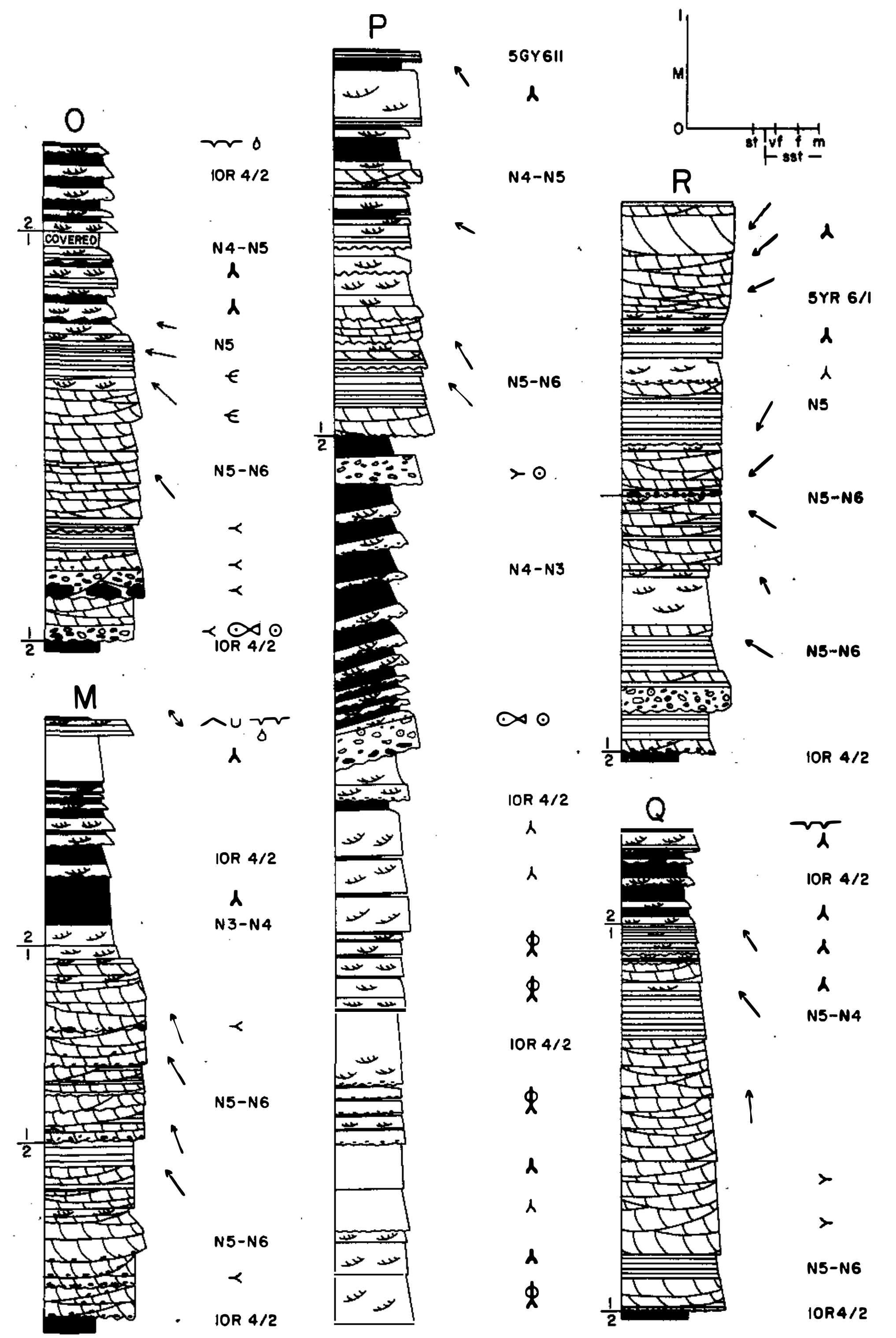
G
J
K
COVERED
COVERED
COVERED
COVERED
5R 4/2
5Y 4/1
N4-N3
N4-N3
N2-N3
N5-5Y 4/1
N4-N5
N4
N4-N5
N4
N5-N4
N4-N3
N5-5Y 4/1
N3-N4
N4-N5
N4
N5-N4
N4-N5
N5-N4
5Y 4/1
2
1


O
10R 4/2
COVERED
N4-N5
N5
N5-N6
10R 4/2
M
10R 4/2
N3-N4
N5-N6
N5-N6
10R 4/2
P
5GY6I1
N4-N5
N5-N6
N4-N3
10R 4/2
10R 4/2
M
0
st
vf
f
m
sst
R
5YR 6/1
N5
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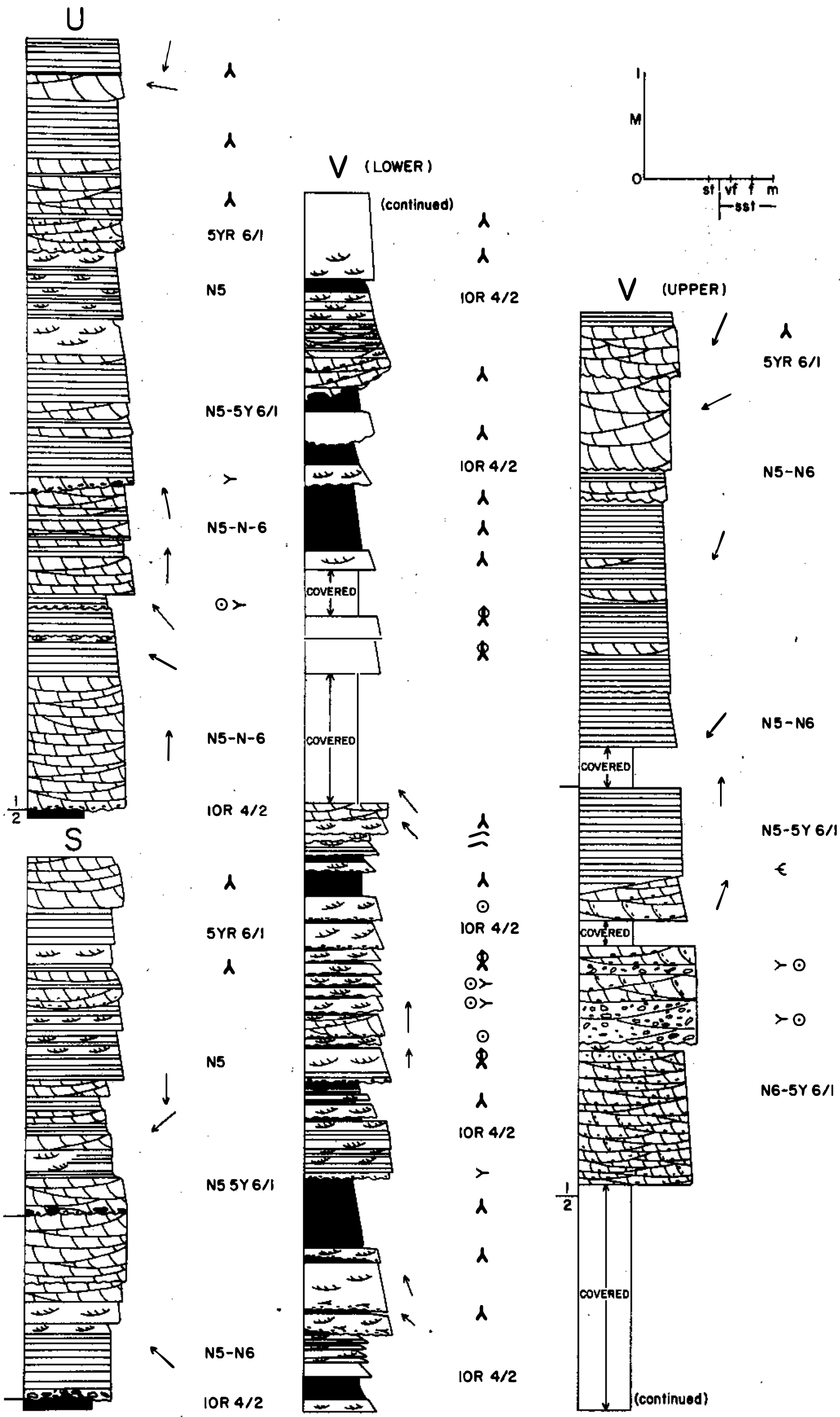
10R 4/2
Q
10R 4/2
N5-N4
N5-N6
10R4/2


U
5YR 6/1
N5
N5-5Y 6/1
N5-N-6
N5-N-6
10R 4/2
S
5YR 6/1
N5
N5 5Y 6/1
N5-N6
10R 4/2
V (LOWER)
(continued)
10R 4/2
10R 4/2
COVERED
COVERED
10R 4/2
10R 4/2
10R 4/2
V (UPPER)
5YR 6/1
N5-N6
N5-N6
COVERED
N5-5Y 6/1
COVERED
N6-5Y 6/1
COVERED
(continued)
M
0
1
st
vf
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m
sst
1/2
1/2

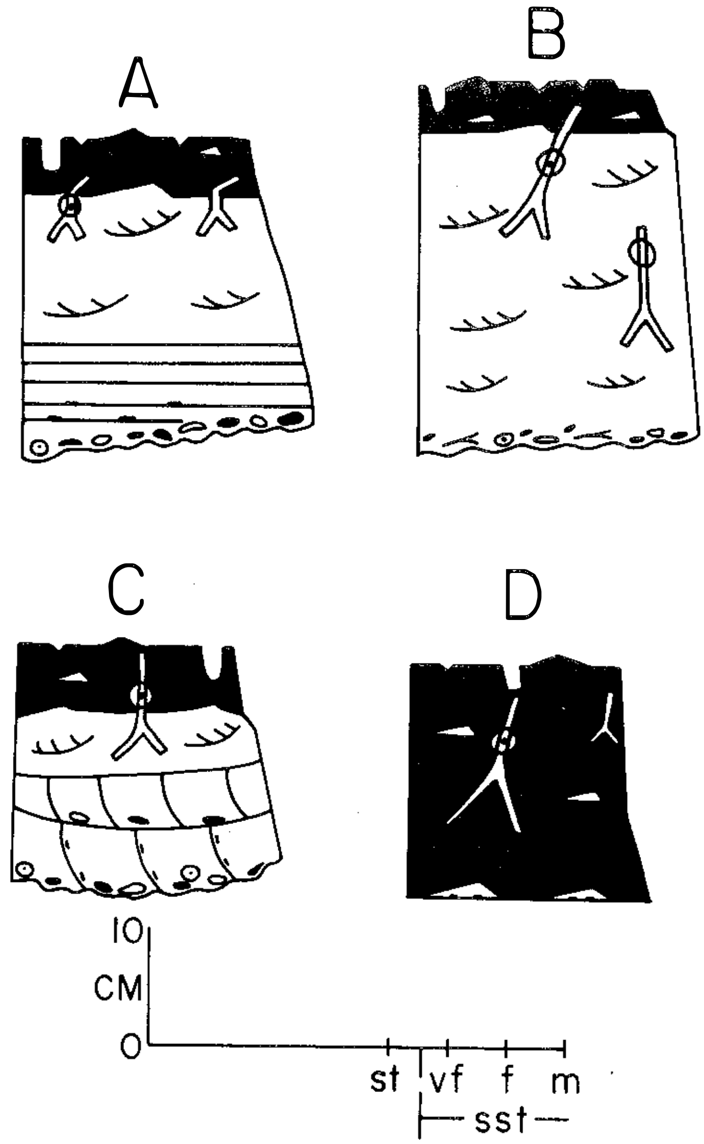
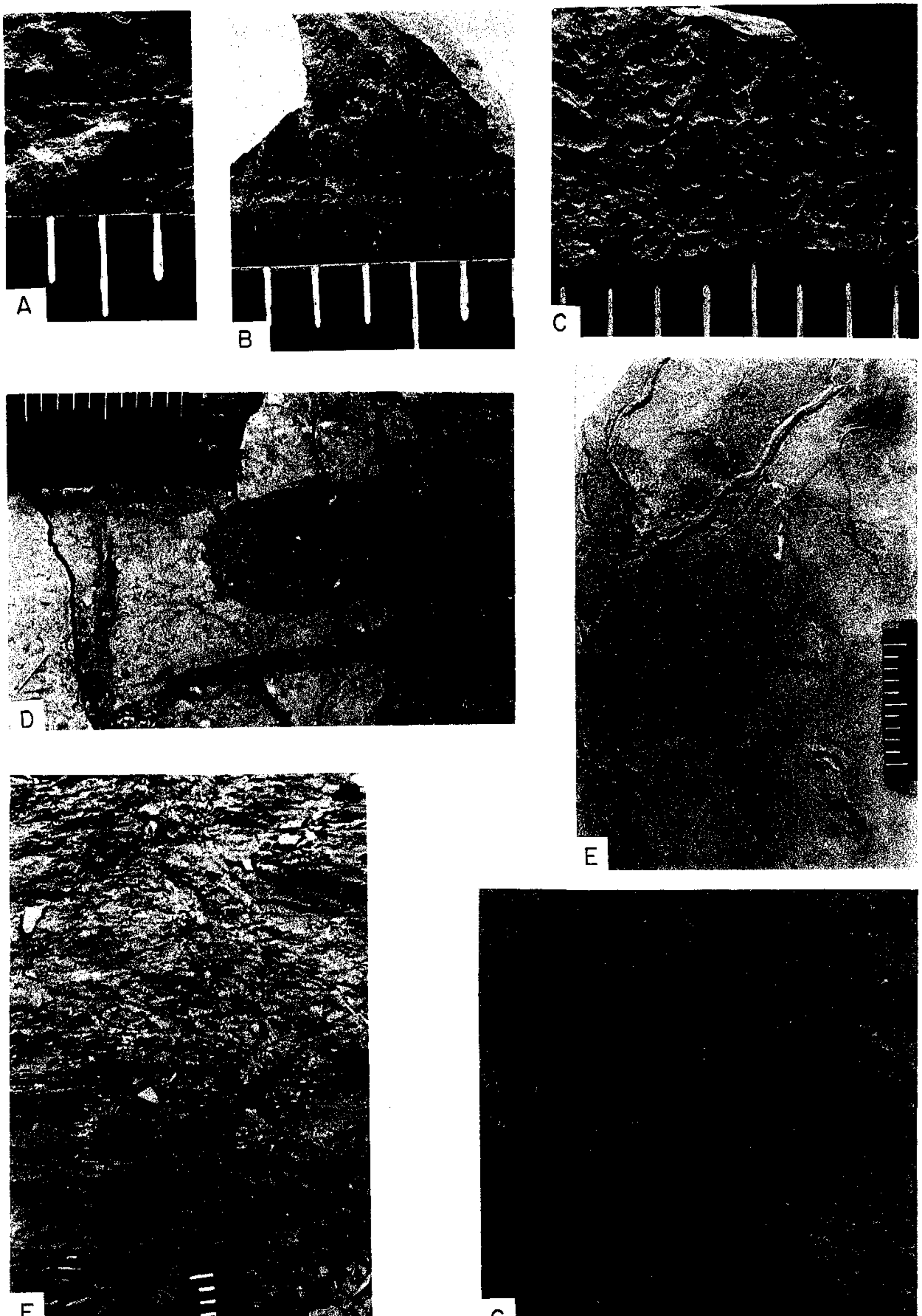


FIG. 4 - Typical small-scale facies sequences comprising facies association 2. See legend.

FIG. 5 - A) Top view of surface trace (facies association 2). B) Basal views of arthropod? trail (f.a.1, channel-fill 1A). C) Raindrop imprints, top surface (f.a.2). D) Large vertical burrows and desiccation cracks (f.a.2). E) Siltstone case of drifted plants on ripple-marked surface (f.a.1, top of 1B). F) Calcareous concretions associated with *in situ* plant roots (f.a.2). G) Desiccation cracks and symmetrical ripple marks (f.a.2). Black board scale is 10 cm in all photos.


A
B
C
D
E
F
G

THE SUBSURFACE ONONDAGA LIMESTONE:
STRATIGRAPHY, FACIES, AND PALEOGEOGRAPHY

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Basin-wide depositional patterns of the Middle Devonian Onondaga Limestone are imperfectly known because its arcuate outcrop belt deceptively exposes only the most shoal-water facies of the formation. The subsurface Onondaga in New York and Pennsylvania, on the other hand, is dominated by moderately deep-water carbonate facies which grade into and are partially time equivalent to the Needmore and Marcellus Shales and the upper Huntersville Chert to the south.

The Onondaga disconformably overlies the Schoharie and Carlisle Center Formations on the eastern side of the basin and the age-equivalent Bois Blanc Limestone (equivalent to the lower Huntersville Chert) on the western side of the basin. In the north-central part of the basin, it unconformably overlies progressively older units from east to west: the Oriskany Sandstone, Helderberg Group, and Bass Islands Group (Upper Silurian). There is, however, no direct evidence for regional subaerial exposure immediately preceding Onondaga deposition.

The Onondaga ranges from 2.5 to 65m (8 to 215 ft) thick and is readily divisible into the Edgecliff, Nedrow, Moorehouse, and Seneca Members. Of these, only the upper part of the Moorehouse Member is present throughout the basin. Other units are absent in places because of nondeposition or submarine scour or because of lateral gradation with the Marcellus Shale.

Onondaga facies are represented by mudstone (fine calcisiltite) or by grainstone, packstone, and wackestone characterized by skeletal constituents of either shoal-water or deep-water faunas. The total absence of ooliths, peloids, carbonate intraclasts, coated grains, and calcified or stromatolitic algae indicate that if particularly shallow, nearshore environments existed, they lay north of the present outcrop belt, perhaps marginal to the Algonquin arch and Adirondack massif. Facies patterns for the Edgecliff, Moorehouse, and Seneca Members reveal successively transgressive paleogeographies. Presumably deep-water facies containing styliolines and delicate brachiopods delineate the epicratonic Appalachian basin in south-central New York and most of Pennsylvania. Facies dominated by crinoids, bryozoans, and robust brachiopods outline an arcuate platform that circumscribed the basin on its northern and western sides. Basin and platform were joined by a south-sloping ramp that was locally dissected by troughs and surmounted by isolated platforms or banks. Basinward, down this ramp, facies change from shoal-water grainstone, packstone, and wackestone, to mudstone and deep-water wackestone and packstone, and finally to gray or black shale.

During deposition of the Onondaga Limestone, water depths increased progressively throughout the basin, as reflected by gradual northward shift of all carbonate facies comprising the successive members and by northward migration with time of the Marcellus Shale. The steadily transgressive Onondaga deposition was terminated by basin-wide stagnation of bottom waters and slow burial by the Marcellus Shale.

SUBSURFACE ONONDAGA BIOHERMAL BANKS:
PALEOGEOGRAPHY, FACIES, AND RESERVOIR FEATURES

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Several gas-producing biohermal banks, popularly known as pinnacle reefs, have been discovered in the Middle Devonian Onondaga Limestone of south-central New York. These subsurface carbonate buildups are distinguished from bioherms known in the Onondaga outcrop belt by their paleogeographic separation from the latter, their far greater size (36 to 63m in thickness and 1200 to 3200m in diameter) and their continued growth throughout Onondaga deposition. Subsurface biohermal banks were initiated as coral-crinoid mounds in the Edgecliff Member and were located on the seaward margins of isolated platforms surmounting the ramp sloping into the basin in south-central New York. Although surrounded by deep-water Moorehouse facies upon subsidence of the platforms, these banks continued vertical and lateral accretion, and persisted even while euxinic Marcellus Shale was replacing upper Moorehouse carbonates to the south. The northernmost build-ups continued growth as Seneca carbonates, overlying the Moorehouse, were further confined by the Marcellus transgression. Bank growth was terminated either by gradual foundering under progressive transgression or by anoxic waters accompanying encroaching Marcellus deposition, but not by terrigenous mud influx. Extinct biohermal banks remained unburied for millenia until gradually onlapped by the Marcellus and Skaneateles Shales.

Study of five subsurface biohermal banks demonstrates their faunal and facies kinship to outcrop Onondaga bioherms, despite the differences mentioned above. Both exhibit broadly domed external structures initiated as bafflestone formed of Acinophyllum-Cladopora thickets, succeeded by alternating Cylindrophyllum and Acinophyllum-Cladopora bafflestones, capped by bryozoan-Cladopora wackestones, and flanked by crinoid-coral rudstone, rich in Emmonsia and Favosites, that grades laterally into deep-water facies. Neither outcrop nor subsurface Onondaga structures were wave-resistant reefs. They formed at considerable water depths probably below the effective photic zone. Algae and stromatoporoids are virtually absent from Onondaga bioherms, despite the equatorial paleolatitude of the northern Appalachian basin, bordering the circumtropical Hercynian Sea, and despite the role of stromatoporoids and calcified algae as prime constructors of other Devonian reef complexes which occupied similar or less favorable paleolatitudes.

These structures are sealed by the overlying Marcellus and Skaneateles Shales which undoubtedly also served as hydrocarbon sources. Most existing porosity consists of primary intraskeletal, intergranular, and growth framework voids. Cementation by calcite spar was halted by invasion of liquid hydrocarbons, presently existing as a tar residue lining the voids. Calcite and chalcendony cementation is virtually

complete in the lower, water-saturated parts of bioherms. All of the seven known subsurface biohermal banks have shown initial production tests of 10 to 18 million cfgd. Wyckoff Reef in Steuben County has produced more than 5 billion cfg since its discovery in 1967.

Conclusions presented in this and the preceding abstract were drawn in part from master's-degree studies at SUNY Binghamton by Mary Rose Cassa (Gulf Research Corp.), Robert M. Coughlin (Shell Oil Co.), and John F. Polasek (Amoco Production Co.). Consolidated Gas Supply Corporation and the American Chemical Society provided funds for these investigations.