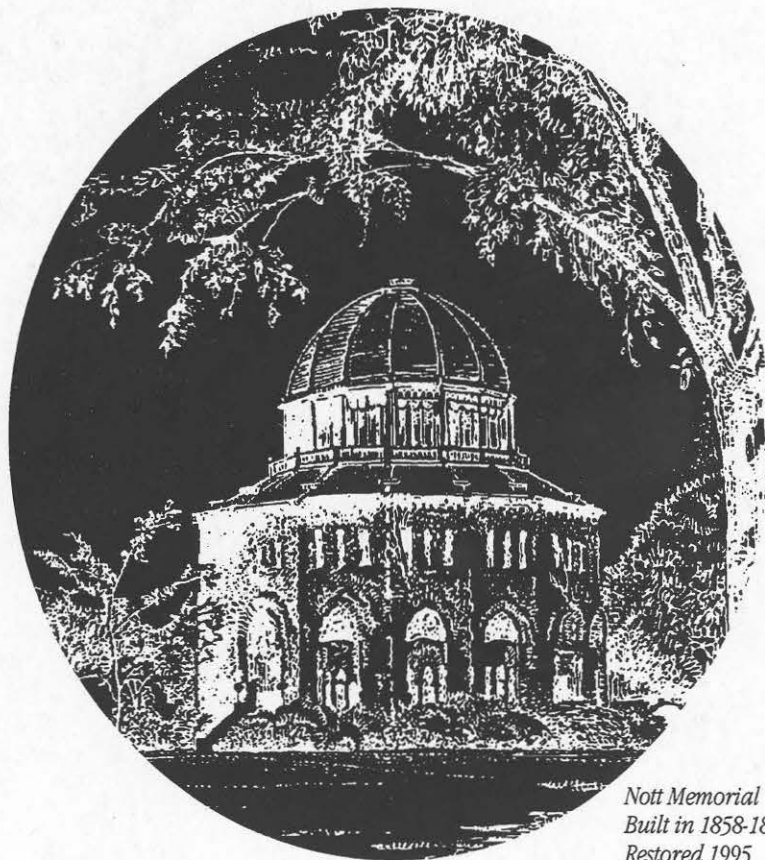


Field Trip Guide for the 67th Annual Meeting of the New York State Geological Association

Edited by
John I. Garver
Jacqueline A. Smith



*Nott Memorial
Built in 1858-1879
Restored 1995*

Hosted by
Union College Geology Department
Schenectady, New York
New York State Geological Survey/Museum
Albany, New York

October 13-15, 1995

Field Trip Guidebook for the 67th Annual Meeting of the New York State Geological Association

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John I. Garver
Jacqueline A. Smith

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For the combined meeting of the
**67th Annual meeting of the New York State Geological Association &
26th Annual meeting of Eastern Section of the American Association of Petroleum Geologists**

Sponsored by

New York State Geological Survey/State Museum and Union College

Hosted by

Geology Department, Union College, Schenectady, NY

This guidebook was published by the New York State Geological Association.
Additional copies may be obtained from the Executive Secretary of the NYSGA:

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Introduction

The Geology Department at Union College is proud to co-host the sixty-seventh annual meeting of the NYSGA and AAPG Eastern Section. Together this combined meeting has brought together academics, students, teachers, petroleum geologists, and environmental geologists. The Geology Department at Union College is happy to be a co-sponsor and co-host of the NYSGA meeting this year because it has been quite some time since we have been able to host such an event. The last time the annual NYSGA meeting was held at Union was 30 years ago when the principal movers in the Department were Philip Hewitt and Leo Hall (see Shaw, this volume). Shortly after this meeting the Geology Department was effectively disbanded and only recently came to life in 1985 after the College received a generous donation from a Geology alumnus. Now, with four permanent faculty, one visitor, and several adjuncts, it is safe to say that we're back for good!

Since the last time NYSGA was held at Union, the Geological Sciences have experienced a major transformation. The first twenty of those years were clearly dominated by the then new plate tectonic paradigm. During the last ten years, and especially the last five years, we have seen quite a dramatic change in the focus of the geological sciences. Due in part to a number of external factors, the geologic community has placed a greater emphasis on issues associated with global change, hydrology, and environmental geology. This shift, has been away from the more traditional areas in geology such as field-based studies in stratigraphy, sedimentology, paleontology, and structural geology. There is little doubt that NYSGA guidebooks reflect the focus and ideas of the times and if one compares the focus of the trips in this meeting to those in the guidebook for the previous NYSGA meeting at Union in 1965, you will see just how much our science has changed.

In assembling the trips for the 67th annual meeting of the NYSGA we tried to balance three important factors: our unique and exciting geologic setting; the breadth of interests of our members; and, importantly, the availability of field trip leaders. Certainly in terms of the first two factors, I feel that we have succeeded and we are looking forward to an exciting meeting!

The geologic setting of the lower Mohawk Valley is ideal for teaching and research. The lower Paleozoic stratigraphy has a regional tilt to the south so one can examine the Grenville-aged rocks of the Adirondacks to the north (Whitney and Olmstead, this volume; McLelland, this volume; Kelly and Hill, this volume), Cambro-Ordovician carbonates and clastic rocks in the Mohawk Valley (Garver, this volume; Kidd and others, this volume), and Silurian and Devonian rocks in the Catskill Mountains to the south (Friedman, this volume; Wolosz and Paquette, this volume; and Ver Straeten and Brett, this volume). The lower Paleozoic stratigraphy in eastern New York is punctuated by thick synorogenic sequences deposited during the Ordovician Taconic Orogeny and the Devonian Acadian Orogeny. To the east of the tilted lower Paleozoic stratigraphy, deformation in the section increases towards the front of the Taconic Allochthon where it and underlying rocks were deformed during the Taconic Orogeny (Hollocher, this volume; Kidd et al., this volume; Friedman, this volume). In the Taconic Allochthon, Cambro-Ordovician deep-water sedimentary rocks have been placed over shallow-water sediments of the Cambro-Ordovician margin.

The surficial and glacial geology in this area is as impressive and diverse as the bedrock geology. Ice dynamics related to the obstructing effects of the Adirondack Mountains resulted in non-uniform migration of ice sheets that coalesced south of the Adirondack massif only to encounter the Catskill Mountains to the south (Dineen and Hanson, this volume). During retreat, glacial Lake Albany (and its relatives) covered much of the Capital District and left an impressive sequence of glacial lacustrine deposits (Wall and LaFleur, this volume). During and after the most recent glaciation, karst and caves formed in the local carbonate strata, resulting in important landform development and interesting problems in hydrogeology (Rodbell and Hays, this volume; Rubin et al., this volume; Rubin, this volume). The long history of industry in the area and demand for space and water have resulted in some unique environment geology (Smith and Eslinger, this volume; Maswick and Snow, this volume; Hewitt, this volume). Finally, the neotectonic setting of eastern New York is examined in a unique trip aimed at understanding the seismic hazards in Columbia County (see Nottis and Cadwell, this volume).

It has become clear that the education of educators is and will continue to be an important priority for both Academe, government, and industry. With this in mind we have included several trips aimed at earth science teachers (Garver, this volume; Hollocher and Hollocher, this volume; and Kelly and Hill, this volume, and others) and we hope that many of the teachers who go on the field trips participate in the special one day workshop which will be held in Monday following the NYSGA meeting

Finally, I would like to personally thank all of the people that have made this guidebook happen. First and foremost are the trip leaders who have spent a considerable amount of time putting together these trips and to my co-editor, Jacquie Smith, who helped with the endless task of reigning in papers and keeping track of things. Thanks are also due to the Organizing Committee and Committee Chairs for the combined NYSGA & AAPG Eastern Section meeting for their support, suggestions, and recommendations. Among them, Ken Johnson (Skidmore College), the General Chair, deserves special praise for keeping us together and focused for the committee meetings which periodically met for more than one year prior to the actual meeting. He also deserves special recognition for finally getting on EMail during that time (you can personally thank him at kjohnson@skidmore.edu). I would also like to thank Bill Kelly (N.Y. State Geological Survey), who is the N.Y.S.G.A. Executive Secretary, for his important contributions which include, but are not limited to advice and planning the logistics of putting this guide together. Finally, Gretchen Turner did a superb job handling the problems that editors face in terms of communication with authors and the nuts and bolts of putting the guide together.

To all who are just about to embark on the 67th annual NYSGA meeting -- have a great time and enjoy the fall foliage!

John I. Garver
President, NYSGA, 1995
Geology Department
Union College
October, 1995

PLEASE READ THIS!

Landowners and outcrop defacement

In preparing this guidebook we asked the authors to do their best to ascertain landowner status at each stop. We have done this because increasing number of problems have arisen between landowners and geologists - in some cases geologists are no longer permitted to visit some localities in the Mohawk Valley. Additionally, a number of classic and important outcrops have been defaced, marked, or heaped with garbage by overzealous visitors who must leave their mark on every outcrop visited. Please pay careful attention to landowner status and cultivate a relationship with the landowner if you plan to use an exposure from year to year. Also remind your trip participants to use have respect for the future geologists who will visit these geological sites. Thank you.

HISTORY OF GEOLOGY AT UNION COLLEGE

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The Early Years

Geology began at Union in 1809 when Thomas Brownell was hired to teach a course in mineralogy. At that time mineralogy was closely allied to chemistry, which was also under his purview. Much of the search for new chemical elements at that time was focused on exotic minerals of unusual composition. Brownell was dispatched to Europe to study and to purchase specimens, instructional aids, and apparatus for teaching the new courses. During the following ten years he added to the mineral collection through his own efforts in the field and donations from alumni and others. When he left to become Bishop of Connecticut the collection, for which he had maintained a partial catalogue, numbered about 2000 specimens.

Joel Nott, one of Eliphalet's sons, was added to the faculty about 1820. He took over teaching mineralogy and chemistry upon Brownell's departure. His close association with geology is evidenced by his inclusion in an expedition to the Michigan Territory in 1821.

The records concerning geological instruction are sparse during the late 1820's and early '30's. Joel Nott seems to have left the college during that time. His brother, John Nott, may have been involved in teaching geological subjects as one of the faculty in Natural Philosophy, but this cannot readily be determined. Geology during the nineteenth century was typically included in the broader field of natural history and at Union College the professor of natural history normally would teach botany, zoology, and geological subjects. Mineralogy, however, was taught by a chemist, at least until the latter part of the century.

In 1834 Benjamin F. Joslin is listed as a professor of natural philosophy, a position he held until at least 1838, but he appears to have been mostly, if not exclusively, concerned with biological instruction. Mineralogy was almost certainly taught in 1833-36 by Chester Averill, an adjunct professor of chemistry and languages, and he was apparently replaced in 1836 by Edward Savage, an assistant professor of languages and chemistry.

Chester Averill died of tuberculosis in 1836, leaving a wife and infant son, Chester Jr.. The younger Averill completed a degree at Union in 1857. He subsequently became a member of the first Geological Survey of California, headed by Josiah Whitney (after whom Mt. Whitney was named). Whitney clearly had an association with Union College, as curator of the college mineral collection, and as an advisor to Eliphalet Nott concerning Nott's investment in the Bristol Mines in Connecticut. That venture was a financial disaster, but apparently not to the detriment of Whitney's relationship with Nott, for Eliphalet wrote an enthusiastic letter supporting Whitney's candidacy for the directorship of the California Geological Survey.

Edward Savage, of whom little record remains, left the college in 1839, to be replaced by Jonathan Pearson, as assistant professor of natural philosophy and chemistry. Pearson is well known for the diary he kept through most of his considerable time at the college, but he was also of great importance as the curator of the college museum, including the mineral collection.

Starting around 1840 geology became a part of the curricular offering, along with mineralogy, and separate textbooks were used for geology and mineralogy. The advent of a curriculum in engineering in 1845 added a strong practical element, to which geology no doubt contributed through study of ore minerals, mining and metallurgy. In the college catalogue of 1852 a Botany and Mineralogy Department is mentioned for the first time, likely a reflection of Pearson's main interests. The curriculum and structure of the college was apparently quite fluid throughout most of the latter half of the century, with course

In Garver, J.I., and Smith, J.A. (editors), Field Trips for the 67th annual meeting of the New York State Geological Association, Union College, Schenectady NY, 1995, p. 1-9.

Figure 1. Harrison Webster. Prof. of Natural Philosophy at Union (1868-1883). Prof. of Geology and Natural History at the University of Rochester (1883-1888). President of Union College (1888-1894).



offerings and departmental designations appearing and disappearing from year to year. However, mineralogy remained an important part of the Science Course, as reflected in descriptions of Analytical Chemistry from the college bulletin:

When [the student] has in this manner acquired sufficient confidence in his skill, he can proceed to the actual Analysis of Minerals, Soils, Manures, (etc.)" and "Mineralogists will have access to the College [mineral] Cabinet, and can also take a full course with the blowpipe, and in Qualitative and Quantitative Analysis.

In 1858, the Wheatley Collection was purchased by Edward Delavan and donated to the college. This important collection has been the core of the departmental collections ever since. The close connection between chemistry and mineralogy continued with the addition of the Wheatley collection, and Charles Chandler (appointed to the faculty in 1857 as an assistant prof. of analytical chemistry) became curator of the museum. In 1865 Maurice Perkins took over from Chandler, including duties as museum curator.

Harrison Webster (1868 - see Figure 1) joined the faculty as a tutor in Natural History in 1868, and soon advanced to a regular faculty position. His responsibility for geology is evidenced by his later assumption of a professorship of geology and natural history at the University of Rochester. Webster was a major actor in the faculty movement opposing President Potter during this time and he left the college for Rochester in 1883. His replacement was another alumnus, James Stoller ('84), first as a tutor then as a professor. Stoller's duties increased with the retirement of Jonathon Pearson in 1885. Webster returned to the College as its president in 1888 (the first non-clergy to hold that position) and he again taught in the department of biology and geology.

Charles S. Prosser

The first full-fledged geologist at Union College was added to the faculty in 1894 in the person of Charles S. Prosser (Figure 2), who as acting Professor of Geology was responsible for the department of Geology and Paleontology.



Figure 2. Charles Prosser. Prof. of Geology at Union College (1894-1899). Prof. and Chair of Geology at Ohio State University (1899-1916). Published early work on Permian and Late Paleozoic stratigraphy of Kansas, Ohio, etc.

While he is probably unknown to even the oldest Union alumnus, Charles S. Prosser was probably the most eminent geologist ever to teach at the college. He was the first professor at the college specifically trained in geology, and was brought to the college in 1894 to start a full-fledged geology program. He was born on March 24, 1860 and raised in the Unadilla Valley of south-central New York. He attended Cornell University where he earned a B.S. in 1883 and a Master's degree in 1886, studying geology as his principle field. He was the first holder of the Cornell Fellowship in Natural History as a beginning graduate student, and was in the first class of initiates to the Society of Sigma Xi, which was founded at Cornell. In 1906 he was awarded a Ph.D. from Cornell.

An avid and careful field observer, he began his career at the United States Geological Survey in the Division of Paleobotany. In 1892 he took a position as Professor of Geology at Washburn College in Topeka, Kansas, where he began a study of Permian rocks. The study of Paleozoic strata occupied him throughout his professional life. Depending upon the rocks available in the immediate vicinity he shifted his focus to different parts of the geologic record, from Devonian at Cornell, to the Permian of Kansas, to the Ordovician of New York and Ohio. In 1894 he answered the call to Union to establish geology. Although his stay was comparatively brief, by the time of his departure in 1899 he had not only provided Union with a strong geology department with a full range of courses, but he had profoundly influenced a small group of young men, at least two of whom (Cumings '97, and Hartnagel '98) would themselves become prominent geologists. His method of teaching was one almost of apprenticeship; in modern terms we would describe it as undergraduate research participation. Field trips and detailed examination of specimens in the laboratory, as well as familiarity with the literature were fundamental to his approach to teaching geology.

Prosser's contributions to geology included many observations and constructions of Paleozoic stratigraphy which were major departures from previous work. He was at various times on the staffs of the State Geological Surveys of New York, Maryland, Kansas, and Ohio, where most of his work was carried out. He was a founding Fellow of the Geological Society of America, and an inspirational mentor to many successful geologists. His view of higher education, and more particularly his approach to teaching geological science would be well heeded by more institutions in the present day. His statements concerning the practice of science are filled with references to the selflessness and hard work necessary to the pursuit of scientific research.

The Turn of the Century

Prosser left Union in 1899 at the behest of Prof. Edward Orton, Sr., who was seeking a successor to the chairmanship of the Geology Department at Ohio State University. The opportunity so presented, especially in contrast to the support for geology forthcoming at Union, made this move inevitable. Within two years he was confirmed as Professor and Head of the Department of Geology, the position he held until his untimely death by his own hand in 1916.

At this point in the history of the department there was a fairly complete curriculum in geology: Geology, Historical Geology, Paleontology, Economic Geology, Areal Geology, Field Geology, and Mineralogy and Lithology. The Wheatley Collection had become part of the Geology Department in 1890, and during his brief stay at the college (until 1899), Prosser put considerable effort into rehabilitating and adding to the collection, especially with paleontologic specimens. Prosser's ambitions for a strong geology department appear to have initially received some support from the trustees and administration of the college, but apparently not to his satisfaction. He left for Ohio State where he soon became chair of the Geology Department.

Stoller resumed responsibility for geology and curation of the collections when Prosser left. Stoller, although primarily trained as a biologist (his Ph.D. Dissertation concerned sowbugs), carried out some very significant early work on the surficial and glacial geology of the Mohawk-Hudson region.

James Stoller

James Stoller (Figure 3) was born on December 11, 1857 and was raised in Johnstown, NY. He attended public schools and the Cazenovia Seminary before entering Syracuse University. He left Syracuse for Union in 1881, and graduated with an A.B. in 1884. The following year he joined the faculty as an Instructor in Natural History. He taught courses in Biology and Geology, and was appointed Assistant Professor of Biology and Geology in 1889. In 1894 he was appointed Professor of Biology following graduate study at Johns Hopkins, Munich, and Leipzig Universities. In 1898 he was awarded the Ph.D. from Leipzig, having completed a dissertation on "The Organs of Respiration of the Oniscidae". This study of the terrestrial isopods commonly known as "sowbugs" or "wood lice" shows his interest in biological matters as well as careful observation and attention to detail. He was appointed chair of Biology and Geology in 1898, and with the arrival a new biology professor in 1919, he became Professor of Geology.

Stoller's investigations on the glacial history of the area from Saratoga to Albany and west up the Mohawk Valley clearly established a sequence of events during the latter part of the glacial epoch. His results, stemming from many years of field observation and careful deduction, were published as Bulletins of the New York State Museum. They include: "Glacial Geology of the Schenectady Quadrangle", "Glacial Geology of the Saratoga Quadrangle", "Glacial Geology of the Cohoes Quadrangle", and "Topographic Features of the Hudson Valley and the Question of Post-Glacial Marine Waters in the Hudson-Champlain Valley." These unassuming titles conceal some of the most fundamental observations concerning the recent geologic past in this area, and remain scientifically useful to the present day.

In 1924, after forty years of service teaching at Union, he was awarded an honorary Sc.D. by his alma mater. He retired the following year, but his presence was felt at the college for many years, as he was a focus for returning members of his class of 1884, who established a fund in his honor. The proceeds from that endowment continue to provide for the purchase of books in geology, and even kept a spark lit during the nearly twenty years when geology was in eclipse. Stoller's divided responsibilities led to a reduction in



Figure 3. James Stoller. Prof. of Biology and Geology at Union College (1884-1925).
Published some of the earliest studies of glacial geology in the Captiol District.

course offerings in geology for the next 20 years, but geology continued as part of the department of biology and geology.

The importance of fieldwork to a geological education had been recognized at least as early as 1889, and during Stoller's years reached a point where an honors course in glacial geology required at least 60 hours of fieldwork, and in 1916-17 an honors course in field geology required at least 120 hours of fieldwork and a detailed report on an assigned area.

The mineral collections continued to be an important educational asset, and were curated by a volunteer, Dr. D.S. Martin from 1908-1917. The developments of X-ray crystallographic methods, beginning with the Bragg's in 1912, reached Union College in 1919, when a series of special lectures in crystallography and x-rays was taught by Albert W. Hull and Wheeler P. Davy.

As Stoller neared retirement, the need for a successor led to the hiring of Edward Staples Cousins Smith (known to the students by various nicknames, e.g. alphabet Smith). Smith was the Geology Department for thirty-five years, from the retirement of Stoller in 1925 until 1960.

Edward Staples Cousins Smith

Certainly the best known geologist in Union's history, "Prof." Smith was mentor to a long string of geology students, a remarkable number of whom went on to eminence in the field. Smith was geology at the college for nearly thirty five years, from his accession to chair of the department in 1925 to the addition of a second regular faculty member in 1957.



Figure 4. Edward S. C. Smith. Prof. and Chair of Geology at Union College (1923-1960).

Edward Staples Cousins Smith was born in Biddeford, Maine on August 23, 1894. He studied chemistry at Bowdoin College, and after his graduation in 1918 he continued with graduate study in geology at the Massachusetts Institute of Technology and Harvard University. After receiving his M.A. degree from Harvard in 1920 he taught for two years at Radcliffe College, which he left to become the new instructor of geology at Union.

Upon Prof. Stoller's retirement in 1925, Smith was promoted to Assistant Professor and chair of the Geology Department. Smith was a geologist both by inclination and education, and he set about creating a comprehensive program in the science.

Smith was promoted to Associate Professor in 1929, and to Full Professor in 1932. Although diminutive in stature he was a "presence" on campus, and many are the stories told by former students and colleagues. While many of these touch on a certain imperiousness, they also reflect the affection which the students held for him, no doubt a re-reflection of his concern and fondness for them.

Smith's contributions to geology, while not extensive in terms of publications, touched upon some very significant areas of geology, especially with regard to the geology of his home state, Maine. It is fair to say that he was first and foremost a dedicated teacher, and not just within the gates of Union College. He had a great concern for the education of the general public, especially regarding the importance of geology to the public at large. His concern for conservation and the environment preceded by many years the fashion which became general a decade or so after his retirement. He was responsible for one of the earliest television series devoted to geology, and was heard over WGY in a series of lectures for the General Electric Science Forum. With co-authors he published one of the earliest books on nuclear energy: "Applied Atomic Power", not only a useful technical compilation, but a helpful work for the general public.

Smith's role in geology was recognized by many of his colleagues. He was a close associate of Christopher Hartnagel, the New York State Geologist, and Rudolf Ruedemann, a noted paleontologist and stratigrapher who worked not only in New York, but studied (with help from Smith) important Cambrian sections in Maine. Ruedemann named a fossil, *Oldhamia smithi*, in his honor. Smith was a fellow of the Geological Society of America and the Mineralogical Society. He was an early organizer and president of the New York State Geological Association.

One of Professor Smith's most interesting contributions was in the field of mineral fluorescence. He and one of his students, William Parsons, did experimental work on fluorescence spectra of minerals. This phenomenon is familiar to many who have visited one of the darkened chamber in a mineralogical museum where "black lights" produce striking colors on a variety of specimens. Smith's collection of fluorescent specimens is still a fascinating part of the mineral collection at Union.

When he retired in 1960 he was honored by his former students, who established a prize fund in his name. The Edward S. C. Smith Geology Prize is awarded to a geology major at the college who shows high professional potential. Although the award was suspended in the early 1970's, as a result of the termination of the geology major, the fund continued to grow. With the re-establishment of geology the prize is once again being awarded to outstanding students in geology. Although he died on Nov. 11, 1971, after the demise of geology at the college, the renewal of the department, and the Smith Geology Prize, somehow brings him to life once again, as a presence among those of us dedicated to geology at Union.

During the 'late 20's and 30's Smith was able to teach a sound curriculum in geology, often with the help of a young instructor, who would be working on a master's degree in geology at the college. This was the only period during which M.S. degrees were awarded in geology at Union, and the contributions made by those few who assisted Smith with instruction are remembered by many of the older alumni in geology. The department can be characterized as a successful, smoothly functioning entity which produced many fine geologists, who left the college with considerable pride in their department. The department structure had finally become formalized about 1920, with courses regularly listed by department from that time to the present. The geology department, however, did not participate in the growth of many of the other departments during that interval, being comprised of Smith and a young visiting lecturer for much of that time. During the thirties the department offered a master's degree, and the lecturer was often a graduate student working on his degree, or a recent graduate of the department.

After World War II

As Smith in his turn neared retirement, a new assistant professor was hired in 1957. Philip Hewitt took over the department in 1960, and Leo Hall was hired as a second full time geologist in 1961, expanding the department to two regular faculty for the first time. Through much of the 1960's these two gradually increased the offerings in geology. The increased interest in geology (partly from increased employment opportunities) coupled with generally increasing enrollments in higher education at that time, encouraged Hewitt and Hall to ask for a further increase in the faculty in the department. This was bolstered by the report of an external examining committee chaired by John Moss, a member of the geology

department at Franklin and Marshall College. The need for an additional faculty member was apparent to the visiting committee, as was a restructuring of the course offerings in geology. On the basis of the small number of geology majors (averaging five for the decade from '55-'65), and declining enrollments in introductory geology due to changes in curricular requirements, the administration refused to increase the size of the department. While this decision is thought by some to have been made by the Board of Trustees, there is little or no evidence that they seriously considered the issue, rather it seems to have been an administration decision. Faced with continuation of what they perceived as inadequate support for Geology, Hewitt and Hall resigned in 1967. Courses were taught to remaining geology majors through an arrangement with RPI until 1971. The geology department was allowed to "run down" as the majors departed and no new majors were added to the program.

With the demise of the geology major, there remained an interest in having geology courses taught as part of the general education program of Union students and for introductory geology for civil engineers, but no investment in a full-fledged department was considered. In 1971 Herman Zimmerman, a marine geologist, was hired to teach introductory level courses in geology and oceanography. For the next 13 years his success was measured, in part, by the number of students who left Union to seek a major in geology at other institutions. Zimmerman was officially a part of the department of civil engineering, the closest entity available to accommodate a geologist. The "geology department" was moved into the second floor of Butterfield Hall when the Civil Engineering department was moved into the first and third floors, following the construction of the Science and Engineering complex.

The return of geology after dormancy

During the late 1970's and early 1980's an effort was made to reestablish a geology department and major at the college. This effort was led by Frank Grigg's, the chairman of the Civil Engineering Department. A group of geology alumni also discussed the possibility of a restart of geology. This movement finally bore fruit through the singular contribution of John S. Wold, a geology major of the class of 1938. In 1983-84, Wold conferred a substantial endowment upon the college with the understanding that it be used for a chair in Geology and for re-establishment of a Geology Department at the college. Interestingly, the report of this gift in "Concordiensis" is rather ambiguous about the purpose of the gift, seemingly a reflection of the ambiguity felt by the administration toward a restart of geology.

Although one might assume that this backing would result in a rapid renewal of geology, such was not to be the case. The college, during deliberations concerning the possibility of restarting geology, requested advice from an external committee. The committee recommendations, perhaps on the basis of their understanding of what the college considered possible for geology, made what can be best described as "minimal" recommendations. In particular little consideration was given to the needs of a new, modern geology department in terms of equipment and space. Even the recommendation of a minimum of three full-time faculty was just that, a bare minimum rather than an estimate of the optimal size. It is clear that the college began the renewal of geology with a substantial underestimate of the costs involved for a quality program. Since the administration was concerned (and so stated) that the new department should be a quality addition to Union, the new department was placed in a resource squeeze, especially regarding space. Much of this was simply a lack of appreciation of what a geology department really needs for facilities. Indeed, this is a problem extending well back into the Department's history. In the late 1890's Charles Prosser clearly had some difficulty in making his needs known to the administration, and certainly Hewitt and Hall likewise.

The Geology Department was officially re-established in 1985, in conjunction with hiring a new assistant professor, Kurt Hollocher. The new department was allocated three full time faculty lines and Hollocher was the second, joining Zimmerman, who chaired the new department. The department began the process of hiring a third faculty member, but the process was interrupted by Zimmerman's announcement of his intention to resign to take a position with the National Science Foundation. A decision was made to hire a visiting assistant professor to temporarily fill the third position while a search was begun for a new department chair. Paul Ryberg joined the department in the fall of 1986 as a visiting professor and the search for a new chair proceeded. The search took two years and ended with the appointment of George Shaw as the John and Jane Wold Professor of Geology and chair of the Geology Department in the fall of 1988. During the 1988-89 academic year the Department hired John Garver as the permanent third faculty member, completing the complement of faculty envisioned for the rebirth of geology at Union.

In 1992, after a four year effort to convince the college of the need for a fourth faculty line, the Department was granted a fourth permanent position. This was filled with the arrival of Donald Rodbell in the summer of 1994. During the 1992-93 academic year the department had the pleasure of hosting Nikolai Sobolev from the Siberian Academy of Sciences. During his visit he taught a course on the Geology of Russia, and gave a number of public lectures. As a result of his visit, Prof. Garver was able to establish important research contacts in Russia, resulting in two field excursions to the Kamchatka Peninsula which has only been recently opened to foreigners. These contacts also resulted in the department hosting another visitor from Russia in 1993-94, Galina Ledneva, a graduate student from Moscow. Overlapping with these two visitors from Russia, we also had a visitor from Romania, Marian Lupulescu. Marian's interest in economic mineralogy was especially appreciated. His curation of the mineral collection, including entering data into a computerised format, will make the collections significantly more useful. He also taught a course in economic geology and led a three week field trip in Romania, attended by two faculty and six students.

In the fall of 1994 George Shaw was asked to chair both the Department of Geology and the Department of Civil Engineering. To compensate the Geology Department for the diversion of faculty time to additional administrative responsibilities, a two year visiting position was granted to the Department. Sharon Locke was hired into this position in the spring of 1994.

During its ten years of new life the Geology Department has grown apace. We are now (at least temporarily) five faculty. The number of geology majors has increased from zero to about twenty. In 1995 eight students graduated in geology, and one in environmental studies with a geology concentration. In the last seven years the Department has obtained grants and contracts totalling more than \$1,000,000.

No history of an academic department would be complete without some discussion of the students. Union's Geology Department has always been small, and the total number of geology alumni, including those of the last few years, is less than 150. However, the quality of the program and students has been high, and the devotion of the graduates to their old department has been admirable. The successes of the alumni have been gratifying, especially as an indication of the quality of geological education provided by the Geology Department and by Union College. Among the graduates are three who became State Geologists (of New York, Pennsylvania, and Wisconsin). About 1/3 of the alumni (~40) went on to earn Ph.D. degrees. About a dozen have been, or currently are, chairs of geology departments at colleges and universities. About 1/3 have enjoyed careers in industry. One currently serves as the Director of the Geological Survey of Canada. The nearly equal division of careers between industry, government and academe demonstrates not only the breadth of interest of our former students, but the breadth of the educational preparation they received, and of which the college can be justly proud.

SEISMIC RISK ASSESSMENT OF COLUMBIA COUNTY, NEW YORK

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INTRODUCTION

In 1994 the New York State Geological Survey began a pilot study to assess earthquake hazards in New York State, on a county by county basis. This effort was modeled after studies done in Portland, Oregon, and Los Angeles, California. The goals of this work were to produce:

- an earthquake ground-motion response map,
- an earthquake-induced landslide-potential map,
- an earthquake-induced liquefaction map, and
- a combined earthquake hazards map.

Our analyses were done as a function of glacial landforms. The maps produced are a result of computer-based analyses of earthquake strong ground-motions, soil depth profiles, soil engineering properties, and topographic slopes. These maps are intended to aid civil engineers, building designers, and emergency managers on the subjects of earthquake preparedness and hazard mitigation.

Columbia County was chosen for the initial study because of its proximity to the Capitol District, and also because of the quantity and quality of geologic data already available. This pilot study was a learning experience. We are still in the process of analyzing the collected data, and the preliminary results are encouraging. During this field trip we will show why such studies are needed, how these studies should be conducted, common problems, and how the results of these studies can be applied.

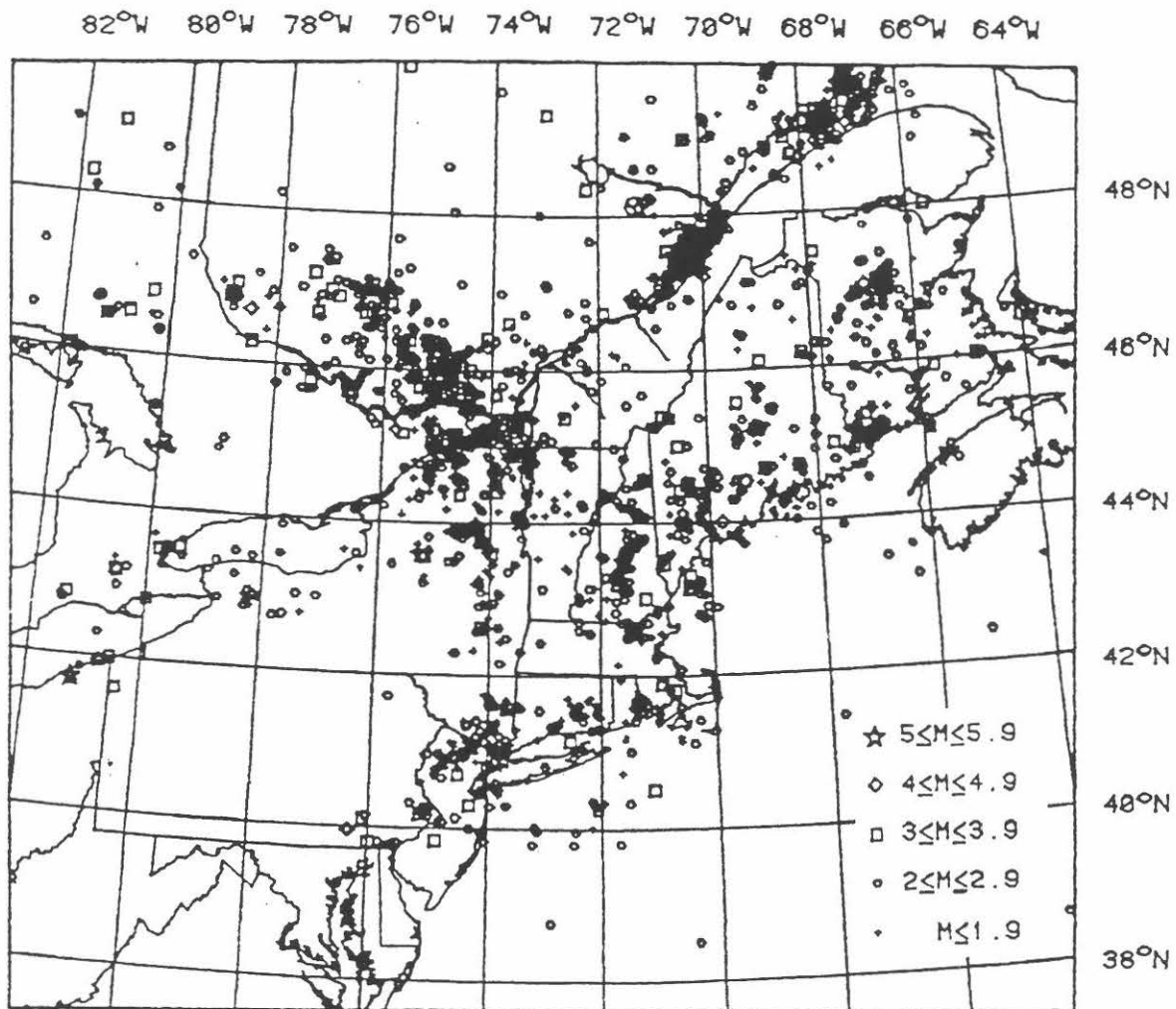
THE EARTHQUAKE THREAT

New York State has experienced 13 or 14 damaging earthquakes since 1737. The estimated or known magnitudes of those events range from 4.5(m_b) to 6.0(m_b)¹. The last damaging earthquake in New York took place at 6:18 a.m. EDT, on October 7, 1983, near Newcomb, New York. That magnitude 5.1(m_b) earthquake caused damage within 6 kilometers of the epicenter, including cracked masonry walls and damaged chimneys at Adirondack camps, rotated tombstones, and small landslides. Aftershocks were recorded for several months. It is believed that this earthquake triggered a small landslide in Columbia County, over 180 kilometers away, that resulted in damage to the foundation and walls of one home.

New York's largest earthquake occurred at 12:38 a.m. EWT (equivalent to EDT), on September 5, 1944, near Massena Center, New York. This magnitude 6.0(m_b) event caused considerable non-structural damage to unreinforced masonry buildings, including schools and governmental buildings, in Massena, New York and Cornwall, Ontario. Over 5,100 (90%) of the chimneys in these communities were damaged or destroyed. Liquefaction, lateral-spread landslides, and small slumps were observed in the Massena area. Power was out in Massena for up to 2 hours after the earthquake, pipes of the community water supply were broken, and street pavement buckled. At least \$17 million of damages resulted. Two people received serious injuries and many others suffered minor injuries. A damaging aftershock occurred 4 days later and additional felt aftershocks continued for 4 months. Seismograph stations in the region recorded aftershocks for up to 4 years following the main event.

Figure 1 shows the distribution of recent (October, 1975-March, 1989) earthquake activity in the New York region. These patterns and clusters of activity have stayed more-or-less stable for over 400 years. However, there is no assurance that areas currently experiencing little or no activity will stay that way. Conversely, areas with seismic

¹ (m_b) is magnitude based on the first three cycles of the P wave, with a period of approximately one second. We do not use the Richter Scale because it is only used in California.



DETAILED INSTRUMENTAL SEISMICITY 1975-1989

Figure 1. Earthquakes of the northeastern United States and southeastern Canada during the period of October 1975 to March 1989. Symbols show the epicenter and magnitude of each event. Figure courtesy of the Weston Observatory, Boston College.

activity may not stay active. The rates of activity in different areas of the northeastern United States and southeastern Canada vary in the historical past. Figure 2 shows that over 70 damaging earthquakes, rated intensity VI-X on the Modified Mercalli scale of 1931² (Wood and Neumann, 1931), have occurred in the northeastern United States and southeastern Canada since that mid-1500's. The 1983 Newcomb and 1944 Massena earthquakes rated intensities VII(MM) and VIII(MM), respectively. The magnitudes of damaging earthquakes in the northeast range from 4.5(m_b) to 7.0(m_b).

Earthquakes that produce slight damage (intensity VI [MM], magnitude 4.5 [m_b] -5.0 [m_b]) occur in the Northeast about once every 6 years. Earthquakes that produce moderate damage (intensity VII [MM], magnitude 5.0 [m_b]-5.5 [m_b]) occur about once every 30 years. Those that produce considerable damage(intensity VIII [MM],

² Modified Mercalli Intensity is a measure of shaking severity based on earthquake effects on people, structures and the natural environment.

magnitude 5.5 [m_b]-6.0 [m_b]) take place about once every 130 years in the Northeast. There is general agreement that earthquakes of magnitude 6.0(m_b) to 6.5(m_b) are possible anywhere and at any time in the region, including New York State. Even larger events are possible in the St. Lawrence River Valley and along the eastern seaboard. Large earthquakes occur frequently enough in our region to be of significant hazard to the concentrated populations, and aging infrastructure and industrial facilities.

EARTHQUAKE HAZARDS

The larger earthquakes of New York State have demonstrated the hazards of strong motion amplification, landslides, and liquefaction. The magnitude 4.5 (m_b) Warrensburg, New York earthquake of April 20, 1931 was a case where a relatively minor earthquake caused moderate (intensity VII [MM]) damage. Warrensburg is situated within the river valley, on lake sands and alluvium directly above bedrock. Amplification of the shaking of the site probably occurred as shear-waves were attenuated, and increased in amplitude, as they passed from bedrock into the unconsolidated materials. The 1931 Warrensburg, 1983 Newcomb, and August 12, 1929, Attica ($m_b = 5.2$) earthquakes all started landslides. The 1944 Massena earthquake is the only New York State earthquake where liquefaction has been documented. Landslides are generally triggered by earthquakes of at least magnitude 4.0(m_b).

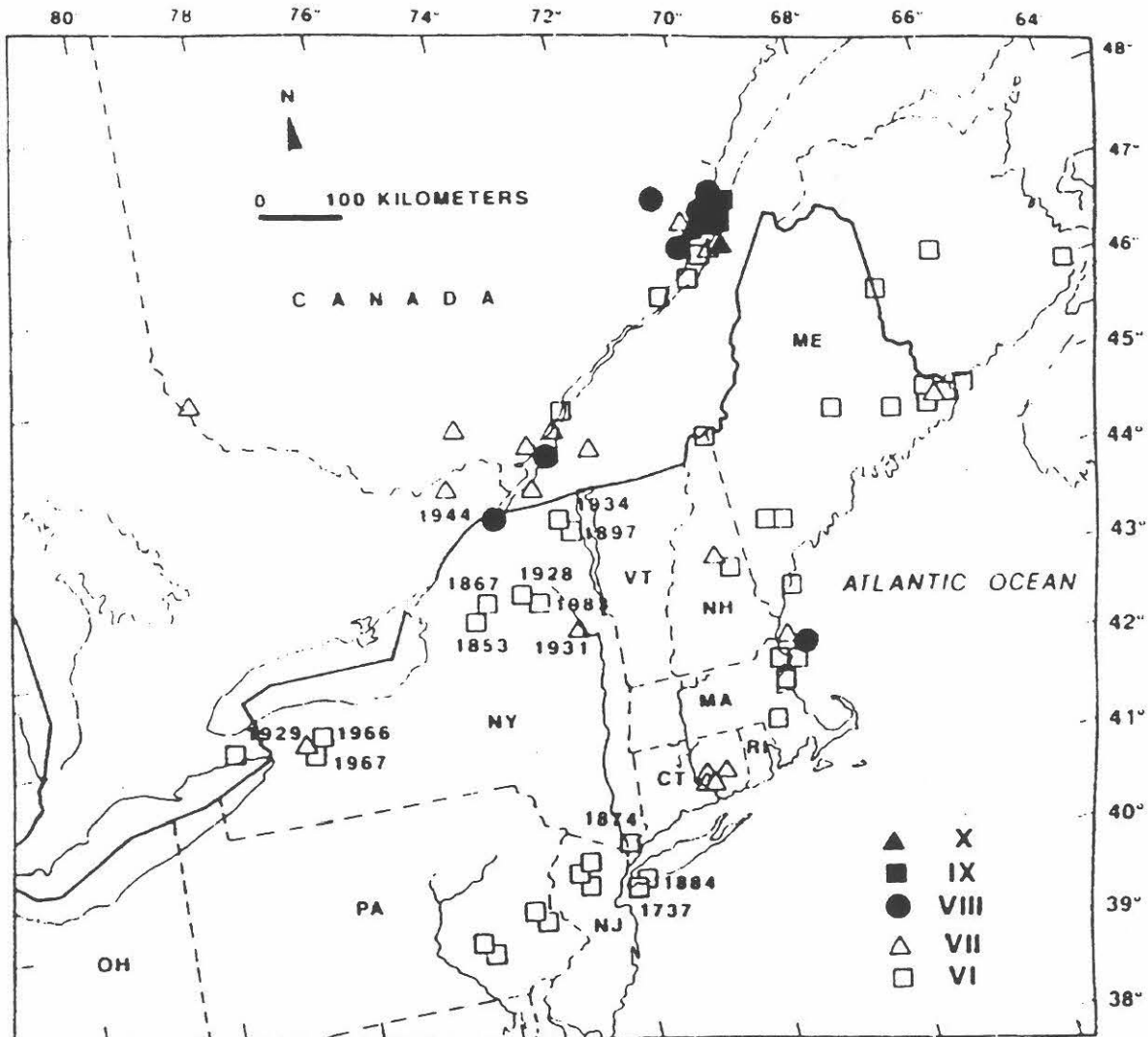


Figure 2. Damaging earthquakes of the northeastern United States and southeastern Canada during the period of 1534 to 1988. Symbols show the epicenter and epicentral intensity of each event. Epicentral intensities are given in terms of the Modified Mercalli Scale of 1931. The year of occurrence appears next to New York State events.

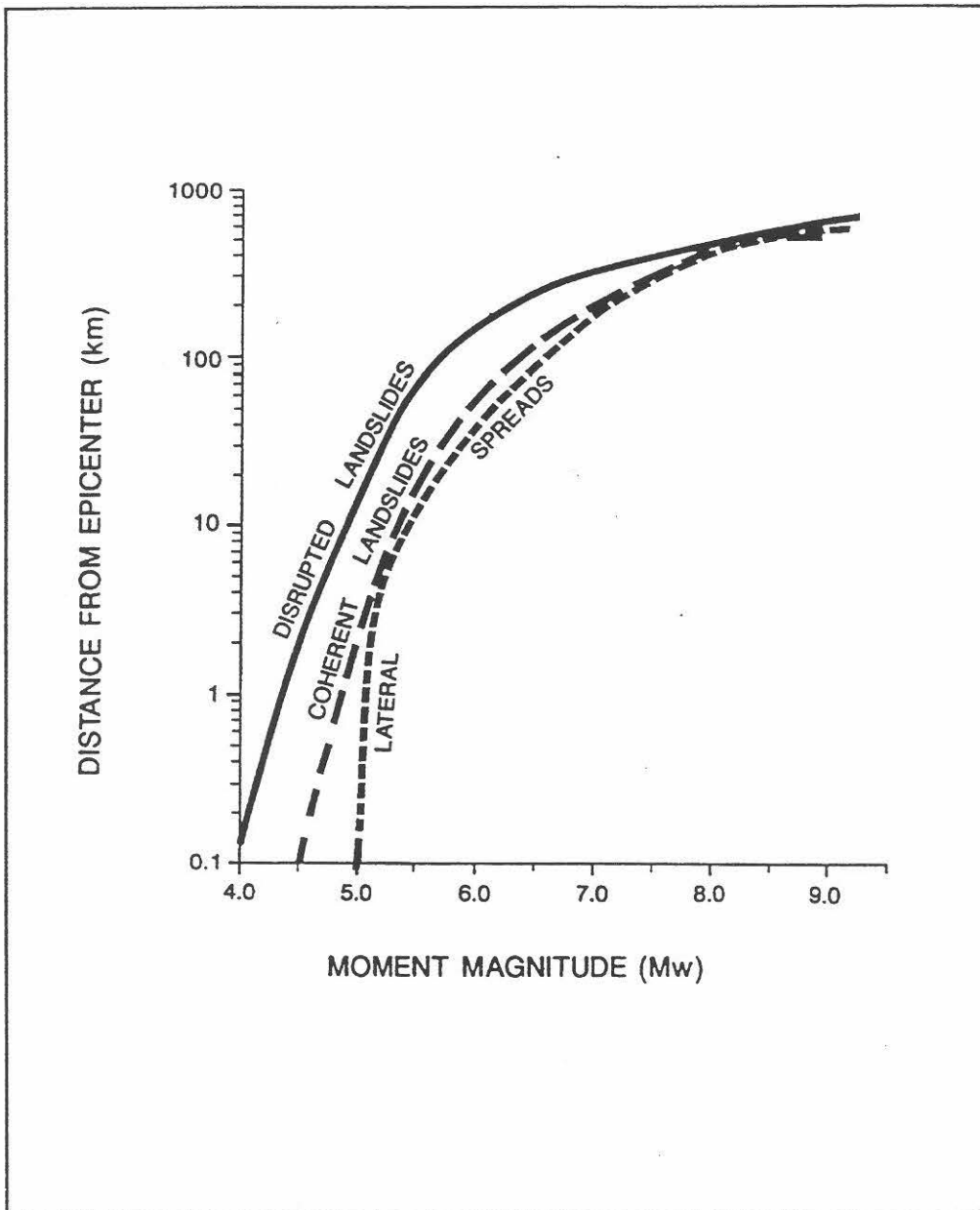


Figure 3. Relationship between earthquake magnitude and the maximum distance that three general types of landslides can occur. Earthquakes in the northeastern United States have and will trigger landslides.

Lateral-spread landslides and liquefaction are usually triggered by earthquakes of at least magnitude 5.0(m_b). The distance away from the epicenter of an earthquake where landslides occur, increases with the magnitude of the earthquake. Figure 3 summarizes data from Figure 1 in Keefer (1984). All of these hazards can result in increased building damage, damage to the infrastructure, and possibly, toxic material spills and fires.

EVALUATION PROCESS

Our first attempt to map ground motion response as a function of glacial landforms used expected values from an average, largest event that can be projected for New York State. That event would be a repeat of the magnitude 6.0(m_b) Massena Center, New York, earthquake of September 5, 1944. We assumed an epicentral distance of 10 kilometers from a Community. That earthquake at a distance of 10 kilometers, should produce a maximum horizontal acceleration of 0.35g (g = acceleration of gravity at the Earth's surface = 981 cm/sec^2) on bedrock. Acceleration values at the surface of the glacial materials were estimated using the program SHAKE91.

SHAKE91 assumes that a soil column can be represented as stacked elastic layers that are connected by springs. Each soil layer is defined by a thickness, shear modulus (obtained from soil tests or shear-waves), damping factor, and bulk density. A soil column is "shaken" using recorded or synthetic acceleration values for earthquakes, and with the assumption that the shaking is only coming from vertical propagation shear-waves. These are simplistic assumptions, but the results of this analysis have been substantiated in actual earthquakes.

The actual recordings of accelerations used in our analyses are from a magnitude 4.0 New Brunswick, Canada, earthquake of March 31, 1982, and from the November 25, 1988, magnitude 5.9, Saguenay, Quebec, earthquake. Eleven acceleration recordings were used, and they were scaled to match our magnitude 6.0(m_b) design event, at a distance of 10 kilometers. When using acceleration recordings, it is important to choose recordings of events that closely match the size, distance, and geological environment of the design earthquake. Earthquake accelerations measured on bedrock, are not frequently recorded in the northeastern United States and southeastern Canada.

To define the glacial landforms, their materials, and their engineering properties, we created or compiled the following databases:

1. 17 digitized surficial geologic 7.5 minute quadrangle maps; 1:24,000 scale;
2. digitized bedrock geologic map from 1970 NYSGS, Hudson-Mohawk Sheet, 1:250,000 scale;
3. water well inventory from 11,000 county property owners;
4. seismic refraction surveys at 50 sites, specifically including individual types of bedrock and surficial materials;
5. soil test data.

Representative soil columns were defined for each glacial unit. The soil column used to characterize glacial landforms is a one layer model over a bedrock half-space.

HIGHER RISK AREAS MAP

The final objective, to produce a map of the county illustrating the regions with a higher risk of damage from ground motion, landslides and liquefaction, is presented in Figure 4.

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**Areas of Higher Earthquake Hazard
Columbia County, New York
1994**

Explanation

■ Areas of landslide or liquefaction potential, combined with intensity IX (MM) or greater shaking. These areas were defined largely using expert opinion in conjunction with computer based geotechnical analysis.



5 0 5 Statute Miles
[Scale bar with markings]

Donald H. Cadwell, Project Manager
Digital data and map created by David A. Gerhard and Mark T. Davin

Prepared for SEMO by the New York State Museum Geological Survey

Figure 4. Areas of higher earthquake risk in Columbia County, NY.

SEISMIC HAZARD ASSESSMENT OF COLUMBIA COUNTY FIELD TRIP ROAD LOG

For the field trip, we will demonstrate a seismic line, using both horizontal and vertical geophones, for P- (primary) and S- (shear) waves in surficial materials. Discussion will be directed toward potential landslides and liquefaction in conjunction with varying watertable conditions. Secondly, we will visit the city of Hudson and discuss the application of seismic hazard data with county & municipal planners, and emergency personnel to potential problems in downtown Hudson: unreinforced structures, old wooden buildings, old stone buildings, and the impact of abundant water on lacustrine clays. Finally, we will demonstrate the use of this seismic-hazard data to county residents, and visit a house damaged as a result of land failure in 1983, coinciding with the 1983 Blue Mountain Lake, magnitude 5.1, earthquake, with the epicenter 180 kilometers away.

The Road Log for this trip starts on Route 9, at the Rensselaer County-Columbia County border. See Figure 5.

<u>ROUTE DESCRIPTION</u>	<u>MILES FROM LAST POINT</u>	<u>CUMULATIVE MILEAGE</u>
Rensselaer-Columbia County border. Proceed south on Route 9.	0.0	0.0
Railroad underpass.	0.2	0.2
Knickerbocker Lake, over hill to left. Kames, kettles and kettle lake.	0.65	0.85
Intersection with blinker light, continue south on Rt. 9.	0.25	1.1
Maple Lane Outwash sands and gravels, prograding southward to delta at Kinderhook, deposited in Glacial Lake Albany.	0.7	1.8
Intersection and light at Rts. 9 & 9H. Bear right onto Rt. 9H.	1.4	3.2
Intersection, light, Keegan Road. Continue south on 9H.	0.3	3.5
Bear right at exit for Kinderhook, Rt 9 South.	1.0	4.5
Bear right onto Rt. 9 south.	0.2	4.7
Turn right into Brosens Garden Market. Pull over to the right of the buildings, out of the way.	0.2	4.9

STOP 1 Demonstration of Seismic line

The Pleistocene equivalent to the Kinderhook Creek deposited the large delta complex into Glacial Lake Albany 18,000 to 15,000 years ago. Abundant meltwater was transported from both the retreating Wisconsinan glacier terminus and the bedrock uplands toward the delta. The present channel of the Kinderhook Creek is incised 17-20 m (50-60 ft) into the delta (Figure 6) and the location of this seismic line is on the floodplain. Water well information suggests bedrock is 12-24 m (35-55 ft) beneath the floodplain.

<u>ROUTE DESCRIPTION</u>	<u>MILES FROM LAST POINT</u>	<u>CUMULATIVE MILEAGE</u>
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Return to Rt. 9, turn left, north.

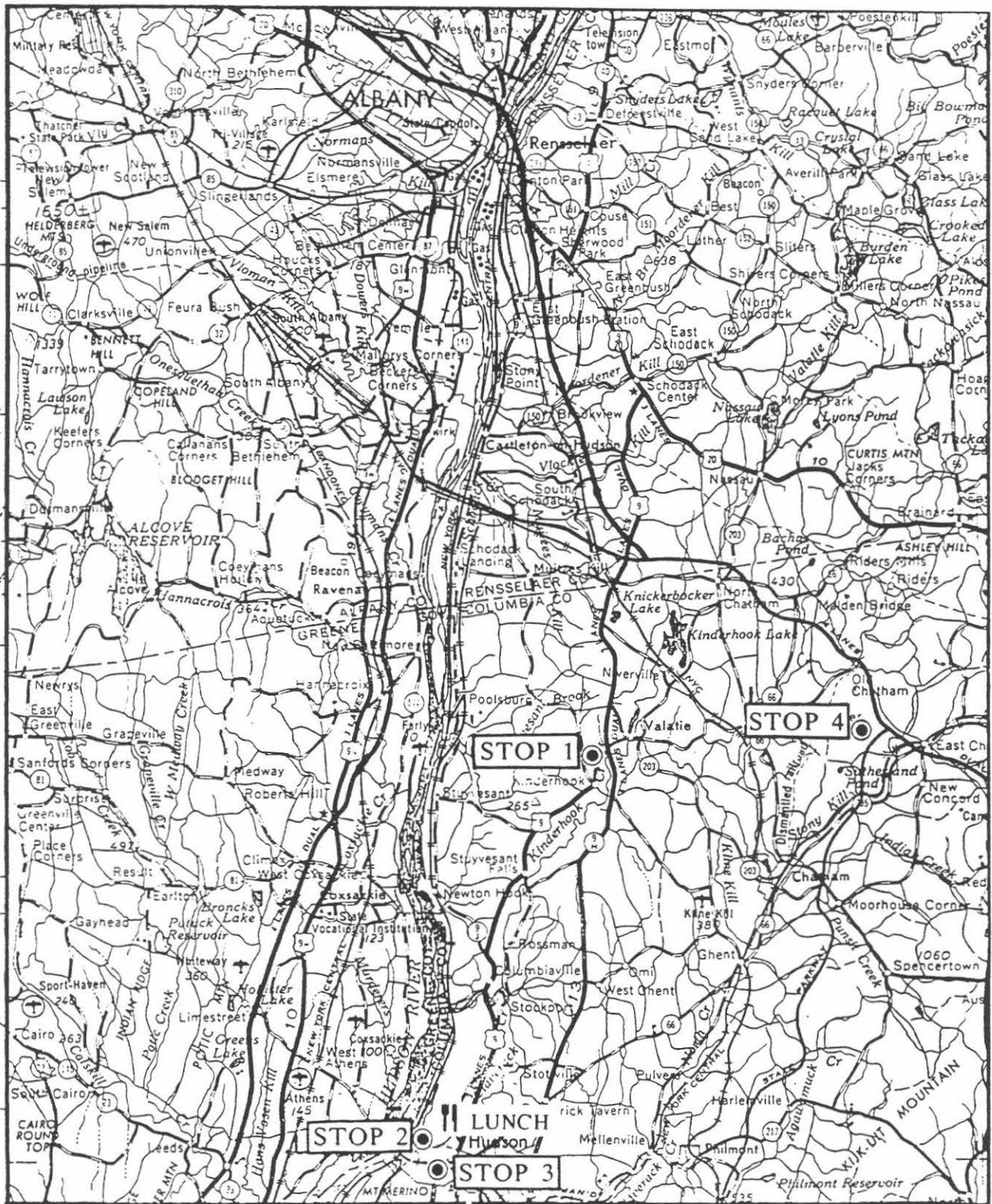
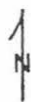


Figure 5. Schedule and location of fieldtrip stops.

0 1 Mile



Map Location

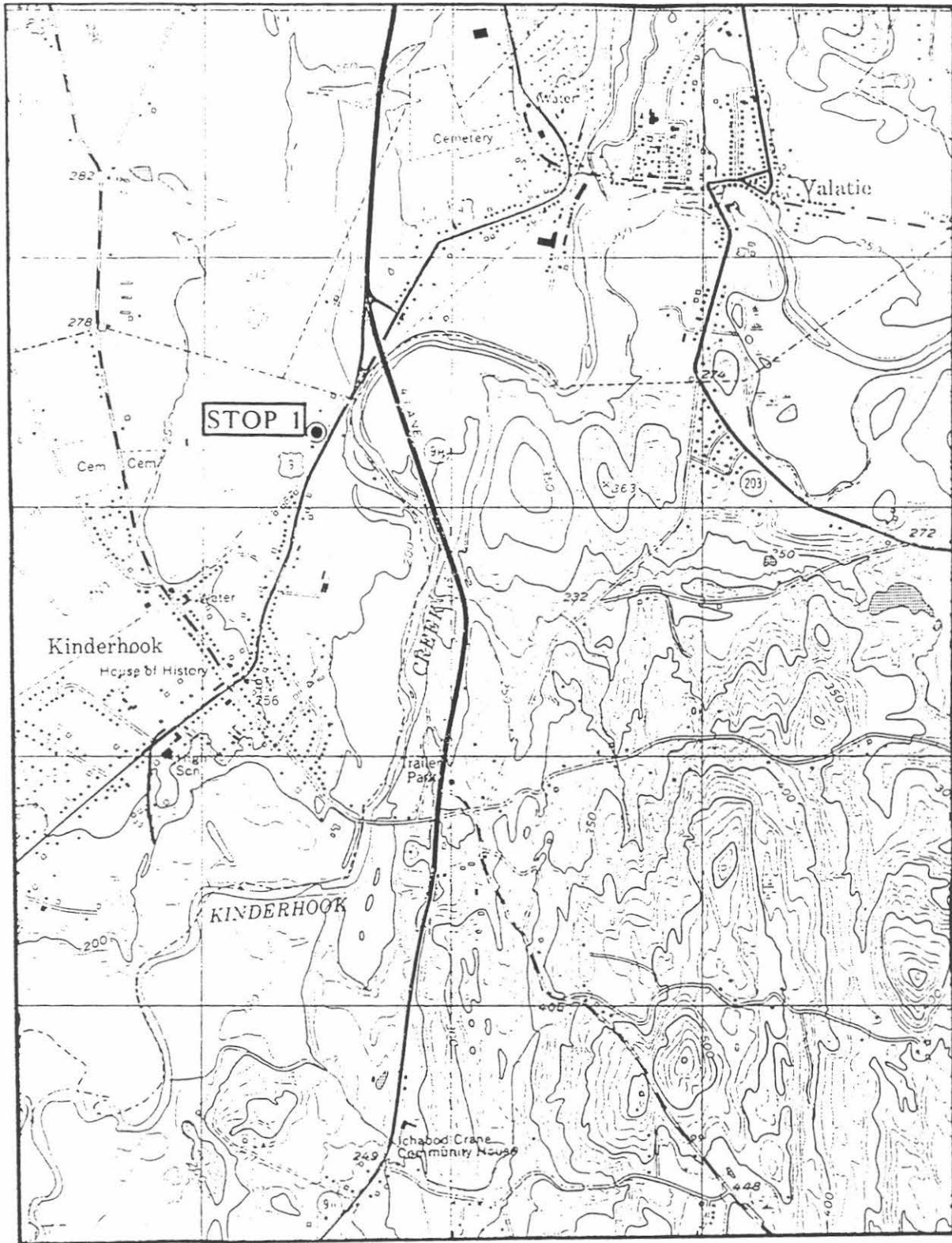
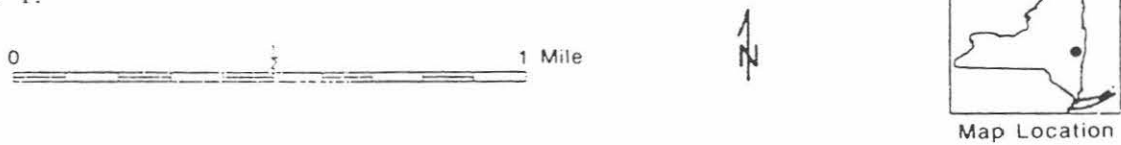


Figure 6. Portion of the Kinderhook 7.5 minute topographic quadrangle indicating location of STOP 1.



ROUTE DESCRIPTION	MILES FROM LAST POINT	CUMULATIVE MILEAGE
Intersection 9H, turn left toward 9H south.	0.1	5.0
Bear right onto 9H south (4 lane highway).	0.3	5.3
Rt. 9H reduces to 2 lane highway.	1.2	6.5
Martin Van Buren Historic Site.	1.7	8.2
Columbia County Airport.	5.0	13.2
Light at Intersection Rts. 9H and 66. Turn right onto Rt. 66, towards Hudson.	1.5	14.7
Cross Claverack Creek and enter Greenport.	1.9	16.6
Light at Healy Blvd. Continue straight on 66.	0.3	16.9
Cross railroad tracks.	0.4	17.3
Enter City of Hudson.	0.1	17.4
Light at junction with Rt. 23b. Turn right onto Rt. 23b.	0.3	17.7
Light at junction of Rts. 23b and 9. Continue straight on Rts. 9 south and 23b.	0.1	17.8
Light at intersection with State Street. Turn Right. Continue on State until 2nd Street.	0.4	18.2
Turn right onto 2nd Street. As we descend the hill, note the old marsh ahead of us.	0.8	9.0
At STOP sign, turn right onto Mill Street.	0.2	19.2
Continue to end of road.	0.1	19.3

STOP 2: Clay Pit

This clay pit, shown in Figure 7, is owned by the City of Hudson, and permission should be obtained prior to visiting the site. Today we will be able to see a small remnant of the original working area. These alternating bands of dark clay and lighter colored silt-clay are rhythmites, rhythmically deposited fine-grained lacustrine sediment. If the dark and light couplet represent an annual winter and summer accumulation, then they are varves. This 10m exposure has couplets 2-5cm thick and could represent 200-300 years of sediment accumulation into Lake Albany. When wet, these clays are very plastic.

1. What do you suppose would happen during an earthquake to the dams of Underhill and Oakdale ponds?
2. What would be the fate of these clay banks?

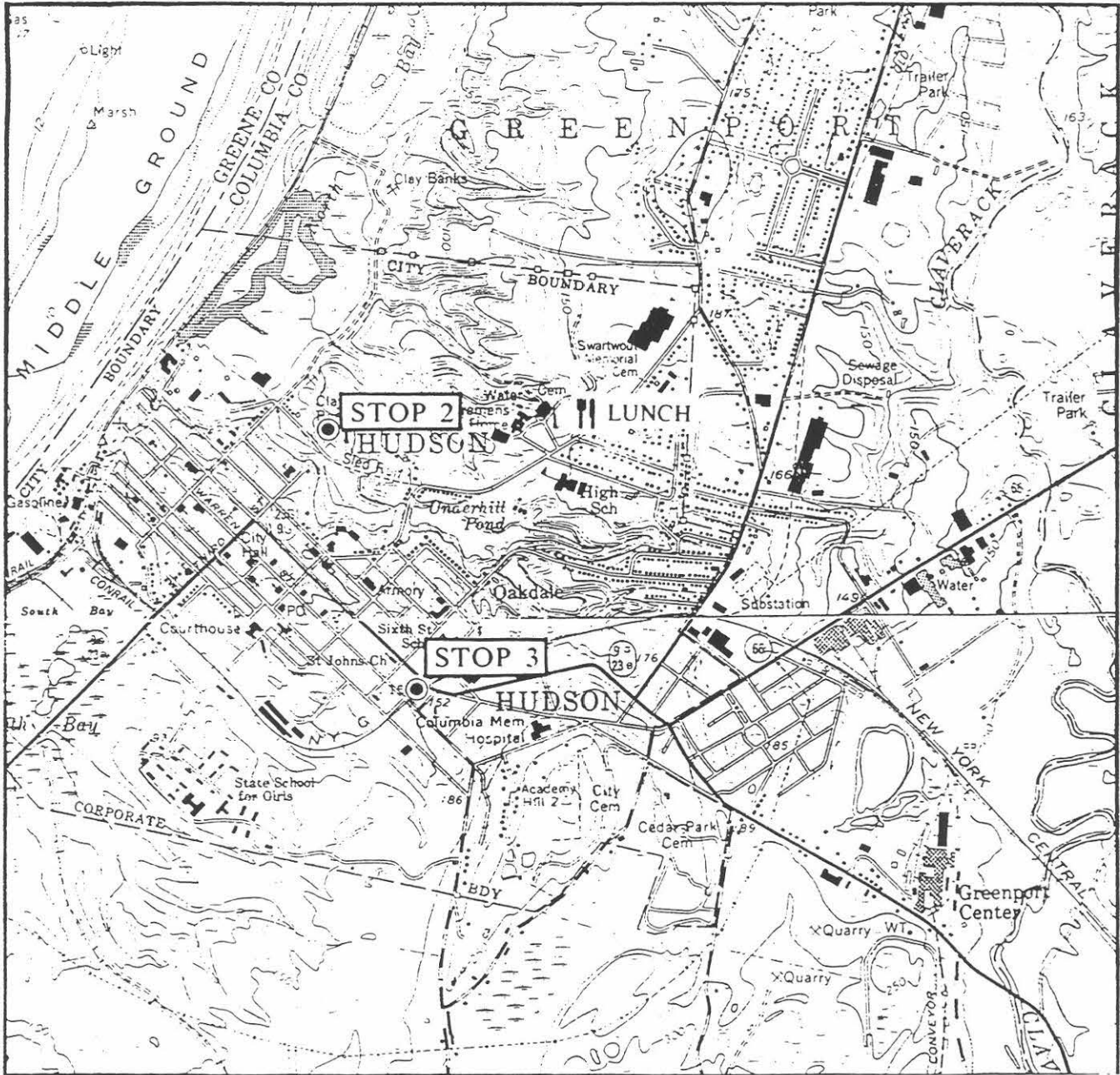


Figure 7. Portion of Hudson North and Hudson South 7.5 minute topographic quadrangles indicating location of STOP 2, STOP 3, and LUNCH.



ROUTE DESCRIPTION	MILES FROM LAST POINT	CUMULATIVE MILEAGE
Turn around and proceed back to 2nd Street. Note the landfill to the right. Turn left onto 2nd Street.	0.1	19.4
Turn left onto State Street.	0.2	19.6
Turn left onto Carroll Street.	0.4	20.0
Turn left onto Short Street. This road becomes Harry Howard Avenue.	0.1	20.1
Note Underhill Pond to right. This is a man-made pond within the gully.	0.1	20.2
Turn left into entrance for Fireman's Home.	0.3	20.5

LUNCH

Note: To eat lunch at the Pavilion it is necessary to obtain permission in advance.
If time permits, the Fireman's Museum is excellent.

Return to entrance of Firemen's Home.
Turn right onto Harry Howard Avenue.

Turn right onto Carroll Street.	0.5	21.0
At light, bear right on State Street.	0.1	21.1
At light, turn left onto 3rd Street.	0.2	21.3
Turn left onto Columbia Street.	0.1	21.4
Turn right on 7th Street. PARK on side of road, Center Square Park.	0.5	21.9

STOP 3: Discussion of buildings, etc., City of Hudson.

See Figure 7 for specific location. Discussion will concentrate on types of building structures, and location of emergency services.

ROUTE DESCRIPTION	MILES FROM LAST POINT	CUMULATIVE MILEAGE
Leave Center Square Park, turn left on Warren Street.		
At blinker and stop sign, turn left on Prospect Street. Pass Columbia Memorial Hospital	0.2	22.1
At blinker and stop sign, continue straight.	0.2	22.3

<u>ROUTE DESCRIPTION</u>	<u>MILES FROM LAST POINT</u>	<u>CUMULATIVE MILEAGE</u>
At yield sign, bear right on Columbia Street.	0.05	22.35
At light at Green Street. Continue straight on Rt. 66	0.15	22.5
Enter Town of Claverack.	1.0	23.5
Light and intersection Rts. 66 and 9H. Continue on Rt. 66 toward Chatham.	1.9	25.4
Enter Hamlet of Ghent. Continue on Rt. 66.	5.8	31.2
Enter Village of Chatham.	5.8	37.0
Light and intersection Rts. 66 and 203.	0.6	37.6
Intersection Rts. 66 and 295. Bear right onto Rt. 295. Do not cross railroad tracks.	0.4	38.0
Underpass for Taconic State Parkway.	2.5	40.5

NOTE: This road log has stopped several miles from the property owner of STOP 4 because this is a private residence and the NYSGA trip has been given special permission to visit. The owners prefer not to be disturbed by an onslaught of people in the future.

STOP 4: Home damaged by land slump.

The damage resulted from land failure on October 7, 1983, coinciding with the 1983 Blue Mountain Lake, magnitude 5.1, earthquake, with the epicenter 180 kilometers away.

<u>ROUTE DESCRIPTION</u>	<u>MILES FROM LAST POINT</u>	<u>CUMULATIVE MILEAGE</u>
At the end of the stop turn around and return to the Taconic Parkway and I-90 access road.		
Intersection with I 90 access road. Turn onto access road for I 90.	0.0	0.0
Stop, get ticket for I 90. Keep LEFT onto I 90 Westbound.	0.7	0.7
Exit at Exit B1, for Albany, Hudson, Rt. 9 and I 90. 8.0 8.7		
Pay toll (cars \$ 0.30 in 1995).	0.5	9.2
<u>Keep right</u> if you want to exit onto Rt. 9. Then turn left on 9 to return to Columbia- Rensselaer County border and start of road log.		

Keep left for I 90 to Albany, Schenectady and to return to Union College.

End of Fieldtrip.

WOLLASTONITE DEPOSITS OF THE NORTHEASTERN ADIRONDACKS

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INTRODUCTION

The presence of wollastonite near Willsboro in the northeastern Adirondacks (Fig. 1, 2) has been known since the early nineteenth century. The earliest reference to it in the geologic literature is by Vanuxem (1821). For over a century, the wollastonite was of little interest except as a mineralogical curiosity. Mining on a small scale began at Fox Knoll near Willsboro in 1938, with the wollastonite being used as a flux for arc welding. In 1951, the Cabot Corporation gained control, and began underground mining in 1960. Interpace Corporation took over and expanded operations in 1969. Product development resulted in uses in ceramic bodies and glazes, as a reinforcing filler in plastics and resins, and as a substitute for short-fiber asbestos. The operation, now known as NYCO, was purchased in 1979 by a subsidiary of Canadian Pacific (US), Processed Minerals Inc. Open pit mining at the Lewis (Seventy Mountain) Mine, ten miles southwest of Willsboro, began in 1980 and in 1982 the underground operation at Willsboro was closed. Both properties are now owned by NYCO Minerals, Inc., a subsidiary of Fording Coal Company of Calgary, Alberta.

The Willsboro deposit was mentioned briefly by Buddington (1939, 1950) and Buddington and Whitcomb (1941); the geology is given in more detail by Broughton and Burnham (1944). Putman (1958) described several occurrences of wollastonite in the Au Sable Forks and Willsboro quadrangles, including those at Willsboro, Deerhead, and Lewis (Figure 2). De Rudder (1962) studied the mineralogy and petrology of the Willsboro ores, and attributed them to contact metamorphism with localized alumina metasomatism. Oxygen isotope work by Valley and O'Neil (1982) demonstrated extensive metasomatism involving meteoric water.

GEOLOGIC SETTING

The Westport metanorthosite dome (Figures 1 and 2) is located east and north of the Marcy Massif. It is overlain on its north and west flanks by interlayered granulite facies metaigneous and metasedimentary gneisses, marbles, and calcsilicate rocks. The wollastonite deposits at Willsboro and Lewis, as well as two undeveloped prospects at Oak Hill and Deerhead, occur within a mappable zone up to 2000 feet thick that extends for at least 14 miles along strike (Figure 2). This ore-bearing zone (OBZ) is characterized throughout by intense foliation and locally prominent lineation. Along the northern flank of the Westport Dome from Willsboro mine to Deerhead, the OBZ directly overlies the metanorthosite of the dome, foliations dip NNE away from the dome, and lineations plunge NW. Southwestward, near Oak Hill and the Lewis mine, dips flatten and lineations become parallel with the regional NNE trend (Whitney and Olmsted, 1993). In this area the OBZ diverges from the dome, although the thickness of intervening rocks is uncertain due to poor exposure and the unknown subsurface configuration of the anorthosite.

Metaigneous rocks within the OBZ occur as sheets and lenses parallel to foliation, emplaced either as sills or as tectonic slivers. They include gabbroic and anorthositic gneisses, amphibolite, and minor charnockite. Interiors of thick gabbroic layers may display relict igneous textures. Metasedimentary rocks consist chiefly of the wollastonite ores, associated garnet-pyroxene skarn, and a diverse suite of granular-textured garnet-clinopyroxene-plagioclase rocks, with minor sphene and apatite. Calcite marble occurs locally, as do very minor amounts of quartzite and metapelite.

The ore at all four known locations occurs as tabular bodies ranging from a few feet up to as much as 80 feet thick. Multiple wollastonite-bearing horizons, separated by gabbroic or anorthositic gneisses and amphibolite, are present at Willsboro (DeRudder, 1962). The orebodies consist of wollastonite-rich ore with garnet-pyroxene skarn

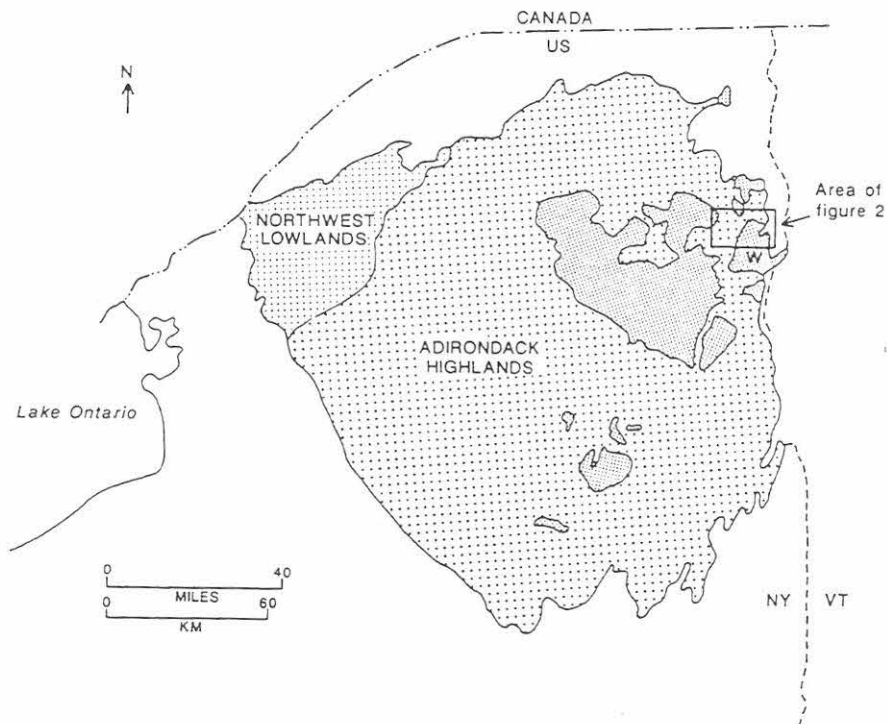


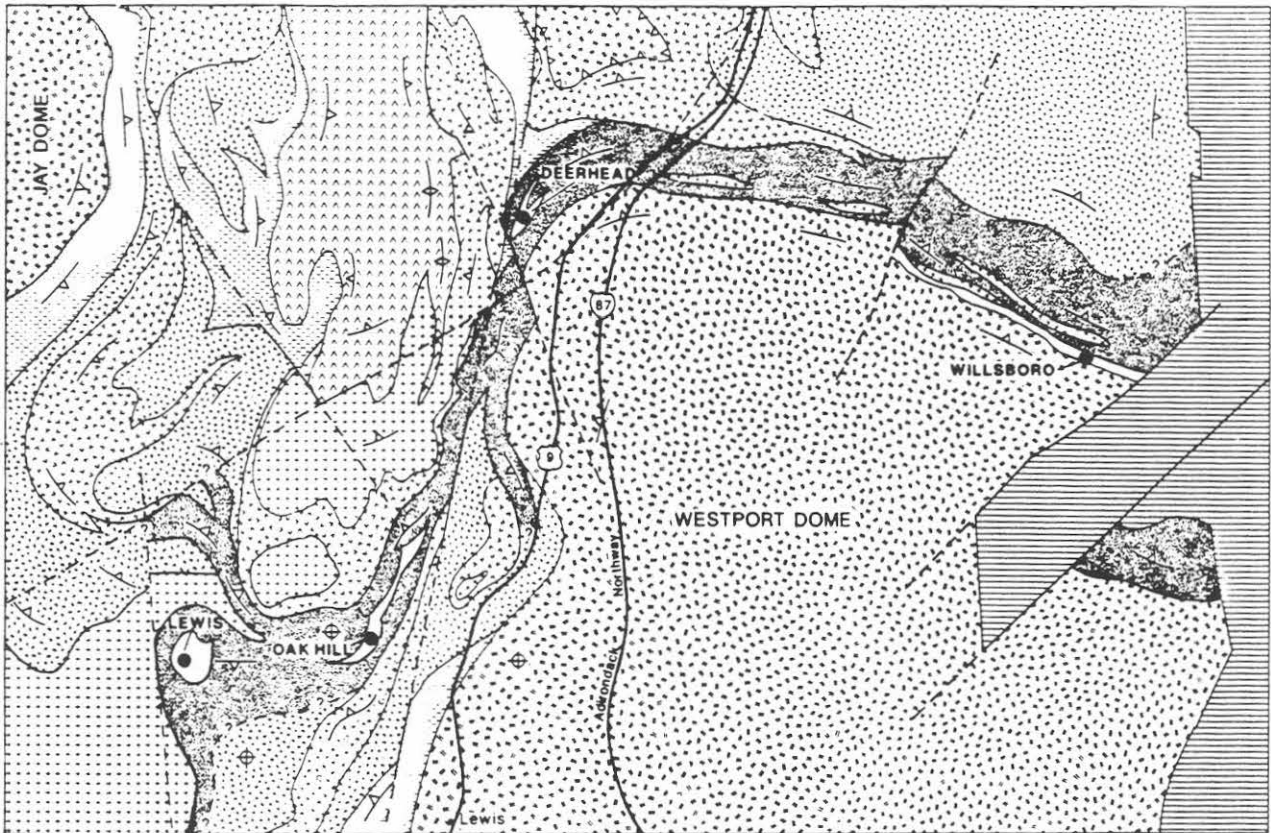
Figure 1. Location map showing the Adirondack region, with major anorthosite bodies in fine stippled pattern. W: Westport Dome. Rectangular outline: area of this trip.

(GPS) layers and lenses ranging from less than an inch to several feet thick. This compositional layering is ordinarily straight and sharply defined; it is probably not an original sedimentary feature but rather a result of tectonically induced metamorphic differentiation during or subsequent to ore formation. More diffuse compositional layering and foliation within the ore locally exhibits complex folding. Where layering is less prominent, garnet and pyroxene may occur in clusters or lenses up to 2 inches across.

MINERALOGY

The ore layers contain the high-variance assemblage wollastonite-grandite garnet-clinopyroxene. Traces of retrograde calcite occur as thin films replacing wollastonite along fractures and grain boundaries. GPS layers within the ore consist chiefly of garnet and clinopyroxene with or without minor wollastonite. Another type of GPS, containing up to several percent of sphene and apatite, occurs at contacts between ore and metaigneous gneisses or amphibolites and, less commonly, as sill- or dike-like bodies within the ore. Minor and trace minerals occurring very locally in GPS include scapolite, plagioclase, clinozoisite, idocrase, and zircon. Discontinuous layers up to several feet thick of nearly pure garnet, or garnet with minor plagioclase and quartz are present at some ore/gneiss contacts. These "garnetites" pinch and swell along strike or form detached lenses that resemble boudins.

Compositions of garnets and pyroxenes in ore and GPS were determined by electron microprobe. Standard polished thin sections were used for these analyses where possible; where the ore was too friable, grain mounts were prepared from hand specimens or 2-4 inch segments of drill core. The pyroxenes lie close to the diopside-hedenbergite join, containing >93% (Di + Hd), with acmite (up to 3.2%) as the most common minor component. The garnets are grossular-andradite mixtures, with >92% (Gr + Ad); almandite (up to 4.9%) and schorlomite (up to 3.1%) are the dominant impurities. Figure 3 shows the range of compositions for the ore and GPS. Compositional variation among grains within a sample can be as great as 20% Ad and 10% Hd for garnet and pyroxene respectively, but individual grains lack detectable internal zoning.



EXPLANATION

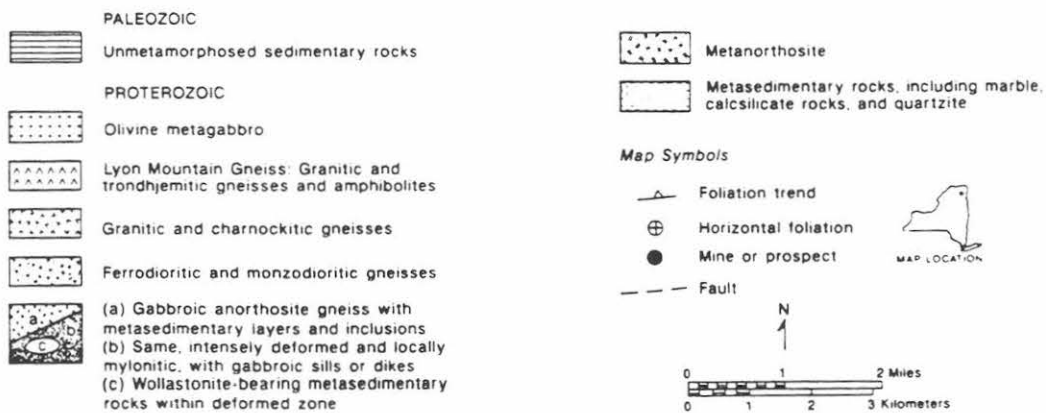


Figure 2. Simplified geological map of the northeastern Adirondack wollastonite district, after Buddington and Whitcomb (1941) and Whitney and Olmsted (1993). The heavy stippled pattern designates the zone of strongly foliated and lineated rocks containing the wollastonite mines and prospects (OBZ). Areas of predominantly metasedimentary rocks within the zone are blank.

GEOCHEMISTRY

Rare Earth Elements (REE)

Thirty-three samples of ore and 20 of GPS from the Willsboro and Lewis deposits were analyzed for rare earth elements (REE) by inductively coupled plasma mass spectrometry (ICPMS). REE were also determined in several metaigneous rocks within and near the ore zone, and in 8 marbles from the northeastern Adirondack area. The analyses reveal a variety of REE distributions in ore and GPS. Three general types can be distinguished; curves A-C in Figure 4a are the averages for these types. All but three of the 30 wollastonite-rich ores display the A pattern; those three have flat or slightly negative Eu anomalies. A similar distribution (A') is found in concordant GPS layers in the ore. A-type REE patterns are associated with relatively andradite-rich garnet (Table 1). The B pattern is typical of sphene and apatite-bearing GPS. Type C distributions occur in a few samples of both lean ore (<50% wollastonite) and sphene-free GPS. Both B and C patterns are correlated with andradite-poor garnet (Table 1).

TABLE 1: MINERAL COMPOSITIONS AND REE DISTRIBUTION

REE pattern	A		A'		B		C	
	AVG RANGE		AVG RANGE		AVG RANGE		AVG RANGE	
Garnet (% Ad)	51	25-93	58	35-83	23	14-41	27	22-33
Pyroxene (% Hd)	43	30-56	38	32-52	30	15-46	40	34-42
n	19		4		6		4	

Major and trace elements

X-ray fluorescence was used to determine the major- and trace- element chemistry of the ores and GPS. Twenty-one samples of ore and 18 of GPS from the Willsboro and Lewis deposits were analyzed. Table 2 shows the average compositions of samples from the four groups defined by REE patterns.

TABLE 2: MAJOR AND TRACE ELEMENTS

REE pattern	TYPE A	TYPE A'	TYPE B	TYPE C
n	18	5	9	7
SiO ₂	47.85	40.85	42.84	44.38
TiO ₂	0.18	0.43	1.94	0.45
Al ₂ O ₃	2.45	6.86	9.22	9.69
Fe ₂ O ₃ (Tot)	6.74	16.06	9.05	8.39
MnO	0.17	0.17	0.21	0.28
MgO	1.33	2.24	3.76	2.41
CaO	40.68	32.92	31.61	33.94
Na ₂ O	0.14	0.21	0.22	0.09
K ₂ O	0.00	0.01	0.01	0.00
P ₂ O ₅	0.01	0.04	0.97	0.05
Rb (ppm)	< 1	< 1	< 1	< 1
Sr	41	10	45	24
Ba	2	2	2	4
Zr	43	108	319	73
Y	7	13	61	79
Nb	2	3	13	2
Ga	8	20	15	15

ORIGIN OF THE ORES

Metasomatism

Wollastonite ordinarily occurs as a contact metamorphic mineral formed by reaction of calcite and quartz. However, the Willsboro-Lewis ores show metasomatism on a large scale. The evidence is outlined in the following four sections.

Mineral assemblages and compositions. If the wollastonite ore had been formed by isochemical contact metamorphism, the absence of either quartz or primary calcite would imply a protolith with precisely the right balance of quartz and calcite. This highly improbable requirement, together with the high variance of the ore mineral assemblage, indicates that metasomatism has occurred (Valley and O'Neil, 1982).

Metasomatic wollastonite, grandite garnet, and pyroxene are commonly observed in hydrothermal skarn deposits (e. g. Einaudi et al., 1981). Zoning is a common feature of hydrothermal garnets (Meagher, 1982), and may arise from variations in pressure and temperature, or changes in fluid composition (Lee and Atkinson, 1985). Varying proportions of andradite and grossular in originally zoned grains may account for much of the variability of garnet compositions in the Willsboro-Lewis ores and GPS (Figure 3). If early, contact metamorphic grossular was present, it is likely that hydrothermal garnet would nucleate on it to form composite crystals. The present lack of zoning in either garnet or pyroxene grains can be attributed to internal homogenization during subsequent granulite-facies metamorphism.

Oxygen isotopes. Valley and O'Neil (1982) determined oxygen isotopes in both the Willsboro and Lewis deposits. They found $\delta^{18}\text{O}_{\text{SMOW}}$ in the wollastonite ore from -1.3 to 7.0‰; as much as 25‰ lower than typical Adirondack marbles. Sharp gradients occur between ore and wall rocks. They (Valley and O'Neil, 1982) showed that the $\delta^{18}\text{O}$ data could not be explained by isotopic fractionation during devolatilization reactions, but required exchange with large volumes of heated meteoric waters at the time of anorthosite intrusion. Valley (pers. comm. 1992) analyzed four of our samples. Three ore samples yielded $\delta^{18}\text{O}$ values of +0.7, +0.7, and +1.3‰, and a GPS sample measured +2.9‰, in agreement with the results of Valley and O'Neil (1982). Eight ore samples from their work were included in this study and yielded type A REE patterns.

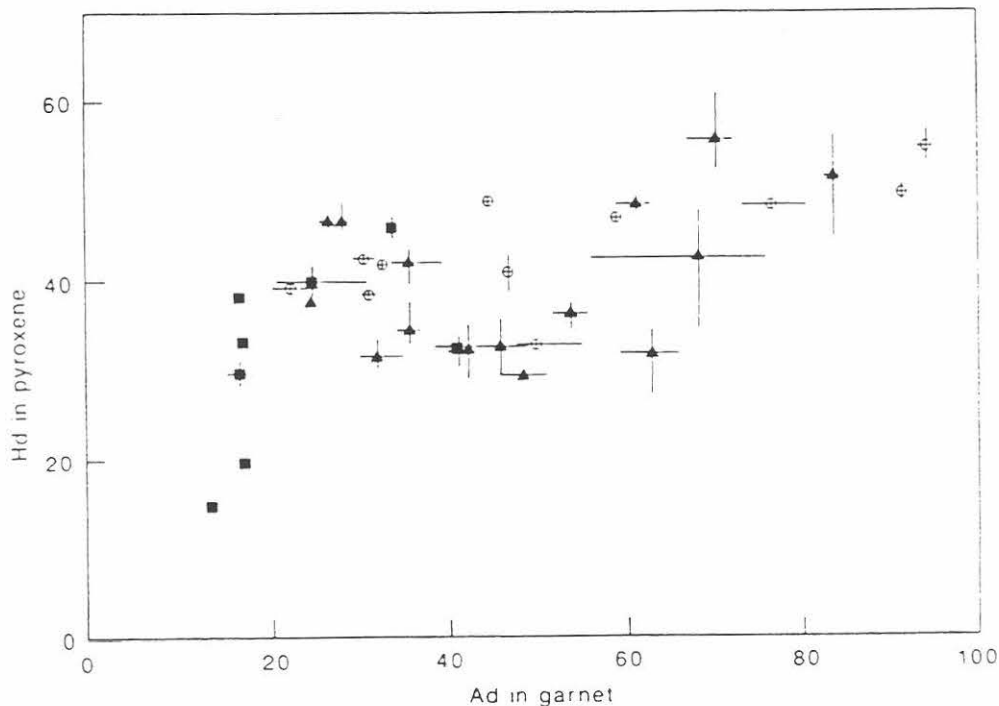


Figure 3. Garnet and pyroxene compositions in wollastonite ores and garnet-pyroxene skarns (GPS) as determined by electron microprobe. Bars show the range of mineral compositions within a thin section or 2-4 inch section of core; individual garnet and pyroxene grains are unzoned. Triangles: Willsboro ore. Open circles: Lewis ore. Filled squares: GPS.

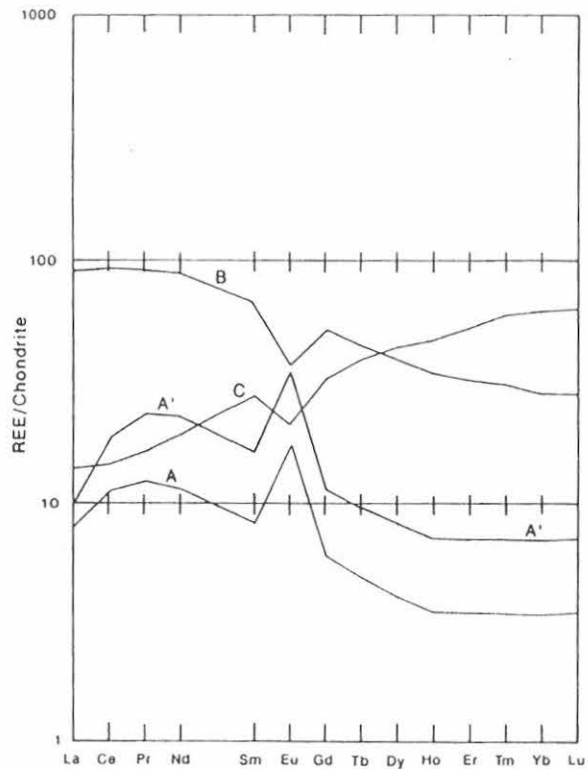


Figure 4A

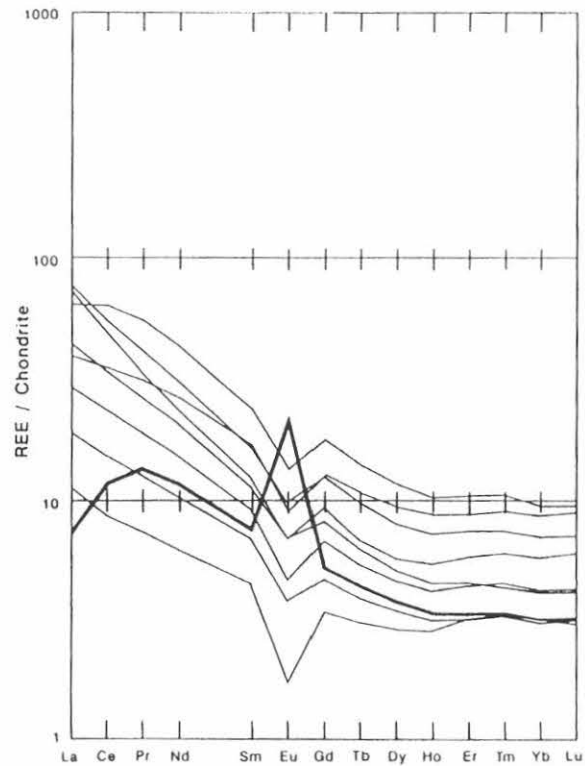


Figure 4B

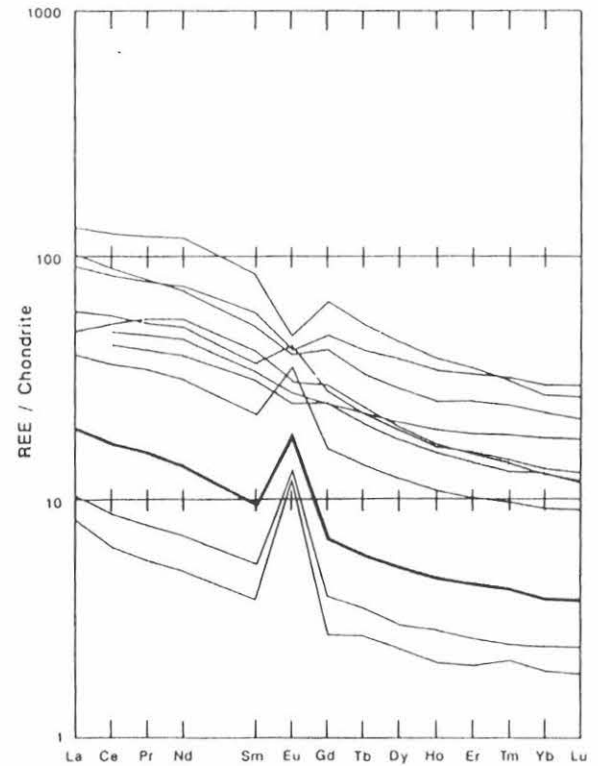


Figure 4C

Figure 4. (a) Chondrite-normalized REE distribution in ore and (GPS). A: Average of 30 wollastonite-rich ores. A': Average of 6 conformable, sphene-free, high-andradite GPS layers in ore. B: Average of 10 sphene-bearing GPS. C: Average of 7 lean ores and sphene-free, low-andradite GPS. Letters correspond to the REE patterns discussed in text.

(b) REE distribution in northeastern Adirondack marbles and siliceous marbles. Heavy line: average wollastonite ore.

(c) REE distribution in metaigneous rocks in and near the OBZ. Upper lines are gabbros, amphibolites, and gabbroic anorthosites; lower lines are anorthosites; lowermost line is Westport Dome anorthosite approximately 10 m below the contact with the ore zone at Willsboro. Heavy line: average of 24 anorthosites and gabbroic anorthosites of the Marcy Massif.

The isotopic requirement for fluids of **meteoric** origin is consistent with the absence of nearby felsic or intermediate intrusive rocks to provide a source of magmatic fluids. Anorthosite and gabbro, both of which have anhydrous primary mineral assemblages, are unlikely to have been the source of large volumes of fluids. In addition, the andradite-rich garnet in the ores and the absence of the graphite and iron sulfides commonly present in Adirondack marbles indicate relatively oxidizing conditions, also consistent with meteoric fluids.

Depletion of Na, K, Rb, Ba, and Sr. Only those elements that can be accommodated in the structures of wollastonite, garnet, and pyroxene are present in significant concentrations in the ores and GPS (Table 2). This is particularly clear for the large-ion lithophile elements (LILE) K, Rb, and Ba, which are present in only negligible amounts. Comparison with Adirondack marbles shows levels of LILE in the ores significantly depleted from a hypothetical marble protolith (Table 3), suggesting metasomatic removal. Strontium is also depleted in the ores relative to marbles, although to a lesser extent. Some of the Sr in the ores is present in secondary calcite and may have been introduced subsequent to ore formation. LILE, Na, and Sr are also depleted by one or more orders of magnitude relative to the levels present in the host gneisses and amphibolites.

TABLE 3: INCOMPATIBLE ELEMENTS IN ORES AND MARBLES

	Marbles NW Adirondacks ¹	Marbles NE Adirondacks	Wo ores
K (ppm)	1594	-----	59
Ba (ppm)	227	170	2
Sr (ppm)	356	-----	41

¹ Data from K. Hauer, (Pers. Comm., 1993)

REE distribution. Assuming a carbonate protolith for the ore, comparison of the REE distributions in ore and GPS (Fig. 4a) with those of northeastern Adirondack marbles (Fig. 4b), confirm substantial metasomatic redistribution of REE. The REE patterns may be influenced by numerous factors including mineral compositions and modal abundance, protolith composition, fluid composition, fluid-rock ratios, and kinetics. Analyses of mineral separates suggest that Type A and A' patterns are largely controlled by the composition of garnet, which contains most of the REE in the ores. Both the size of the Eu anomaly and the Ce/Yb ratio are highly correlated with the andradite content of garnet, probably because the relatively large Eu^{+2} and light rare earth (LREE^{+3}) ions are more readily accommodated in the larger X sites of andradite relative to grossular (Novak and Gibbs, 1971). Interaction of the metasomatizing fluids with strongly Eu-positive anorthosite (Fig. 4c) may also have contributed to the Eu anomaly. The largest REE ions, La and Ce, tend to be excluded even from andradite, resulting in their depletion relative to Pr and Nd. Concentration of garnet from the ore into conformable GPS layers by tectonically induced metamorphic differentiation produces the similar A' pattern.

The very different REE distribution (Type B) found in sphene- and apatite-bearing GPS probably resulted from localized metasomatism of mafic igneous rocks in contact with ore; compare the B pattern in Figure 4a with the REE distribution in mafic metagneous gneisses in the ore zone (Fig. 4c). In these rocks, LREE are retained in sphene and apatite while the heavy rare earths (HREE) remain in the relatively grossularitic garnet. The third major type of REE distribution (C) may originate from contact metamorphic calcsilicates and siliceous marbles containing abundant grossularitic garnet. The garnet retains the HREE and negative Eu anomaly of the protolith; subsequent overgrowths of metasomatic andradite from REE-poor hydrothermal fluids are insufficient to obscure the earlier pattern.

Sequence of ore-forming events

Origin of the ores by hydrothermal metasomatism requires a heat source to provide the minimum temperatures (ca. 450°C) for formation of wollastonite and to drive the hydrothermal circulation. This requirement, and the close spatial association between the ores and the Westport Dome (Fig. 2) strongly indicate that the ore is

coeval with emplacement of the anorthosite, in agreement with the conclusions of earlier workers (Buddington 1939, 1950; Broughton and Burnham 1944, DeRudder 1962). Moreover, access of large volumes of dominantly meteoric fluids implies a relatively shallow depth of emplacement (Valley and O'Neil 1982, Valley 1985). Access of fluids would also be facilitated in an extensional tectonic setting. Massif anorthosites are widely believed to be associated with extensional tectonics (Ashwal, 1993). Whitney and Olmsted (1993) and Fakundiny and Muller (1993) have argued that the Adirondack anorthosites were emplaced in an extensional setting that included large listric or detachment faults. We speculate that the present OBZ was the locus of one or more such faults. A similar association of extensional faulting with magmatic doming has been proposed by Lister and Baldwin (1993) for some metamorphic core complexes. Major low-angle extensional faults can provide channels for circulating hydrothermal fluids (Reynolds and Lister, 1987; Kerrich and Rehrig, 1987). When the Westport Dome was emplaced, hydrothermal circulation driven by heat from the intrusive may have followed the low-angle faults, fed from the surface by meteoric water penetrating along associated high-angle normal faults (Figure 5). Where the faults intersected or followed reactive carbonate units, infiltration metasomatism produced wollastonite and andraditic garnet, accompanied by exchange of REE and oxygen isotopes.

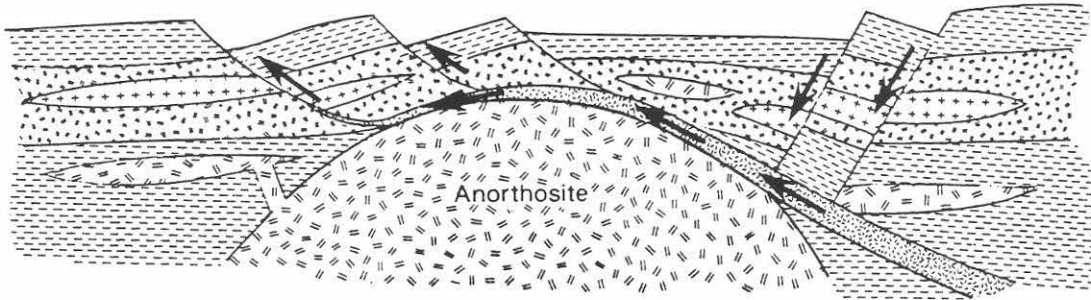


Figure 5. Cartoon illustrating the hydrothermal system inferred to be responsible for the wollastonite ores. Low-angle extensional faults developed coincident with anorthosite intrusion, and coalesced to form a zone along the flanks of the anorthosite dome. Surface-fed hydrothermal circulation (heavy arrows) in the fault zone, driven by heat from the intrusion, formed ore by infiltration metasomatism where the zone intersected or followed carbonate units.

Continuing or subsequent deformation produced the foliation in the ore and concentrated garnet and pyroxene into conformable GPS layers and lenses by mechanical metamorphic differentiation. Garnetite endoskarns (Einaudi et al., 1981), at contacts of the ore with sills or tectonic slivers of igneous rock, were boudinaged and disrupted. Subsequently, tabular bodies of mafic rocks, some of which crosscut foliation in the ore, were intruded during a second period of igneous activity. Where these were in contact with wollastonite ore, localized metasomatic reactions have replaced them with a second generation of GPS having Type B REE signatures inherited from the igneous protolith. The timing of this second metasomatism is unknown; it may have occurred as late as the subsequent granulite facies metamorphism in the Adirondack highlands (Bohlen et al, 1985), which postdates the anorthosite by as much as 50-80 ma (McLelland and Chiarenzelli, 1990a, b).

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ROAD LOG

The route of the trip and the locations of the six planned stops is shown in Figure 6.

At Exit 31 of the Adirondack Northway (I87), turn right (E) on Route 9N and go 0.2 miles; park on right opposite large roadcut.

STOP 1: METANORTHOSITE OF THE WESTPORT DOME

Massif anorthosites are plagioclase-rich igneous rocks. Their origin is controversial; the debate is summarized in Ashwal (1993). Most current models involve variations on this theme: Large volumes of mantle-derived basaltic magma fractionate in the upper mantle or lower crust toward an aluminum-rich residuum. Appreciable contamination may take place from partially melted lower crust. As the magma rises through the crust, plagioclase crystallizes to produce a crystal-rich magma or mush which is emplaced in the middle or upper crust. Further fractionation may follow emplacement; iron-enriched residual liquids may sink or be expelled by filter pressing.

In the Adirondack anorthosites, plagioclase occurs in two forms: dark-gray megacrysts and smaller, light-gray to white interstitial grains. Compositions range from An₄₂ to An₆₀, most commonly An₄₈ to An₅₅. Antiperthitic megacrysts are locally common. The interstitial feldspar, which may include some K feldspar, consists of varying proportions of crushed megacryst ("protoclastic") plagioclase and feldspar crystallized from residual liquid. In addition to plagioclase, the principal minerals of igneous origin are orthopyroxene, clinopyroxene, and iron-titanium oxides. Metamorphic minerals include garnet (in the more iron-rich varieties), hornblende, and biotite.

The Westport Dome is the largest anorthosite body in the Adirondacks outside the Marcy Massif. This exposure, near the southern end of the dome, illustrates the outcrop-scale lithologic variability common in Adirondack anorthosites. Gabbroic anorthosite is present in the eastern third of the cut; anorthosite in the remainder; textures and abundance of megacrysts vary widely. Note the vertical, ENE-trending ductile shear zone and the small, deeply weathered fault near the middle of the cut.

Turn around and go right onto the northbound entrance ramp for the Northway. At approximately 5.5 miles, turn right into rest area and park.

STOP 2: ANORTHOSITIC GNEISS

Cuts along the E side of the parking area expose strongly foliated anorthositic gneiss. The foliation here ranges from nearly horizontal to gently north-dipping; a NNE-trending lineation, common throughout the northeastern Adirondacks, is visible on some surfaces. Intense ductile deformation is common along the margins of the dome but is also present in the interior in zones such as this one. The timing of this deformation is unclear save that it postdates consolidation of the anorthosite. Walk north 0.1 miles along the road. On the left, blasted outcrops of anorthositic gneiss contain an irregular, tabular body of garnet-two pyroxene-plagioclase gneiss with minor K feldspar. "Jotunites" such as this one may be residual liquids from the anorthosite.

Begin road log at this point (mile 0.0). Caution: road signs may not correspond to the road names shown on the topographic maps.

- | | | |
|------|------|---|
| 0.0 | 0.0 | Return to the vehicles and drive N out of the rest area onto the ramp at Exit 32. |
| 0.3 | 0.3 | Turn right on Essex County Route 12 (Jersey Street). |
| 0.95 | 1.25 | Outcrops on left are moderately deformed anorthosite containing lenticular dark patches rich in pyroxene that may be the remnants of large pyroxene oikocrysts. |
| 2.35 | 3.60 | Crest of small rise overlooking the Champlain Valley with Lake Champlain and the Green Mountains of Vermont visible in the distance. |

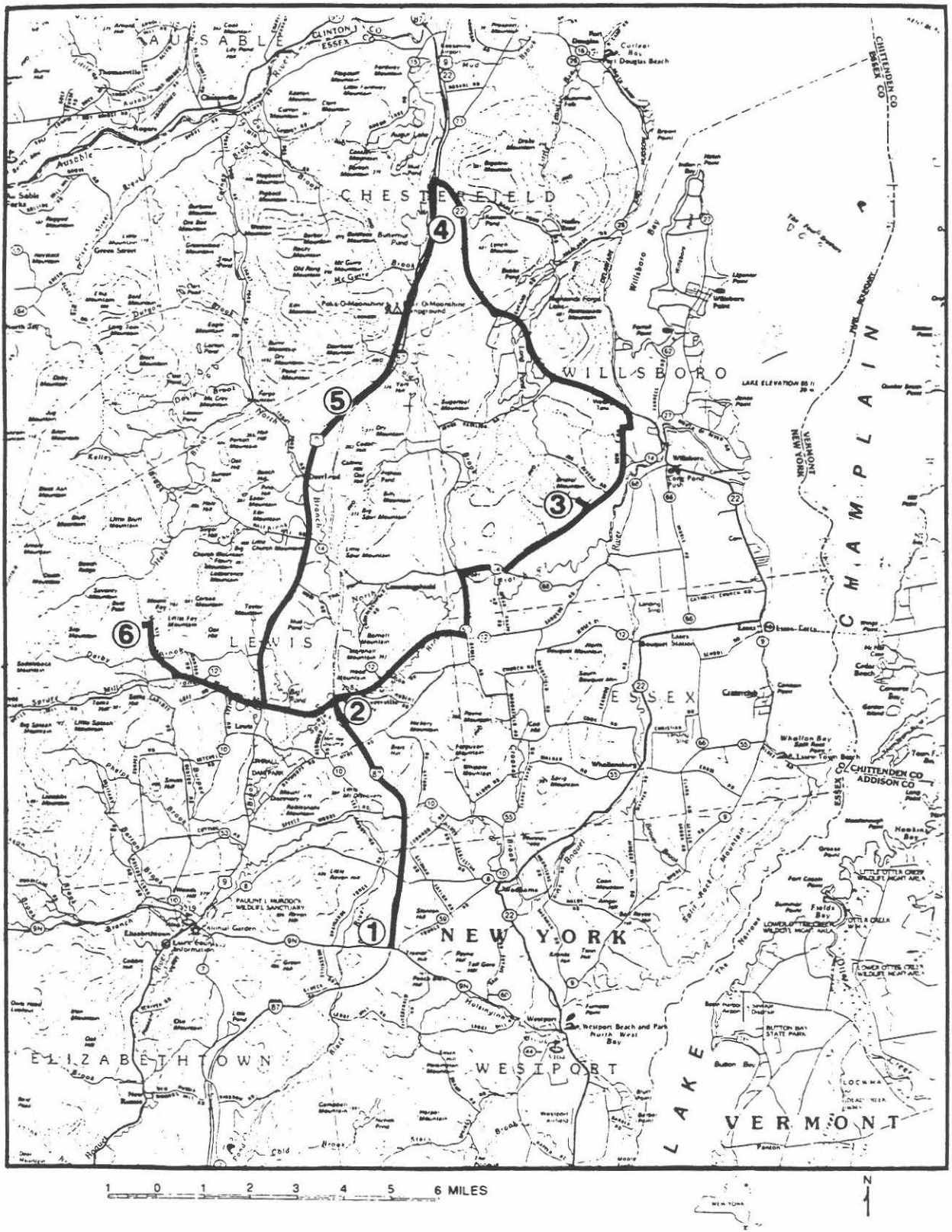


Figure 6. Map of parts of the Willsboro, Au Sable Forks, Elizabethtown and Port Henry 15' quadrangles showing the route of this field trip. Scale: 1:150,000.

- 0.05 3.65 Turn left (N) on Reber Valley Road.
- 1.40 5.05 Turn right on Mountain View Road.
- 1.05 6.1 Turn left at intersection; this is still Mountain View Road.
- 0.3 6.4 To the right, a fine view of the Green Mountains, from Mt. Mansfield (L) to Camel's Hump (R).
- 2.1 8.5 Willsboro Mine road (unpaved) on left. This is private property; be sure to obtain permission from NYCO before entering. Proceed through the gate and about 0.4 miles up this road to the lower portal of the Willsboro Mine.

STOP 3: WILLSBORO MINE

The Willsboro deposit is located near the base of the ore-bearing zone (OBZ), separated from the underlying metanorthosite by, at most, several tens of feet of amphibolite. Average strike of foliation and compositional layering within the OBZ is N 60°-65° W with considerable local variation; dips range from 15°-35° north. DeRudder (1962) cites drill core evidence for as many as seven ore layers, with maximum thickness of the principal ore horizon on the order of 50 feet. Numerous small brittle faults and unmetamorphosed mafic dikes (Isachsen et al., 1988) cut the ore.

We will examine exposures at both the lower and middle portals of the now-abandoned underground mine. The area around the middle portal was formerly operated as an open pit. Observe the straight, sharply defined garnet-pyroxene rock (GPS) layers in the ore as well as diffuse compositional layering within the ore; the latter shows complex folding in a few places. One GPS layer appears to crosscut foliation and compositional layering at a low angle. At contacts of the ore with gabbroic anorthosite or amphibolite, discontinuous layers and lenses (boudins?) of pale orange-brown garnet rock ("garnetite") are locally present. If time permits, we will traverse part of the OBZ overlying the ore, to examine a complex section of interlayered rocks including garnetite, amphibolite, gabbroic anorthosite gneiss, mafic calcsilicate rocks, and olivine metagabbro.

Return to the gate and resume road log (8.5 miles)

- 1.0 9.5 Turn left (N) on Fish and Game Road. Outcrops of unmetamorphosed carbonates in the woods just east of this intersection attest to the presence of a N-S normal fault, mapped by Buddington and Whitcomb (1941), that separates the Precambrian and Paleozoic rocks in this area. Going North on Fish and Game Road, the outcrops on the left are Precambrian marbles and gneisses.
- 1.6 11.1 Turn left on Route 22.
- 2.4 13.5 Long Pond on the left. The outcrops along this road are gabbroic and anorthositic gneisses with some interlayered metasedimentary rocks, occupying a structural saddle between the Westport dome to the south and the Port Kent dome to the north.
- 4.7 18.2 Turn left (S) onto US Route 9.
- 1.0 19.2 Park on right shoulder.

STOP 4: MARBLE XENOLITH IN JOTUNITE

Climb the bank on the right side of the road. At the top, climb over the fence and examine the rocks at the top of a large roadcut on the northbound lane of I-87. **Please exercise great caution near the edge of the cut cliff facing the Northway. Also, please do not climb down to road level, as it greatly distresses the State Police.**

The white rock is a large xenolith of calcite marble surrounded by mafic gneiss. The marble contains varying amounts of phlogopite, diopside, chondrodite, grossular, graphite, and sulfides. Phlogopite also occurs in clots a few centimeters across. A few lens-like masses of amphibolite are folded in a manner that suggests the

"fishhook" forms commonly found in other intensely deformed marbles. Diopside-plagioclase rock with very calcic plagioclase occurs near contacts with the host rock; identical rocks are found in the OBZ upsection from the Willsboro Mine.

The host rock is strongly foliated and lineated garnet-pyroxene-K feldspar-plagioclase gneiss, similar to the jotunitites at Stop 2. This is one of many locations where carbonates in contact with rocks of the anorthosite suite have not developed large amounts of wollastonite. The outcrops on both sides of Route 9 at this location are also jotunitites; those on the east side contain a layer or lens of garnet-rich gabbroic anorthosite gneiss, as well as masses of quartz (quartzite xenoliths?).

Continue south on Route 9.

- 1.2 20.4 Roadcuts beneath and to the south of the I-87 overpass are strongly foliated gabbroic anorthosite gneiss with garnet- and pyroxene-rich zones.
- 0.3 20.7 The cliffs on the right, on the east face of Pokomoonshine Mountain, are granitic gneiss; the outcrops at road level are gabbroic anorthosite gneiss.
- 0.4 21.1 Entrance to Pokomoonshine State Park on right. Continue South on Route 9.
- 0.5 21.6 Outcrops of fault breccia on right. A NNE-trending brittle fault roughly follows the road here.
- 1.6 23.2 **Alternate stop if time permits.** Park on the left, just North of a small outcrop of gabbroic anorthosite gneiss. Cross a narrow strip of woods to the southbound lane of the Northway. The roadcuts here are anorthositic gneisses with marble inclusions. Again, **please remain on the top of the outcrop, and do not cross the road.**
- 0.6 23.8 Park on right shoulder.

STOP 5: GABBROIC AND ANORTHOSITIC GNEISSES

Two fault-shattered outcrops on the West side of the road and one on the East contain very strongly foliated gabbroic and anorthositic gneisses. We map these as part of the OBZ, at the NW corner of the Westport Dome. The gabbroic gneisses contain millimeter-scale layers of garnet-pyroxene rock. Marble is present at the base of the outcrop on the East side. The more northerly outcrop on the West contains an undeformed hornblende-quartz-feldspar pegmatite. If wollastonite-rich layers are present in this section of the OBZ, they are not exposed.

- 1.6 25.4 Trout Pond Road on right. Continue on Route 9.
- 5.1 30.5 Turn right onto Wells Hill Road.
- 2.0 32.5 Bear right on Seventy Road.
- 0.9 33.4 Gate to Lewis Mine. **Get permission from NYCO before proceeding.** The large open pit of this operating mine is a short distance up the gravel road.

STOP 6: LEWIS MINE

While the immediate geologic setting of the Willsboro deposit is well known (Putman, 1958; DeRudder, 1962; Olmsted and Ollila, 1988), that of the Lewis deposit is less clear due to lack of natural exposures in the immediate vicinity. The orebody strikes roughly E-W and dips gently south, approximately parallel to the topographic surface, on the south limb of an open, E-W trending anticlinal crossfold. On the west side of the open pit, where the ore is close to 80 feet thick, it is overlain by strongly foliated charnockite; to the east the overlying rock, where exposed, is mainly anorthositic gneiss. Throughout much of the present mine area, the ore was exposed at the erosion surface. When the overburden was removed to begin mining, a karst-like surface was present, owing to the fact that wollastonite is one of the few silicate minerals that are appreciably water-soluble. Based on drilling data and temporary exposures within the mine, the footwall appears to be amphibolite and gabbroic anorthosite gneiss.

The hills north of the mine are gently N-dipping gabbroic anorthositic gneiss, underlain by deformed olivine metagabbro of the Jay Mountain body; both of these units are structurally above the ore-bearing horizon. Unlike Willsboro, where the ore is at most a few tens of meters above the anorthosite of the Westport Dome, our mapping suggests that the ore at Lewis is underlain by a considerable thickness of mixed gneisses and metasedimentary rocks. At one point during mining, a 4-6 m wide unmetamorphosed diabase dike was exposed.

Throughout much of the Lewis orebody, ore is interlayered with straight, sharply defined, discontinuous GPS layers parallel to foliation in the ore. We interpret this as tectonic layering. The regional NNE lineation can be seen locally on foliation surfaces in the ore here and at the Oak Hill prospect two miles to the east. A few thin, discontinuous layers of pyroxene-plagioclase-sphene granulite, rimmed by garnet-rich GPS, are present locally within the ore. Toward the north end of the present pit, layering and foliation are less prominent and the ore is leaner, rich in pyroxene and grossularitic garnet. At the northwest corner, a folded dike of GPS clearly crosscuts foliation.

After leaving the mine, return to the intersection of Wells Hill Road and Route 9. Cross Route 9 and continue east 1.6 miles to Exit 32 of the Northway (I-87).

END OF TRIP

GLACIAL DEPOSITS IN THE MOHAWK AND SACANDAGA VALLEYS OR A TALE OF TWO TONGUES REDUX

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PURPOSE

The purpose of this field trip is to examine the late glacial deposits of the Mohawk and Sacandaga Valleys, particularly the recessional and interlobate moraines preserved in the vicinity of Gloversville, NY. The glacial sediments exposed in the area were deposited during the waning stages of the last ice sheet, when the local topography split the thinning and wasting ice sheet into the Mohawk and Sacandaga glacial tongues. Sediment-laden meltwaters from these lobes deposited a complex of moraines and glacial lake sediments from Gloversville to Galway. The three-dimensional distribution of sedimentary facies record the demise of these tongues and provide a framework for one of the most productive aquifers in NYS.

This study focuses on three 7-1/2 quadrangles: Gloversville, Broadalbin, and Northville. Adjacent 7-1/2 minute quadrangles studied include Amsterdam, Tribes Hill, Galway, Jackson Summit, and Peck Lake (Figure 1).

INTRODUCTION

The study of the glacial stratigraphy of the Mohawk and Sacandaga Valleys has enjoyed a resurgence in the past thirty years after a lull of nearly forty years. Brigham (1928) published a map and lengthy interpretation of the glacial deposits of the central and western Mohawk Valley. This map and its associated text stood as the "last word" until Robert LaFleur and his graduate students began to re-examine glacial deposits and glacial history of the area in the early sixties (LaFleur, 1965; Yatsevitch, 1968). Yuri Yatsevitch completed a Masters Thesis in 1968 that outlined the framework of the late Wisconsinan history of the valleys. LaFleur (1969, 1975, 1979, 1983) built upon Yatsevitch's thesis, paying particular attention to evidence for a Late Wisconsinan readvance (the Yosts Readvance) and the wide-spread distribution of a dark gray, clay-rich till (the Mohawk Till). Later, Dineen and Hanson continued LaFleur's and Yatsevitch's work while mapping for the New York State Geological Survey's Surficial Mapping Project (see Cadwell and Dineen, 1986). They refined the geologic mapping and stratigraphy in the eastern Mohawk Valley while Jack Ridge refined the mapping and stratigraphy in the western Mohawk (Dineen and Hanson, 1985; Muller and Ridge, 1986; Ridge, 1991; Dineen et al., 1992).

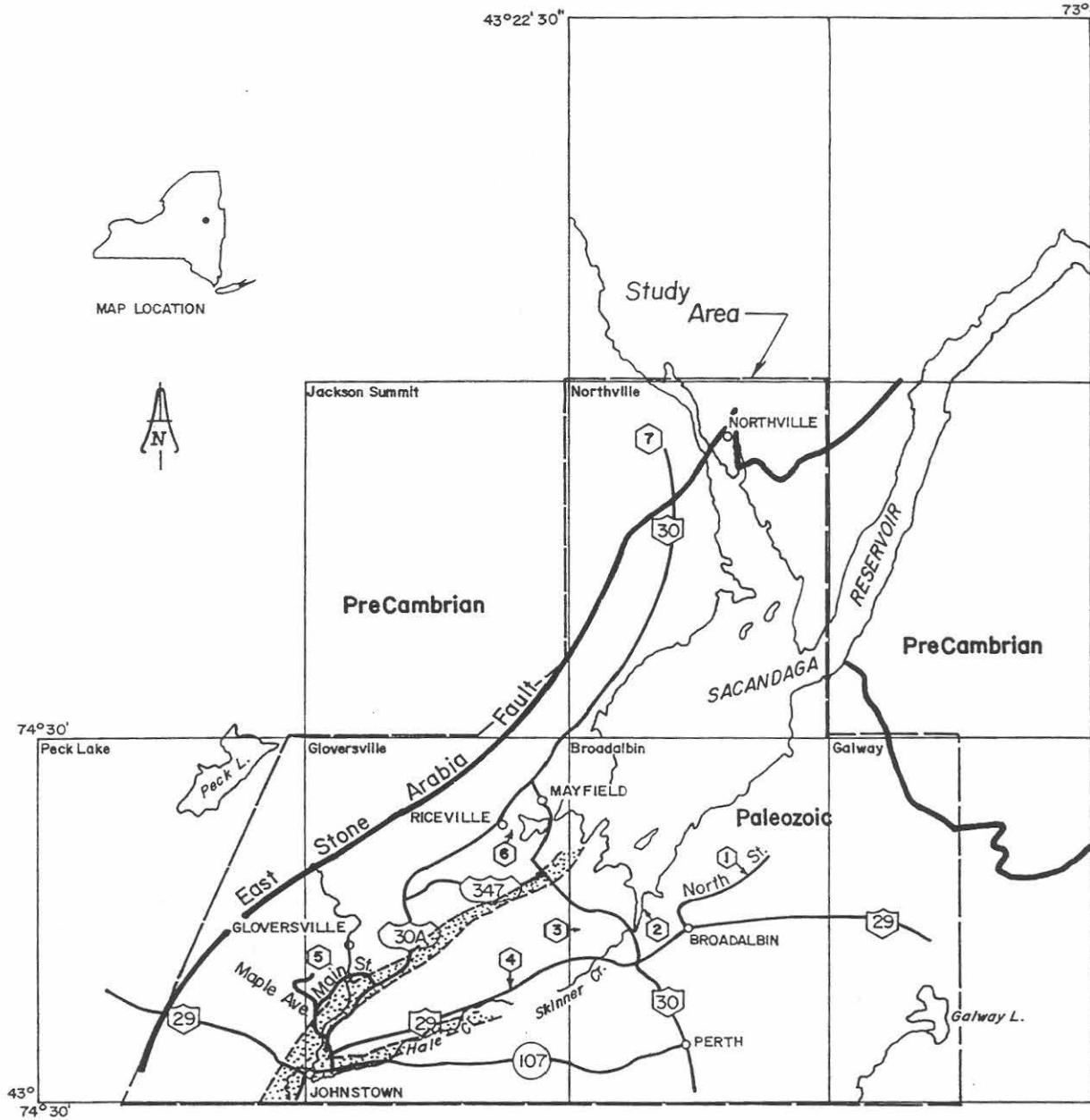
SUMMARY OF EASTERN MOHAWK AND SACANDAGA VALLEY GEOLOGY

Bedrock and Structure






The eastern Mohawk Valley is underlain by lower Paleozoic sedimentary rocks, including shales, dolostones, limestones, and sandstones. These rocks dip off of the dome of the Adirondack Mountains. They have been broken into a series of half-grabens, which strike north-northeast and dip down to the west (Fisher, 1980). The grabens are bounded on their eastern edges by normal faults that are, from east to west: the Hoffmans, Tribes Hill, Fonda, Noses (south of Johnstown), and East Stone Arabia (north of Johnstown) faults (Roorbach, 1913; Fisher et al. 1970; Fisher, 1980).

The present topography is controlled by bedrock hardness and structure. Sandstone, dolostones, or gneiss underlie the high, eastern edges of the half-grabens (Roorbach, 1913). Shaley dolostones and shales underlie their low western edges.

The Sacandaga Valley is a half-graben underlain by Paleozoic shales and sandstones. The uplands that border the basin on the west, north, and east are underlain by PreCambrian gneiss and quartzites (Fisher et al, 1970).



Generalized Bedrock Geology and Field Trip Stops

-  NYS Route
-  County Route
-  Field Trip Stop
-  Thalweg of Buried Preglacial Channel
-  Contact Between PreCambrian & Paleozoic Rocks

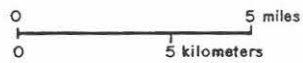


Figure 1

The PreCambrian rocks have been uplifted by the Fonda and East Stone Arabia faults.

Cenozoic Drainage

During the Cenozoic Era, stream systems carved primary, strike streams into the lower, western portions of the half-grabens. Shorter streams drained the dipslopes along the higher portions of the half grabens. The preglacial Sacandaga River was a large strike stream that drained the Sacandaga basin (Fig. 1), and flowed south into the preglacial Mohawk River (Brigham, 1929; Arnow, 1951). The preglacial Mohawk River drained east across the bedrock structure into the Hudson Lowlands.

Regional Glacial Movement

The preglacial strike valleys were oriented at nearly right angles to the movement of the Mohawk Sublobe of the Hudson Glacial Lobe, and were subparallel to the movement of the Adirondack Sublobe of the Hudson Lobe. The strike valleys channeled the Adirondack Sublobe and the high edges of the half-grabens split the Adirondack Sublobe into multiple subordinate ice tongues, including the Sacandaga tongue. Interglacial and glacial sediment have been preserved from glacial erosion by the high eastern buttresses of the half-grabens resulting in the accumulation of thick, complex sequences of Pleistocene deposits.

GLACIAL DRIFT IN THE MOHAWK AND SACANDAGA LOWLANDS

The products of the Wisconsin stage glaciation preserved in the Mohawk and Sacandaga Valleys include striae, drumlins, eskers, recessional moraines, interlobate moraines, and outwash systems. The glacial drift was deposited in the subglacial, ice marginal, and proglacial environments.

The texture of the glacial drift provides clues to glacial movement. The grain sizes and clast lithologies of the glacial drift are influenced by the bedrock lithologies that were eroded by the glacier. Crystalline and metamorphic rocks, and sandstone form sandy glacial deposits in response to glacial grinding. Glacial erosion grinds carbonates, phyllites, schists, siltstones, and shales into fine-grained debris. The drift in the Sacandaga basin is a bouldery, silty, sand. Gravel and clay are not important components of the Adirondack Till. The shales and carbonates in the Mohawk Valley were milled into a bouldery, gravelly, silty, clay forming the Mohawk Till (Fig. 2a). Drumlins and thick ground moraine were deposited in the lower areas of the grabens. The higher portions of the half-grabens were blanketed by thin ground moraine. The drift texture was subsequently modified through sorting by mass wasting, wind, and water. The topography controlled the relatively intensity of the various drift-modifying processes.

Features that form beneath the ice are subglacial. Subglacial features or deposits record the movement of a glacier. Striae and drumlins record the direction of ice flow.

Striae are glacial scratches on rock or compact, fine-grained sediment surfaces (such as till or lake clay). They are parallel to ice movement. **Drumlins** are ellipsoidal, streamlined hills that are orientated with their long axes parallel to the ice direction. Their tapered or "pointed" ends indicate the direction that the ice is moving towards. They are composed of concentric laminae of till or contain cores of older till, stratified drift, or bedrock with an outer skin of till. **Eskers** are elongated ridges of sand and gravel that were deposited by streams flowing within, under, or on the glacier. They record the orientation of the ice margin.

Ground moraine records the previous extent of the ice. It is an irregular blanket composed of a mixture of compact, fissile till, loose till, and stratified drift. Exposures of the compact till break into pencil-sized fragments along shear planes (fissility). Shear planes are also present in underlying sediments (see Stop 1). The ground moraine contains loose to firm till, with lenses of varying proportions of silt, sand, and/or gravel as well. The loose till contains significantly less clay than the compact till. A boulder pavement often occurs at the top of any underlying till unit. The bedrock underlying the ground moraine is often polished and striated.

Ice Marginal features form along the edge of the ice and include recessional moraines, interlobate moraines, and outwash systems.

Recessional moraines are large "dumps" of sediment forming arcuate ridges. They include till and stratified drift that accumulate along the margins of the ice sheet or lobe during hesitations in glacial retreat.

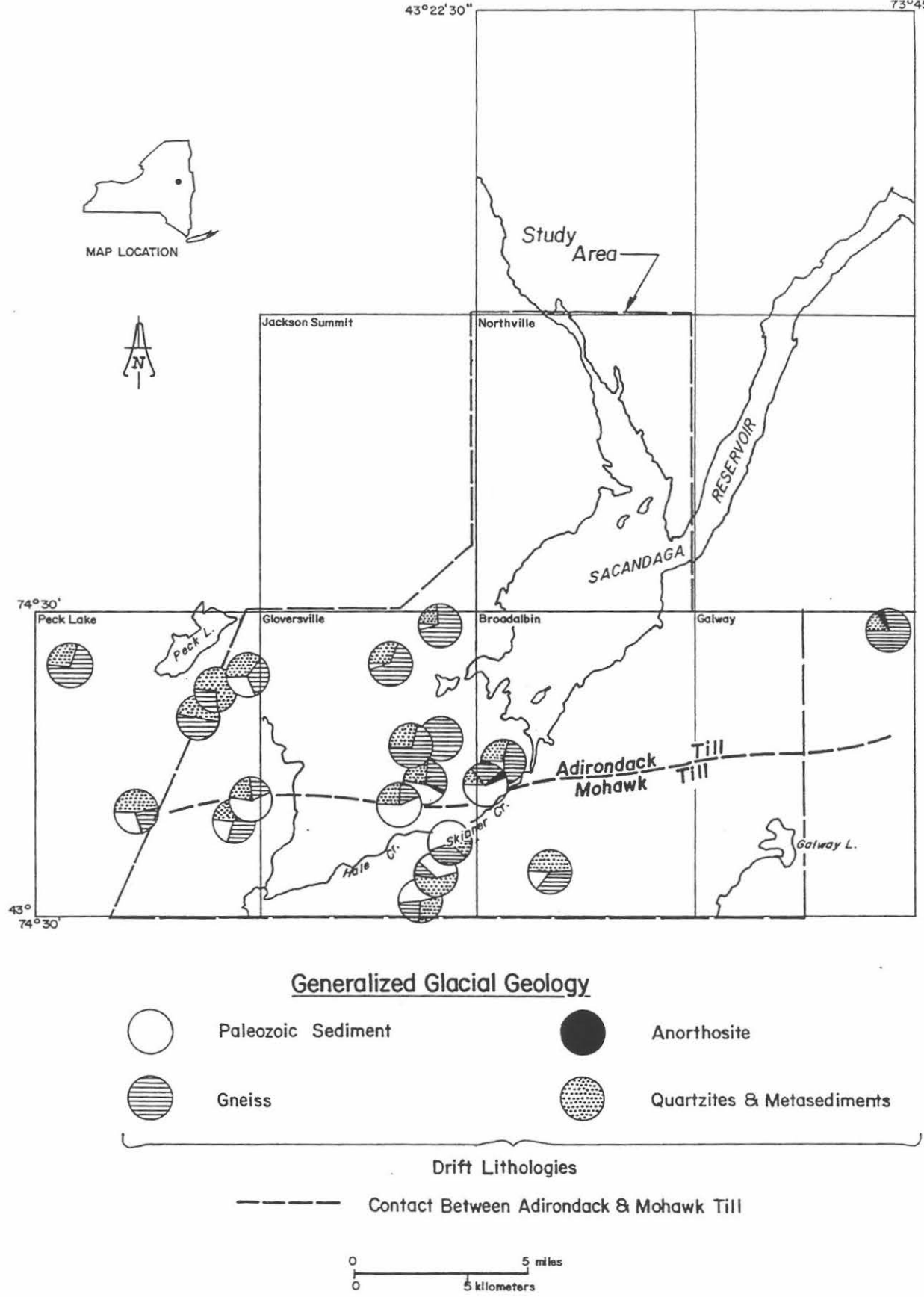
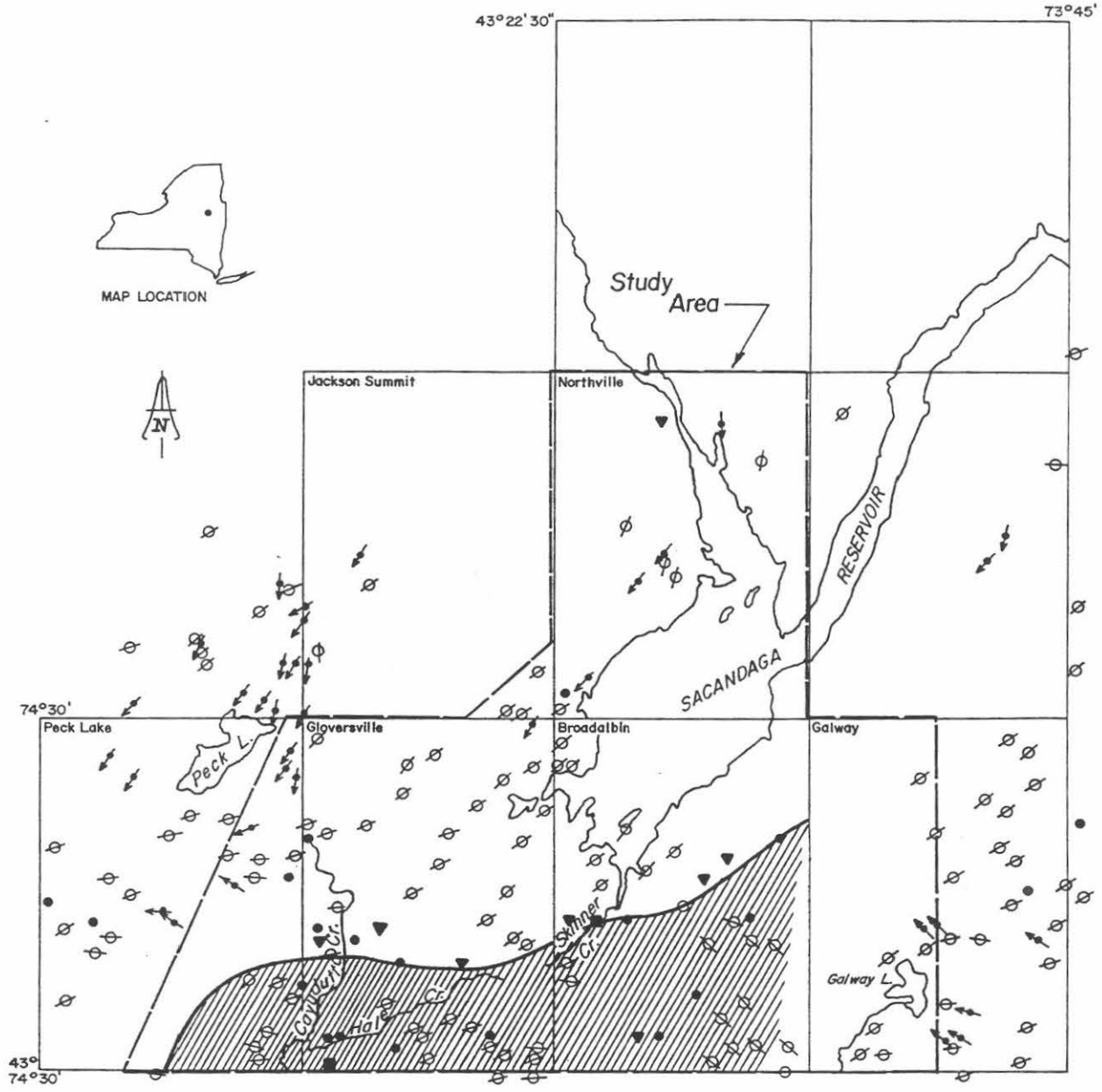


Figure 2a



More Generalized Glacial Geology

- ⊖ Drumlin Orientation
- Buried Soil Zones
- Striae Location
- ▼ Exposure w/ Till-Over-Outwash or Lacustrine
- Well w/ Till-Over-Outwash or Lacustrine
- ▨ Area w/ Stratified Drift Under Lodgment Till (Yost Readvance)

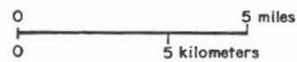


Figure 2b

Interlobate moraines are wide ridge complexes deposited between two or more glacial lobes. The ridges are usually hummocky or irregular, with numerous pits or depressions that mark the locations of melted ice blocks. The moraines are composed of variable quantities of stratified drift and till.

Proglacial deposits form beyond the ice margin. **Outwash** is deposited in meltwater stream systems that carried meltwater away from the ice. The outwash abruptly terminates along the ice margin at the head of outwash. Outwash also includes alluvial fans that originated at the ice margin. **Glacial lake sediments** are deposited in lakes adjacent to, under, and in front of the ice. The glacial lake sediments underlie planar areas or terraces, and are composed of fine sand, silt, and/or clay.

ICE MOVEMENT IN THE MOHAWK AND SACANDAGA VALLEYS

Drumlins and other ice movement indicators were mapped in the Mohawk and Sacandaga Valleys based on topographic maps, airphoto interpretation, and field observations. The drumlin data was supplemented by observations of rocdrumlins (composed almost entirely of glacially streamlined bedrock), flutes (low, elongated, smooth bedrock ridges), grooves (elongated, streamlined depressions), and roche moutonnee (whale-back shaped rock ridges). Flutes, grooves, and rocdrumlins were observed most frequently on the upper slopes of the grabens.

The drumlins and striae in the Sacandaga Valley are oriented east of north (Fig. 2b). They record the northeast-to-southwest ice movement of the Adirondack Sublobe, down the Sacandaga Valley, and across gneiss and anorthosites in the eastern Adirondack Mountains. The drumlins and striae in the Mohawk Valley record ice movement that was predominantly flowing from east to west, with a radial deflection of the Mohawk Sublobe to the northwest. The ice had flowed across Paleozoic shales, graywacke, and carbonates.

The drumlins are cored with Adirondack Till and sandy stratified drift in the Sacandaga Valley. These drumlins are oriented northeast to southwest (some drumlins appear "flipped" on the topographic maps, with the rounded end on south side!). Drumlins contain Mohawk Till in Mohawk Valley. They are oriented east-west, with some drumlins oriented northwest-southeast south of Perth Moraine. The Mohawk drumlins also contain some stratified drift.

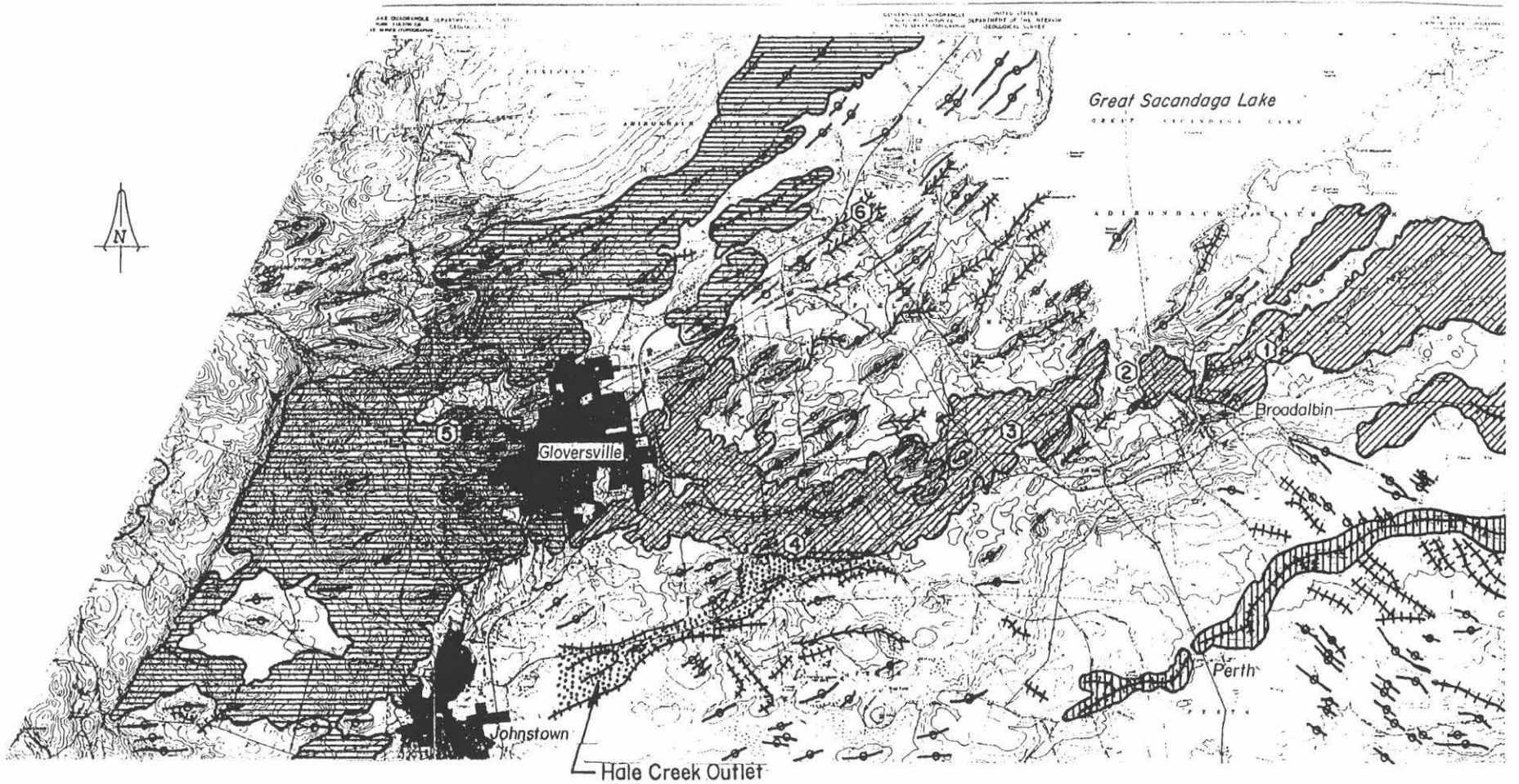
MORAINES IN THE MOHAWK AND SACANDAGA VALLEYS

Morainal features in the Gloversville vicinity include recessional and interlobate moraines. The locations of the moraines were controlled by the ice margin positions, which were controlled by the topography. Recessional moraines were deposited on the valley sides along the edges of the glacial tongues and in the valley bottoms, along the end of the glacial tongue. The glacial tongues advanced or remained further down valleys and receded along the uplands. Several recessional moraines were deposited in the vicinity of Gloversville. These include the Jackson Summit Recessional Moraine, Perth Recessional Moraine, and Gloversville Kame Complex (Fig. 3). The Broadalbin Interlobate Moraine was deposited along the suture between the Mohawk and Adirondack glacial tongues.

The Jackson Summit and Perth recessional moraines illustrate the influence of topography. The Jackson Summit Recessional Moraine is a 5 to 50-meter thick mass of sand and sandy till that is draped on the scarp of the East Stone Arabia Fault (Fig. 3; Dineen and Hanson, 1985). The moraine becomes thicker and its top surface is higher in elevation from southwest to northeast. The Perth Recessional Moraine extends from Perth to Galway Lake and consists of a 5- to 15-meter high narrow ridge of interbedded compact silty clay till and massive to planar bedded gravelly sand outwash (Dineen and Hanson, 1985).

The Gloversville Kame Complex (GKC) and Broadalbin Interlobate Moraine (BIM) form a zone of morainal topography that extends from the base of the East Stone Arabia fault block through Gloversville to Broadalbin and Galway. The Gloversville Kame Complex is a nested series of sandy moraine loops. The GKC and BIM form a series of concentric recessional moraine-interlobate moraine couplets from Clip Hill (at the East Stone Arabia Fault scarp) through Gloversville to the present-day Lake Sacandaga shoreline at Munsonville (Fig. 3).

The BIM is a wide ridge of silty sand and gravel that exhibits rapid changes in texture and bedding. It is 15 to 50 meters thick, and extends from Gloversville to Galway Lake. The moraine contains numerous loose, silty flow tills, reworked tills, and mud flows. The tills are interbedded with planar to cross-stratified sand with silt and gravel. Cross-laminated sand and silt dominate the distal portion of the moraine. Cross-bedded sands with silt and flowtills dominate the proximal portion of the moraine. The deposits are block faulted. The Town of Broadalbin,



Recessional and Interlobate Moraines in the Sacandaga Basin

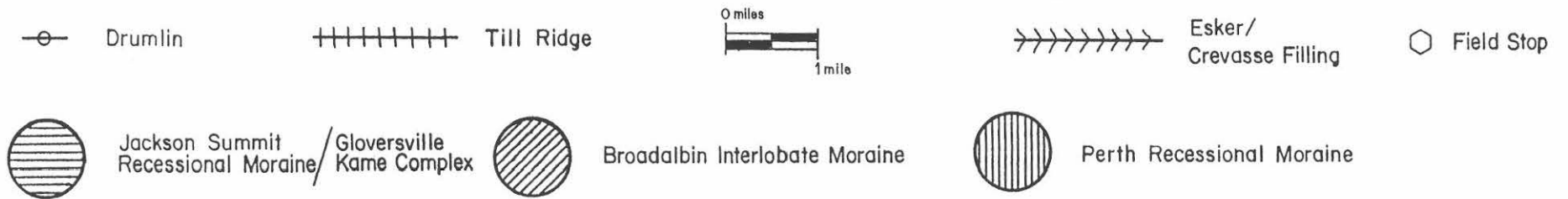


Figure 3

Herba, Rex Excavating, and Twin Cities Sand and Gravel pits are exposures in the proximal-through-distal BIM (Stops 1, 3, 4, and 5 in the field trip log, below). The southeastern edge is mantled with till that was deposited by the Yosts Readvance.

Glacial Lakes Gloversville and Sacandaga occupied the Sacandaga Valley and Glacial Lake Schoharie occupied the Mohawk Valley (Brigham, 1928; Yatsevitch, 1968; LaFleur, 1965, 1969, 1975, 1979). The lake sediments have a sandy texture in the Sacandaga Basin, where the Adirondack Till was reworked by water. The lake sediments have a clayey texture in the Mohawk Valley, because of reworking of the Mohawk Till.

The moraine systems are surrounded by fine sands that were deposited in the ice marginal lakes. The lakes included Glacial Lake Gloversville (GLG) at elevations from 870 to 840 feet above sea level (asl). GLG drained through an outlet near the Sammons Cemetery (south of Johnstown) into 690 to 600-foot asl Lake Schoharie in the Mohawk Valley. The subaqueous fan exposed in the Twin Cities Sand and Gravel pit at Gloversville (Stop 5) was deposited in GLG. Glacial Lake Sacandaga had water levels from 840 to 800 feet asl. It drained through the Hale Creek outlet into Glacial Lake Schoharie. As the ice retreated into the upper Sacandaga basin, ice-contact subaqueous fans were deposited in Glacial Lake Sacandaga. The Scotia Pit (Stop 2) at the lake shore near Broadalbin is an exposure of a fan.

Eskers and crevasse fillings occur in the area of ground moraine and lake plains that lies between Gloversville and Mayfield. The eskers (and, to a lesser extent, the crevasse fillings) contain trough cross-bedded sand with some gravel and silt, till lenses, and a mantle of till and/or aeolian sand. One to three meters of wind-blown silt and silty fine sand mantles many of the deposits in the area. Frost-heaved stones are common in the aeolian deposits. The interior of the eskers is exposed at Stop 6.

ACKNOWLEDGEMENTS

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GLACIAL DEPOSITS IN THE MOHAWK AND SACANDAGA VALLEYS OR A TALE OF TWO TONGUES REDUX

FIELD TRIP LOG

Miles from Start	Miles from Last Stop	
0.0	0.0	Leave the main entrance of Union College. Turn right onto Union St. and proceed west.
0.5	0.5	Turn left onto Erie Blvd.
1.1	1.1	Turn right onto the entrance ramp of I-890 (west) and proceed west.
-	-	Enter the NYS Thruway (I-90) and proceed west.
-	-	Crossing the western edge of the Hudson Lowlands.
-	-	Crossing the trace of the Hoffmans Fault.
16.7	16.7	Exit the Thruway onto NYS Route 30 (north). Proceed north through the City of Amsterdam. We are proceeding across ground moraine of Mohawk Drift. The hills are drumlins, many of which have stratified drift cores.
-	-	Crossing the Perth Recessional Moraine.
26.6	26.6	Fulton County Route 155/ Main Street Broadalbin. Proceed straight (east) into the Village of Broadalbin on Route 30. Route 30 turns left.
28	28	Turn right (north) onto North Street.
28.5	28.5	Turn right onto Union Street, proceed east.
29.6	29.6	Stop 1: Town of Broadalbin Pit-on the left side of Union street, next to the Broadalbin Town Hall.

Stop 1: Town of Broadalbin (see Figures S-1 and S-2a).

This exposure is in the southeast, Mohawk Drift portion of the BIM. The eastern slope was the ice-proximal side. The area was overridden from the southeast by the Yosts Readvance. The overriding ice plowed into the moraine, shearing and displacing the drift to the northwest. The deposit later collapsed by normal faulting and slumping when the Yosts ice melted. The outwash and ice-contact stratified drift exposed in the pit was deposited by meltwater flowing east to west along the margin of the Mohawk tongue. This conclusion is based on cross-bedding and climbing ripple orientations.

In 1984, the Section exposed at the Town of Broadalbin Town Pit was as follows:
6 m: Top of section.

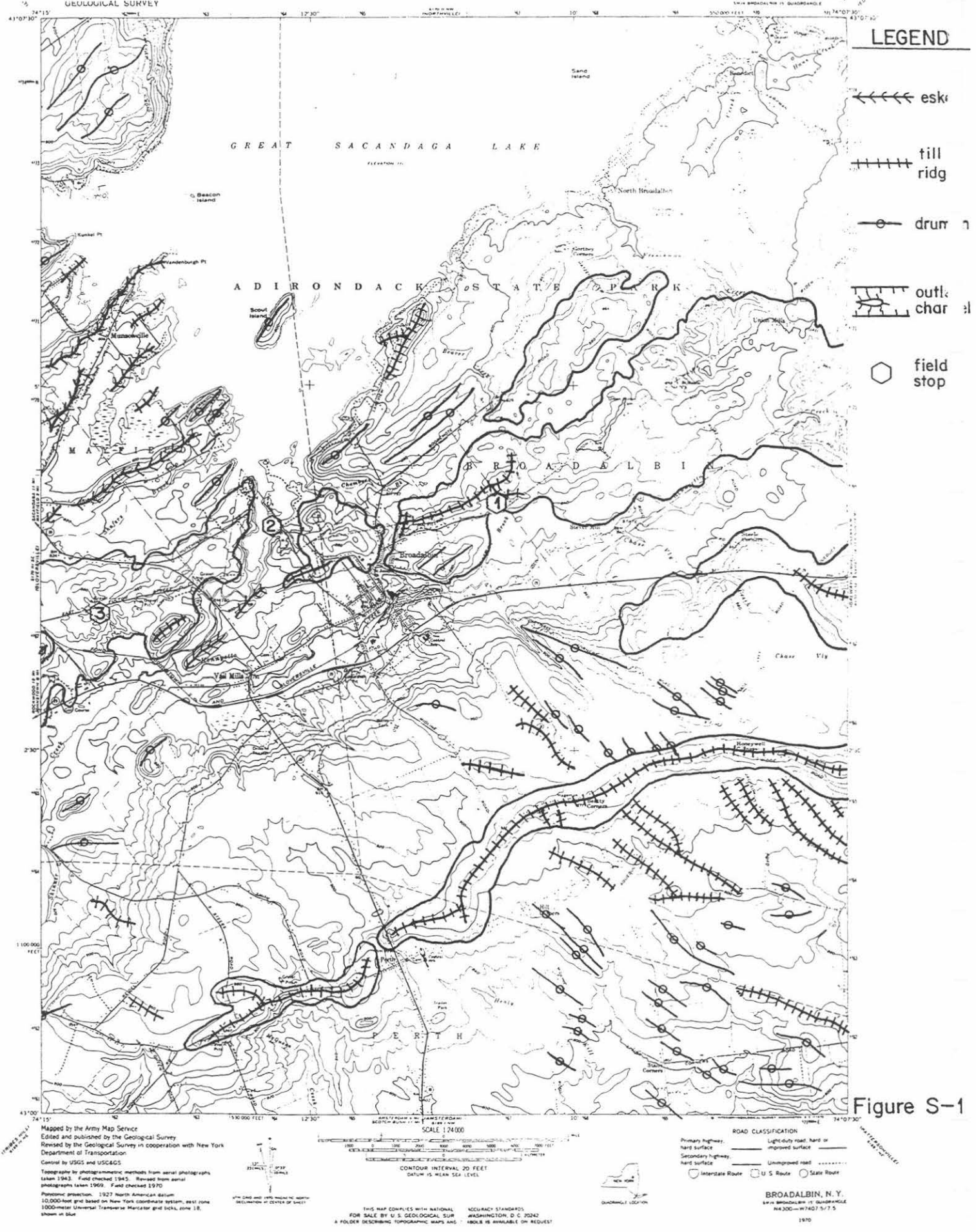


Figure S-1

4.5 to 6 m: 10 YR 6/4 Fine sand with little gravel, in planar beds and climbing ripple laminae. Contorted to the east. Upper 50 cm is massive and fissile.

4.5 m: Disconformity.

3.5 to 4.5 m: 10 YR 6/2 Climbing-ripple and ripple-trough-laminated fine sand with lenses of gravel near its base.

3.5 m: Disconformity.

2.5 to 3.5 m: 10 YR 6/4 Trough cross-laminated, fine to coarse sand with fine gravel in 1 to 3 m wide, 1 to 1.5 m deep channel fills inset into ripple laminated very fine sand. The unit is contorted and overturned to the northwest and gravity faulted to the east. The upper 50 cm is fissile and sheared. The unit contains lenses of coarsening-up, silty, sandy matrix-to clast-supported diamicts.

2.5 m: Disconformity.

Base to 2.5 m: 10 YR 6/4 Planar cross-laminated fine sand with lags of gravel on truncation surfaces. Ripple laminated to the east. Faulted down to the southeast.

A pit 70 m west of this Stop contained 5 meters of contorted, ripple-bedded sand. The upper portion of the sand was deformed by shearing, and was overturned to the northwest. The sand was overlain by 1 to 2 meters of compact, fissile, sandy matrix-supported diamict.

Leave Stop 1, turn left and proceed west on Union Street.

30.7	1.1	Turn south (left) onto North Street.
30.9	1.3	Turn right (north) onto North Main Street.
31.9	2.3	Turn left onto North Second Street, Stop 2 is immediately on the right.

Stop 2: Scotia Sand and Gravel (see Figure S-1):

This subaqueous fan contains planar, trough, and ripple-laminated sand and silty sand. Cross beds suggest that the sand was deposited against the ice by meltwater flowing from the east and southeast. The Sacandaga tongue lies to the northwest, in the vicinity of the present shoreline of the Sacandaga Reservoir.

Leave Stop 2 and proceed south on North Second Avenue.

32.7	0.8	Turn left onto Main Street (Fulton County 155).
33.6	1.7	Turn right onto NYS Route 30 (north).
34.4	2.5	Turn left onto Sand Creek Road.
35.3	3.4	Stop 3 Herba Pit is on the left, the Mayfield Landfill is on the right.

Stop 3: Herba Pit (see Figures S-1 and S-2b).

This exposure is in the northwestern portion of the BIM, on the Adirondack Drift side. The ice was on the northwestern side of the moraine. Many flow tills and a few compact lodgement tills are observed in this pit.

In 1983, the Section exposed at the Herba Pit was as follows:

9.7 m: Top of section.

9 to 9.7 m: 7.5 YR 6/8 Sand and silt matrix-supported diamict. Some cobbles are present.

8.6 m to 9 m: 10 YR 6/3 Trough cross-bedded, cobbly gravel.

6 to 8.6 m: 10 YR 7/6 Lenticular, trough cross-bedded, cobbly sand and gravel.

4.8 to 6 m: 10 YR 5/4 compact, sandy matrix-supported, massive to laminated diamict. Lenses of fine sand are present.

3.9 to 4.8 m: 10 YR 7/2 Interbedded sandy massive to planar-bedded diamict with greasy, rotten shale clasts. The lower 20 cm is sheared.

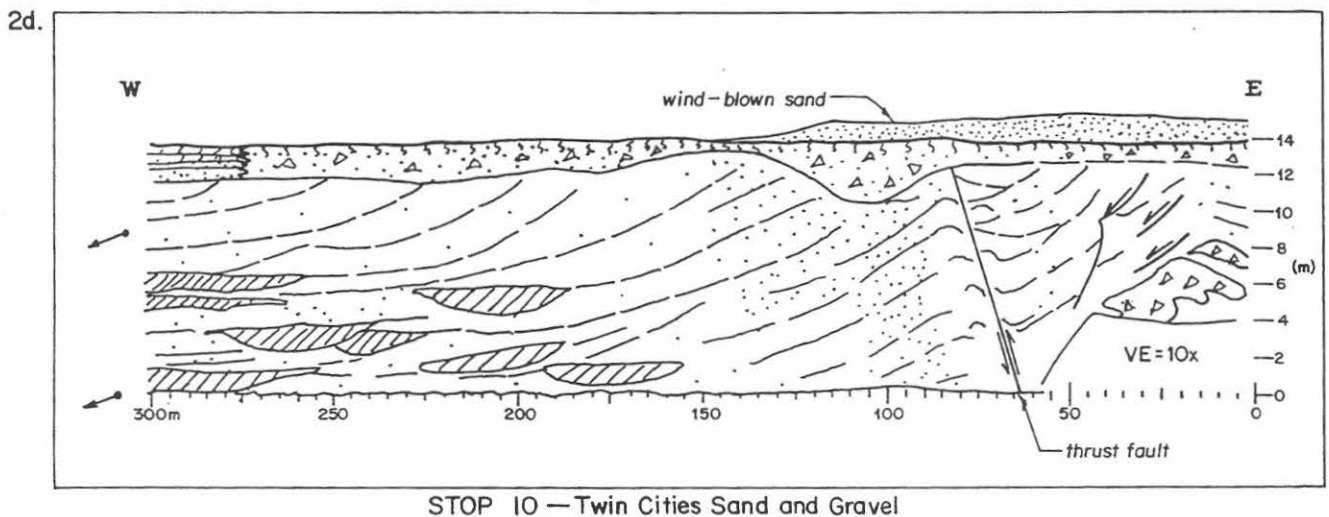
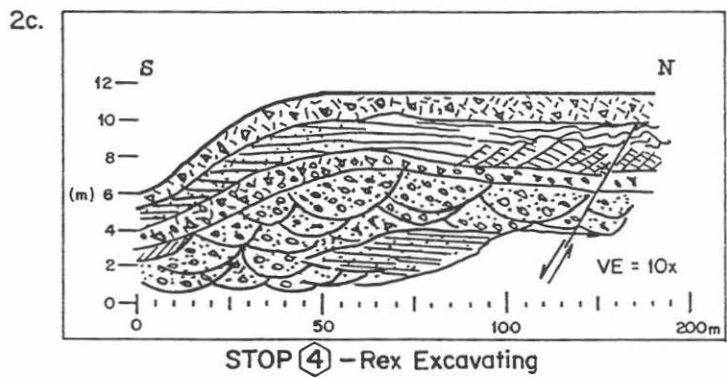
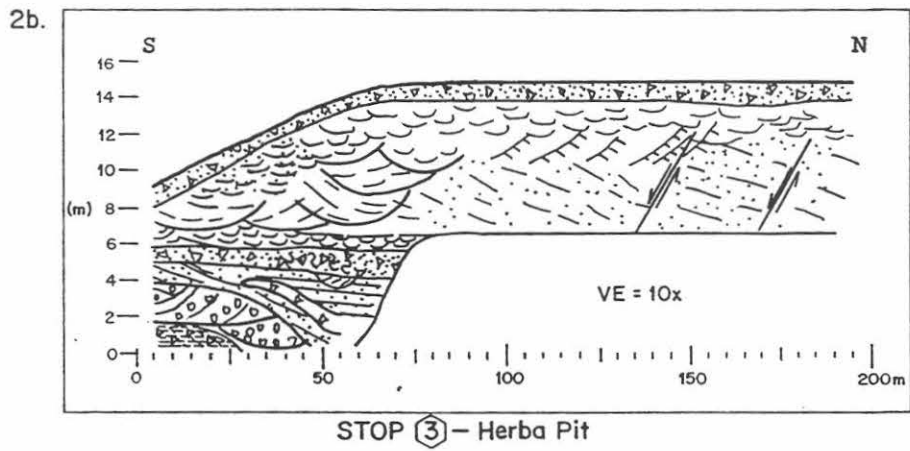
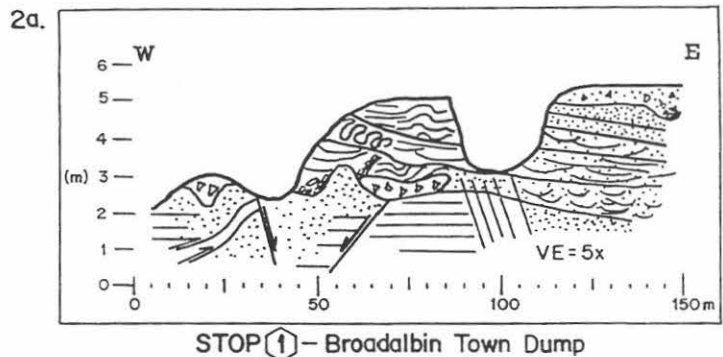
3.9 m: Truncation surface;

Base to 3.9 m: 10 YR 7/2 Cross-bedded cobbles and sand. The cross beds dip 10 S80E.

Leave Stop 3, turn right onto Sand Creek Road, proceed east.

36.1	0.8	Turn right onto NYS Route 30 (south).
37.3	1.2	Turn right onto NYS Route 29 (west).
41	3.7	Stop 4 on the right side of the road.

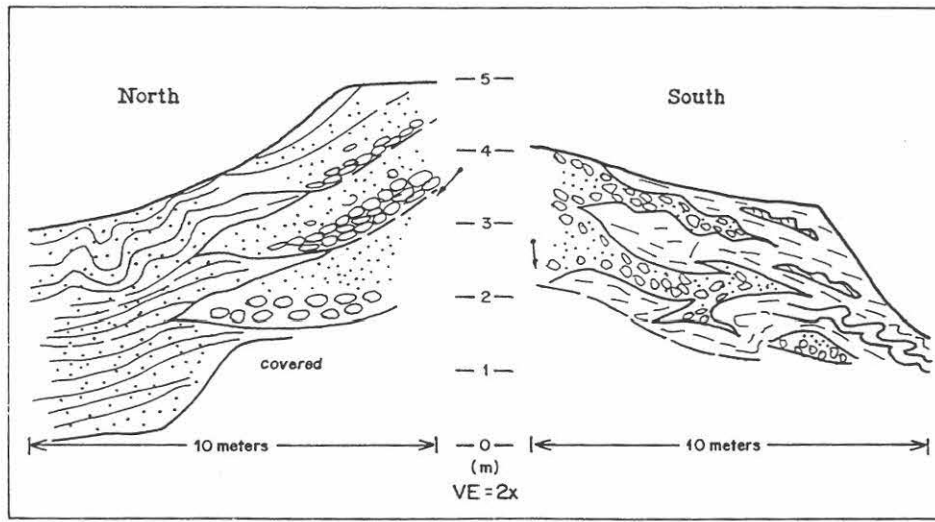
Stop 4: Rex Excavating (see Figures S-3 and S-2c).



Field Sketch of Exposures

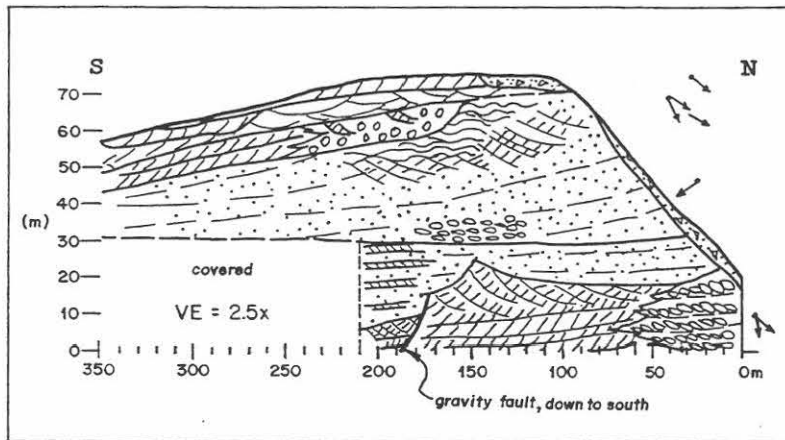
Figure S-2

2e.



STOP ⑥ - Mayfield Pits

2f.



STOP ⑦ - Bradt Pit

LEGEND (for Exposure Sketches 2a-2f)

DIAMICTON:			massive		sand & gravel		imbricated		starved ripples
			reverse graded		crossbed direction		graded		contorted
	SZ		graded		trough direction		reverse graded		planer crossbeds
			stratified		climbing ripples		trough crossbeds		planar laminae
VE	vertical exaggeration		silt & clay matrix		ripple troughs		symmetrical ripples		
			sand matrix		clay & silt				

Figure S-2 (cont.)

This exposure is in the southwestern portion of the BIM, on the Mohawk Drift side. Southeast was the ice-proximal side. The pit contains several till and lacustrine sequences that suggest the ice margin was oscillating. It was overridden by the Yosts Readvance. A similar exposure lies 1.7 km to the east.

In 1984, the Section exposed at the Rex Excavating Pit was as follows:

- 11.3 m: Top of section.
- 9.7 to 11.3 m: 10 YR 4/2 silt matrix-supported diamict. The unit is faintly laminated with sand stringers at the base. The base has a sheared contact with:
- 8 to 9.7 m: 10 YR 6/4 Irregularly-laminated fine sand and silt. Cobbles of rotten shale are present.
- 7.7 to 8 m: 10 YR 4/6 Laminated, silt matrix-supported diamict.
- 6.6 to 7.7 m: 10 YR 6/3 Planar-laminated fine sand with cobbles. Unit is locally cemented.
- 2 to 6.6 m: 10 YR 5/3 Planar cross-bedded, gravelly sand with a trace of boulders. The cross beds dip 10 to 15 N60W.
- Base to 2 m: 10 YR 5/3 Silt matrix-supported diamict with striated boulders.

Leave Stop 4, turn right onto NYS Route 29 (west).

- | | | |
|------|-----|--|
| 45.2 | 4.2 | Cross NYS Route 30A onto Briggs Street (City of Gloversville). |
| 45.7 | 4.7 | Turn right (north) onto North Perry Street. |
| 46.6 | 5.6 | Veer left onto Maple Avenue, proceeding northwest. |
| 47.4 | 6.4 | Stop 5 on the right. |

Stop 5: Twin Cities Sand and Gravel (see Figures S-3 and S-2d)

This pit is in the distal portion of the BIM and contains distinct subaqueous fan features (foreset beds and starved ripples). Both gravity and thrust faults are common in the eastern (proximal) portion of the pit. Aeolian sand mantles the upper 4 m of the pit. The upper cross-bedded sand is the leading edge of an aeolian dune. The upper sandy diamict is wind-blown silt and sand.

In 1984, the Section exposed at the Twin Cities Sand and Gravel Pit was as follows:

- 14 m: Top of section.
- 12 to 14 m: 10 YR 5/4 Cross-laminated fine sand.
- 10 to 12 m: 10 YR 6/3 Laminated to massive (west to east) sand matrix-supported diamict. This unit fills in depressions in the lower unit. The top contact has a soil zone.
- Base to 10 m: 10 yr 6/4 Planar cross-bedded fine sand with lenses of trough laminated medium to coarse sand. Faulted and interbedded with contorted lenses of silty matrix-supported diamict to the east.

Leave Stop 5, turn left onto Maple Street.

- | | | |
|------|-----|---|
| 48.2 | 0.8 | Turn right onto North Perry Street. |
| 48.6 | 1.2 | Turn left (east) onto Townsend Road. |
| 50.1 | 1.5 | Turn left onto NYS Route 30A (north). |
| 57.5 | 8.9 | Turn right onto NYS Route 30 (south). Stop 6 is immediately on the left. |

Stop 6: Mayfield Pit (see Figure S-3 and S-2e).

This pit is in an esker that pokes above the Glacial Lake Sacandaga plain.

In 1984, the Section exposed at the Mayfield Pit was as follows:

- 6.5 m: Top of section.
- 5 to 6.5 m: 10 YR 7/3 Beds of contorted laminated fine to medium sand to thin rhythmically-laminated, very fine sand to silt.
- 2 to 5 m: 10 YR 7/3 Fining-upward, planar-bedded, imbricated cobbles to silt, beds dip 10 S50W.
- Base to 2 m: Covered.

Leave Stop 6, turn left onto NYS Route 30 (north).

- | | | |
|------|------|--|
| 68.8 | 20.2 | Stop 7 is on the left side of the road. |
|------|------|--|

Field Trip Stops

LEGEND

- ←←←← esker
- ++++ till ridge
- drumlin
- ⊞ outlet channel
- ◻ field stop

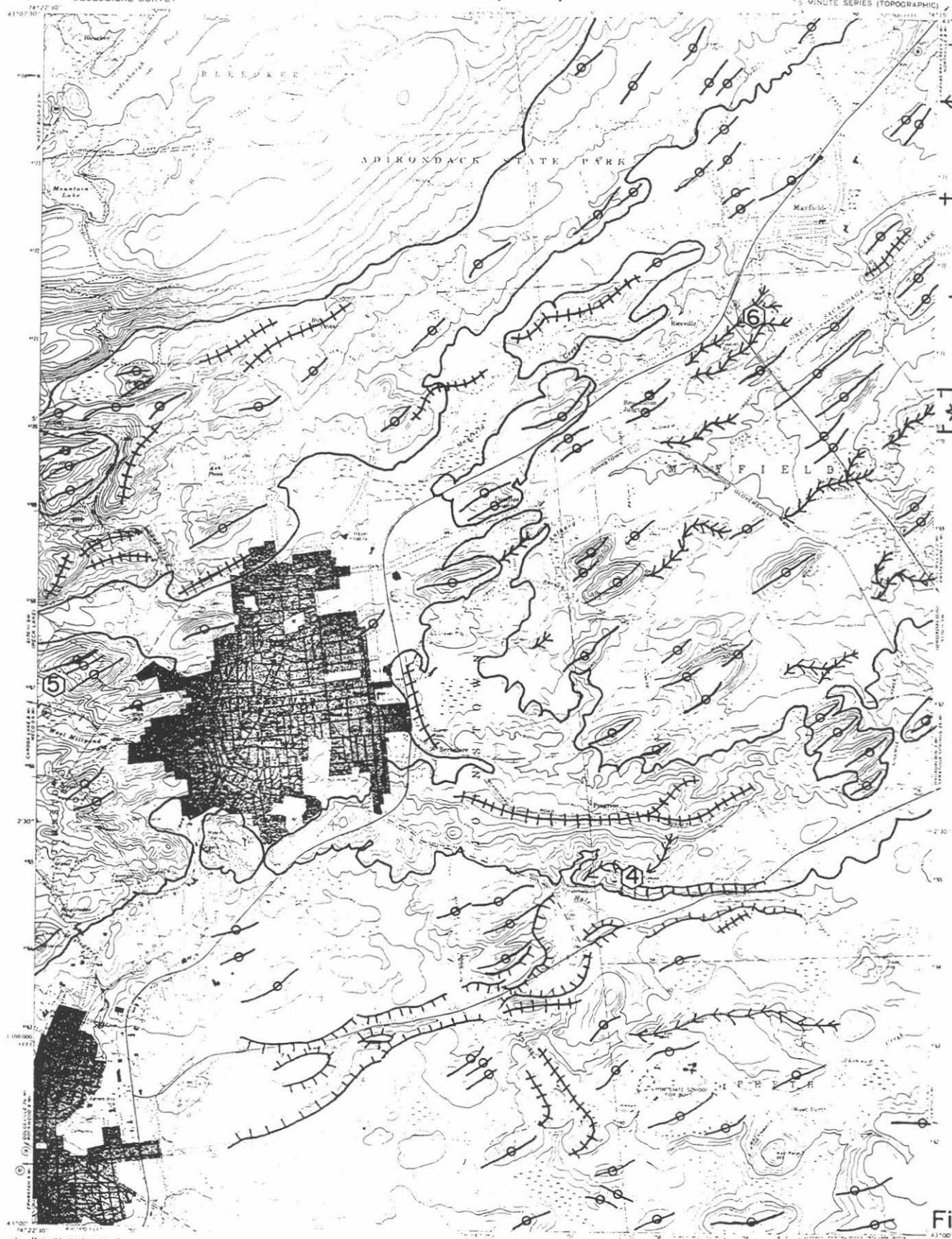
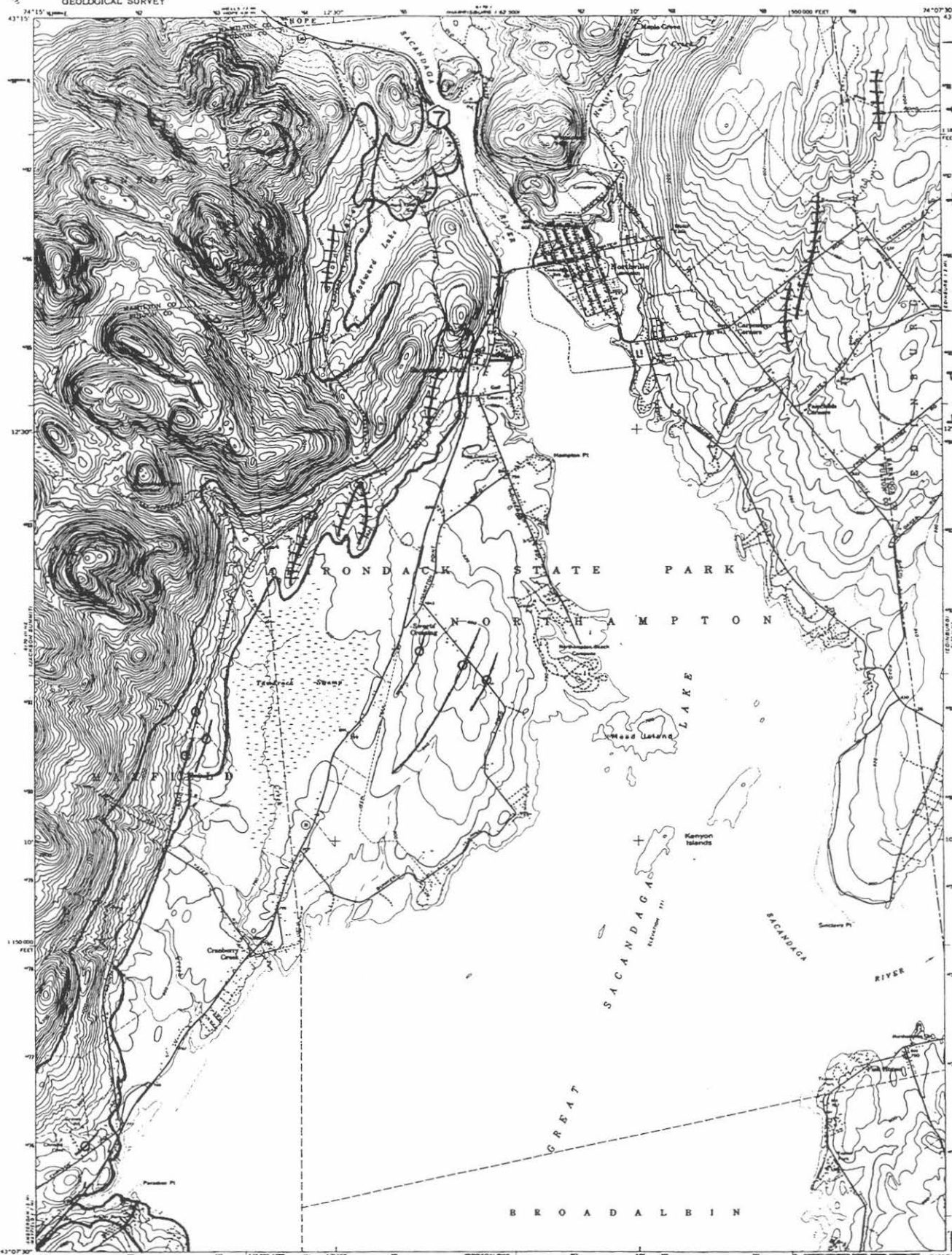


Figure S-3

Mapped by the Army Map Service.
Edited and published by the Geological Survey.
Reviewed by the Geological Survey, in cooperation with the
New York Department of Transportation.
Copies in 1950 and 1962.
Topography by photogrammetry, from aerial photographs of the
1940s. Contour interval 20 feet. Elevation in feet.
Photographic base 1:50,000. Contour interval 20 feet.
Projection: Universal Transverse Mercator.
Scale: 1:50,000. Contour interval 20 feet.
When in blue.

ROAD CLASSIFICATION
 Primary road ——— Light duty road, hard or
 surface
 Secondary road ——— Improved surface
 Tertiary road ——— Unimproved road
 U.S. Route ——— State Route

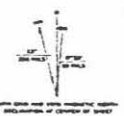
GLOVERSVILLE, N. Y.
14100-14115



LEGEND

- esker
- till ridge
- drumlin
- outlet channel
- field stop

Mapped by the Army Map Service
Edited and published by the Geological Survey
Revised by the Geological Survey in cooperation with New York
Department of Transportation
Control by USGS and USCGS
Topography by photogrammetric methods, from aerial photographs
taken 1943-1944. Field checked 1965. Revised from aerial
photographs taken 1968. Field checked 1970.
Physical projection, 1927 North American datum.
30,000-foot grid based on New York datum system, and also
1000-meter Universal Transverse Mercator grid lines, zone 18,
shown in blue.



SCALE 1:24,000
CONTOUR INTERVAL 20 FEET
NATIONAL GEODESIC VERTICAL DATUM OF 1929

ROAD CLASSIFICATION
Primary highway, hard surface
Secondary highway, hard surface
Light-duty road, hard or improved surface
Unimproved road
Interstate Route U.S. Route State Route



THIS MAP COMPLETES WITH NATIONAL MAP ACCURACY STANDARDS
FOR SALE BY U.S. GEOLOGICAL SURVEY, RESTON, VIRGINIA 22082
A POLYMER BINDING TOPOGRAPHIC MAPS AND STRIPS IS AVAILABLE ON REQUEST

NORTHVILLE, N.Y.
843075-874075/7.5

Figure S-4

Stop 7 Bradt Pit (see Figures S-4 and S-2f)

This pit is a deep cut into a kame delta that was built into a 900 ft asl glacial lake, possibly an ice-marginal portion of Lake Gloversville. The exposure was 65+ m high and 350 m long in 1984!

In 1984, the Section exposed at the Bradt Pit was as follows:

65 m: Top of section

50 to 65 m: 10 YR 5/4 Shallow, trough cross-bedded, fine to medium sand, some gravel grading down into fining upwards, 10 YR 6/2 planar cross-bedded boulders through coarse sand, dipping 5 to 10 S60E.

45 to 50 m: 10 YR 6/4 Planar cross-bedded, fine to medium sand and ripple-laminated, gravelly, fine to medium sand, dipping 15 S30W.

30 to 45 m: 10 YR 6/2 Coarsening upwards, planar cross bedded, cobbly, gravelly, fine to coarse sand, dipping 15 S50W. Cobbles are subangular to subrounded.

15 to 30 m: 10 YR 6/4 Planar cross-bedded, fine to medium sand with trace cobbles. Beds dip 10 to 15 N60W.

Base to 15 m: 10 YR 6/2 Imbricated, open-work, subangular to subrounded boulders in 1 to 1.5 m cross beds. Imbrication dips N20E.

End of field trip, proceed homeward on Route 30 to the NYS Thruway.

LITHOFACIES AND STRUCTURE OF THE TACONIC FLYSCH, MELANGE, AND ALLOCHTHON, IN THE NEW YORK CAPITAL DISTRICT

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INTRODUCTION

The Taconic Allochthon of eastern New York is bordered to the west by a zone of mid-Ordovician shale and greywacke turbidites, which are appropriately characterised by the term "flysch". These synorogenic deep-water clastics are now interpreted to represent the fill of the migrating flexural basin created by the advance of the Taconic thrust sheets onto the Cambro-Ordovician passive margin of eastern North America (Rowley and Kidd, 1981; Bradley and Kusky, 1986). The flysch is markedly diachronous, with a thick basal carbonaceous shale unit, the Utica Shale, and extends many hundreds of kilometres west of the present margin of the Taconic Allochthon. The Allochthon consists almost exclusively of sedimentary rocks that represent a sample of the continental rise part of the Cambro-Ordovician passive margin, and of the latest Precambrian-earliest Cambrian clastics of the late-stage rift fill, and rift to passive margin transition; all are strongly folded and have been transported westward on a complex system of thrusts at least 150 to 200 km relative to the North American craton (Bradley, 1989). In the New York Capital District, a zone about 16-20 kilometers wide of the Ordovician flysch adjoining the western margin of the Allochthon has also undergone strong deformation, including widespread conversion of once stratified rocks to melange, associated with the later stages of the Taconic Orogeny. It is the purpose of this guide and field trip to examine these rocks, to attempt to illuminate the structures which they contain, to try to roll back at least some of the nomenclatural and stratigraphic confusion they have suffered, and to place them in the larger regional context of the Champlain Thrust system.

GENERAL GEOLOGICAL SETTING OF THE FIELD TRIP

Before confusing readers and trip attendees with the "stratigraphic" terminology, we set out the distribution of basic rock types and their structural condition (refer to Figures 1 and 2, the geological maps). West from a NNE-trending line through a point on the Mohawk River in Niskayuna, just east of Schenectady, regionally flat-lying [very gently-dipping] Paleozoic strata are exposed, which consist, in the immediate area of the field trip, of the medial Ordovician flysch, that is greywackes and shales, in varying proportions. East of this boundary, the western limit of Taconic deformation, deformed medial Ordovician rocks occur in a zone 16-20 kilometers wide, bounded to the east by the [also] NNE-trending western border fault of the Taconic Allochthon, the Taconic Frontal Thrust. The deformed rocks of this zone dominantly consist of highly disrupted shales and greywackes, and these are appropriately termed shale-matrix melange. The western side of the deformed zone consists of a belt about 5-6 kilometers wide of folded and internally faulted rocks that are still largely bedded (we term this the Vischer Ferry Zone), and which can be interpreted as expressing a zone of increasing strain transitional from the undeformed flat-lying strata in the west to the highly-strained, disrupted melange in the east. Within the melange, there are lens-shaped belts of less-deformed material, both of shale-dominated, and of greywacke-dominated protolith, and ranging in structural condition from merely folded, to "broken formation", transitional to melange. These less-deformed lenses range up in size to regionally mappable; the most prominent in the area of the field trip is the bedded greywacke and shale of the Halfmoon Greywacke Zone (see Figures 1 and 2). South of the Capital District,

¹ now at Telegrafenberg C2, 14472 Potsdam, Germany

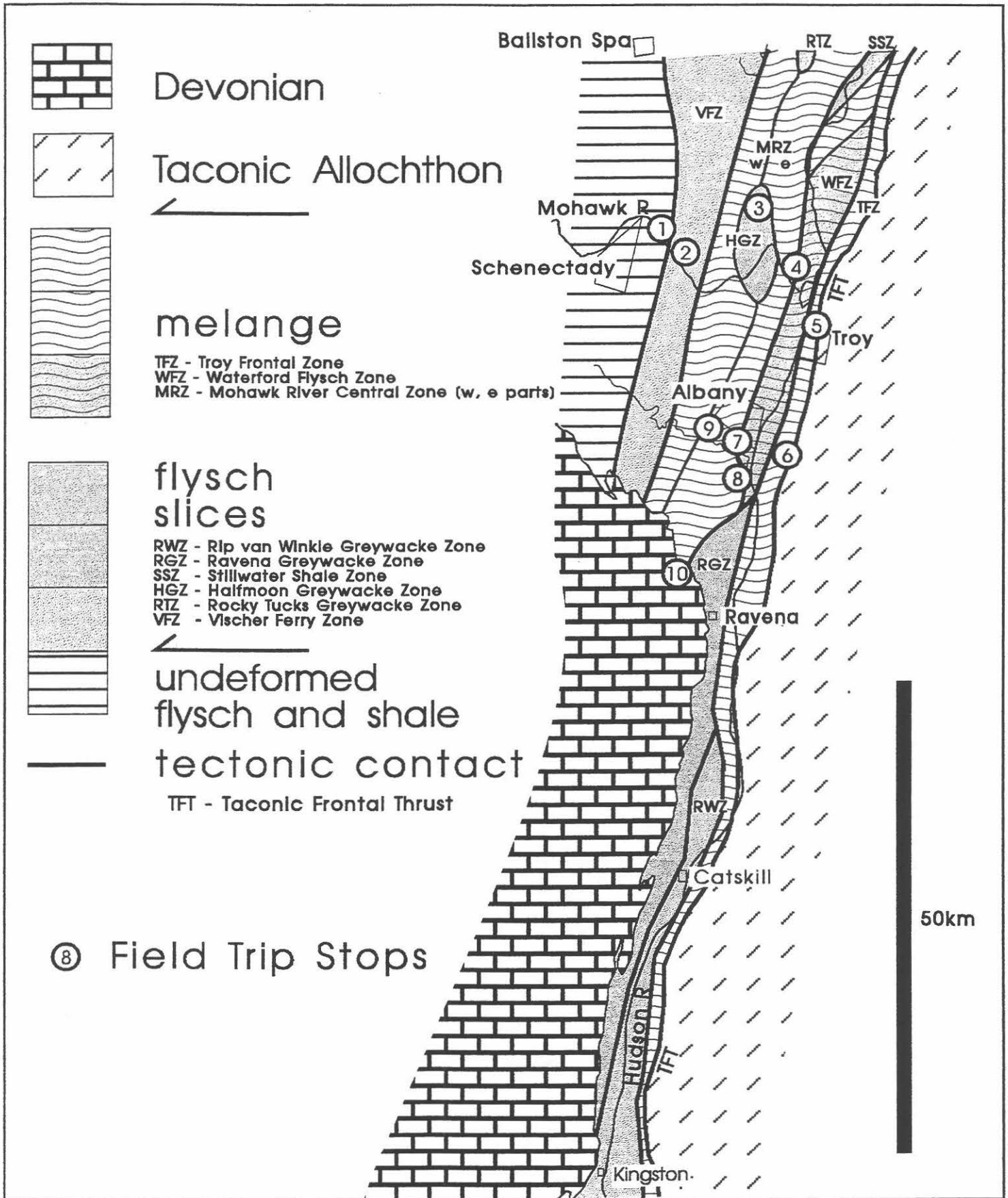


Fig 1. Tectonic units in the Saratoga Lake - Capital District - Kingston segment of the central Hudson Valley

along the Hudson valley as far as Kingston, most of the exposed width of the flysch and melange belt is occupied by two of these bedded belts, with a narrow melange belt separating them, and another bordering the eastern side (Figure 1). The change of structural style, from melange-dominated in the north, to "fold/thrust"-dominated to the south, was probably controlled by the change from shale-rich to greywacke-rich strata, which in turn must have been a product of sediment supply and local basin geometry.

PREVIOUS WORK

Detailed and systematic studies of the medial Ordovician rocks of the Hudson River Valley are contained in Ruedemann's reports of mapping and stratigraphic and structural studies, *Geology of Saratoga Springs and vicinity* (Cushing and Ruedemann, 1914) and *Geology of the Capital District* (Ruedemann, 1930), which even now include the most detailed published maps of this area. Ruedemann identified, and marked on his maps, the western limit of deformation in the medial Ordovician flysch, meaning folding and pervasive faulting and melange formation, and devoted considerable space in his text to the structural condition of the deformed zone, which occupies most of the width of the exposure of these rocks in the Hudson River lowlands in this area. It is unfortunate that this pioneer structural work has been submerged by the choice of the compilers of the last several geological maps of New York State (Fisher et al., 1970, Rogers et al., 1990) and of the Albany County area (Fickes, 1982) not to indicate that the medial Ordovician flysch in most of this area is significantly deformed, in strong contrast to these strata farther west, and to have continued use of a stratigraphic nomenclature that, among other defects, actively works to obscure this fact. It is ironic that Ruedemann himself remains responsible for creating much of the stratigraphic nomenclature and confusion in the first place!

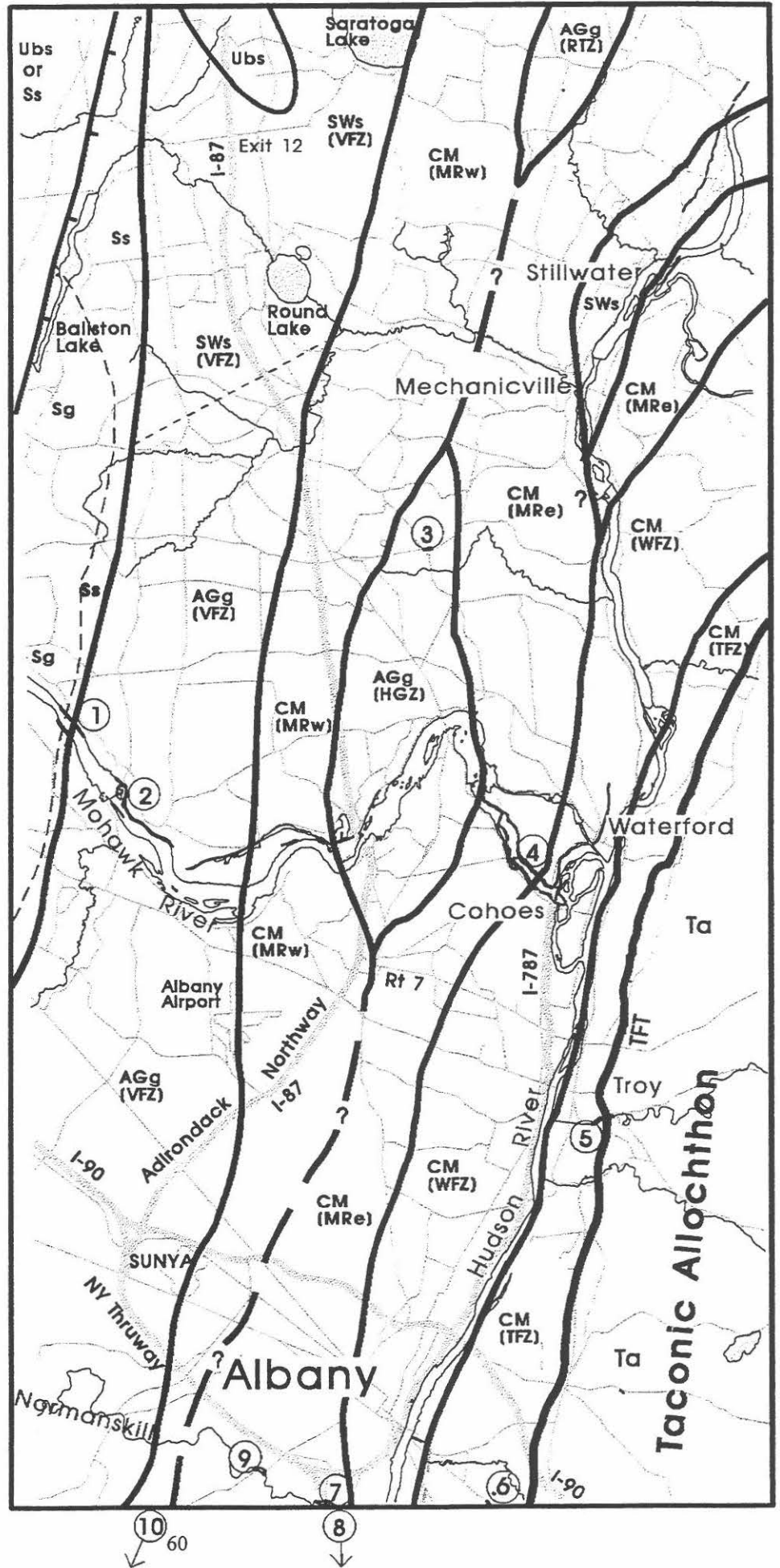
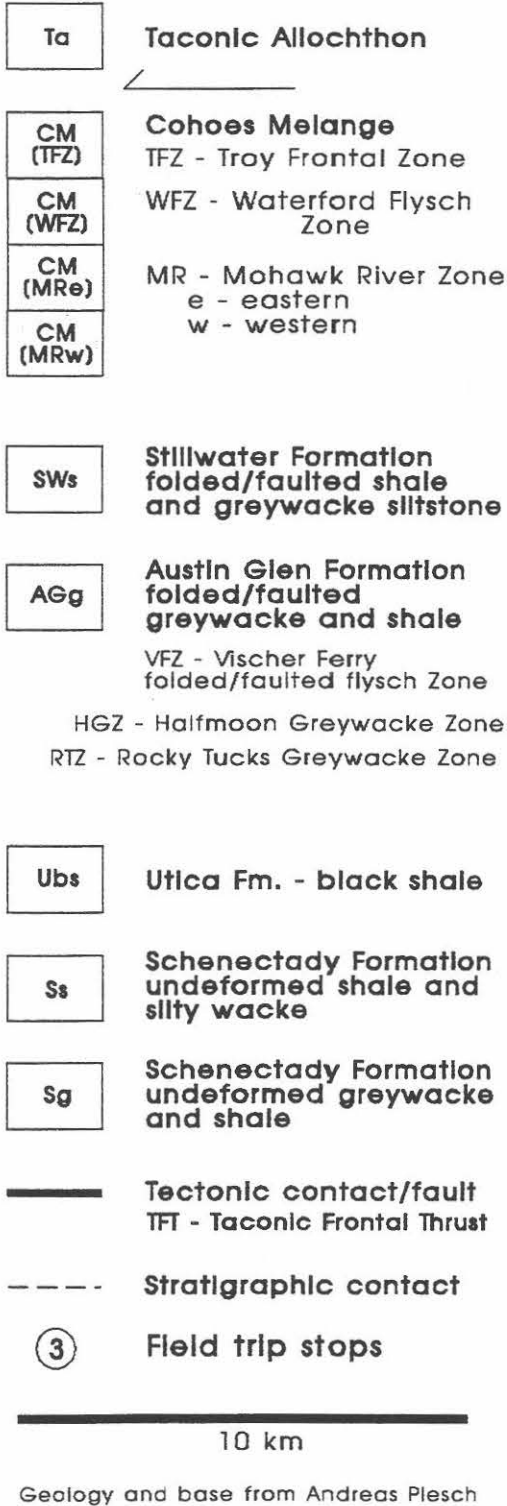
Bird (1963, 1969) first clearly documented the regional extent of the melange, and its general structural significance in its relation to the emplacement of the Taconic Allochthon, although we reject the then-prevalent notion of gravity sliding for the emplacement mechanism of the Allochthon and the formation of the melange. More recent work in the Hudson Valley flysch by Vollmer (1981a), Bosworth and Vollmer (1981), Vollmer and Bosworth (1984), and Plesch (1994) includes detailed mapping of the area between Ravena and Saratoga Lake, and compilation of outcrop mapping farther south to the area of Middletown (in Plesch, 1994). This work reveals, more clearly than that of previous workers, the abundance of melange in the Capital District (see Figure 1), and that this is a product largely of subsurface shear strain, not of superficial slumping, and that the shearing was accommodated by significant brittle fracturing, largely on a small scale, besides more ductile behaviour.

These melange zones of the Hudson Valley are a part of the southern extension of the Champlain Thrust and its subsidiary faults. Vollmer (1981a) and Vollmer and Bosworth (1984) pointed out that melange produced by this thrusting is unconformably overlain in the southern Capital District of New York by the earliest Devonian Helderberg Group carbonates. The unconformable contact constrains the formation of the melange, and the thrusting which produced it, to be a product of the Taconic Orogeny, and hence a purely Ordovician event. This unconformable relationship applies to the deformation in the westernmost (and hence youngest) part of the Taconic fold and thrust belt, implying that similar melange now east of the outcrop of the unconformity is also a product of Ordovician tectonism. The Champlain thrust system links the deformed flysch of the central Hudson Valley to similar sections in southern Quebec; it is a purpose of this guide and field trip to point out that there is much greater similarity between the marginal zone of the Taconic fold and thrust belt in these two areas than might be inferred from the existing published maps which, despite similar poor outcrop, clearly indicate the importance of melange and the extent of the marginal deformed zone in Quebec (St. Julien and Hubert, 1975; St. Julien et al., 1983; Avramtchev, 1989), but do not in New York.

STRATIGRAPHIC TERMS

Names and assigned ages of rocks in the deformed belt of the Ordovician flysch of New York are badly confused, both because of the structural complexity and because of the application, by Ruedemann and subsequent workers, of biostratigraphic names to inadequately defined lithic units. To the west of the deformed belt, where strata are close to flat-lying, the black shales of the basal part of the foreland basin sequence are termed Utica shale [Canajoharie shale has been biostratigraphically distinguished as slightly older than the type Utica, but is lithologically not distinguishable from it]. This shale is overlain by rapidly coarsening-upward flysch, which is termed Schenectady Formation in the eastern Mohawk Valley. West of the central Mohawk Valley this is replaced by the Frankfort Formation which is not that different lithologically, although on average containing somewhat thinner-bedded greywackes, and in part somewhat

Fig 2. Geology of the Ordovician flysch and melange, Albany-Saratoga Lake area, and field trip stops



younger because of the westward-younging diachronous flysch fill of the axis of the foreland basin (Rowley and Kidd, 1981). While it is probably least disruptive, and partly justified by historical practice, to continue to have separate names for these two areas of equivalent and similar strata, it is not justified to place Frankfort under Schenectady, a practice started by Fisher (1977, 1980), because the maps (Fisher, 1980) and the outcrops show them to be lateral equivalents with each starting directly on the Utica shale. Also, thin-bedded, shale-rich sections indistinguishable from the Frankfort are found within the Schenectady Formation, and coarser greywacke-bearing sections indistinguishable from the Schenectady are found in the Frankfort. As soon as one enters the deformed belt of flysch near Schenectady along the Mohawk River, things get (for stratigraphic terms) much worse!

The greywacke turbidite-shale flysch strata, when they are encountered in a structurally undisrupted but folded condition in the deformed zone of the Hudson Valley, are at present commonly known as Austin Glen Member [of the Normanskill Formation] [or Austin Glen Greywacke] from a definition by Ruedemann (1942) using a locality near Catskill (Figure 1). Regrettably, no section is available that shows the stratigraphic base or top of these strata. Furthermore, they are not distinguishable, despite the claim to the contrary by Rickard and Fisher (1973), from the coarser parts of the Schenectady Formation, which does have a defined base, and top (albeit erosional). Identical strata where they are unquestionably in the stratal sequence of the Taconic Allochthon, stratigraphically overlying the Mount Merino chert, are termed Pawlet Formation in northern New York and adjacent Vermont (Shumaker, 1967; Rowley et al., 1979), but have been termed Austin Glen farther south, towards and beyond the Capital District. Ruedemann (1901a; 1930) was previously responsible for starting use of the term Normanskill Formation mostly [judged by outcrop area] to describe identical rocks, using a type locality to be seen on this trip (Stop 8; Figure 2), in the southern part of the City of Albany. This term has become so biostratigraphically and chronostratigraphically ensnared, and applied indiscriminately by Ruedemann, and by others (e.g. Ruedemann, 1942; Rickard and Fisher, 1973; Berry, 1962, 1963; 1977) to rocks which are utterly different lithologically [specifically the red Indian River Slate, and the black and green Mount Merino chert, both of the Taconic Allochthon stratigraphic sequence] that it should be abandoned by those wishing clearly to identify still stratified greywacke-shale rock units in the deformed zone. Thus Austin Glen Greywacke, at least specified as a member of the Normanskill "Formation", is a term also contaminated by this association, and unwanted attendant biostratigraphic implications.

It might be less confusing to most geologists, not natives to the area, to use Schenectady Formation for the greywacke-shale flysch, whether these strata are folded or not. The alternative, besides creating yet another name, is to elevate the Austin Glen to separate Formation status for the folded greywacke-shale facies of the flysch, as long as this is understood to include explicit and complete divorce from Normanskill biostratigraphic and age associations. We favour the latter proposal, and use it below, because it will allow clear separation of deformed from undeformed rocks on future maps. We acknowledge that this promotes continued use of one nominally redundant lithostratigraphic name, and that the Austin Glen cannot, because of its structural setting, be given a "proper" type section with base and top both included.

Shale-dominated rocks in the deformed belt are mostly in the form of melange. Intact or nearly intact stratified shale and thin-bedded silty to fine sandy greywacke, and these rock types in the form of "broken formation" transitional to melange, are found also, forming lens-shaped belts. These rocks, including the melange, have been inflicted with various stratigraphic terms indiscriminately, without regard to their structural condition. We think it is important, in order to understand them, to distinguish between still-bedded strata and the melange; for that reason alone we reject the application to the areas of melange of stratigraphic terms based on stratified type sections.

One term, Snake Hill Shale, has perhaps been most widely applied to the shaly rocks of the deformed belt, although we regard it as particularly inappropriate because, at the type locality, on Saratoga Lake, the rocks consist of medium-bedded greywackes, some quite calcareous and containing abundant brachiopod fauna, with lesser shales interstratified. Apart from the presence of the fauna, the arenites are otherwise very similar to those in the Schenectady Formation, and the Austin Glen Formation. The unusual lithology is only seen at a few other places in the area, and the fact that it is unusual, besides the fact that the stratified, fossil-bearing part of the section at Snake Hill is not dominantly shale, makes Snake Hill an entirely unsuitable term for the large areas of shaly rocks forming the bedrock of the deformed flysch zone of the Capital District, whether they are in the form of bedded strata or in the form of melange. Another term used by Ruedemann (1930) is Canajoharie shale [now designated the basal Utica shale], which is not appropriate because of the widespread presence of thin greywackes in these rocks, where they are not melange, in all but one area, and the fact that they consist, in all but the same one area, of grey shales, significantly unlike the black, carbonaceous shales of the Utica Formation at Canajoharie, and elsewhere.

There are localities where black graptolitic shales occur in the belt of melange and highly deformed flysch, including one just southeast of Snake Hill, but these are slivers and blocks in grey shale matrix melange. We suggest that neither of these terms is appropriate for even the minority of bedded shale and thin silty greywackes in the deformed belt, and certainly not for the melange. There is one area where black, non-greywacke-bearing shale occurs in the western belt of folded and faulted flysch, between Ballston Spa and Saratoga Lake, and including a very prominent outcrop of flat-lying black shale in the median of Interstate 87 about a mile north of exit 12 (see Figure 2). Poor overall outcrop prevents determination of whether this area of Utica Shale is exposed due to updoming of Utica from beneath the grey shales and wackes of the Stillwater and Austin Glen of the folded and faulted flysch Zone, or due to local overthrusting of the Utica over those grey shales and wackes.

The Schenectady Formation contains sections that are thin-bedded and shale-dominated, and this unit, with a facies designation (shale facies, as opposed to greywacke facies) might be an appropriate way to designate the areas of little-disrupted shale-dominated flysch in the deformed belt. Alternatively, in keeping with the proposal to formalize Austin Glen Formation for the folded greywacke-shale facies, a separate [and new] name would be needed; we favour this alternative and suggest Stillwater [Shale] Formation for the good exposures of this unit around that town (Figure 2), and specifically along Schuyler Creek, up to about 1.5km west of the Hudson River.

We propose that the term Snake Hill be restricted to occurrences of the brachiopod-rich wacke/shale facies seen at Snake Hill, and that it be specifically designated Snake Hill facies of the Austin Glen Formation, because of the rarity of the occurrences, and the inclusion of most of them as blocks or slices in melange.

MELANGE

Specific lithounit terms for parts of the melange have only been given by two authors. Zen (1961) proposed "Forbes Hill Conglomerate" for a very specific pebbly olistostromal unit only found (as originally defined) adjacent to the northernmost Taconic Allochthon. This conglomerate in its type area and location is not at all a typical occurrence of the melange, and also has a poorly exposed type locality. Bird (1969) used this term to apply to the melange in general; we think that this usage ought to be dropped for this context, since most of the melange is emphatically not conglomerate. Bird (1963, 1969) also used the informal lithological descriptive term "wildflysch" to characterise the melange; this term is not at all inappropriate. Fisher (1977) introduced "Poughkeepsie Melange", based on a few outcrops that are still not adequately understood in terms of relationships to their surroundings (i.e. not mapped in detail). Because we think that these two names have significant defects, we propose a new lithostratigraphic name, the Cohoes Melange, for the melange of the deformed flysch belt of the Hudson Valley. We propose as a type locality the excellent cliff and riverbed outcrop on the north bank of the Mohawk River from Cohoes Falls to the end of outcrop in the riverbed below the dam and spillway east of the Waterford-Cohoes Route 32 highway bridge (to be seen at Stop 4; Figures 2 and 5). This section contains both "exotic" and "non-exotic" types of melange (see below). Because the melange is a product of disruption of stratified rocks by structural processes, it is not feasible to define a type section that includes top and base; for that reason it is not appropriate to define it as a conventional "Formation". However, it is possible to specify a well-exposed, accessible type locality, and to define clearly the lithologic characteristics and contents. We base these on the mapping and detailed descriptions of Plesch(1994), and Vollmer (1981a).

The detailed mapping of Plesch (1994) identified two main varieties of melange in the Capital District. One contains only greywacke blocks in grey phacoidal shale matrix, both components being ultimately derived from the bedded flysch. The other variety has a more complex derivation, with blocks that are "exotic", at least in comparison with the exclusively mundane greywacke blocks of the first-mentioned variety, and with at least two types of shale for the phacoidal matrix. We emphasise that the term "exotic" is used in a relative sense, and that no blocks in the New York Taconic melange are exotic in the sense that term is often used elsewhere, for example to denote blueschist and eclogite blocks in the Franciscan of California. With the single exception of the basaltic pillow lava of Stark's Knob [near Schuylerville], all known "exotic" blocks in the Taconic melange are sedimentary rocks, and most of the types can be matched with lithic units of the Taconic Allochthon stratigraphic sequence. In the melange belt of the Capital District, exotic blocks are seen in the largest [and best] exposures to be arranged in distinct zones, often in a systematic assemblage. However, it is impractical to distinguish on maps all the zones with exotic blocks versus those without such blocks, partly because of the generally poor outcrop, and because many of the alternating zones are too narrow to show; because of this, we do not propose separate lithostratigraphic names for exotic-bearing and non-exotic-bearing melange.

Within the exotic melange, there are several distinctive block/slice lithologies, some of which have (in some places) been given specific names. The most widespread of these lithic types are cherts of dark grey, black, and green aspect, and which are identified confidently as samples of the Mount Merino (chert) Formation, a unit native only to the Taconic Allochthon stratigraphic sequence (Rowley et al., 1979). These cherts, to be seen at Stop 8 (Figure 9), and closely related black argillites/slates, contain the most prolific and best preserved graptolite faunas (of *Nemagraptus gracilis* age) obtained in the belt of deformed flysch and melange. Unfortunately, it has been presumed by Ruedemann (1930), who described the faunas, and by others (e.g. Berry 1963, 1977; Rickard and Fisher, 1973) that the Mount Merino chert was a lithostratigraphic member of the flysch of the Hudson Valley, a notion propelled by the inclusion of the chert in the Normanskill "Formation". We maintain that the contacts of these black cherts and slates are everywhere tectonic within the belt of deformed flysch greywackes, shales, and melange. Given their origin from the allochthonous strata of the Taconic Allochthon, they are evidence for out-of-sequence and structurally late formation of at least some of the melange; Taconic slices were emplaced over flysch and then both were mixed together by out-of-sequence thrusting and melange formation. The other unit given a specific name in this area is the Ryesdorph Hill conglomerate of Ruedemann (1901b, 1930). This carbonate conglomerate/breccia, to be seen at stop 6 (Figure 7), also has suffered the presumption that it was part of the stratigraphic sequence of the flysch. We are of the opinion that it is a block in the exotic melange, and that the youngest fauna in this block is only a constraint on the maximum age for the formation of the melange in which it is contained. Other occurrences of carbonate blocks, mostly breccias, are not known to contain faunas with the exceptional age range Ruedemann painstakingly extracted from the Ryesdorph Hill locality. Other rock types found in the exotic melange include distinctive rusty-weathering sideritic carbonate mudstone, whose source is unknown, but which is unlikely from comparison to be the Burden Iron Ore found south of Hudson within the Taconic Allochthon stratigraphy (Hofmann, 1986). In the melange matrix, bright green shale, which is not found in the melange that bears only greywacke blocks, is also an "exotic" lithotype; it always (reliably) accompanies other exotic lithotypes. A few occurrences of brachiopod-rich wackes, like those seen in stratified rocks at Snake Hill on Saratoga Lake, also occur in the exotic melange, particularly in Cohoes Gorge where (Riva, J., pers. comm, 1983) they also contain the trilobite *Cryptholithus tessellatus*, and in part of the western long road cut on the new Route 7 in Latham.

Discussions of age relations of the flysch and melange, from fossils (Berry, 1962, 1963, 1977; Rickard and Fisher, 1973), derived either from within blocks in melange, or from bedded sequences in the deformed flysch belt, have not been particularly conclusive, partly because of a failure to establish a clear lithostratigraphic framework, and partly because the highly tectonised condition and structural complexity of this belt of rocks was not sufficiently recognised. We suggest in particular that the Hudson Valley flysch and melange is not as old as previously inferred, because *N. gracilis* faunas only occur in blocks of Taconic Allochthon-derived black chert and slate. Also we suggest that none of the presently known faunas in the greywackes necessarily demand (but do not rule out) out-of-sequence thrusting or deposition in basins ("lower slope basins") stratigraphically upon previously deformed flysch and melange.

Several major belts of melange occur in the Capital District (Figures 1 and 2), based on the outcrop maps of Plesch (1994) and Vollmer (1981a). We term these belts "zones", not slices, because they are highly likely to contain significant faults, and are not just bounded by faults. The widest zone occupies the center of the deformed flysch belt; we term this the Mohawk River Central Melange Zone. It is divided by the Halfmoon Greywacke Zone (intact folded Austin Glen Formation) into eastern and western parts. It is impossible to separate these parts or to locate the boundary precisely where the Halfmoon Greywacke Zone is not present, but we think there is good evidence (see Plesch, 1994) that this boundary is a fault that must continue to north and to south of the Halfmoon Greywacke Zone, in particular to link with the Rocky Tucks Greywacke Zone along strike to the north. The Central Melange Zone contains significant occurrences of exotic fragments, across the full width of its outcrop. The zone of bedded shaly rocks around Stillwater appears to further divide the eastern Central Melange zone in the northeastern portion of Plesch's map. The western margin of the Central Melange adjoins the belt of less-deformed folded and faulted flysch (Figure 2). The eastern margin is formed by the 1-2 kilometer-wide belt of mixed broken formation, small intact bedded blocks, and melange, mostly formed of, or derived from, shaly to fine arenite flysch, and which lacks "exotic" components; this belt we term Waterford Flysch Zone. The other mappable melange belt is the one that is clearly localized adjacent to the Taconic Frontal Thrust, and which truncates some zones of the bedded flysch and melange to its west. This melange we term Troy Frontal Melange Zone, for its well-exposed section in the gorge of the Poestenkill in Troy (to be seen as Stop 5). The Troy Frontal melange is prominently "exotic"-bearing, although there appears to be an exotic-poor, greywacke block-dominated zone in the hundred meter width adjacent to the Taconic Frontal Fault, at least in the two

exposures of this interval to be seen on the trip [Stops 5 and 6]. The melange at Poughkeepsie noted by Fisher is possibly equivalent, at least in part, to this one.

STRUCTURE OF THE MELANGE AND FLYSCH AND ITS ORIGIN

Structural features of the Taconic melange in eastern New York have been most recently studied by Vollmer (1980, 1981a, 1981b), Bosworth and Vollmer (1981), Vollmer and Bosworth (1984), Bosworth (1989), and Plesch (1994). The melange formed by the progressive deformation of synorogenic flysch as an accretionary thrust wedge advanced over the Ordovician North American continental margin, creating, and supplying sediment to an active submarine foreland-type basin. The flysch was derived from, and was progressively accreted to and overridden by the Taconic Allochthon, resulting in the formation of belts of tectonic melange.

Three principal mechanisms are believed to have operated to form the melange: folding, boudinage and disruption of graywacke-shale sequences due to viscosity and ductility contrasts; imbrication and out-of-sequence thrusting resulting in intercalation of sedimentary units; and tectonization of olistoliths, and possibly slumps, derived from exposed fault scarps. Exotic clasts in the melange are probably both of sedimentary and of structural origin. In some locations pebbly mudstones are preserved without extensive deformation, suggesting that some of the exotic melange could have formed by later deformation of such deposits. In other cases exotic fragments appear to have been introduced by structural imbrication.

The state of consolidation of the rock during melange formation may have varied considerably, but it is thought that melange formation was dominated by lithified rock tectonic processes. Folded veins, brittle failure at fold hinges, axial plane cleavage cross-cutting veins, consistent orientation of axial planes, and a general absence of criteria indicating soft sediment deformation were found by Vollmer (1981a). In particular, steeply plunging isoclinally folded calcite veins suggest that much of the strain was accommodated after the rock was consolidated enough for extensional vein formation to occur (Vollmer, 1981a). Extensive veining is presumably related to high pore pressures, however, and only a limited degree of lithification may have been required for brittle failure to occur.

A progressive change in the orientation of fold axes was documented by Vollmer (1980, 1981a, Bosworth and Vollmer 1981, Vollmer and Bosworth 1984) in Albany County (stops 7-10). Away from the allochthon near-horizontal fold axes trend NNE. As the Allochthon is approached, fold axes swing 90 degrees to steep ESE plunges. This change is accompanied by increased fold tightness, stratal disruption, and development of a phacoidal fabric. The amount of rotation also appears to be lithology dependent, as folds in the presumably more viscous graywacke beds appear to have undergone less rotation. These observations suggest that fold rotation was progressive and was related to increasing strain (Vollmer 1981a).

The characteristic fabric of the melange is a phacoidal cleavage, which is intimately associated with high strain, bedding disruption, and fold rotation. The adjective [phacoidal] refers to the lensoid character of the shale chips and other fragments that define the fabric. Greenly (1919), who first applied the word melange to a rock body, used this term in his description of melange fragments. The term "scaly" is used to describe similar fabrics elsewhere (for example, Moore 1986), however, the term "phacoidal" more accurately describes the fragment shapes which define the fabric here (Caine, 1991). Studies of the Taconic melange fabric by Vollmer (1981b, Vollmer and Bosworth 1984), Bosworth (1989), and Caine (1991) show a complex three-dimensional network of anastomosing seams and conjugate shear planes. Although the development of this fabric is not fully understood, it appears to be only developed within high strain zones, and is considered to be a shear zone fabric. Similar fabrics are described from melanges worldwide. Outcrops at Normansville (Stop 9; Figure 10) show good examples of the phacoidal cleavage with associated stratal disruption and folding.

Some of the most complex deformation in the area was documented by Vollmer (1981a) within the Normanskill gorge (stop 7), the type locality of "Normanskill Formation". There, directly under the 9W overpass, a large rootless antiformal syncline is exposed (Figure 9). Flute casts clearly demonstrate its downward-facing character. Downward facing bedding/cleavage relationships can also be found. Vollmer (1981a) modeled the deformation there as either the refolding, or slumping, of a sequence of variably plunging folds with extensive stratal disruption. This location is clearly a singularly poor choice for a stratigraphic type section.

The upper limit to the age of melange formation is constrained by the unconformity exposed at Feuri Spruyt (Creek) near South Bethlehem (stop 10; Figure 11) which was described by Ruedemann (1930) and Vollmer (1981a). Here highly disrupted melange is unconformably overlain by carbonates of the lower Helderberg group. Although it could conceivably be argued that the contact at this locality is structural, the regionally extensive melange belt passing underneath the carbonates requires an unconformable relationship.

Late quartz veins in the melange rather consistently show slickenside fibres plunging to $120^{\circ} \pm 15^{\circ}$ (Plesch, 1994), which is an orientation clearly identified with the Champlain thrust (e.g. Stanley and Sarkisian, 1972); in contrast, slickenside fibres on thrusts and flexural slip surfaces in the Devonian Helderberg Group between Ravena and Catskill, just south of the Capital District, trend $090^{\circ} \pm 15^{\circ}$, the local Acadian Orogen orientation. This data also suggests most of the structure in the melange belt was formed in the Taconic Orogenic event.

The belts of Taconic melange are thus envisioned as broad shear zones, with dominantly "hard-rock" structures, that formed thrusts within the synorogenic flysch during the emplacement of the Taconic Allochthon. The east to west decrease in deformation intensity at the western margin of the deformed belt of flysch might be interpreted as reflecting a progressive decrease in strain rate as the Taconic thrusting ceased, although this cannot be conclusively demonstrated. Because of the evidence for out-of-sequence thrusting in the Taconic belt, it could alternatively be a petrified representative sample of the progressive incorporation of flysch into the accretionary thrust complex, a rate inferred rather speculatively by Bradley and Kusky (1986) to have been of the order of 2 cm/yr.

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FIELD TRIP LOG

	Increment	Mileage Cumulative
From Union College, follow Union St east to Rte 146, or Lenox Rd and Nott St [6] or Rosa Rd/Providence Ave [8] east to Rte 146 and turn left to go north on Rte 146 Balltown Road		
Mohawk River bridge heading north on Rte 146	0	0
Junction 146 and River Road; turn right view over Mohawk to flat Schenectady Fm in south bank [0.6]	0.4	0.4
Junction with Grooms Road - keep right	1.3	1.7
Turn right onto Brian Drive	1.7	3.4
Park at end of road	0.15	3.55

Stop 1 - View of flat-lying Schenectady Fm. flysch and shale outcrop on south bank of Mohawk River from end of Brian Drive.

Outcrops are scarce in the belt of flat-lying to gently dipping greywackes and shales [flysch] east of the trace of the Saratoga-Ballston Lake normal fault. The best are in the south bank of the Mohawk River, and are not readily accessible to large groups. At this stop (see Figure 2 for location) a shaly unit of the Schenectady Formation flysch can be seen across the river underlying medium-bedded greywackes and shales with gentle west dips and no signs of internal deformation.

These rocks are autochthonous and have not been transported by thrusting during the Taconic Orogeny. About 300 meters to the east of this location, along the southern shore, there is an abrupt contact of these flat-lying strata with significantly folded and faulted thin-bedded greywackes and shales, defined by the first thrust fault encountered in a west-to-east transect. This is not clearly apparent from this viewpoint, but a representative outcrop of the folded and faulted flysch is seen at the next stop.

Return to River Road, turn right	0.15	3.7
Crossroads at top of rise, turn right	1.4	5.1
Park in area on right before gate 0.2 5.3 walk down road to Stop 2 - folded and faulted flysch of Vischer Ferry Zone		

Stop 2 - Folded and faulted flysch of Vischer Ferry Zone - outcrop along north shore of Mohawk River near Vischer Ferry power plant/Lock 7

The rocks here form part of the Vischer Ferry Zone of folded and faulted flysch (Figures 1 and 2). Thin-bedded greywackes and shales, with subvertical dip, are exposed in the discontinuous roadcut going down the paved road and branching off onto the dirt track down to the shore of the river. At the shore,

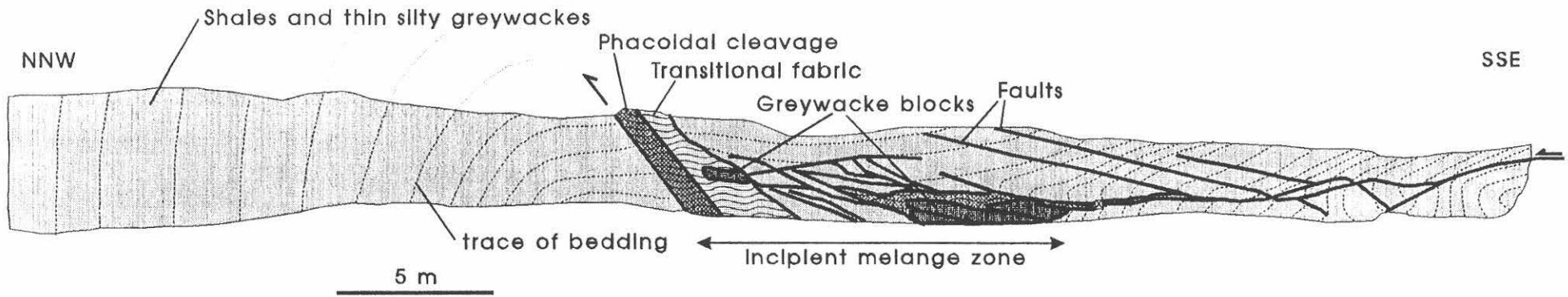


Fig. 3. Sketch section of outcrop [stop 2] adjacent to Vischer Ferry power station, north bank of Mohawk River (after Bosworth, 1989). Melange developing on small thrusts in thin-bedded shaly flysch facies.

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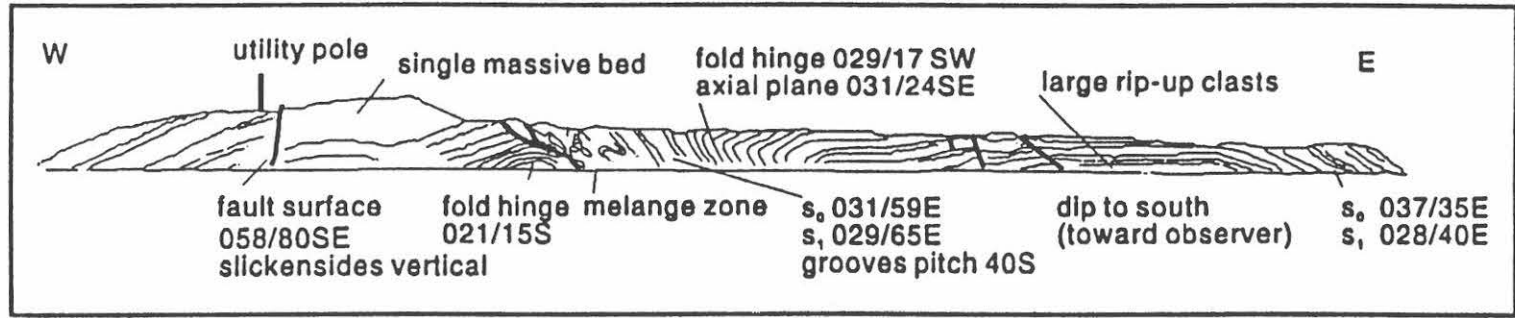


Fig. 4. Sketch section of road cut [Stop 3] on north side of Route 146, Clifton Park; folded Austin Glen Fm. with small melange zone on thrust.

similar thin-bedded flysch and shale comprises the outcrop extending west from the path to the abutment of the power plant (Figure 3). This outcrop displays asymmetrical folds and small faults connected with east-over-west thrusting. A detailed description of this outcrop appears in Bosworth (1989), who pointed out the incipient development of a melange-type fabric in the fault zone (Figure 3), mainly through fracture development and incremental slip on the fracture surfaces. Farther down the dirt track to the south-east, a small hillock and its shoreline on the river under the power transmission line exposes more subvertical, west-younging thin to medium-bedded greywackes and shales.

While the strata here are in part shale-dominated, enough arenaceous greywacke occurs here and in the surrounding areas for the Vischer Ferry Zone to be mapped as Austin Glen Formation across the Mohawk River. To the north, however, near Round Lake (Figure 2), the Zone changes to largely shale, lacking significant arenaceous wackes, and there it is specified to consist of Stillwater Shale. This change is gradational, and thinning and fining of the arenaceous component of the flysch is occurring, within the Zone, across the Mohawk River; we choose for consistency to include anything containing arenaceous greywackes within the Austin Glen Formation.

Return to junction with River Road	0.2	5.5
Proceed straight across, north along Sugar Hill Road		
Turn right onto Rt. 91, Grooms Road	1.8	7.3
Go east on Grooms Road to junction with Rt 9 [2.7 light; 3.8 cross over Northway]	5.5	2.8
Proceed across Rt. 9 following Rt. 91; turn left onto Rt. 236	0.2	13.0
Turn left onto Fellows Road	1.3	14.3
Stop sign at Rt.146, go straight across and	1.2	15.5
Park on right just after crossing	.05	15.55
walk to east along Rt. 146 inside the guard rail to roadcut Stop 3.		

Stop 3 - Thick-bedded Austin Glen Formation flysch of Halfmoon Greywacke Zone, folded, with melange on small thrust fault - road cut on north side of Route 146 east of Clifton Park

This outcrop (Figure 4) mostly exposes structurally intact medium to thick-bedded greywackes and lesser shales in a broadly antiformal structure, hence the popular name for this locality, Clifton Park Anticline. However, in the center of the outcrop there is a thrust fault, which can be identified from the zone of overturned steeply-dipping thinner greywackes and shales that form the lower limb of a close fold in the immediate hanging wall of the fault.

This fold might be a tightened hanging wall ramp anticline. To the west [structurally below] this fold structure, there is an approximately 3 meter wide zone consisting of broken blocks of greywacke in a disrupted shale matrix, which is melange derived from the adjacent materials, and defines the fault zone. We think that this fault is an excellent example that shows in miniature how the larger zones of melange formed.

This outcrop is part of a large structural slice, which we term Halfmoon Greywacke Zone (HGZ), that consists of relatively intact, medium-thick bedded greywackes like these, surrounded both to east and west by highly disrupted melange, including some materials exotic to the immediate greywacke-shale association. This HGZ slice consists along the Mohawk River of a disrupted large-scale synform; however we do not favour the klippen interpretation shown for it on the New York State Geological Survey maps since 1970; in particular there is no evidence for a westerly dipping, nor a relatively west-downthrown fault/fault zone on its eastern side.

Excellent sedimentary structure assemblages may be seen in the greywackes in this outcrop indicating deposition by turbidity currents, and local reworking by bottom currents [beds completely occupied by climbing cross-laminations], perhaps the tails of turbidity current events. Claims of the existence of shallow water sedimentary structures, particularly mudcracks, have occasionally been made for the Schenectady Formation, and for greywackes of the same structural slice as this outcrop near the Mohawk river (Rickard and Fisher, 1973); we think that these reports are mistaken, and the structures are probably either syneresis cracks, or tectonic structural features. All greywacke-bearing outcrops we have seen in the Hudson Valley and in the Schenectady Formation have turbidite characteristics, certainly deposited below wave base, and likely in substantially deep marine water given the analogy with modern arc-passive margin collisions, such as Timor-Australia.

Turn around, take right at stop sign onto Rt. 146 west.		
Turn left at light at junction with Rt 9	1.6	17.2

Proceed south on Rt. 9 [Western Exotic Melange just to west near Sitterly Rd 1.2mi; 2.3 Grooms Road; 3.4 Crescent-Vischer Ferry Road; 3.6 center of Mohawk bridge]

Turn left onto Cohoes-Crescent Road	3.9	21.1
Proceed along south bank of Mohawk River into Cohoes [0.4 view over river to outcrops of eastern Melange; 0.9 Crescent Dam and power plant; 2.5 Cohoes Falls on left; 3.0 Harmony Mills; 3.6 railroad crossing]; turn left at intersection with Rt. 32	3.6	24.7
Cross Mohawk bridge [0.2]; take left turn at light Clifton St.	0.4	25.1
Right turn onto Grove Street	0.1	25.2
Left turn onto Grace Street	0.1	25.3
Left turn onto Columbia Street	0.3	25.6
Park near end of street [turn around first]	0.1	25.7

Walk south from end of street into open ground; angle 45 degrees right and cross main path along cliff top; find path down cliff close to where bushes start [distance to the top of this path from end of pavement on Columbia Street is less than 100m]- Stop 4 - Cohoes Melange outcrop, eastern Mohawk River Central Melange Zone; down path and along base of cliff to Cohoes Falls.

Stop 4 - Cohoes Melange outcrop [eastern Mohawk River Central Melange Zone], cliff on Waterford side of Mohawk River below Cohoes Falls

This outcrop and its continuation up and downstream constitutes the best, and most continuous section through the melange of the Hudson Valley. Structural features seen here, and elsewhere in this belt indicate a "hard-rock" origin for the melange, particularly the striations visible on many of the characteristic "phacoidal" or lenticular splinters of the shale, and the polished nature of these surfaces in places when freshly exposed. Clear evidence of olistostromal origin of units with blocks in matrix is found in a few places, but the large majority of the melange in the region does not show features that require a surficial mass-wasting origin. As long as shear strains were high, imbrication and mixing of materials could be achieved in large ductile thrust fault zones, as long as some of these were out-of-sequence thrusts, in order to introduce originally overthrust materials [from the Taconic Allochthon] among the medial Ordovician flysch-derived melange matrix. The latest structures here and elsewhere in the melanges consist of quartz veins, some fibrous, and slickensided, which in many places, including here, show a consistent 120° plunge; this orientation is known to be associated with the Champlain Thrust in Vermont, which is an along-strike continuation of the faults defined by the Hudson Valley melange zones. This fault displacement direction contrasts with the ~090° plunge of thrust-related slickensides in the Devonian Helderberg Group to the south between Ravena and Catskill.

The section here consists (Figure 5) of alternating zones of exotic-containing and less- to non-exotic-bearing melange. A large isolated wacke block is seen part-way down the cliff path; lower down, several large carbonate breccia blocks are exposed, possibly derived from a Taconic Allochthon source, or from normal fault scarps in the outer trench slope prior to arrival of thrusting and melange formation. Just east of the bottom of the path is a large block of black argillite, which is in our opinion derived from the allochthonous strata of the Taconic Allochthon, specifically from the upper Mt. Merino-basal Pawlet Fm. interval, which contains, regionally, black graptolitic argillites with *Nemagraptus gracilis* faunas like the one collected and identified by Riva [pers. comm., 1981] from this block. Smaller blocks of chert, also originally from the Mt. Merino Fm. of the Taconic Allochthon, occur at the base of, and above Cohoes Falls. The remaining materials in this part of the section consist dominantly of greywacke and shale [flysch], although there is also a persistent occurrence of greenish shale seams and films in all areas where the exotic melange occurs, as well as scattered small blocks of sideritic mudstone whose stratigraphic source is unidentified, and both of which are not seen in areas just containing greywacke and shale and their melange derivatives. Farther east in this outcrop than the black argillites, there are a few occurrences of blocks/lenses of a fossiliferous wacke, very similar to that exposed in a partly-disrupted bedded section at Snake Hill on Saratoga Lake.

This brachiopod/trilobite fauna contains *Cryptolithus tessellatus* [Riva, pers. comm, 1981]. Because of its local occurrence, we suggest that this lithology represents input to the flysch trough from a

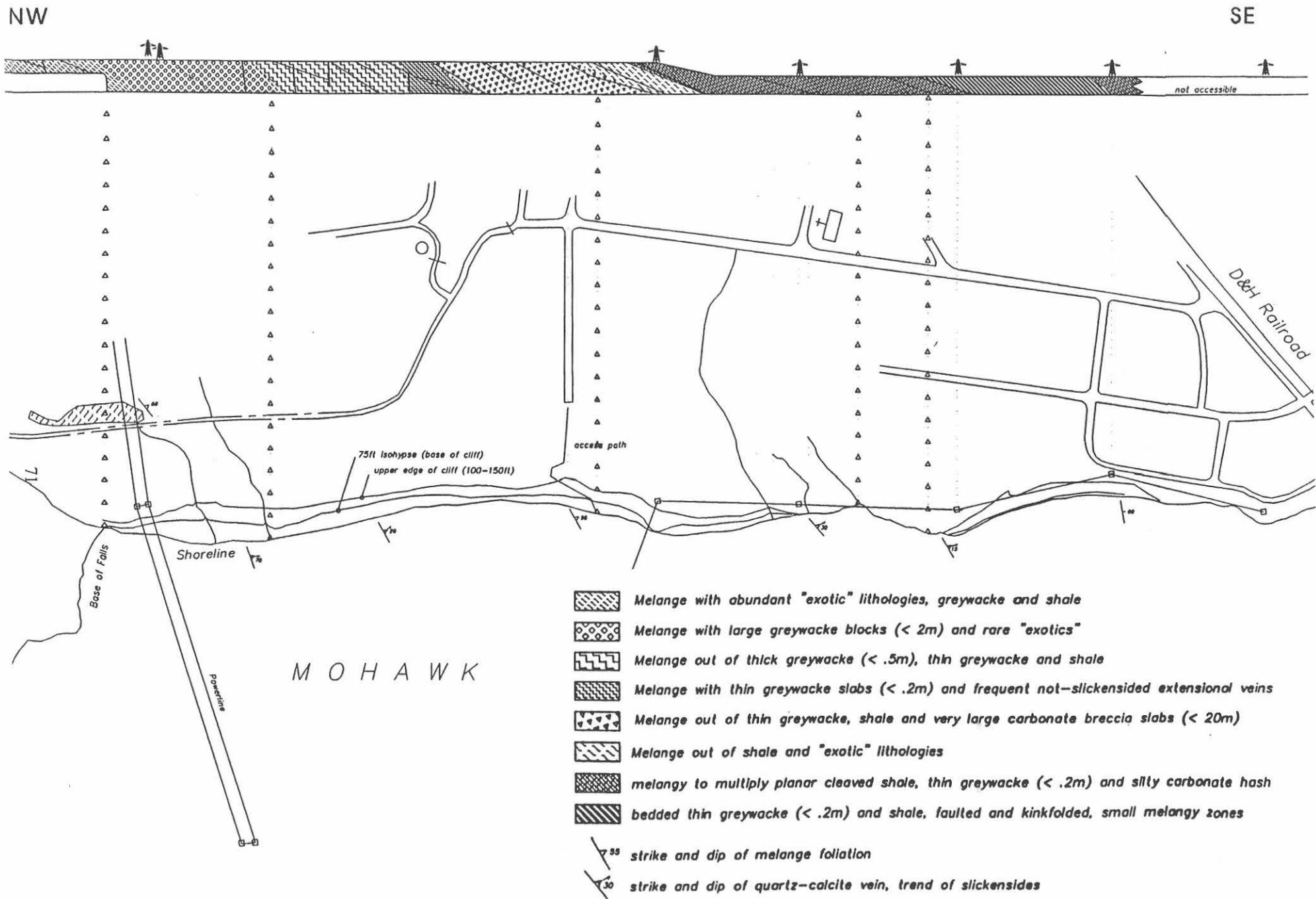


Fig. 5. Sketch section and location map from Plesch (1994) of north cliff of Mohawk River downstream from Cohoes Falls [Stop 4]. Type section for Cohoes Melange, including "exotic"-clast facies, in Mohawk River Central Melange zone.

local across-axis source, unlike the majority of the flysch, which clearly has a southerly derivation and along axis transport, from the paleocurrent evidence.

If the water level is very low, excellent exposures of incipiently to strongly disrupted shales and silty wackes of the Waterford Flysch Zone may be seen just below the dam located east of the Waterford-Cohoes road bridge; these represent a discontinuous belt of in part less-deformed rocks forming a slice, or set of imbricate slices, on the eastern margin of the Mohawk River Central Melange. Access to this outcrop, at low water, may be obtained from either the north or south side of the Mohawk River, east of the bridge.

Return to Rt. 32 reversing directions above; turn right	0.6	26.3
Turn left at light onto Interstate 787 [2.6 cross under Rt. 7]	0.3	26.6
Take exit for Rt. 378 East	6.0	32.6
Cross Menands Bridge over Hudson River; curve to to left to light; go straight	1.0	33.6
Intersection with Rt. 4. Fork left to follow Rt. 4 north [shallow right fork to follow Rt. 4 at 0.3 and 0.6]	0.1	33.7
Turn right at Canal Ave [currently not marked - just before small bridge, one way sign] [if you miss this turn, take next available right; at intersection with Hill Street turn right, pick up log at bridge/Canal Ave]	1.3	35.0
Turn half right onto Hill Street	0.3	35.3
Turn left onto Linden Ave	0.1	35.4
Turn left into parking lot next to road	0.2	35.6
Walk down path to Stop 5 - Poestenkill Gorge, Taconic Frontal Fault and Troy Frontal Melange Zone		

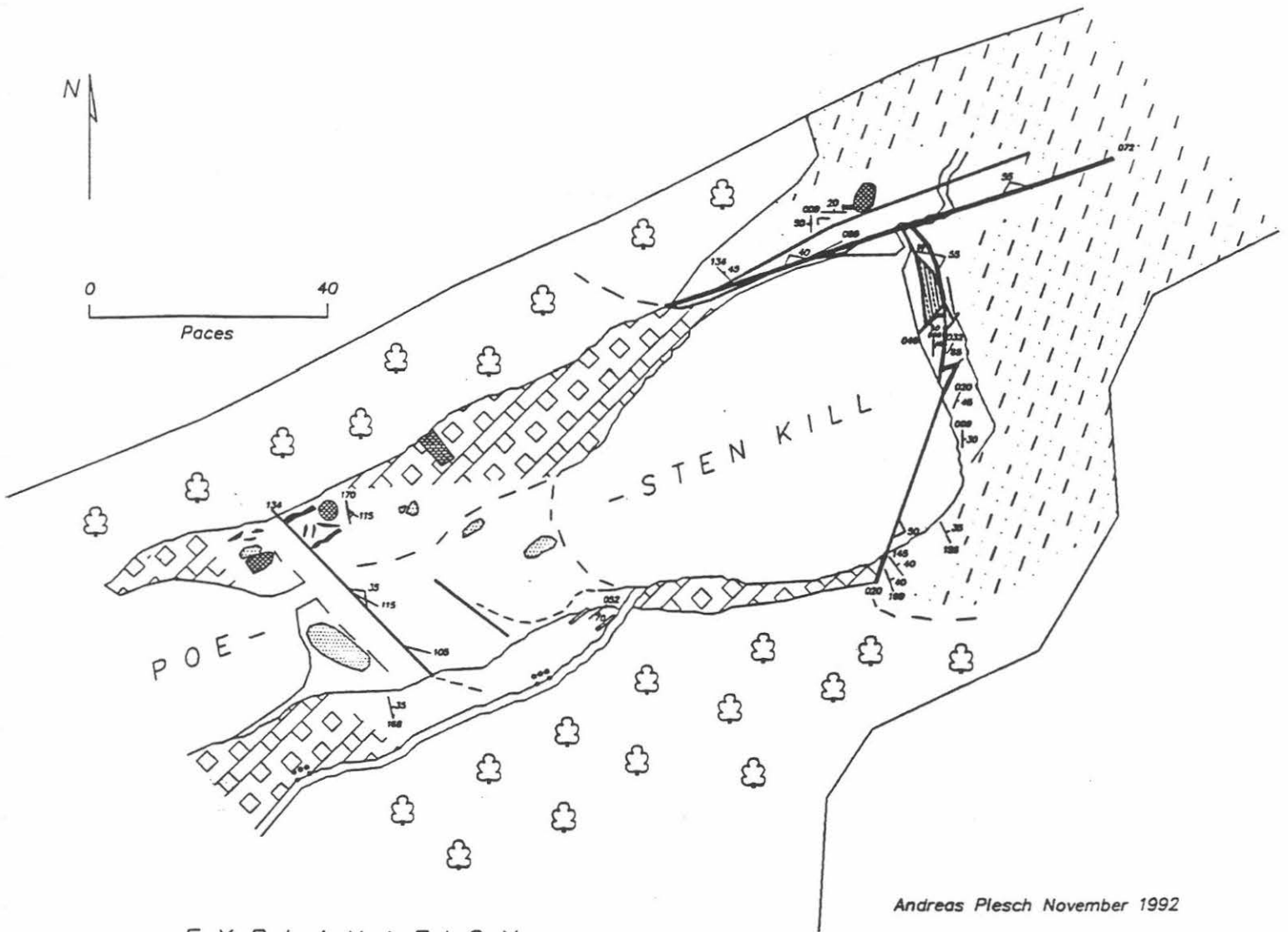
Stop 5 - Poestenkill Gorge, Taconic Frontal Fault and Troy Frontal Melange Zone

This is one of the few exposures, and certainly the best (Figure 6), of the Taconic Frontal Fault anywhere along its ~200 km length. Green arenites and silty wackes and slates of the Bomoseen Formation are in fault contact here with greywacke block-rich grey shale melange of the Troy Frontal Melange Zone. The original (presumed thrust) fault contact has been cut and displaced by a later steep strike-slip fault at this locality, so that most of the exposed length is of the latter; the thrust may be seen under and to the south of the waterfall. It dips quite steeply [$\sim 55^\circ$], as is also the case at the few other exposures of this fault; a possible reason for this includes [back] rotation by later thrust ramps. It is the case that outcrop evidence of thrust sense of shear for this fault, here and at the few other localities where it is exposed, is not overwhelmingly convincing. An alternative explanation for the steep dip, also lacking convincing outcrop evidence, is that the fault is extensional, or has a component of late normal displacement, and formed late in, or long after, the Taconic orogenic event. Even if extensional, this in no way contradicts the regional relationships demonstrating the overthrust origin for the Taconic Allochthon and the belt of melange and deformed flysch.

Leaving parking lot, turn right		
Intersection with Hill Street; turn left	0.2	35.8
Go up hill to sharp intersection with Campbell Ave; turn sharp right	0.9	36.7
Intersection with Rt. 4, Vandenburg Ave, turn left onto Rt. 4 at next light at top of hill, take center lane for straight ahead Proceed south on Rt. 4 [2.0 RPI Tech Park; 3.7 Rt. 43 enters from left; 4.4 Rt. 43 leaves to right; 5.6 cross Interstate 90]	1.1	37.8
Turn right at intersection with Rt. 151 at Couse Corners	6.2	44.0
Park well off on hard shoulder near east end of large roadcut	1.0	45.0

Stop 6a - Troy Frontal Melange Zone in road cut on Rt. 151 west of Couse Corners

This cut (for location, Figure 7) exposes in all the north side and most of its length on the south side a similar greywacke block-rich shale matrix melange as seen at stop 5, the Troy Frontal Melange Zone. The exposure lies a few hundred meters west of the [here unexposed] trace of the Taconic Frontal Fault. It is close to a unique locality first investigated by Ruedemann (1901b), the carbonate conglomerate of Rysedorf Hill [see below, stop 6b] which is the main reason for stopping here in addition to stop 5.



EXPLANATION











-  taconic lithologies, greenish siltstone, partly with quartzite lenses well bedded with slumping folds, tightly folded
-  greywacke-rich melange, phacoidally cleaved shale matrix chaotic structure
-  interbedded grey and black shale, flysch facies folded with melange zones
-  largest individuals of greywacke blocks in melange
-  slickensided quartz-calcite veins with strike and dip of vein and trend of lineation
-  strike and dip of general foliation in melange fabric
-  strike and dip of bedding planes
-  smaller veins (mostly slickensided)
-  area of broad (10cm) extensional quartz-calcite veins, frequently pegmatitic and open
-  large holes in bedrock

Fig. 6. Sketch geological map of Poestenkill Falls, City of Troy [Stop 5]. Exposure of Taconic Frontal Fault and Troy Frontal Melange Zone.

Within the greywacke-shale melange here, late quartz veins like those seen earlier cut all other structures. One block of wacke about 20 meters from the east end of the cut on the north side shows a slate with cleavage oblique to bedding on part of one margin, and this cleavage is clearly cut by the lenticular phacoidal melange matrix fabric. Some of the wacke blocks in this cut have angular, fracture-controlled margins, others have a more rounded and smoothed appearance, possibly indicating they were not fully consolidated upon incorporation. The westernmost outcrop along the south side of Rt. 151 here shows a quite different assemblage, with both dark grey/black and green shale melange matrix, and only modest-sized [~10-40 cm] sideritic mudstone blocks; it is identical to parts of the exposure in Cohoes Gorge, seen at stop 4. The contact with the greywacke block-rich melange to the east is not exposed, but is inferred to be a significant fault. This melange type forms the matrix on the west side of the exposure of the Rysedorph Hill conglomerate, located a few hundred meters to the north.

Turn right onto Olcott Lane	0.1	45.1
Park on right- ask permission at house on west side of street	0.2	45.3

Stop 6b - Rysedorph Hill Conglomerate. ASK PERMISSION at the house on the west side of Olcott Lane BEFORE entering this property. Please DO NOT hammer this outcrop.

A limestone conglomerate, from which Ruedemann obtained a fauna ranging in age from the Cambrian through the medial Ordovician, forms a unique exposure on part of this hill. Walk up the slope within the woods from the unsurfaced Olcott Lane, starting before the first bend to the east (Figure 7). Dark and green shale melange with small sideritic mudstone blocks is found, mostly as loose shale chips and fragments, on the slope. The limestone conglomerate and breccia is found forming a knoll, and also as loose fragments, near the top of the first slope. Much written about this famous [notorious?] locality presumes it to be in stratigraphic arrangement with the surrounding shales and greywackes. As you have seen from the roadcut, this is an untenable hypothesis, and the conglomerate/breccia must form a large block, in the exotic matrix melange, or on the contact of this melange with the block-rich greywacke melange to the east. Nonetheless, the large age range of the fauna in this small occurrence of limestone conglomerate needs explanation, even though it is now contained in a block within melange. We suggest that the limestone conglomerate/breccia may be explained by talus accumulation below a normal fault scarp produced by flexural extension in the outer trench slope of the Taconic foredeep basin [Bradley and Kidd, 1991], and the subsequent picking up of a sample of this breccia into the advancing accretionary thrust wedge bordering the Taconic Allochthon. Strata containing physically similar carbonate conglomerates/breccias do occur within the intact stratigraphic sequence in the Taconic Allochthon, but they nowhere have extended faunal age ranges, and none are known with ages younger than early Ordovician.

Return to Rt. 151, turn right	0.2	45.5
Turn left onto Sherwood Ave	0.3	45.8
Turn right onto Columbia Turnpike, Rt. 9 & 20	0.7	46.5
Proceed straight down hill into Rensselaer;		
last light before road curves to right is at:	1.5	48.0
Proceed ahead onto ramp to Dunn Memorial Bridge over Hudson		
River; at western side take exit marked for Interstate 787 South		
[0.7 from last light]		
South on I-787 to next exit, marked for Port of Albany [at 1.5]		
End of ramp, at light, joins Rt. 32; turn left	1.8	49.8
South on Rt. 32; at light turn part right onto Old South Pearl Street	0.6	50.4
Go to dead end of street, turn, and park	0.25	50.65

Walk up path on left side of culvert at end of street to railroad embankment, turn right, and walk up trackbed 0.1mi to Stop 7 - outcrops of melange and flysch in cut and bed of Normanskill [Ruedemann's "type" section]

Stop 7 - Outcrops of melange and flysch in old railroad cut and bed of Normanskill [Ruedemann's "type" section]

These outcrops form the "type locality" of Ruedemann for the "Normanskill Formation", mainly because of the fauna he obtained here at Kenwood, and "nearby", including Stop 8 at South Glenmont. As you can see, structural disruption is prominent, and the bedded sections seen between faults have no defined, or definable, top or base. In addition, the use and abuse of the term Normanskill since Ruedemann,

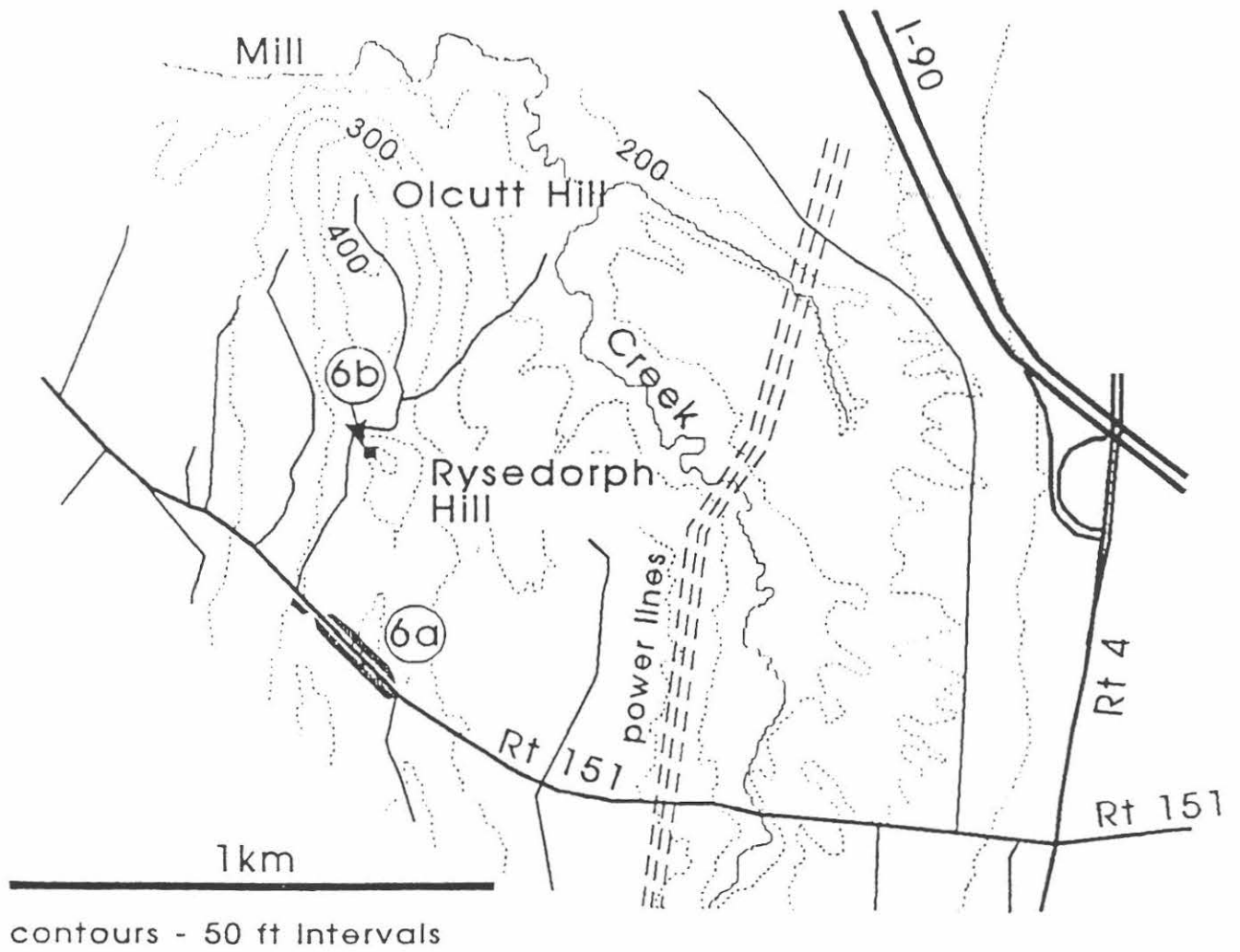


Fig 7. Location map for stop 6a and 6b, Route 151 and Rysedorph Hill

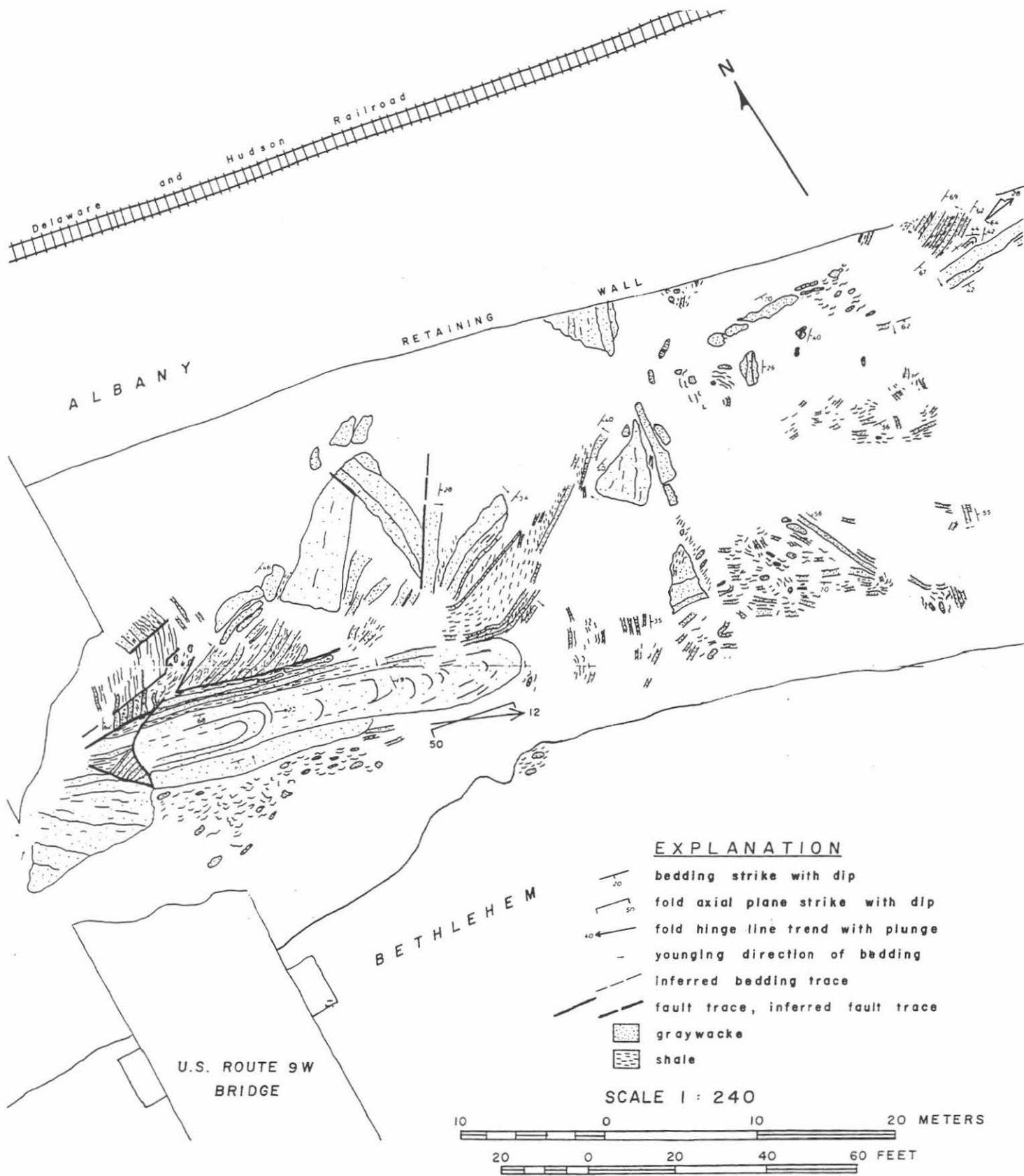


Fig. 8. Outcrop map [from Vollmer, 1979] of part of the Normanskill at Kenwood [Stop 7], type locality of the Normanskill "Formation" of Ruedemann.

especially as a time-stratigraphic term, makes it inappropriate to use it for the general lithic assemblage of flysch greywacke and shale, which is, despite the structural disruption, quite well exposed here. In the creek bed, if the water level is not too high, a downward-younging bed in the crest of an antiform may be seen (Figure 8), with the turbidite sole structures exposed on a surface defining the fold hinge. This structural complexity is unusual for the bedded greywacke-shale sections in the Hudson Valley; Vollmer suggested that the stratigraphic inversion represented by the downward-facing nature of this fold may have been produced by slumping, with the fold being the regional set later superimposed on the already inverted section. This outcrop maps near the eastern margin of the Mohawk River Central Melange Zone (Figure 2).

Return to Rt. 32, turn right	0.25	50.9
Proceed south; intersection with Rt. 144; continue straight [onto 144]	0.6	51.5
Park at area on right just before road crosses bridge above railroad tracks as it curves to left and downhill	1.4	52.9
Outcrop in road cut, by road and to east of bridge up to power line, [and along railroad tracks].		

Stop 8 - South Glenmont locality of allochthonous Mt. Merino Cherts

Black, grey, and green cherts, some with burrow mottling, as seen in this set of exposures, are characteristic small and large blocks in the exotic melanges. However, they are always blocks, or fault-bounded slices, in this context. Identical chert lithologies are found in stratigraphic sequence with other rock types only within the Taconic Allochthon. There they occur only at one position in the stratigraphic succession, at the top of varied continental rise-deposited sediments derived from the Cambro-Ordovician passive margin of North America, and immediately below black graptolitic shales with *N. gracilis* graptolite faunas, which are in turn overlain by greywacke-shale turbidites of the Pawlet Formation. This locality yielded to Ruedemann the most abundant fauna found in the greywacke-shale area of the "Normanskill", and has as a result strongly influenced the presumed age of the flysch of the Hudson Valley. We are of the opinion that all occurrences of these cherts are highly allochthonous, representing tectonically separated, and perhaps erosionally isolated, pieces originally emplaced as integral parts of the Taconic Allochthon. We think that all ages derived from the faunas in the cherts, and the related black graptolitic slates, are therefore older, by at least some small amount, than any of the greywacke-shale flysch of the Hudson Valley. Prominent down-dip plunging folds (Figure 9) are seen in several parts of this set of outcrops; these probably represent rotation of originally shallow-plunging folds by heterogeneous simple shear associated with thrusting, as indicated by Vollmer and Bosworth (1984) for the more strongly deformed flysch and melange generally. This locality maps close to the eastern boundary of the Mohawk River Central Melange Zone (Figure 1).

From parking area, return north along Rt. 144 to intersection with Glenmont Road, Rte 910A - turn left	0.3	53.2
Intersection with Rte 9W; turn right onto 9W going north [1.3 and 1.6 intersections with Rt. 32; 1.9 cross Normanskill bridge; 2.4 entrance to I-787; 2.5 entrance to NYS Thruway I-87].	1.5	54.7
Light at junction with Southern Boulevard, go straight [go straight at light where Rt. 9W angles to right]	2.8	57.5
Junction with Rt. 443 Delaware Avenue; turn left	0.5	58.0
Turn left onto Mill Road	0.4	58.4
Park at old bridge	0.2	58.6
Go down to east of old bridge to riverside outcrop of Stop 9.		

Stop 9 - Normansville melange and broken formation.

This outcrop in the north bank of the Normanskill exposes shaly rocks with greywackes that can be demonstrated to be in part a broken formation (Figure 10), with partly traceable sedimentary layering, rather than a melange with completely disrupted layering. Additionally, this locality contains some folded quartz veins. The outcrop also maps within the Central Melange Zone.

Return to Rt. 443; turn right	0.2	58.8
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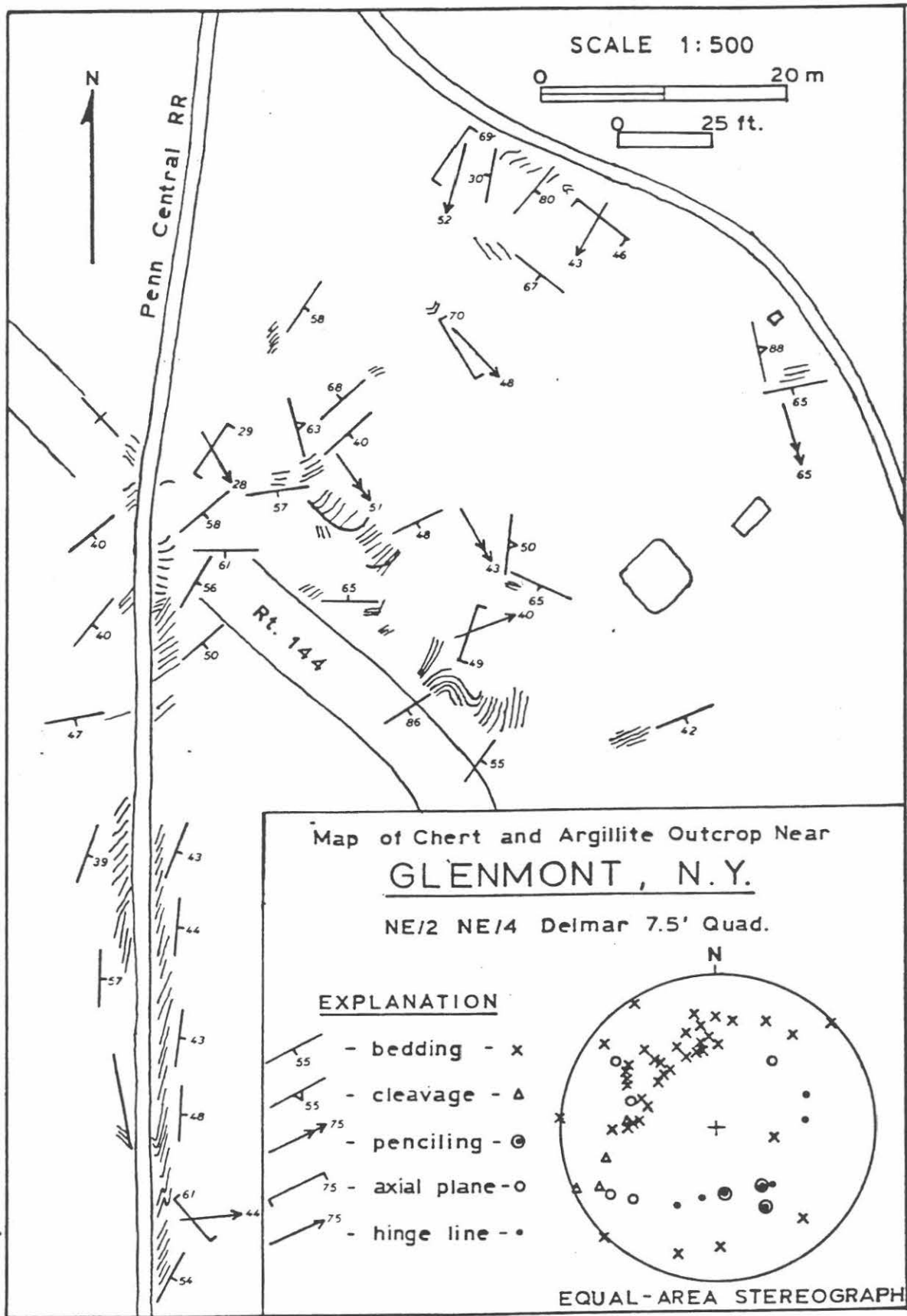


Fig 9. Outcrop map of South Glenmont locality [Stop 8], exposures of Mount Merino Chert, part of a large slice or block in Mohawk River Central Melange Zone (east part). From Vollmer (1979).

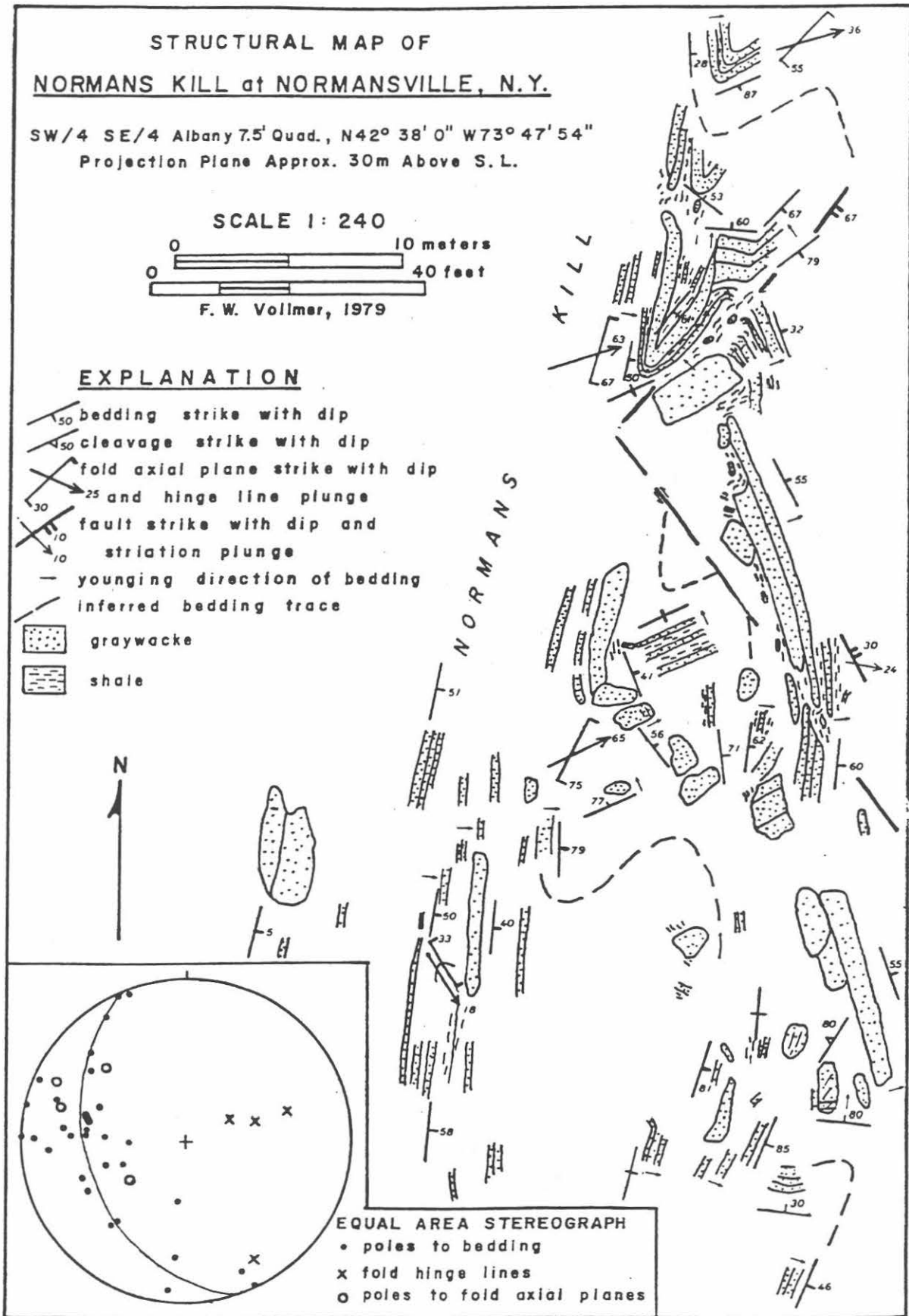


Fig. 10. Map of outcrop at Normansville on the Normanskill [Stop 9]. Broken formation transitional to melange; within Mohawk River Central Melange Zone (east part).

Turn right at light onto McAlpin Street	0.4	59.2
Continue straight onto Southern Boulevard, and Rt. 9W		
Go straight past Thruway and I-787 entrance	1.0	60.2
Follow Rt. 9W to Beckers Corners, intersection with Rt. 396, turn right	4.1	64.3
Turn left in South Bethlehem onto Rt. 101	2.7	67.0
Park in visitor lot on right	0.5	67.5

ASK PERMISSION inside Callanan office building across road; it is preferable to request permission in advance by calling 518-767-2222, or writing Callanan Industries Inc, Corporate Headquarters, South Street, South Bethlehem, NY 12161

Stop 10 - Feuri Spruyt and Callanan Quarry - unconformity with Helderberg Limestones overlying melange. PERMISSION REQUIRED - request permission in advance by calling 518-767-2222, or writing Callanan Industries Inc, Corporate Headquarters, South Street, South Bethlehem, NY 12161

This locality (Figure 11) shows carbonates of the basal Helderberg Group [early Devonian] unconformably overlying the folded greywackes and shales, and melange of the Ordovician. The boundary between the Ravena Greywacke Zone and the eastern side of the Mohawk River Central Melange Zone passes beneath the unconformity just west of the quarry (Figures 1, and 11). In particular, the unconformable relation to the melange demonstrates that the fault zone which created the melange is an Ordovician structure, and predates entirely, at this position, the Acadian structures that fold the overlying Devonian strata south of this place (Bosworth and Vollmer, 1981). The map relationships, and unfaulted structural condition of the Helderberg Group, west of this locality, shows that all the width of the melange belt comprising the western and eastern exotic melange, which goes beneath this unconformity, must also be an Ordovician, and Taconic-age structure.

Return to Rt. 9W reversing directions above; go straight across continuing on Rt. 396	3.2	70.7
Intersection with Rt.144; turn left	2.2	72.9
Entrance to NYS Thruway; turn left	0.4	73.3

End of road log. For Albany, take I-787 exit 23; Albany exit 24 for I-87 Adirondack Northway and airport; Schenectady exit 25 to return to Union college area.

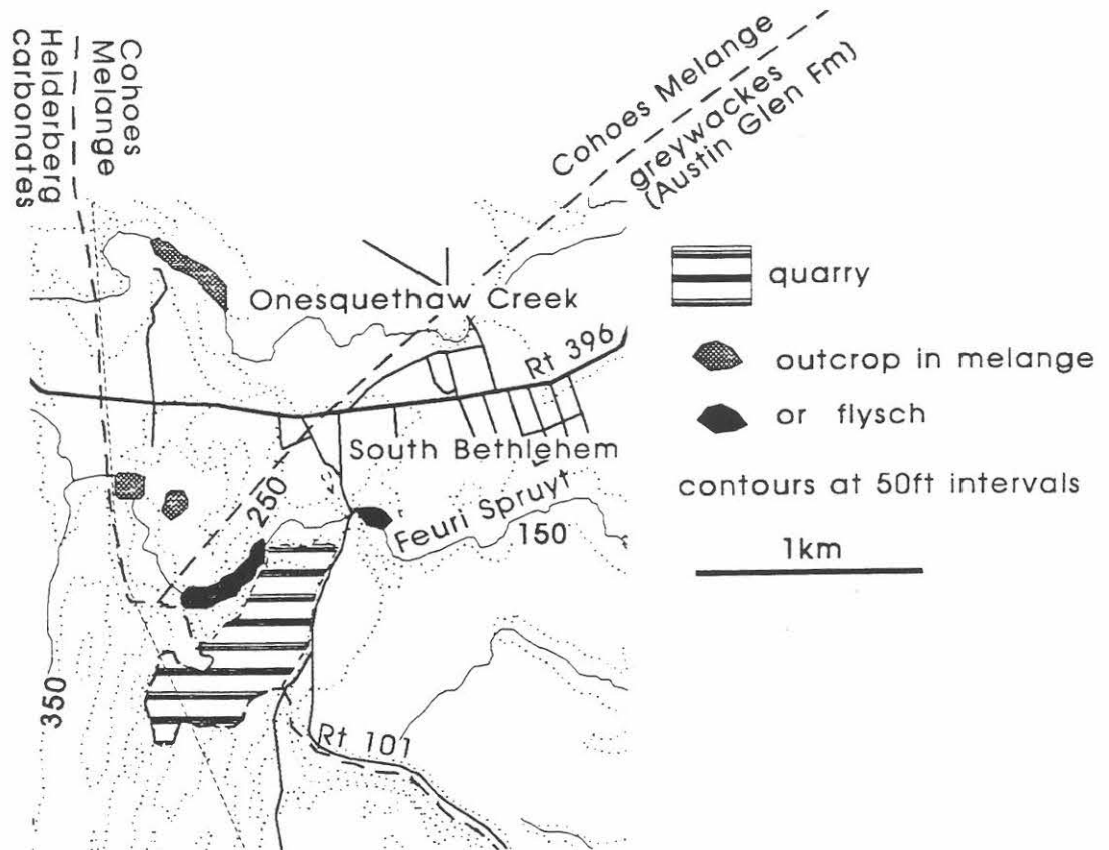


Fig 11. Sketch location map of Callanan Quarry, South Bethlehem and adjacent Feuri Spruyt Creek [Stop 10].

GEOMORPHOLOGY, PALEOCLIMATOLOGY AND LAND USE CONSIDERATIONS OF A GLACIATED KARST TERRAIN; ALBANY COUNTY, NEW YORK

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ABSTRACT

Karstified terrains in Albany County, New York provide evidence of paleoclimatic conditions during Pleistocene glaciation. Caves and the areas surrounding them preserve important clues that can be interpreted to reconstruct portions of the geomorphic and paleoclimatic history of a region. Knowledge gained from these investigations is of increasing importance as development pressures extend into these forested, agricultural and rural areas. This paper focuses on the geology, karst hydrology, and glacial features of two areas not previously described. As a result of karst processes and anthropogenic diversion of the headwaters of the Onesquethaw Creek, the Hollyhock carbonate aquifer and its receiving stream (i.e., Onesquethaw Creek) are extremely sensitive to contaminant inputs. Thus, we have incorporated a section specific to land use considerations and concerns. This ongoing work is an extension of previous work addressing the interaction between karst and glaciation (Rubin, 1991b; Palmer et al., 1991a, 1991b).

Two karst areas will be visited on this field trip. The first stop, at the Hollyhock Hollow Sanctuary, will feature a cave system that 1) developed prior to the advance of the late Wisconsinan (Woodfordian) Laurentide ice sheet; 2) was enlarged by meltwater invasion; and 3) remains active today. The second stop will be at Joralemon Park, where active and relict caves are used to interpret the interaction between caves and glacial processes. Solutionally enlarged fractures and relict caves and swallow holes in the groundwater basin are integral, functioning components of the epikarst, funnelling runoff and infiltrating meteoric waters to deeper conduit flow routes.

INTRODUCTION: HOLLYHOCK HOLLOW SANCTUARY

The Hollyhock Hollow Sanctuary and surrounding area represent a unique geologic and hydrologic setting. The Hollyhock carbonate aquifer supports no surface drainage today, except for the Onesquethaw Creek which is seasonally pirated underground for ≈ 1 km (0.7 mi). All drainage is subsurface in limestone bedding planes, joints, faults and conduits, largely within the Manlius and Coeymans Limestones. Structural deformation here within the Hudson Valley Fold Thrust belt has folded and faulted these limestone units. This deformation is locally responsible for orienting subsurface flow paths

In Garver, J.I., and Smith, J.A. (editors). Field Trips for the 67th annual meeting of the New York State Geological Association, Union College, Schenectady NY, 1995. p. 81-107.

and was in place prior to Wisconsinan glaciations. During glaciation, subglacial meltwater carved shallow bedrock channels in the Coeymans Limestone. Several relict caves and perhaps the only known meltwater-carved limestone pedestals in such a setting attest to the unique geology of this karst basin. Scallops (solutional pockmarks in cave walls) in swallow holes document former rapid surface and groundwater flow here. Some of this meltwater was pirated into now relict swallow holes over many thousands of years. This same flow created or enlarged limestone cave conduits which today still drain the carbonate groundwater basin that extends to the north and northwest. These enlarged joints and conduits provide a route for possible rapid transport of contaminants.

Location stops for this field trip are organized sequentially for walking tours.

HOLLYHOCK HOLLOW STUDY AREA BOUNDARIES

The Hollyhock Hollow study area is located in Albany County, New York along the Onesquethaw Creek, a tributary to the Hudson River. The larger study area is bounded by the Onesquethaw Creek to the southwest, the Manlius/Coeymans escarpment to the north and northeast, extends a short distance east of Route 102, and may extend beyond Rowe Road to the west. A smaller area within this broad area, extending outward from the Hollyhock Hollow Sanctuary property, is the focus of the field trip stop. The study area is located in the Towns of New Scotland and Bethlehem.

GEOLOGY

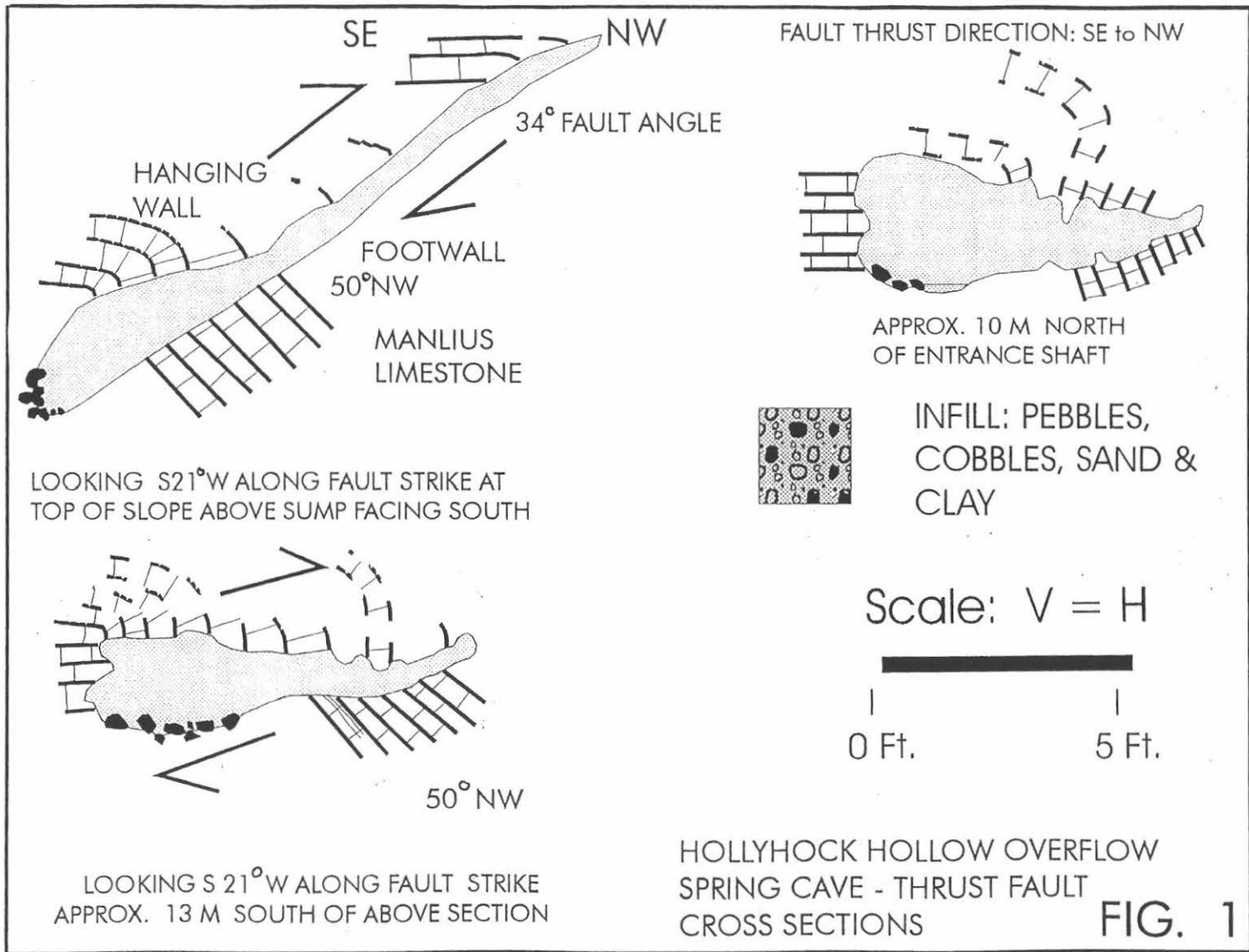
The Hollyhock carbonate aquifer is developed predominantly in the Lower Devonian Manlius and Coeymans limestones, with small outcrop areas of the Kalkberg and New Scotland formations. The massive and blue-gray, thick-bedded Coeymans Formation is the uppermost bedrock unit exposed throughout most of the drainage basin. Infiltration of meteoric water through vertical fractures recharges the underlying carbonate aquifer. These fractures, and those extending into the underlying Manlius Limestone, comprise the epikarst - the unsaturated upper part of the percolation zone. Locally, within the Hollyhock Hollow Sanctuary, these fractures have historically been exploited for dimension stone. Thin "millimeter" beds are characteristic of the Manlius Limestone.

The Hollyhock Hollow Nature Sanctuary, Joralemon Park and surrounding areas are situated within the Hudson Valley Fold-Thrust Belt (HVB) and exhibit faulting characteristic of this belt. Marshak (1986, 1990), Marshak and Engelder (1987), and Cassie (1990) discuss structural deformation within parts of the HVB. Marshak (1990) addresses the spectacular cross section of a detachment fault, a duplex, and a ramp anticline in the Feura Bush Quarry situated north and probably within the Hollyhock groundwater basin.

While evidence of faulting is present in the area and does locally disrupt the regional bedrock dip, it is hypothesized that the overriding control on groundwater flow in most of the Hollyhock carbonate aquifer is the local dip. Preliminary geologic mapping in the vicinity of Hollyhock Hollow reveals bedrock strikes and dips ranging between N2°E and N34°E, and 4°SE and 14°SE, respectively.

HOLLYHOCK HOLLOW OVERFLOW SPRING CAVE

A large groundwater basin is tributary to the Hollyhock Hollow Overflow Spring. The overflow spring flows only during periods of heavy snowmelt or storm water infiltration. The gated entrance shaft (\approx 9 m; 30 ft) is developed along a vertical joint that was filled with boulders and rubble sometime in the last 250 years, then partially capped with concrete. It was opened in 1994 by cavers in cooperation



with New York Audubon. At the base of the entrance shaft, a low, wet, and muddy cave may be followed for approximately 38 m (125 ft) to a sump. The presence of a rounded 36 cm (14 in) diameter quartzite boulder near the farthest point of penetration in this overflow cave indicates high discharges and velocities through large conduits. The nearest likely source of this boulder is from the Rienow Swallow Hole (see below), some 600 m (2,000 ft) to the north.

The Hollyhock Hollow Overflow Spring Cave has developed along a thrust fault and related fractures in the Manlius Limestone (Fig. 1), striking north-northeast and upthrown to the northwest, indicating the possible importance of faulting and fault-related fractures for cave development and groundwater flow in the HVB. Conduit enlargement preferentially occurred along a structurally weakened zone of increased permeability. In order to assess the importance of the structural geology on groundwater flow, excavation efforts are underway in an effort to gain entry into upgradient conduit portions of the carbonate aquifer.

Based on observations made by New York Audubon staff, Rubin (1994a) calculated an estimate of groundwater discharge from this overflow spring. While further hydrologic work is required to refine the discharge estimate, some 1.3 m³/s (47 cfs; 21,000 gpm) flowed during a 2-day storm event (calculated using the Hazen-Williams equation for turbulent pipe-full flow conditions), indicating a groundwater basin contributing flow from at least 9 km² (3.5 mi²). Tracer tests are required to define the extent of the groundwater basin.

The formation of this floodwater overflow spring could have resulted from one or more of the following factors: 1) the relative inefficiency of the occluded base flow springs situated south of this overflow spring (see Spring Zone section); 2) the increased discharge from a thin soil-mantled and well-karstified carbonate unit; or 3) the invasion of large quantities of glacial meltwater through swallow holes and fractures.

Scallops (solutional pockmarks) on the walls of Wiltsie's Cave and the Rienow Swallow Hole (see Rienow Swallow Hole and Wiltsie's Cave sections) document even larger supercritical turbulent paleoflows (to 55 cfs) into this groundwater basin (Rubin, 1994b) that, when considered with other inputs from throughout the broad watershed, attest to the maturity and drainage efficiency of the basin. Thus, groundwater from a large drainage area flows into the Hollyhock Hollow Sanctuary. Land uses far beyond the Hollyhock Hollow property boundaries have the potential to degrade groundwater and surface water quality at and downgradient of the sanctuary.

Geomorphic And Karst Considerations

How long has the Hollyhock Hollow Overflow Spring Cave operated as a floodwater overflow route? Dineen (1987) and Isachsen et al. (1991) document that this area has been modified by stream incision and physical weathering during the Cenozoic era (last 65 my). Dineen (1987) determined that present-day drainage trends in the Hudson Valley were established before Wisconsinan glaciations, sometime prior to 70,000 years ago. Evidence that today's drainage was in place prior to inundation by Woodfordian ice includes 1) glacial striations near Clarksville cut across the channel of the Onesquethaw Creek, and 2) insufficient time for denudation and erosion rates to incise the channel to the striated level during post-Woodfordian time. Wisconsinan drainage along Onesquethaw Creek was probably little different from what it is today. Thus, the physical setting was in place for the piracy of meteoric water into limestone fractures and subsequent flow to the base level Onesquethaw Creek.

²³⁰Th/²³⁴U dating of speleothems from Schoharie County caves provides evidence of initial cave development in the region prior to 350 Ka, the limit of this dating method. Based on evidence of well-adjusted drainage patterns in the Cenozoic era, it is possible that cave development has been ongoing over

the last several million years. $^{230}\text{Th}/^{234}\text{U}$ disequilibrium series dating of speleothems from Schoharie County caves reveals dates between 277 Ka and 165 Ka, and greater than 350 Ka (Dumont, 1995 and Stein-Erik Lauritzen [cited in Palmer et al., 1991a]) indicating active karstification throughout the region for hundreds of thousands of years. Rubin (1991b) provides additional evidence that preglacial caves were modified and enlarged by the invasion of glacial meltwaters. Thus, the physical setting for development of the floodwater overflow route was probably in place prior to the last glaciation. The presence of boulders and large cobbles in the overflow route, in the absence of a likely surficial input location under present day climatic conditions, suggests that enlargement of this route occurred either during Wisconsinan glaciations or before, when even greater flows occurred through the cave system.

RIENOW LEGACY

The karst hydrology (e.g., karren, caves and lack of surface drainage) clearly documents that a well-developed carbonate aquifer and related cave system underlies a many square kilometer area. To date, only small caves have been entered at various points throughout the basin. For the flashy high-discharge Hollyhock Hollow Overflow Spring to function, an extensive and well-integrated cave system must be present. Entry into the system (assuming much of it is physically traversable) will permit additional aquifer characterization; geomorphic, structural, and hydrologic assessment; and water quality evaluation within some of the groundwater basin. At this time, digging efforts are focused on the Rienow Swallow Hole.

Geologists and cavers attempting to gain access to the Hollyhock carbonate aquifer may not be the first. A detailed and tantalizing two-page New York State Historic Trust form, filled out by Professor Robert Rienow (deceased former owner of the property) in 1973, has one sentence describing "caves" on the 56 hectare (138 acre) Hollyhock Hollow Farm. Specifically,

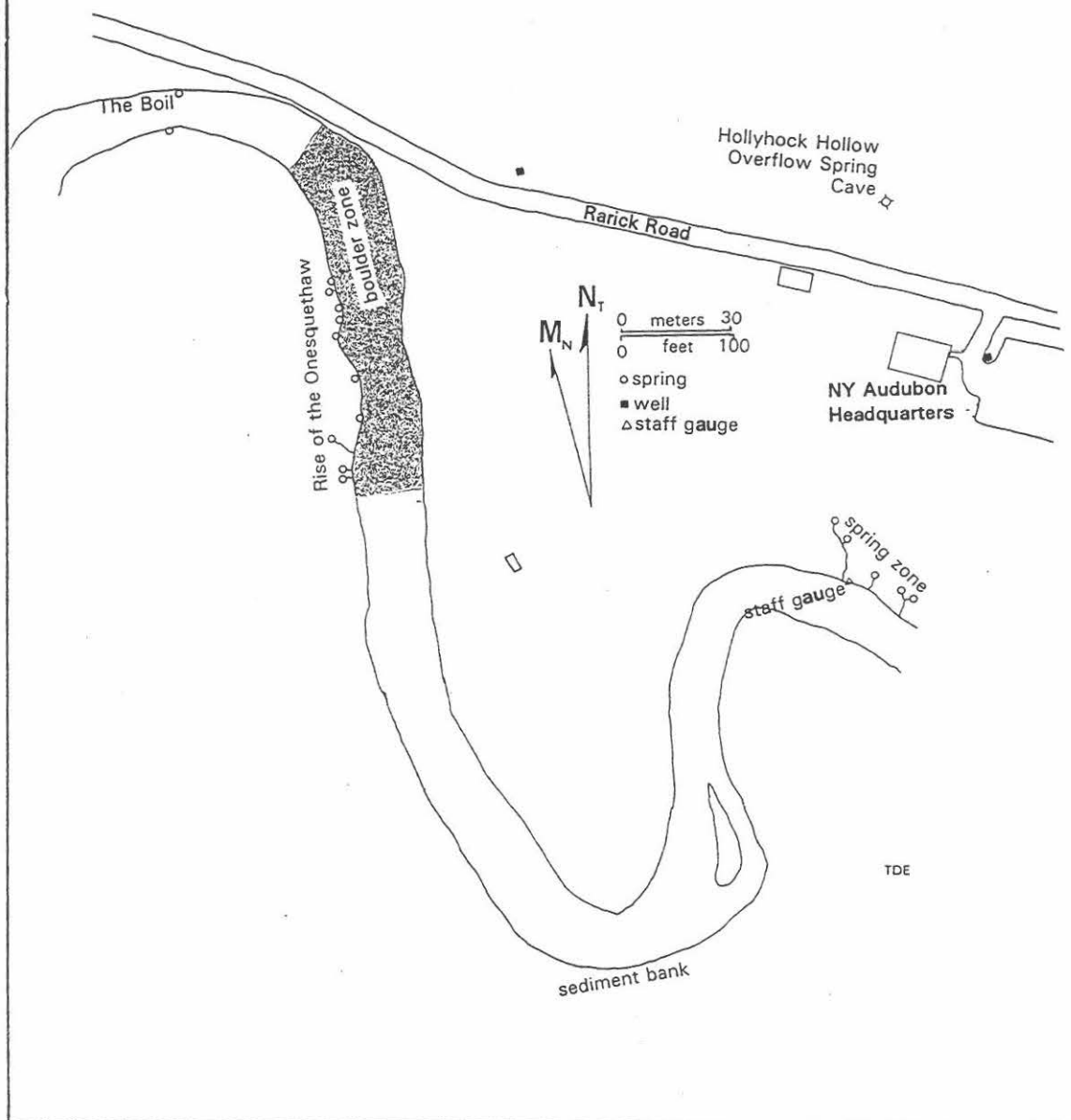
"Of historic interest, as well as geologic, the entire farm is honeycombed with caves which as late as 1940 still had rope ladders for access 50 feet down (now closed)."

This reference, especially when taken in conjunction with other factual references provided on the form, strongly points to the presence of a large, formerly enterable, cave on the property. While Professor Rienow did not describe the location of the access point, we know that it could not have been Wiltsie's Cave since access did not require rope ladders, it was not on his property, and it was not closed. Furthermore, the Hollyhock Hollow Overflow Spring Cave near Rarick Road does not fit the description, would not be considered accessible by most and does not suggest the term honeycombed. Other access options include assorted boulder piles (possibly covering a shaft entrance) and a second sinkhole (with a small cave) east of the Rienow Swallow Hole.

SPRING ZONE: RESURGENCE POINT FOR THE HOLLYHOCK CARBONATE AQUIFER

During periods of low and moderate flow, the Hollyhock carbonate aquifer and integrated cave system resurge or drain through five alluviated springs, the number depending on the amount of water incident to the cave system and the elevation of backflooding behind the occluded bedrock outlet (Fig. 2). While these springs still flow during periods of high flow, their efficiency is exceeded, causing water to backflood within the cave system, and sometimes discharge out the overflow spring. The springs are situated downhill and south of the headquarters building and to the east and west of the staff gauge. The five low and moderate flow springs probably issue from a single bedrock outlet now occluded by a combination of till, deltaic, and perhaps lacustrine deposits. These springs are distinct from springs on the creek's western side that form The Rise of the Onesquethaw (see section below).

Figure 2 - Springs along the Onesquethaw



STAFF GAUGE AND RATING CURVE DEVELOPMENT (WATER QUANTITY)

The importance of baseline Onesquethaw Creek water quantity and quality data may one day be critical, should surface water withdrawals or impoundments or waste-water/leach field (surface and subsurface) additions be proposed along the Onesquethaw Creek Corridor. As developmental pressures build in this area and along the Onesquethaw Creek Corridor, it is becoming increasingly important to understand the flow dynamics of the Onesquethaw and its tributary aquifers and their assimilative capacity for contaminants. While a detailed, but short-term (15+ month), record of streamflow was kept farther upstream in Clarksville in the mid-1980s (Rubin, 1991a; 1992), installation of the Hollyhock Hollow staff gauge represents the first effort to maintain long-term discharge records along the Onesquethaw Creek.

This gauging station is in a critical location since the Onesquethaw Creek is a losing stream upstream of The Boil (see below), with flow occurring in solutional conduits during much of the year. The staff gauge was installed in November 1994 in cooperation with New York Audubon, which records stream stage daily. A rating curve is being developed. Continuous recording of stage height, coupled with hydrologic data from the cave system, should permit detailed analysis of the relationship between storm pulses from surface runoff versus episodic pulses transmitted through limestone conduits. In this manner, aquifer dynamics and contaminant transport could be characterized, as could monitoring schedules.

On June 28, 1995, during a period of low flow, the total discharge of the Onesquethaw Creek 76 m (250 ft) downstream of the Hollyhock Hollow staff plate was gauged at only 1.4 l/s (0.05 cfs; 22 gpm). This leaves little water available for assimilation of contaminants stemming from leach fields and other sources in the groundwater basin. It is likely that Onesquethaw Creek discharge is even lower during droughts (e.g., July 1995). An estimate of the peak flow at this location was made by measuring the maximum elevation of debris (2.4 m; 8 ft), surveying the channel cross section and gradient, using a Manning's η of 0.045 and utilizing the Manning equation. Peak flow of the Onesquethaw, 76 m downstream of the Hollyhock Hollow staff gauge, is on the order of 100 m³/s to 113 m³/s (3,500 to 4,000 cfs).

ONESQUETHAW CREEK DELTA BUILT INTO GLACIAL LAKE ALBANY

A large failing sediment bank is exposed upstream of the staff gauge (Fig. 2). It is composed of deltaic material (e.g., gravel, sand, sandy loam) in episodically well-exposed bottom set, foreset (dipping $\approx 20^\circ$ SW), and topset beds. The elevation of the foreset/topset contact occurs near 310 ft. msl, coincident with one of the glacial Lake Albany levels cited by Dineen (1986). Some 180 m (600 ft.) downstream of Rt. 102, a massive gray clay bed with pebbles and small cobbles is exposed, also indicating lacustrine deposition.

Dineen (1986) concludes that Lake Albany is at least 14,000 years old in the Albany area. This time period is long enough for the Hollyhock Hollow Overflow Spring Cave to have formed behind a deltaic or lacustrine sediment occlusion to the original cave mouth, or both. However, the presence of a heavy, rounded glacial boulder far in the cave may indicate earlier development, when conduit discharges were greater than today due to upgradient glacial meltwater invasion.

CALCITE-CEMENTED GLACIAL DEPOSITS

Along the southwestern bank, as well as in the channel, of the Onesquethaw Creek, there are a number of exposures of either deltaic deposits or glacial outwash cemented with calcite. They have the appearance of a conglomerate or a puddingstone. The calcite was precipitated from groundwater seepage

from the adjacent limestone bench. Degassing of carbon dioxide from surfacing groundwater caused supersaturation with respect to calcite. Lithification of this conglomerate occurred post glacially.

THE BOULDER ZONE AND THE RISE OF THE ONESQUETHAW

Some 300 m (1,000 ft) upstream of the staff gauge, the gradient of the stream increases sharply. This zone of 111 m (365 ft) is armored with large boulders. A short distance upstream, at The Boil, and immediately downstream of some small fault bend folds that indicate the presence of a basal thrust fault, the Manlius streambed strikes to the NNE and dips between 8° and 12° SE. Thus, the steep gradient of this stream reach may be structural in nature. The unusually high abundance of boulders in this reach, rather than at its base, may reflect a reduction in stream power (during a period of high discharge glacial meltwaters) coincident with the surface of glacial Lake Albany.

A spring zone, composed of ten or more springs (Fig. 2), issues from the southwestern stream edge within the boulder zone. A combination of the factors discussed above has resulted in groundwater surfacing here. Some of these springs continue to flow during extremely dry periods, even when the streambed upstream is dry for over 1 km. Although groundwater tracing is required for verification, this spring zone almost certainly comprises the resurgence of the pirated Onesquethaw Creek (see Creekbed Upstream Of The Boil below), making it The Rise of the Onesquethaw.

THE BOIL

A boil of water rises in the Onesquethaw Creek just upstream of The Boulder Zone (Fig. 2). The boil fountains above creek level to approximately 0.3 m (1 ft) during periods of moderate flow. The source of the upgradient water remains to be traced. We believe that it is the outlet for upstream water that sinks into fractures and bedding planes, perhaps in an area of exposed and lithified mud cracks. Alternately, the source of The Boil's hydraulic head may be the carbonate aquifer to the north. Thin-bedded Manlius limestone can be seen along the edge of the creek just downstream of The Boil.

CREEKBED UPSTREAM OF THE BOIL

The bed of the Onesquethaw Creek upstream of The Boil, for approximately 1 km (0.7 mi), carries water for only part of the year. A similar situation occurs upstream of Clarksville, where the Onesquethaw Creek loses all its surface flow seasonally into the Onondaga Limestone (Rubin, 1991a and this volume). During drier periods, all surface flow is pirated into solutionally enlarged joints and bedding planes in the streambed. Far upstream, large and deep fissures in the limestone funnel surface water into the subsurface. The farthest downstream point of water infiltration is a function of fracture and bedding plane enlargement and their hydraulic efficiency, the number of fractures integrated with the conduit system, and the discharge of the creek. Obviously, the location of stream gauging activities must be selected with knowledge of the karst hydrology.

Following the piracy of surface waters, groundwater flow then occurs through conduits (i.e., caves) until its hypothesized resurgence down-dip at the spring zone referred to above as The Rise of the Onesquethaw.

THE PEDESTAL AND MELTWATER CHANNELS

The best of several examples on the New York Audubon property, this stream-lined and smoothed pedestal is believed to be physically contiguous with the underlying bedrock, being apparently attached and dipping approximately 5°SE. It stands some 2.4 m (8 ft) above the surrounding topography, appearing to be an erosional remnant. We hypothesize that the base of an ice sheet rested on top of this pedestal with meltwater coursing against its base and nearby in sub-parallel meltwater channels. Note the smoothed, rounded and gently sloped base and northern face that is interpreted as being worn by subglacial meltwater.

A number of shallow relict channels stand tribute to widespread and substantial subglacial meltwater flow throughout Hollyhock Hollow and much of the surrounding area. Dineen (1986) describes stagnant ice associated with the Schoharie ice margin as riddled with tunnels that drained into glacial Lake Albany. Similar tunnels related to Dineen's Alcove ice margin, the Delmar margin, or meltwater flows under glacier ice at different times apparently drained meltwater southeast off the Helderberg Escarpment. These overland flows may account for the extremely thin soil mantle present.

RIENOW SWALLOW HOLE

The Rienow Swallow Hole is located approximately 0.6 km (2,000 ft) north of the Hollyhock Overflow Spring Cave. The karst hydrology of the aquifer indicates that the Rienow Swallow Hole is physically connected to the Hollyhock Hollow Overflow Spring Cave via a conduit, being situated upgradient of the sump currently stopping exploration. Snowmelt and intense storms provide local runoff that flows into this swallow hole.

Excavation and examination of this swallow hole and debris fill provide valuable insight into the paleohydrologic flow dynamics once operable here. Moderately rounded quartz sandstone and gneiss cobbles and boulders in the clay fill denote glacial transport from the north. Well-defined, solutionally carved walls within the Rienow Swallow Hole provide evidence of paleoflow conditions no longer active today. Small scallops (solutional pockmarks) present near the base of the northwestern bedrock wall provide definitive evidence that turbulent water once flowed rapidly into the sinkhole and into a shaft or passage capable of receiving this water and bringing it into the underlying carbonate aquifer. Palmer (1991) documents that conduit enlargement requires a minimum of 5,000 to 10,000 years; thus the flow conditions necessary to produce the observed scallops were present for an extended period of time. The invasion of glacial meltwaters may have substantially enlarged a preglacial cave system that was already graded to the Onesquethaw Creek base level.

Determination Of Paleoflow Into The Rienow Swallow Hole

Measurement of scallop wavelengths may be used in the calculation of paleo and recent flow velocities and discharges (Blumberg and Curl, 1974 and Curl, 1974). A number of scallops were measured (ranging between 3.6 cm (0.12 ft) and 5.5 cm (0.18 ft) in length) along with the dimensions of the solutionally enlarged walls. This information was then assessed in order to estimate the flow conditions present at the time of formation. In order for the scallops to form in the base of the Rienow Swallow Hole, unrestricted flow must have occurred into the subsurface. Blumberg and Curl (1974) derived a universal constant for the scallop Reynold's number used in these calculations, based on plaster model studies, of 2200. Rubin (1991a), based on research specific to limestone in nearby Clarksville Cave, determined that the scallop Reynold's number may actually not be a constant, but instead may best be characterized by a range of values. Empirical observation of the flow dynamics in Clarksville Cave, coupled with characterization of flood-return intervals within the catchment basin, suggest that a scallop

Reynold's number on the order of 3300 might best fit the cave-specific conditions. Using a scallop Reynold's number of 3300, the paleo flow velocity and discharge into the Rienow Swallow Hole were determined to be on the order of 1.7 m/sec (5.6 ft/sec) and 0.7 m³/s to 1.6 m³/s (26 to 55 cfs), respectively (Rubin, 1994b).

The quantity of water flow indicated by Rienow Swallow Hole scallops is roughly equivalent to the maximum estimated discharge at the Hollyhock Hollow Overflow Spring. This indicates that the combined paleoflow from throughout the catchment basin was much greater during periods of glacial melting than today.

Excavation of the southern end of the Rienow Swallow Hole revealed a lens-shaped shaft extending some 2 m (7 ft) to loose fill. The central long axis of the shaft is \approx 0.8 m (2.5 ft), with a short axis central width of \approx 36 cm (14 in). Immediately north of this shaft, a 13 cm (5 in) wide solutionally enlarged joint opens to an underlying room or passage. White calcitic flowstone within and near this shaft was deposited by infiltrating meteoric waters after massive inflows ceased. This shaft may one day provide access into the Hollyhock carbonate aquifer.

Even greater flows are indicated by scallops present in a bedrock edge at the southern end of the Rienow Swallow Hole. The scalloped edge occurs near the top of the sinkhole, indicating periods of excessive flows when the discharge into the sinkhole and receiving cave could not handle the quantities of water present. At these times, turbulent water apparently poured into the swallow hole, partially cascading over this edge and downhill toward the Onesquethaw Creek.

Speleothem Recovery From Excavated Fill In The Rienow Swallow Hole

Excavation of sediment and rock debris from the Rienow Swallow Hole revealed numerous broken speleothems. Initially, this was thought to be evidence for a former cave entry point here, with the formations being indicative of vandalized and discarded material from within the cave. However, further excavation found speleothems, at random angles, within hard-packed clay and beneath \approx 550 kg (1200 lbs) limestone blocks. The clay packing against speleothems clearly exceeds 55 years of natural packing of infilled debris.

These speleothems denote a complex geomorphic history, all within this small, but significant, swallow hole. The types of formations found include much finely-layered flowstone, small stalagmites, stalactites, and varieties formed in supersaturated pools. The bottom portion of a broken stalagmite measuring \approx 30 cm (1 ft) wide at its base, 23 cm (0.75 ft) high, and with an upper diameter of \approx 12 cm (0.4 ft) was recovered. This specimen reveals a history of forming on an inclined mud slope (in air-filled cave passage), incorporation of a fragment of flowstone into its structure during formation, and solutional dissolving of its center sometime after formation. Other smaller recovered specimens reveal central drip holes surrounded by concentric layered calcite through which water flows to form stalactites. These broken speleothems formed in a cave environment.

Broken speleothems within the now collapsed cave passage provide evidence for variable paleoflows and climatic conditions, as well as multiple glaciations. One 20 cm² (8 in²) sample reveals a cave history of 1) open, unrestricted water influx for at least 10,000 years and development of a short bedrock-roofed cave passage, probably from glacial meltwaters and responsible for scallop formation; 2) cessation of water responsible for conduit formation, perhaps during an interstadial period; 3) flowstone deposition over red clay fill within a cave passage; 4) formation of small stalagmites; 5) deposition and buildup of flowstone around the stalagmites; 6) crystalline growth in lily pad patterns in a small calcite-supersaturated pool (requiring a lengthy and extremely stable cave environment); 7) passage collapse and infill, probably due to glacial loading and the rapid influx of stream-borne sediment.

This sample was found in a vertical position in hard-packed clay. Another sample, possibly from a drapery-like formation, has truncated flowstone layers, with subsequent deposition, revealing a series of at least four erosion phases. Growth of this sample may have been halted by cold phases associated with various glacial periods. $^{230}\text{Th}/^{234}\text{U}$ dating of this sample may date interglacial and interstadial periods. Steadman's (pers. comm.) $27,350 \pm 750$ yrs BP ^{14}C age on a wood rat bone sample excavated from a Clarksville area cave may 1) closely limit the maximum age for the initial advance of the late Wisconsinan ice margin, and 2) correlate with an interstadial speleothem erosional phase.

WILTSIE'S CAVE

Even short caves can provide valuable information that can be used to reconstruct the geomorphic history of an area. The entrance to Wiltsie's Cave lies approximately 60 m (200 ft) north of the Rienow Swallow Hole, and also probably served as a swallow hole for glacial meltwaters (Fig. 3). The large size of this relict cave, now receiving only small quantities of local surface runoff and infiltration, further supports the formation or enlargement by subglacial meltwater invasion argument. The cave descends as a joint-aligned (S32°W) vadose canyon, with ceiling heights and passage widths up to 7.6 m (25 ft) and 3 m (10 ft), respectively. Flowstone covers much of the walls. The passage descends in a series of steps through several chert beds in the upper portion of the Coeymans Limestone and down into the Manlius Limestone. Near the cave's southwestern end, and perhaps 12 m (40 ft) below the ground surface (not surveyed), the rectangular passage cross section changes to a keyhole shape. The top tubular portion of this keyhole-shaped cross section may reflect the elevation of a former water table in the aquifer. Figure 3 shows the distinct change in passage trend (to S13°W) toward the southwestern terminus, aligned with the strike of the bedrock. Palmer (1991) established that groundwater flow in the vadose zone is controlled by bedrock dip (with localized joint control), and follows strike in the phreatic zone. Southwestern continuation of this passage is blocked by flowstone, but a narrow joint some 12 m (40 ft) northeast of the flowstone occlusion, drops vertically to a small stream that flows at the base. Assuming that the short section of tubular passage (i.e., the top portion of the keyhole), does reflect a former water table; the deep solutionally-enlarged joint represents stream piracy to a new and lower base level.

Efforts by Nick Viscio and others to extend this cave beyond the base of its southwestern pit and into the larger portion of the carbonate aquifer in the late 1970s did not meet with success. The entrance to the cave was physically closed in 1984.

OTHER KARST FEATURES IN THE GROUNDWATER BASIN

Other small caves and karst features within the carbonate hydrologic unit provide information on relict and active flowpaths within the epikarst. These include a number of very large sinkholes present within the groundwater basin. Some have deep solutionally enlarged joints that provide infiltration pathways into the epikarst and underlying carbonate aquifer. Many are situated north and north-northwest of Hollyhock Hollow in a limestone area of about 13 km² (5 mi²). This area contains many karst features too numerous to discuss here. However, some mention of the significant ones is warranted.

Two small caves are known north of Hollyhock Hollow, Hole in the Wall Cave and Brokendown Cave. The former is a small cave located about 1,950 m (6,400 ft) north of the overflow spring in a large limestone quarry north of Albany County's filtration plant. This is the "Quarry Near Feura Bush" of Marshak (1990). Hole in the Wall Cave, a small relict cave, is exposed high up on the quarry's southern wall. The cave formed in the Manlius Limestone and served as a vadose infeasor to a downgradient segment of a larger cave system. The initial passage trend is S24°W. The passage averages 0.6 m (2 ft) wide and 0.9 m (3 ft) high for approximately 2.7 m (9 ft) before turning to the

WILTSIE'S CAVE

Albany County, New York

Brunton-and-Tape Survey
7 August 1972
by Ernst H. Kastning and George Hrepta

Redrafted August 1995 by E.H. Kastning
from detailed manuscript map at 1:120 scale

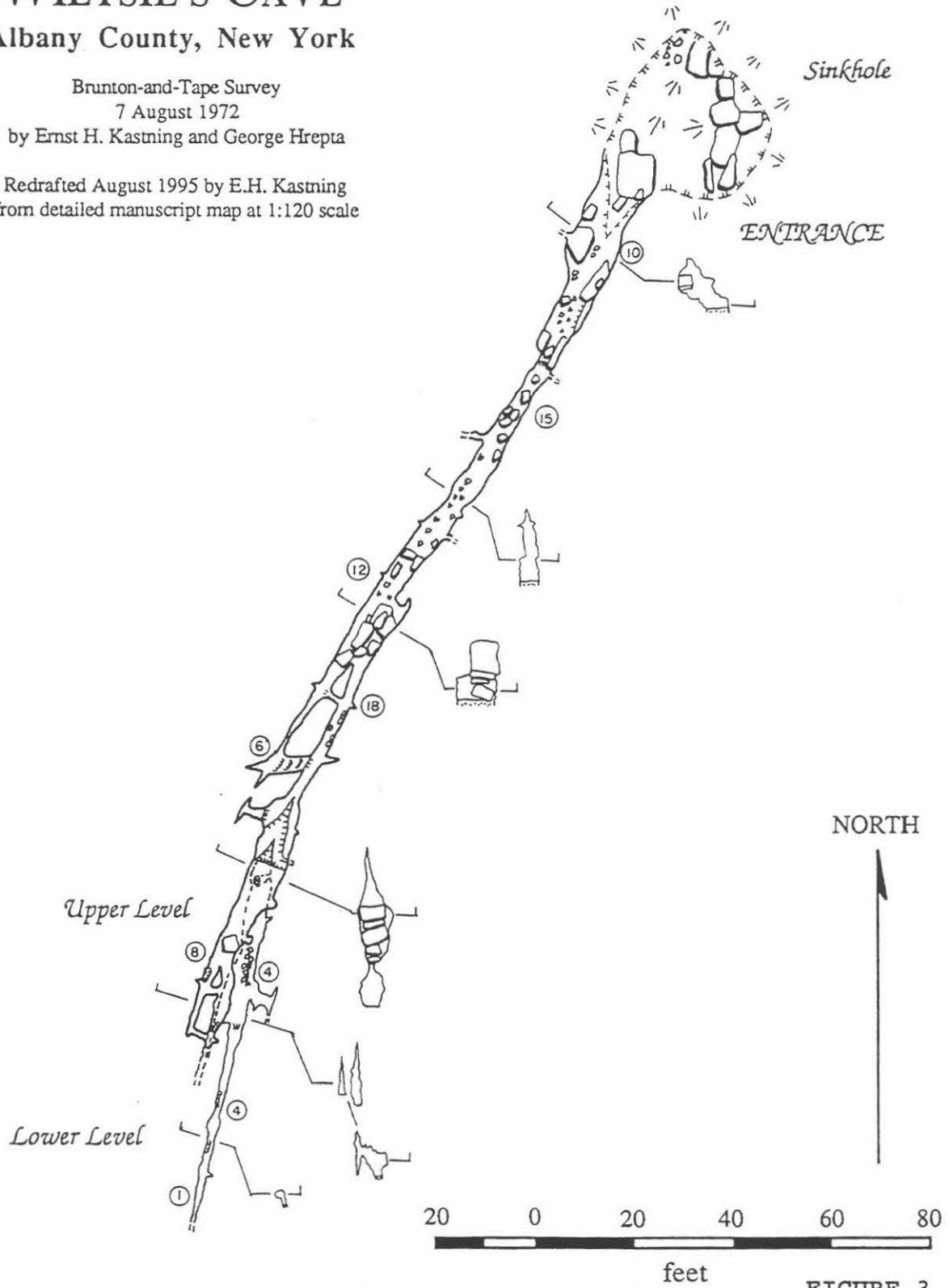


FIGURE 3

southeast. A calcite mineralized bedding fault is present in the cave, but it did not control passage development.

Brokendown Cave is located about 370 m (1,200 ft) south of Hole in the Wall Cave in the base of a small sink in a large shallow depression. The cave is morphologically complex and contains about 75 m (250 ft) of passage. It is formed along two (2) parallel joints with connections between the joints at two or more levels. Overall, the cave gives the impression of a vertical maze.

About 600 to 1,200 m (2,000 to 4,000 ft) west of Brokendown Cave and 1,500 to 2,300 m (4,900 to 7,400 ft) north of the overflow spring are three large closed depressions. Each of these receives runoff from outside the topographic limits of the depressions. The western-most of these sinks is said to contain a 12 m (40 ft) deep pit called Dribblemouth Pit. Efforts to find this pit in 1993 and 1994 proved unsuccessful. The depression with the largest drainage area is the northern-most one.

About 3,200 m (10,500 ft) north-northwest of the overflow spring is an occluded lateral resurgence near the base of the Helderberg Escarpment. Little is known about this feature due to access difficulties. The lack of resurgences in the immediate vicinity seems to indicate that water sinking here is flowing down-dip to the southeast. It may represent one of the farthest up-dip recharge points for the Hollyhock carbonate aquifer. About 450 m (1500 ft) west of this resurgence is a small cave called Tin Can Cave. It is at the bottom and on the south side of a closed depression about 11 m (35 ft) deep and 150 m (500 ft) in diameter. The cave is fed by a small lateral resurgence with a limited drainage area.

AQUIFER CHARACTERISTICS AND EVIDENCE FOR CONDUIT PATHWAYS FROM WELLS

Data obtained from well drilling in the area provide information on the nature of the carbonate aquifer. A number of wells have produced little or no water, often causing property owners to drill numerous wells. Wells along Rarick Road in fractured limestone are reported to have yields of approximately 0.1 l/s (2 gpm), with one yielding in excess of 0.6 l/s (10 gpm) (Banahan, pers. comm.). Some property owners along Rarick Road report sulfur water, probably stemming from pyrite in the Snake Hill shale.

Two wells in close proximity, situated approximately 600 m (2,000 ft) northwest of the New York Audubon headquarters building, penetrated approximately 6 m (19 ft) of "mason sand" and 29 feet of limestone prior to penetrating two foot conduits filled with sand. Banahan (pers. comm.) reports yields in excess of 0.6 l/s (10 gpm) from these conduit wells. Sand filling of this or these conduits indicates pre-Holocene formation with infilling by deltaic sands, perhaps those that only a short distance to the southeast formed a delta into glacial Lake Albany.

Another exception to problems found obtaining water in the aquifer is the New York Audubon headquarters building well (situated in the parking lot area). The well is 16 m (54 ft) deep with a static water table of 12.87 m (42.22 ft) below the top of casing on June 28, 1995. Analysis of a short-term (160 minute) pumping test conducted on the aquifer (using the Theis nonequilibrium well equation and type curves) indicates an aquifer transmissivity on the order of 110 m²/d (9,000 gpd/ft), enough to provide a small municipality with water. Assuming no hydraulic boundaries are encountered during longer-term pumping, this well may be capable of providing a sustained yield of 300 m³/d (55 gpm; ≈ 79,000 gpd). A log of the well is not available, but it may be inferred that it is completed in 1) the highly fractured Manlius Limestone below the elevation of the cave system; 2) deltaic and glacial sediments present in a shallow incised and filled bedrock-flanked channel of the Onesquethaw Creek; or 3) a segment of the lower end of the cave system, backflooded behind the now sediment-occluded cave mouth. Another well on New York Audubon property, situated approximately 120 m (400 ft) to the northwest, at approximately the same surface elevation, is 10 m (32 ft) deep, with a water table at 6.80

m (22.32 ft) below the top of casing. Groundwater in this well probably flows through the fractured Manlius Limestone to either the Hollyhock Overflow Spring Cave or directly to the Onesquethaw Creek.

Five wells were drilled to depths of 49-67 m (160-220 ft) around Wiltsie's Cave in an effort to find water for Hereford cattle (Banahan, pers. comm.). Occasional voids up to 0.15 m (0.5 ft) were found prior to reaching the black, middle Ordovician, Snake Hill shale (designation of Isachsen et al., 1991) between 34 and 37 m (110 and 120 ft) below ground surface. Not surprisingly, groundwater was not found in randomly drilled wells in the Coeymans and Manlius Limestones, as groundwater converges in karst settings toward conduits where the hydraulic head is lowest. Quinlan and Ewers (1985) have estimated the odds of encountering a dissolutional conduit in a karst aquifer by drilling a well at about 1:2600. If access to the cave system and its stream are gained through either the Rienow Swallow Hole or Wiltsie's Cave, a productive well location will be radio-located.

Drilling by Banahan (pers. comm.) northeast of Wiltsie's Cave revealed a 1.7 m (5.5 ft) conduit (i.e., cave passage) with in-washed gravel in limestone at a depth of 16 m (52 ft) below the ground surface. This well has continuously supplied water to his barn since 1983, except under drought conditions when groundwater tributary to the well temporarily ceased flowing for 16 days in July 1995. Two other carbonate wells situated within approximately 120 m (400 ft) south of the barn conduit well encountered sediment plugged conduits from 15 to 17 m (50 to 56 ft) below the ground surface. This plugged conduit(s) provides evidence for either 1) a now abandoned segment of a cave system within the Hollyhock carbonate aquifer that carried groundwater sometime prior to the most recent deglaciation, or 2) a partially sediment occluded, but still functional, conduit that carries groundwater alongside sediment fill. An attempt to flush and blow out these sediments failed. The elevation of this conduit (the more likely situation) or conduits may coincide with the Manlius/Coeymans contact.

The location of the conduit or conduits found in these three wells may imply that subsurface drainage trends further down-dip (4-14° SE) and southeast of Wiltsie's Cave and the Rienow Swallow Hole prior to encountering a structural barrier or fault plane that shunts groundwater flow southwest to the Hollyhock Hollow base and overflow springs. Northward projection (\approx N21°E) of the steeply dipping thrust fault along which the Hollyhock Hollow Overflow Spring Cave formed places it somewhere near these wells. This suggests that either 1) the 50°+NW dipping limestone beds of the fault's footwall form a structural barrier to southeastern groundwater flow, and/or, 2) as observed in the Hollyhock Hollow Overflow Spring Cave, groundwater flow and conduit development occurs preferentially along the thrust fault plane and associated fractures.

PALEOCLIMATE

Evidence was found documenting that the Hollyhock carbonate aquifer formed prior to the late Wisconsinan (prior to ca. 24,500 yr B.P.; Miller and Calkin, 1992) Woodfordian ice advance. The history portrayed in the Rienow Swallow Hole and Wiltsie's Cave provides evidence that the Onesquethaw Creek channel and regional drainage patterns were adjusted to today's base level prior to cave formation and glacial modification. Interpretation of physical features within the Hollyhock Hollow Sanctuary suggests that a cave system formed in pre-Woodfordian time and was subsequently enlarged by subglacial meltwater invasion. The lengthy time (at least 10,000 years) required to enlarge fractures into conduits, such as in the Rienow Swallow Hole, argues for long-term seasonally warm (above 0° C) climatic conditions. Perhaps these conditions occurred during recessional stages of glaciation. Cave enlargement is interpreted as occurring under subglacial conditions, when a very different flow regime was present. Seasonal meltwater beneath glacier ice may have flowed down a series of bedrock troughs and into several swallow holes as it drained toward the already well-adjusted Onesquethaw Creek.

The pedestals, scallops in the Rienow Swallow Hole and Wiltsie's Cave, and meltwater channels provide evidence for widespread and long-term surface water flow throughout the Hollyhock area. The present catchment is not sufficient to provide the required water quantities. A physically enlarged glacial ice catchment is inferred, with large seasonal flows under warm based ice. Even the minimum rates of cave development, for caves requiring water from beyond the physically available catchment (i.e., from under glacial ice extending to the northwest), argue for extended periods of seasonal warmth during glaciation. Similar seasonal variations occur today in Castleguard Cave that extends under the Columbia Ice Field in the Canadian Rockies.

Speleothems recovered from sediments in the Rienow Swallow Hole provide evidence for cessation of glacial meltwater inflow for extended periods during interglacial or interstadial periods. Drier physical and climatological conditions permitted stalactite, stalagmite and flowstone formation in this shallow cave passage. Several erosion phases are indicated by truncated flowstone growth and the incorporation of a broken flowstone fragment into another speleothem. $^{230}\text{Th}/^{234}\text{U}$ dating of speleothems may elucidate the geomorphic chronology.

LAND USE CONSIDERATIONS

Carbonate aquifers and their receiving streams (i.e., the Onesquethaw Creek) are very sensitive to contaminant inputs and require special land use consideration. Carbonate aquifer hydrology is very different from porous media (i.e., soil) and fractured bedrock aquifers with slow laminar groundwater flow, instead being characterized by rapid non-Darcian (i.e., turbulent; non-laminar) subsurface flow through conduits (i.e., caves) with no natural filtration of contaminants. In the classification of carbonate aquifers of Quinlan et al. (1992), the vulnerability or susceptibility of field trip karst aquifers to groundwater pollution is hypersensitive.

Whereas subdivision and development within karst basins has historically occurred on an individual application basis, a more broad-based master planning process is needed to maximize protection of groundwater and surface water resources. Planning in environmentally sensitive areas should take into account the likely cumulative contaminant loading into the karst system, and a reasonable measure of it and its receiving stream's assimilative capacity. The development of an area must be within the natural constraints of its geology and hydrology (Rubin, 1990, 1992, 1994c, 1995).

Knowledge of flow dynamics in the epikarst is critical in land use planning above hydrologically sensitive carbonate aquifers. Surface runoff and infiltration into Wiltsie's Cave and the Rienow Swallow Hole (relict caves no longer receiving the water responsible for forming them) verifies the integration and continuum of groundwater flow between relict and deeper, higher discharge drainage routes. Thus, solutionally enlarged fractures and relict caves and swallow holes in the groundwater basin are integral, functioning components of the epikarst, funnelling runoff and infiltrating meteoric waters to deeper conduit flow routes.

Significant quantities of waste water and other contaminant additions to Albany County carbonate aquifers may contribute to groundwater and surface water degradation. Because the Hollyhock carbonate aquifer has 1) little or no soil-mantle; 2) a well-developed epikarst; 3) an efficient carbonate aquifer; 4) much of its drainage basin (and thus its contaminant assimilative capacity) beheaded at the Wolf Hill Dam; and 5) an influx of periodic nutrient, pesticide, and pathogen additions from the stressed and threatened Helderberg Lake, it is worthy of special land use and, ultimately, zoning consideration. We recommend that long range plans for development for this area not be adopted until the karst hydrology present is further investigated, critical area maps are made available, and potential downgradient impacts are addressed.

INTRODUCTION: JORALEMON PARK

At Joralemon Park and vicinity, entrances to active (Hannacroix Maze and Merritt's Cave) and relict (Joralemon's) caves will be visited. Damming of a former surface drainage route by glacial sediment, as well as erosional derangement of surface streams, have resulted in the enlargement of the Hannacroix Maze floodwater cave. All surface flow from throughout the basin is deranged from an earlier flow route and is now pirated through Hannacroix Maze. Prior to deposition of the sediment dam that now forces surface water from throughout the basin to be pirated through Hannacroix Maze, surface flow occurred through Joralemon's Cave. Nardacci (1994) has documented much of the karst hydrology of this stop.

JORALEMON PARK LOCATION

The Hannacroix Maze karst is located along Route 102, slightly northwest of the village of Ravena, Albany County, New York (see Fig. 8 and Road Log). A portion of the study area occupies a section of Joralemon Park, owned by the Town of Coeymans.

GEOLOGY

Three rock units outcrop in Joralemon's Park. All cave development has occurred in the fossiliferous, light bluish-gray Onondaga Limestone. The Onondaga is subdivided into subunits based on the presence or absence of black chert beds. Fracture enlargement and conduit development occurs in all subunits, but the more massive chert-free subunits tend to exhibit larger conduit development. Chert beds sometimes temporarily perch groundwater flow until a fracture ultimately permits downward flow into a less cherty subunit. Cave development in the Onondaga Limestone proceeds through all subunits, with the possible exception of where the hydrologic base level lies above the base of the unit. The Onondaga Limestone is recognized as one of the best cave-forming rock units in New York State.

The Onondaga Limestone is underlain by a thin bed of the Schoharie Formation which is a silicious, clay-rich, dolomitic limestone and a poor cave former. It is dark bluish gray in color, weathering by solution of the lime into a brown porous sandstone. This unit may serve as the base level control for downward cave formation in Joralemon Park. The Schoharie Formation is underlain by the thick, impermeable Esopus Shale. The Esopus is composed of dark brown to black shale and siltstone. It is often recognized in exposures by the *Zoophycus* trace fossil that resembles a rooster's tail. The Schoharie Formation and the Esopus Shale belong to the Lower Devonian Tristates Group.

The Tristates Group lies stratigraphically above the Helderberg Group. The lowest two units of the Helderberg Group, the Manlius and Coeymans limestones, were observed at the Hollyhock Hollow Sanctuary.

HANNACROIX MAZE

Hannacroix Maze (Fig. 4) is developed in the lower subunits of the Onondaga limestone above the contact with the Schoharie Formation. It is a low, wet network of joint-controlled passages (Nardacci, 1994), presenting a classic example of Palmer's (1975) floodwater maze. A detailed map of the area (Fig. 5) shows a sediment ridge preventing the small stream and runoff in the area from flowing freely to the south, resulting in water ponding to the northeast of Hannacroix Maze. While there is no dramatic elevation change as surface water flows south, it is funneled into a relatively narrow valley at the foot of which is the ridge containing Hannacroix Maze and, farther south, Merritt's Cave. Hence, during floods water has sufficient head to be injected into all available joints, thus producing a maze pattern.

HANNACROIX MAZE

Albany Co., NY

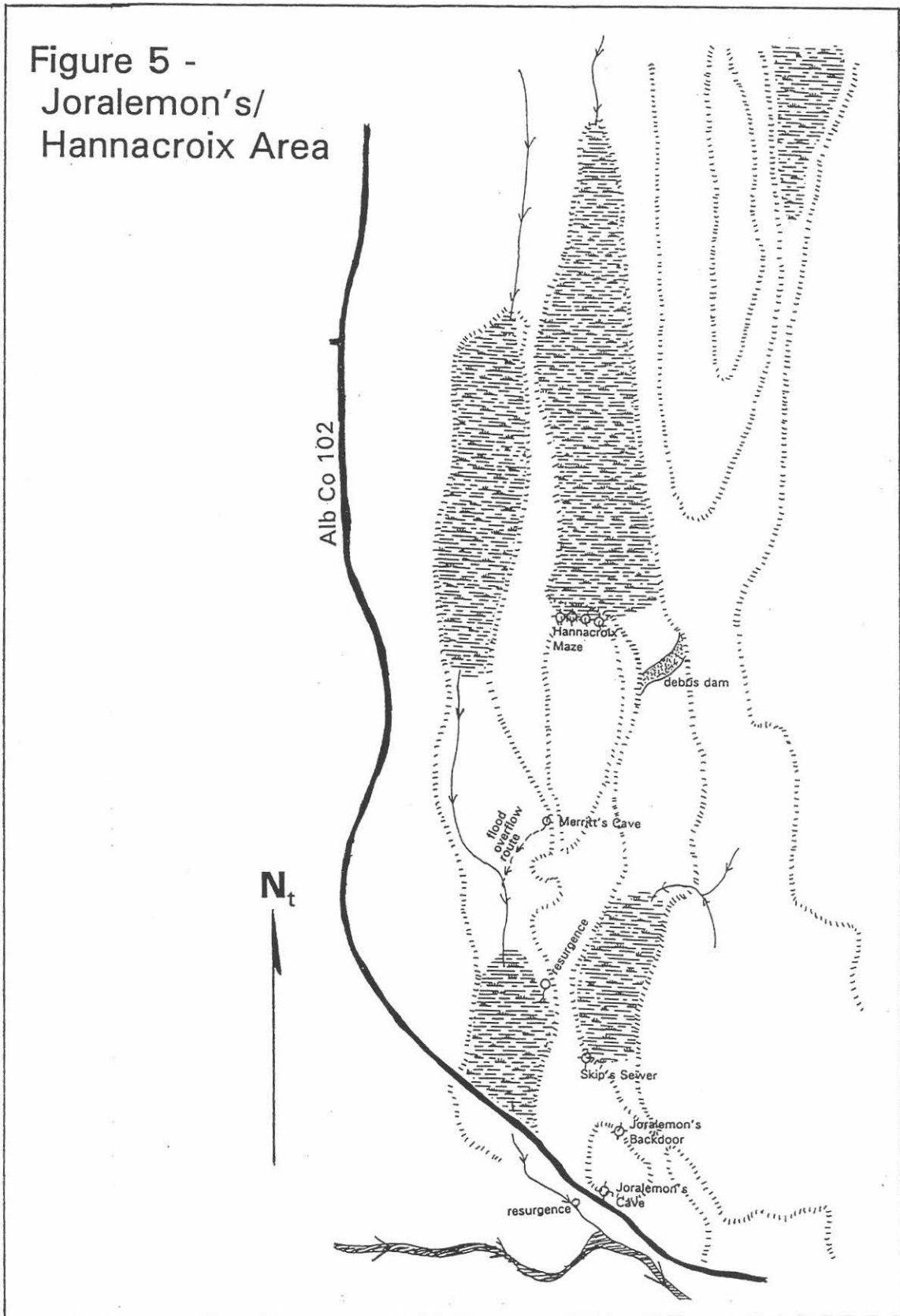
Grades 3, 4, & 5 in larger sections. Grade 1 & 2 elsewhere

T. Engel, D. Gregg, W. Gregg, D. Hauser, G. Hrepta, P. Rubin, N. Thompson



Figure 4 - Hannacroix Maze

Figure 5 -
Joralemon's/
Hannacroix Area



Although the passages of the cave are small - with heights frequently less than one meter and their widths rarely exceeding that - there is evidence for turbulent flow during floods. Scallops on the walls commonly show lengths of less than 5 centimeters (2 in), and huge quantities of flood-borne organic debris such as bark, tree limbs and leaf fragments are found in the extremities of the passages and jammed into ceiling fissures.

Most of the development and enlargement of Hannacroix Maze probably occurred during and after Woodfordian glaciation. Deposition of the small sediment ridge southeast of the Hannacroix Maze resurgence area may have occurred during Woodfordian glaciation, since most surficial soils reflect the most recent glacial processes. A minimum of 14,700 years (following retreat of Woodfordian ice) (DeSimone and LaFleur, 1985) have been available for cave enlargement and perhaps more if subglacial meltwaters also contributed to enlargement of the pre-existing fracture network. Evidence of such subglacial meltwater flow is provided by Dineen (1986), Rubin (1991b and this volume), and at the Hollyhock Hollow Sanctuary (this paper). Both Woodfordian and Holocene enlargement of Hannacroix Maze and Merritt's Cave is more likely than solely post-glacial development because solutional cave formation requires a minimum of 10,000 years (Palmer, 1991) before passageways obtain sufficient size for human entry. Additional passage cross-sectional size requires additional time. Holocene climatic conditions have resulted in frequent seasonal flooding of the cave system. Both Hannacroix Maze and Merritt's Cave usually flood annually.

The age of the cave has not been determined, although $^{230}\text{Th}/^{234}\text{U}$ disequilibrium series dating of speleothems may help establish a minimum date of formation. Reconnaissance to date has not revealed allochthonous glacially-derived sediments, but they may be present under the surficial ooze. Furthermore, the presence of such deposits may simply signal Holocene inwashing.

MERRITT'S CAVE

Merritt's Cave, like Hannacroix Maze, has not yet been fully explored and surveyed. It almost certainly connects with Hannacroix Maze. Like Hannacroix, it is a maze cave, although its entrance resembles a talus cave. On this western edge of the ridge, numerous boulders have slumped and been moved by glacial ice from the bedrock, creating numerous false entrances and squeezes that obscure the actual entrance to Merritt's Cave. One section of the cave that can easily be entered is a high narrow canyon. Scallops in the walls average less than 5 cm (2 in) in length, indicative of rapid flow. Like Hannacroix, Merritt's Cave contains numerous small side passages, some of which pinch out rapidly or occlude in breakdown, and some of which may eventually be pushed by explorers to extend the cave's length. By considering the relative physical positions of Hannacroix Maze (600 m; 2,000 ft) and Merritt's Cave (275 m; 900 ft) and their lengths, there may be hundreds of meters of unentered passage in the cave system.

Merritt's Cave is a floodwater overflow spring resurgence. Water flowing out the entrance must rise about two meters. In normal flows, a small stream is seen inside the cave. This flows into an inefficient drain just inside the eastern-most entrance. In low- to moderate-flow conditions, the water resurges at an occluded spring about 240 m (800 ft) south of Merritt's. Four other occluded springs are also present in the area of this resurgence. All five springs may discharge from a single conduit. As flow increases, each successive spring to the north and elevationally higher begins to flow. During extreme flood, the highest and most northerly spring flows.

A sixth resurgence is present some 130 m (425 ft) south of Merritt's. It resembles a sinkhole in shape. At very high flow, water rises from a pool in the depression and overflows to the west.

As already noted, Merritt's Cave serves as a floodwater resurgence for Hannacroix Maze. About 15 m (50 ft) downstream of the Merritt's Cave entrance, surface flow sinks into the subsurface. At certain flows, water coming from Merritt's entirely sinks at this point. As flow increases beyond the capacity of this sink, the water flows west, then south on the surface. Of interest is a low bedrock dam at the bend in the water course. In times of very high water, water is impounded behind this dam, composed of faulted, $\approx 30^\circ$ SE dipping, Schoharie Formation beds. Beyond this, the stream flows south to a tributary of Hannacroix Creek.

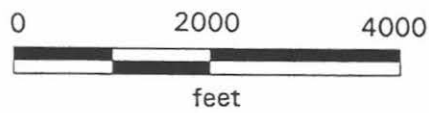
JORALEMON'S CAVE


Joralemon's Cave and Joralemon's Backdoor (Fig. 7) are of particular scientific value because together they preserve what may be the oldest remaining geologic record of the drainage patterns, base level, and rates of regional erosion present many tens or even hundreds of thousands of years ago in the Joralemon Park area (Fig. 5). The way in which their rock layers are carved tells a story of the evolution of the landscape we see today. Joralemon's Cave and Joralemon's Backdoor are two segments of a single cave, now separated by sediment. When this cave formed, water that flowed through it did so at the lowest topographic drainage level. The surface topography and drainage routes present at this time were quite different from what they are today. We believe that the limestone knoll that Joralemon's Cave formed in once extended farther to the east and west. The age of the cave's formation and downward erosion of the limestone to the east is not known. Although erosion (glacial and non-glacial) and downcutting have left it as an abandoned cave segment high up in a ridge, its physical setting was such that it once occupied a low valley. Runoff from higher elevations to the north flowed downhill until sinking into fractures and sinkholes draining into Joralemon's Cave.

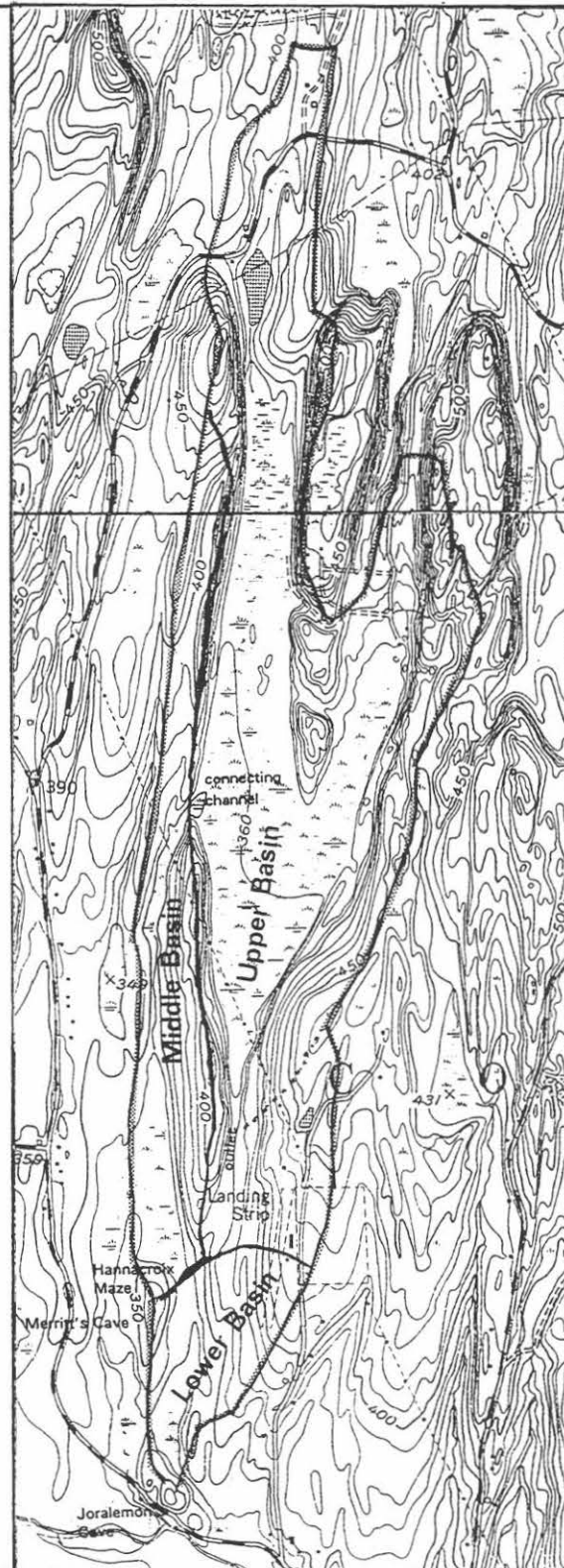
Preserved in a resistant and faulted Onondaga Limestone knoll, Joralemon's Cave is a large-diameter (6^+m^2 ; 65 ft^2), relict cave segment partially filled to an unknown depth with sediment. It is deranged from present day drainage, much of which now flows through Hannacroix Maze and Merritt's Cave. Scallop marks on the walls of Joralemon's Cave provide evidence of rapid flow to the ancestral Hannacroix Creek. Using the methodology described previously (see Determination of Paleoflow into the Rienow Swallow Hole), scalloped walls in Joralemon's Cave indicate paleoflows varying between 4.2 and $10.5\text{ m}^3/\text{s}$ (150 and 370 cfs) with velocities of ≈ 0.7 to 1.7 m/s (2.3 to 5.6 ft/sec). The exposed cross-sectional area of Joralemon's Cave indicates active flow for a minimum of 10,000 years, possibly tens of thousands of years. Deposition of the glacial sediment dam southeast of Hannacroix Maze is believed to be a factor that resulted in the derangement of surface flow formerly draining to Joralemon's. This derangement contributed to the ultimate abandonment of Joralemon's Cave. Water-worn bear and muskrat bones excavated from Joralemon's Cave indicate Pleistocene stream-borne deposition. Holocene age bones of northern wood rat, frogs, turtles, and many other species have also been found in the cave (Steadman, pers. comm.).

Determination of the age of formation of Joralemon's Cave presents more of a challenge than the Hannacroix Maze/Merritt's cave system. Since all present drainage occurs through conduits elevationally lower than Joralemon's (i.e., Hannacroix Maze/Merritt's system) and Joralemon's Cave is deranged from today's drainage patterns, it is our interpretation that cave formation and drainage through Joralemon's Cave predates that of Hannacroix Maze. In keeping with this interpretation, the elevation of Joralemon's Cave might be used to assess a former hydrologic base level or infer development during another glacial or interglacial period. Sediment infilling in the back of Joralemon's Cave reveals that the conduit dips downward toward Joralemon's Backdoor. The last chapter in the flow history of Joralemon's Cave is recorded in the sediment infill that now floors the cave and blocks the former connection route between Joralemon's Backdoor and Joralemon's Cave, perhaps coincident with the final retreat of Altonian (pre-Woodfordian) ice from this area.

Figure 6 - Surficial
Drainage Basins for
the Joralemon's/
Hannacroix Area



basin boundary 



JORALEMON'S CAVE JORALEMON'S BACKDOOR

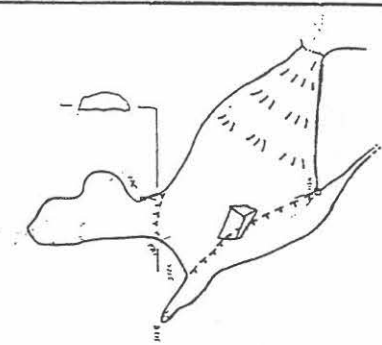
Shown in their relative positions

Grade 5

T. Engel

F. Torncello

M. Torncello



Joralemon's Backdoor



Joralemon's Cave

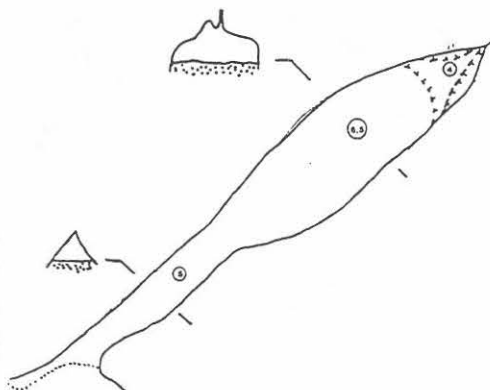


Figure 7

Development and enlargement of Hannacroix Maze and Merritt's Cave can be inferred to have occurred predominantly after the abandonment of Joralemon's Cave, perhaps with initial formation coincident with the drop in base level that ultimately led to the abandonment of Joralemon's Cave. Woodfordian glacial sediment dams block drainage east of the Hannacroix ridge (the low limestone ridge containing Hannacroix Maze and Merritt's Cave), furthering development of Hannacroix Maze and Merritt's caves. While the surface catchment tributary to Hannacroix Maze is on the order of 2.4 km² (0.9 mi²) (Fig. 6), it is important to recognize that the subglacial catchment basin tributary to these caves was once greatly expanded. Thus, flow beneath warm-based ice straddling the Hannacroix Cave ridge, the Joralemon's knoll and the ridge east of Joralemon's knoll converged down to a narrow, restricted outlet area.

The rates of dissolution required to develop a conduit the size of Joralemon's Cave suggests periodic floodwaters flowed through the cave for a period in excess of 10,000 years. Furthermore, the quantities of water needed to largely fill (or at least obtain the discharges indicated by scallops) Joralemon's Cave require a drainage area larger than that available today or a greater source of water, or both. Both factors can be associated with a subglacial catchment basin. Thus, we infer that climatic conditions remained stable for thousands of years, with seasonal warming, during glaciation.

JORALEMON'S HYDROLOGIC CHRONOLOGY

There are several possible interpretations that can be put forth regarding the ordering of events that shaped the landscape we see today in the Joralemon's/Hannacroix area. We offer the following proposal.

Figure 6 shows the surficial drainage basin for Hannacroix Maze as well as a southern portion extending south to Joralemon's. For the sake of discussion we have divided the basin into three sub-basins: upper, middle, and lower.

At one time water may have flowed directly from the upper basin into the lower basin through the pass at the very southern end of the upper basin. At this time, water was flowing through Joralemon's Cave, probably aided by subglacial meltwaters. During this same time, some water was also flowing in the middle basin. However, due to the small size of this basin only a fraction of the discharge currently seen was flowing here. We believe that a small "ancestral" Hannacroix Maze formed at this time. The larger passages in this cave such as the Fungus Footpath (see Fig. 4) may have formed at this time. It is interesting to note that the major secondary carbonate deposits in the cave are found only in these larger passages.

At some later point, the upper basin ceased to flow into the lower basin. This may have been caused by glacial damming of the outlet from the upper basin some 600 m (2,000 ft) northeast of Hannacroix Maze or by stream piracy into the middle basin by headward erosion or by both. Stone walls found in the former outlet of the upper basin contain a high percentage of glacial erratics. If we assume that the material used to build these walls was from the immediate vicinity, then glacial damming seems highly likely. The sediment dam (Fig. 5) southeast of Hannacroix Maze is located at the junction of the upper and lower basins. It is approximately at 335 ft msl, an elevation of one stage of glacial Lake Albany (Dineen, 1986). The dam has likely been reworked by water that still flows during periods of high runoff.

More water in the middle basin resulted in backflooding and enlargement in both Hannacroix Maze and Merritt's Cave. Even now water levels in Merritt's may vary during the course of the year by as much as 2.4 m (8 ft).

The greatly reduced flow in the lower basin found a new outlet just north of Joralemon's Backdoor. There appears to have been a gradual headward abandonment of insurgences in this area to its current location at Skip's Sewer (Fig. 5). This water may resurge in the pool across Rt. 102 from the entrance to Joralemon's Cave.

ACKNOWLEDGEMENTS

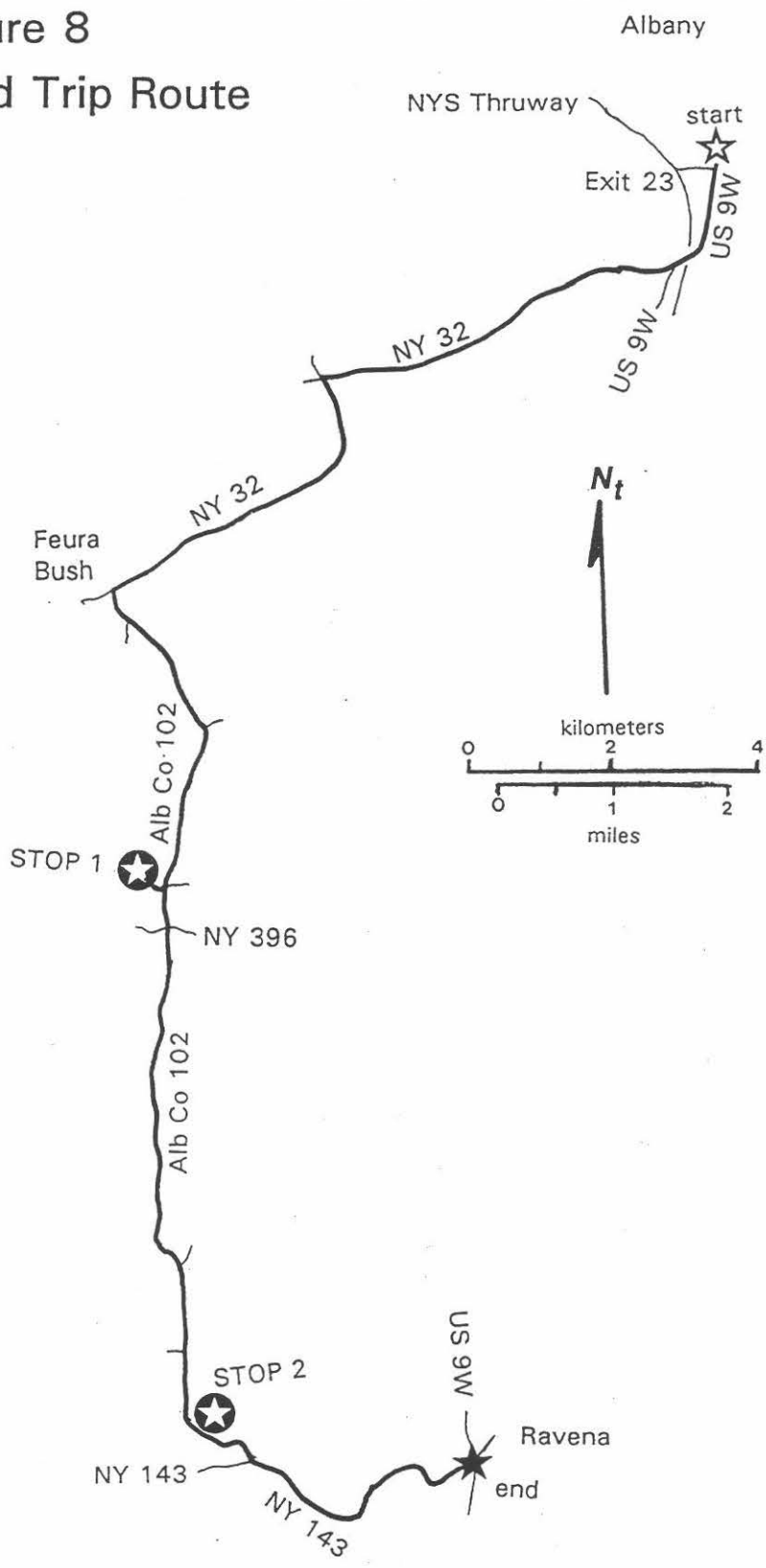
Heartfelt thanks are extended to Tom Uhl, Paul Woodell, Daniel Low and the many northeastern cavers and New York Audubon staff members who have contributed to the various activities associated with this ongoing study. Different aspects of the project have included stream gaging, stream monitoring, leveling, surveying, drafting and many fine hours of digging with friends. Phil Bodanza deserves special recognition for fabricating and installing the staff gauge support structure and the gate on the Hollyhock Hollow Overflow Spring Cave. We are especially grateful to The Audubon Society of New York State, Inc. and the Town of Coeymans for granting access to their lands. Thanks are also extended to Ernst Kastning for providing the Wiltsie's Cave map and Bill Banahan for sharing important well information.

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- , 1994c (Nov. 25), Planning considerations above Town of Bethlehem carbonate aquifers; comments on master plan: Letter to Town of Bethlehem Town Board, 6 p.
- , 1995 (Aug. 15), Water quality protection of the Fox Creek: A water supply resource for the Village of Schoharie; comments specific to the proposed subdivision of the Lawyer property; Letter to Town of Wright Planning Board, 7 p.

Figure 8
Field Trip Route



ROAD LOG: HOLLYHOCK HOLLOW SANCTUARY AND JORALEMON PARK

<u>Total Miles</u>	<u>Miles From Last Point</u>	<u>Route Description</u> (see Figure 8)
0.00	0.00	<u>START</u> - Jct US 9W and Thruway Exit 23. Proceed south on 9W.
0.45	0.45	Cross Normanskill Creek on US 9W.
0.80	0.35	Jct NY Rt 32 and US 9W. Continue south on US 9W and NY Rt 32.
1.20	0.40	Rts 32 and 9W split. Veer right on NY Rt 32.
4.45	3.25	Turn left following NY Rt 32.
6.20	1.75	Cross Vlomankill.
7.60	1.40	Turn left on Albany County Rt 102 (Old Quarry Rd.).
8.10	0.50	Exposure of Snake Hill shale behind house on right.
9.30	1.20	Feura Bush Quarry uphill and on right.
10.00	0.70	Roadcut through Coeymans Limestone.
10.70	0.70	Turn right onto Rarick Rd.
10.90	0.20	<u>STOP #1</u> - Turn left into the New York Audubon parking lot. The Hollyhock Hollow Nature Sanctuary is open to the public. Turn right onto Rarick Rd. and return to Alb. Co. 102.
11.10	0.20	Turn right onto Albany County Rt. 102.
11.30	0.20	Cross the Onesquethaw Creek.
11.50	0.20	Jct with NY Rt 396. Continue south on Rt 102.
11.60	0.10	Cross Feuri Spruyt Creek.
12.35	0.75	Esopus Shale outcrop on right side of road. Continue S on Rt 102.
15.20	2.85	Large rock of Onondaga Limestone on right.
15.80	0.60	Joralemon Park tennis courts on right.
16.15	0.35	<u>STOP #2</u> - Pull off of Rt 102 into area on right. This is a Town of Coeymans park. Continue south on Rt 102.
16.65	0.50	End Albany County Rt 102. Turn left onto NY Rt 143.
19.15	2.50	<u>END</u> - Jct US 9W and Rt 143. Start of field trip is about 12 miles north on Rt 9W.

FLYSCH AND MOLASSE OF THE CLASSICAL TACONIC AND ACADIAN

OROGENIES: MODELS FOR SUBSURFACE RESERVOIR SETTINGS

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ABSTRACT

This field trip will examine classical sections of the Appalachians including Cambro-Ordovician basin-margin and basin-slope facies (flysch) of the Taconics and braided and meandering stream deposits (molasse) of the Catskills. The deep-water settings are part of the Taconic sequence. These rocks include massive sandstones of excellent reservoir quality that serve as models for oil and gas exploration. With their feet, participants may straddle the classical Logan's (or Emmon's) line thrust plane. The stream deposits are Middle to Upper Devonian rocks of the Catskill Mountains which resulted from the Acadian Orogeny, where the world's oldest and largest freshwater clams can be found in the world's oldest back-swamp fluvial facies. These fluvial deposits make excellent models for comparable subsurface reservoir settings.

INTRODUCTION

This trip will be in two parts: (1) a field study of deep-water facies (flysch) of the Taconics, and (2) a field study of braided- and meandering-stream deposits (molasse) of the Catskills. The rocks of the Taconics have been debated for more than 150 years and need to be explained in detail before the field stops make sense to the uninitiated. Therefore several pages of background on these deposits precede the itinerary. The Catskills, however, do not need this kind of orientation, hence after the Taconics (flysch) itinerary, the field stops for the Catskills follow immediately without an insertion of background information.

FLYSCH OF THE TACONICS

The first part of this field trip will examine classical sections of the Appalachians, specifically Cambro-Ordovician basin margin and basin slope facies of the Taconic sequences of rocks generally known as flysch. The term flysch is a corruption of the German verb *fliessen*, which means to *flow*. This term was applied because the outcropping parts of the shaley flysch in Austria were especially prone to slope failures. The strata named flysch are *a thick succession of marine sedimentary strata consisting of repetitively interbedded alternating and laterally persistent sands (and/or coarser sediments) and shales found in the interior of a fold-mountain chain*. These deep-water deposits are part of the Taconic Sequence (Fig.1). The term Taconic Sequence refers to basin strata correlative with various formations of the shelf (pericontinental) strata assigned to the Sauk Sequence and lower part of the Tippecanoe Sequence which spread across New York (Sloss, 1963; Guo, Sanders and Friedman, 1990) (Fig. 1).

The Appalachian Basin is a multi-stage foreland basin. A foreland basin is defined as a *sedimentary basin located between the front of a mountain range and the adjacent craton, and related to overthrusting at the convergent plate margin* (Friedman, Sanders, and Kopaska-Merkel, 1992). Thrusting is the active basin-forming mechanism.

The rocks of the Taconic Sequence are part of the Taconic allochthon, a remnant of an accretionary prism that includes latest Precambrian (?) -Early Cambrian rift facies and overlying Cambrian through early Middle Ordovician passive margin, deep-water sedimentary rocks. This sequence was pushed westward at least 30 km onto the founded autochthonous platform during the Middle Ordovician Taconic orogeny (Jacobi, 1981; Rowley and Kidd, 1981).

While deposition of shallow-water carbonate sediments was going on in a shelf or pericontinental sea throughout New York State, in a direction that during the Early Paleozoic was south, but is now east, deposition of both terrigenous and carbonate, but for the most part terrigenous sediment, was taking place in a slope-rise environment (Friedman, 1972; 1979;

Friedman, Sanders, and Martini, 1982). The shelf-edge setting was that of a passive margin which the Taconic tectonic event terminated. This event began in early Middle Ordovician times and lasted through the early Silurian. It involved collision of the North American continental margin with a belt of oceanic island arcs above an east-southeast dipping subduction zone (Rowley and Kidd, 1981; Hiscott et al., 1986)(Figs. 2, 3).

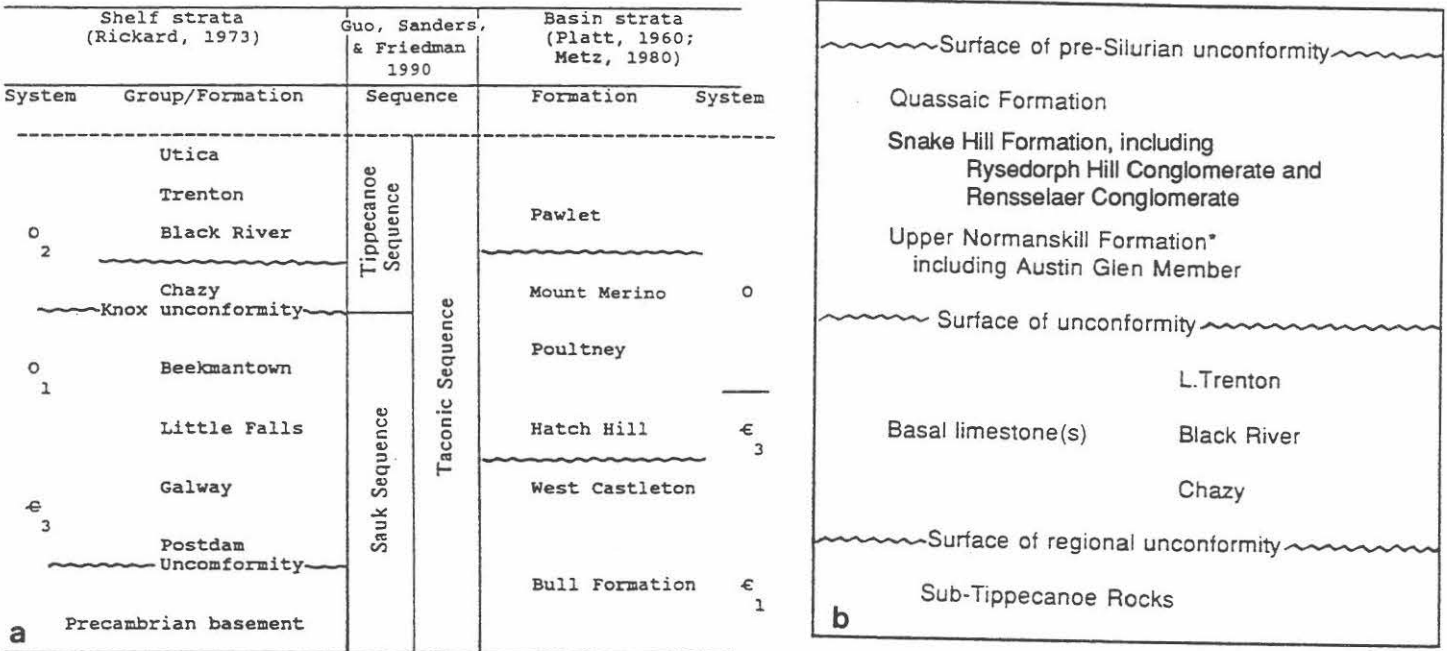


Figure 1. (a) Stratigraphy of Cambro-Ordovician in northern New York State (after Guo, Sanders, and Friedman, 1990). (b) Names of formations in Tippecanoe Sequence, eastern New York State (modified from Sanders, 1995).

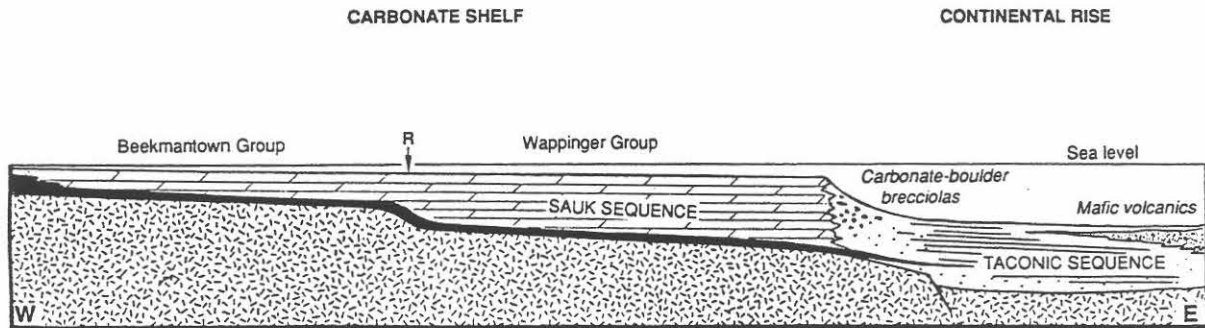


Figure 2. Reconstructed paleogeomorphologic profile-section across the Early Paleozoic passive margin in North America. Short line segments in random pattern, Precambrian basement; black basal quartzose sandstone (Upper Cambrian Potsdam on the W and Lower Cambrian Poughquag on the E). Dolostone pattern for Sauk Sequence includes limestones (now calcite marbles) on the seaward part of the former carbonate shelf (modern E; Early Paleozoic S) (Friedman, Sanders, and Guo, 1993).

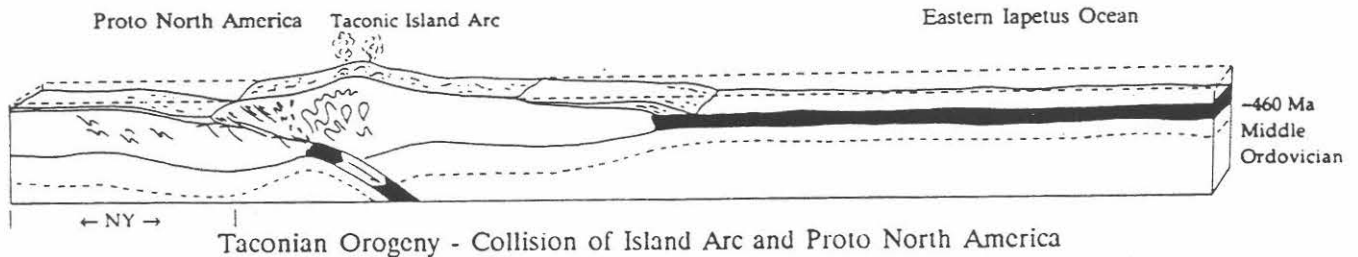


Figure 3. Block diagram showing the collision between the North American platform and island arc. This collision is the Taconic Orogeny (Ysachsen et al., 1991).

The stratigraphy of the northern Taconic sequence is shown in Figure 1. Previous workers considered the Cambrian interval which we shall study in the field as West Castleton Formation (Fig. 1)(Bird and Rasetti, 1968; Keith and Friedman, 1977; Friedman, Sanders, and Martini, 1982), but this formation is now mapped as part of the Hatch Hill Formation (Fisher, 1984; Landing, 1993). The rocks are part of the structurally lowest, youngest, less deformed and least metamorphosed Giddings Brook slice of the Taconic allochthon.

Between Early Cambrian and Early Ordovician times the shelf to basin transition was east of Rutland, Vermont. Tectonic movements shoved Cambrian and Ordovician rocks of slope, rise and basin facies across the shelf facies so that today the exposures at and near Troy, New York, are basin or basin margin (rise facies with shelf facies of Cambrian and Ordovician age occurring to the west) (Friedman, 1972) (Figs. 4, 5).

During Cambrian-Ordovician time, most of the North American continent was a shallow epeiric or epicontinental sea, like the present-day Bahama Bank, now often referred to as the Great American Bank. At the eastern edge of this shallow sea, that is at the eastern edge of this continent, a relatively steep slope existed down which terrigenous and carbonate sediment moved by slides, slumps, turbidity currents, mud flows, and sandfalls to oceanic depths to come to rest at the deep-water basin margin (rise), where a shale facies was deposited (Sanders and Friedman, 1967; Friedman, 1972, 1979; Keith and Friedman, 1977, 1978; Friedman and Sanders, 1978; Friedman, Sanders, and Martini, 1982). Shale also formed much of the basinal facies in the deep water beyond. Because allochthonous transport has been inferred for large blocks of rocks presently exposed in and near Troy, where our field study begins, the evidence on the ground shows that the area is the site of Cambrian and Early Ordovician rocks of basin margin (rise) and deep basin facies [shales deposited in the Middle Ordovician (Schenectady) west of Troy are autochthonous basin facies]. Thus deep-water basin margin (rise) and basinal facies can be visited in and near Troy, whereas to the west carbonate shelf facies are exposed that are analogous to those of the west shore of Andros Island on the Great Bahama Bank (Fig. 5). The paleoslope was probably an active hinge line between the continent to the west and the deep ocean to the east, similar to the Jurassic hinge line of the eastern Mediterranean between carbonate shelf facies and deep-water shales (Friedman, Barzel, and Derin, 1971). Such hinge lines in the early geosynclinal history of mountain belts are fixed by contemporaneous down-to-basin normal faulting (Rodgers, 1968, quoting Truempy, 1960), as probably occurred with the rocks in the Troy area. Later thrusting to lift the deep-water facies across the shelf facies along hinge-line faults resulted in the contiguity of the two facies. This later displacement was so great that the Cambrian and Early Ordovician deep-water sediments were shifted far west of their basin margin. The original basin margin (rise) was located near the present site of the Green Mountain axis.

Preceding plate collision, as the continental margin approached the subduction zone, the seaward part of the carbonate shelf floundered, probably due to "normal faulting caused by plate flexure with down bending" (Hiscott et al., 1986). As

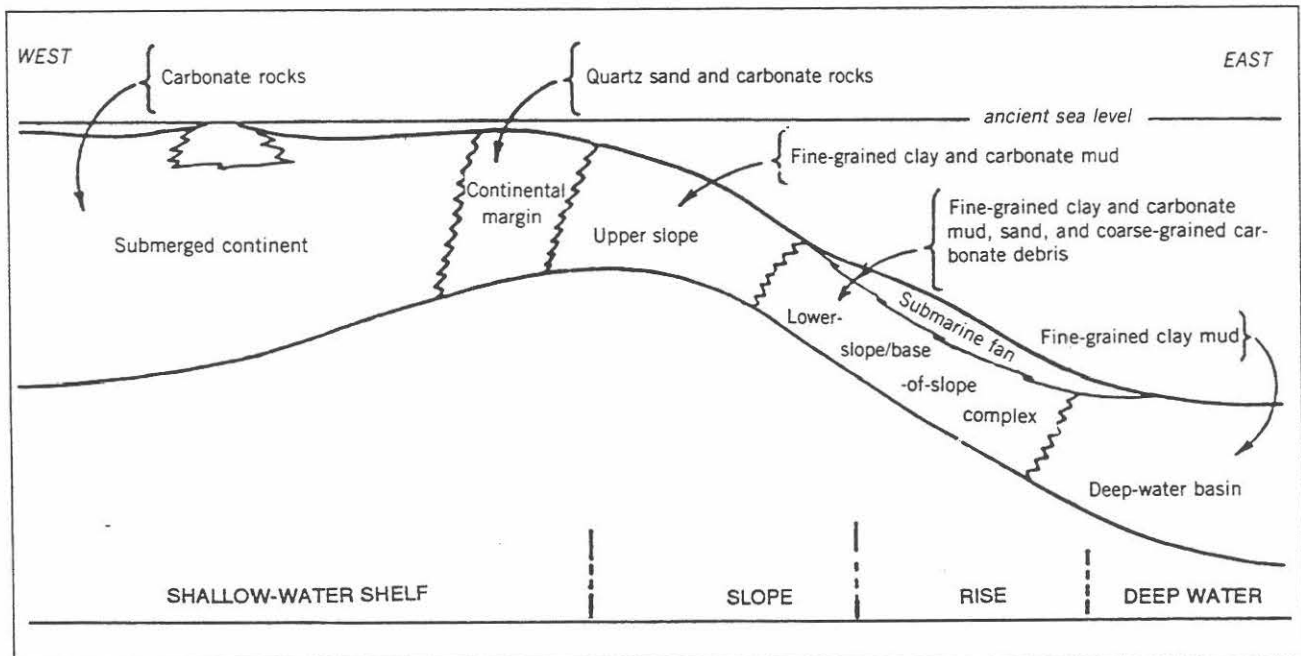


Figure 4. Schematic profile from basin margin to shelf showing depositional environments and characteristic sediments for Proto-Atlantic (Iapetus) Ocean during the Early Paleozoic (after Keith and Friedman, 1977; Friedman, 1979; Friedman, Sanders, and Kopaska-Merkel, 1992).

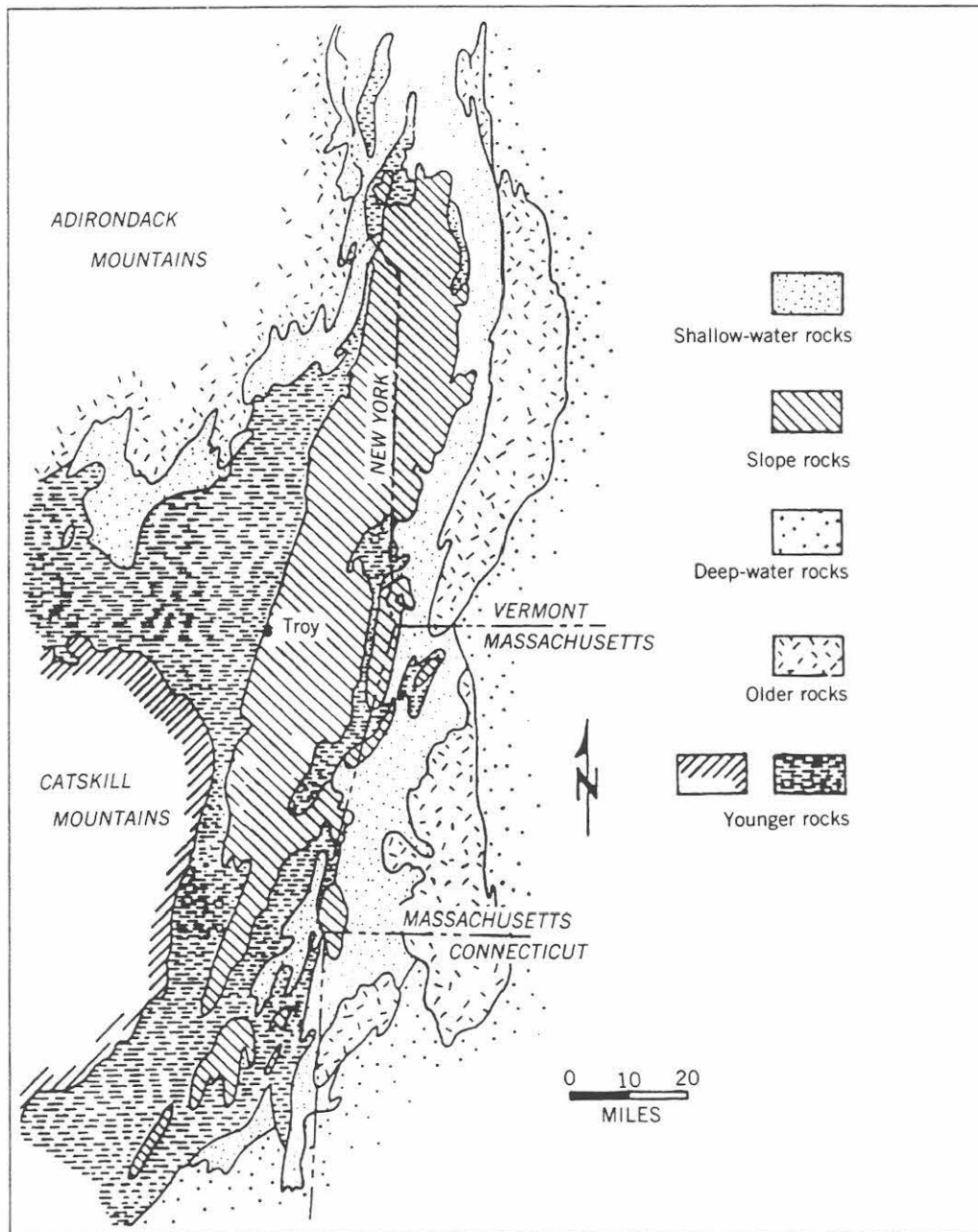


Figure 5. Generalized geologic map of Taconic area showing general rock types of Sauk and Taconic Sequence (Fig. 1). The fine stipple shows shallow-water rocks from the submerged continent, coarse stipple shows deep-water rocks from the ocean basin, and diagonal lines show rocks from slope and rise. The older rocks shown on map are Precambrian and the younger rocks are part of the Silurian-Devonian Catskill Complex (after Keith and Friedman, 1977; Friedman, 1988b).

convergence continued, stacked thrust-sheets overrode the outer parts of the continental terrace and resulted in rapid flexural downwarping of the ancestral continental slope and rise (Price and Hatcher, 1983; Quinlan and Beaumont, 1984; Tankard, 1986). The downwarped continental margin bounded on the oceanic side by the rising Taconic orogenic belt created the elongate Appalachian foreland basin (Fig. 3).

As already explained, the strata of deep-water setting are part of the Taconic Sequence (Fig.1). These rocks have received the attention of geologists for more than 150 years, and because of their exceedingly complex structural and stratigraphic relations have been the object of considerable debate. In fact, approximately 150 years ago Ebenezer Emmons' (1799-1863) advocacy of the Taconic System (1842, 1844, 1848, 1855) and the division of thought on this problem resulted in the famous duel between James Hall (1811-1898) and Emmons which ultimately forced Emmons to leave New York State. A court decision involving several of the most well-known geologists of the last century assured Hall's victory by forcing Emmons out of New York; he settled in North Carolina, where he became state geologist, away from his Taconic rocks. A field view of

Emmons' classical book on his beloved Taconic Mountains is shown in Figure 6. Ironically, in death Hall and Emmons were reunited. They are buried almost next to each other in the Albany Rural Cemetery.

Strata of the Taconic Sequence extend from north to south approximately 150 miles (Fig. 5), and for the most part within New York State are composed of shales and sandstones. Carbonate rocks are minor by comparison, but are important as they reflect depositional conditions. Although the stratigraphy and tectonics of the area have been the subject of considerable controversy (a debate that has become known as the "Taconic Problem"), stratigraphic succession and structure have more recently been clarified (Zen, 1967; Bird and Rasetti, 1968; Rowley and Kidd, 1981; Friedman, Sanders, and Martini, 1982; Isachsen et al., 1991).

Environmental reconstruction for the Cambrian part of the Taconic Sequence in eastern New York State indicates a depositional environment analogous with a modern continental rise or more specifically with a slope-fan-basin-plain model (Fig. 4) (Keith and Friedman, 1977, 1978). Terrigenous and carbonate sediment was removed from the Cambrian shelf and deposited with muds of the slope, now slates and siltstones, by a variety of processes at work on the slope and within submarine canyons. In addition, carbonates accumulated on the slopes. These sediments can be divided into seven main lithofacies, each bearing the imprint of the principal process or processes involved in its deposition. These include: (1) microcrystalline or cryptocrystalline limestone (micrite, a lithified marine carbonate deposit); (2) carbonate-sandstone clast conglomerates (inferred products of debris flow), (3) blocks of bedrock (olistoliths); (4) massive, coarse sandstones (apparent deposits of liquefied cohesionless particle flow), (5) graded sandstones (presumed turbidites), (6) parallel-laminated sandstones (probable turbidites), and (7) current-ripple-laminated sandstones (thought to be the products of reworking by contour-following bottom currents or submarine overbank levee deposits). All of these processes were working together or in opposition. Analysis indicates that only the lower slope and base-of-slope portion of the early Paleozoic continental margin has been preserved in the Taconic Sequence (Keith and Friedman, 1977, 1978). We shall discuss these lithofacies.

Microcrystalline or Cryptocrystalline Limestone (Micrite)

This lithofacies is composed of beds of dense, texturally simple, peloidal, intraclastic microcrystalline or cryptocrystalline limestone resembling micrite that in thin section locally shows neomorphism, where the original cement has become recrystallized. Beds of this lithofacies are found at several of the sections to be seen on this field trip and comprise a significant amount (approximately 20%) of all the lithofacies south of Schodack Landing. Single and multiple beds are found interbedded with beds of shale (Fig. 7). These beds show some pull-apart or boudinage structure and, locally, slump folds.

This peloidal and intraclastic microcrystalline or cryptocrystalline limestone resembling micrite is widely considered to be the most common kind of carbonate deposit in both modern and ancient submarine environments and originates as magnesian calcite (Friedman, 1964; Ginsburg et al., 1971; Alexandersson, 1972; Schroeder, 1973; Friedman et al., 1974; MacInyre, 1977, 1985; Longman, 1980; Stenoft, 1994; Friedman, 1985, 1995; Reitner et al., 1995). Peloids are ubiquitous throughout. This fabric was precipitated as a submarine cement and not deposited mechanically as a lime mud (Friedman,



Figure 6. View of Taconic Mountains in Ebenezer Emmons' classical book on the Taconics (1843; p.75). It is labeled "part of the Taconic Range from Stone Hill". (By Nancy A. Payzant, adapted from lithograph by Ebenezer Emmons in Friedman, 1988b).



Figure 7. Field photograph of cycles of alternating microcrystalline or cryptocrystalline limestone (micrite) and calcareous shale of the Hatch Hill Formation of the northern Taconics of eastern New York, south of Hudson, N.Y. (Stop 9).

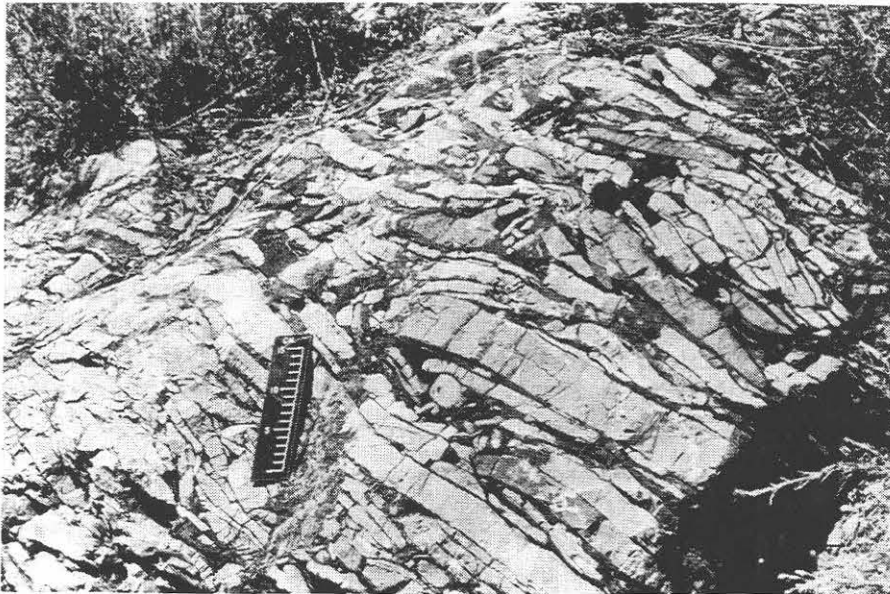


Figure 8. Carbonate-clast conglomerate of shallow-water origin displaced by debris flow from shelf margin into deep water. Hatch Hill Formation (Cambrian) south of Hudson, N.Y. (Stop 9).

1985). In places a bed will contain fossil fragments. Modern analogues for such carbonate deposits are lithified carbonate sediments of Atlantis Seamount, Mid-Atlantic Ridge (Friedman, 1964) and steep slope limestones of the Tongue of the Ocean, Bahamas (Grammar et al., 1993). These deposits are analogous to those of steeply dipping slope deposits documented from the fossil record. Carbonate slopes from the Permian of west Texas and New Mexico (Yurewicz, 1977; Ward et al., 1986; Garber et al., 1989), the Lower Ordovician of Utah (Wilson et al., 1992), the Devonian of Western Australia (Playford, 1980; Playford et al., 1989), the Triassic Dolomites of northern Italy (Bosellini, 1984; Harris, 1988), the Cretaceous of east-central Mexico (Enos, 1986), and the Miocene of the Gulf of Suez (Haddad et al., 1984) all exhibit primary depositional slopes of 30-40°. In addition to slope declivity, the geometry and thickness of beds as well as the dominant texture of the slope deposits in the Tongue of the Ocean, Bahamas, are also similar to these ancient examples (Grammar et al., 1993). Steep-slope profiles similar to those observed in outcrop are also frequently observed in seismic profiles from ancient carbonate platforms in the subsurface (e. g. Sarg, 1988).

By analogy with both modern and ancient submarine carbonate facies cited above, these limestones are interpreted as deposits that cemented in place on a steep slope. The correlative sediments on the shelf were sands and muds, not carbonate

sediments, because where beds of these cemented carbonate slope deposits broke up clay or sand, now sandstone and shale, are wedged between the breaking-up slabs.

Carbonate-Sandstone Clast Conglomerate (Figs. 8-11)

Monomictic carbonate to polymictic carbonate-sandstone conglomerates occur throughout the Taconic Sequence; a significant percentage of sand particles or clay may be present in some beds (Figs. 8, 9). The carbonate clasts are the most common; they have a general preferred orientation parallel to the bed boundaries, where they are exposed, but some clasts in a bed are oriented up to 90° to the general trend.

The shape of the clasts ranges from irregular to tabular and from angular to rounded. Some elongated clasts are bent (Fig. 10). The fragments show considerable variation in size, up to 60 cm thick and 240 cm long. *Deposits composed of rubble of carbonate rocks, usually angular, interstratified with dark-colored shale* are known as **brecciolas**. The brecciolas have been interpreted as products of turbidity currents, gravity slides, and debris flows. Such brecciolas formed along hundreds of kilometers of the original eastern edge of the carbonate shelf. They mark the former margin of the basin during the Cambrian and Lower Ordovician periods. Slides, slumps, turbidity currents, mudflows, and sand falls moved down the steep unstable slope beyond the shelf edge. Brecciolas and other gravity-displaced deposits can originate at the shelf margin or in deeper water on the slope.

Large clasts occur as slabs and consist of laminated to bedded finely textured limestone. The alignment of long axes of clasts with bedding planes (Fig. 8) suggests laminar flow conditions typical of plastic debris flows. In places coarse heavy blocks or slabs accumulated near the base of a debris flow and smaller lighter fragments at its top (Fig. 11). Such a slump deposit, showing graded bedding which may be 2 to 3 m thick, may have accumulated in a few hours. Figure 11 shows such a brecciola deposit which is about 1.2 to 1.5 m thick that accumulated probably in a matter of hours, whereas the overlying sediment of comparable thickness took thousands of years or even longer to accumulate. The contrast in timing between the two kinds of flow is extraordinary.

Some of the conglomerates are parts of turbidite beds 3 m thick in which the matrix at the base is coarse sand and the top of the bed is fine-textured shaly matrix.

In contrast to the relative calm of the deep water to the east and the high-energy shallow water to the west, the steep slopes of the continental margin were subjected to periodic violent activity. Sediments building on the edge of the continent or on the slope sometimes broke loose; the resulting slumps sent debris cascading down the steep slopes to form fans of coarse rock fragments. The debris moved with tremendous force; we know from modern ocean studies that catastrophic subsea avalanches can pack a tremendous wallop, flowing with extraordinary speed down the slopes and through submarine canyons.



Figure 9. Rubble of incoherent slump or debris flow composed of boulders of limestone, sandstone, and chert. This rubble, known as brecciola, originated in shallow water behind shelf edge and was displaced into deep-water, dark-colored shales. Note calcite-healed fractures in view. Boulder in center is approx. 30 cm across. Hatch Hill Formation (Cambrian), Campus of Rensselaer Polytechnic Institute (Stop 1).

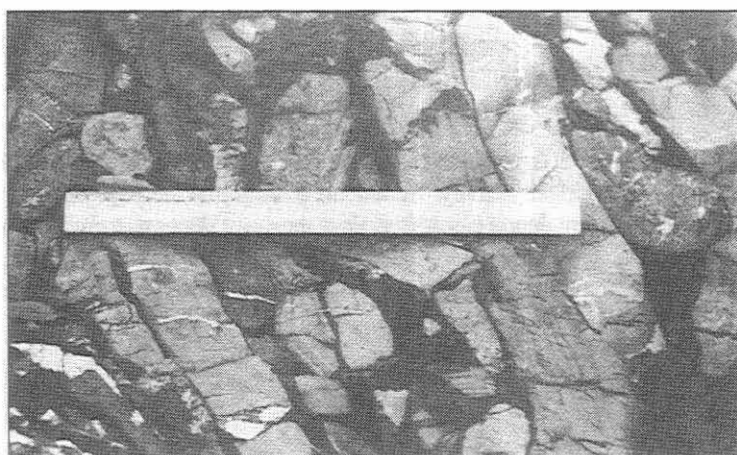


Figure 10 a,b. Slump rubble of skeletal and pelletal limestone (micrite) resembling breccia displaced into deep-water, dark-colored shale. This limestone rubble is known as brecciola. Note that tabular slabs have been bent and twisted. Slab in upper middle view of photograph Fig. 10a is 30 cm thick. Bent slab in lower middle of same photograph is 1 m long and 5 cm thick. Hatch Hill Formation (Cambrian), Troy, N.Y. (Stop 2).

The recognition of a debris-flow model for many of these conglomerates having a lack of organized internal structure has become well established (Fraser, 1989; Friedman et al., 1992; Savage, 1984). A **debris flow** is defined as a *flowing muddy mixture of water and fine particles that supports and transports abundant coarser particles* (Friedman, et al., 1992, p. 236).

Such a model does fit well with the large clasts in a clay matrix. The range of composition of the clasts can be easily accounted for, as being derived from the shelf buildup, or the basin-margin or slope beds. Some conglomerates appear to be quite local in origin, and interbedded with beds similar to the source beds for the clasts, which also seems compatible with a debris-flow model. The upward decrease in clast size in some beds (Fig. 11), with the pervasive preferred orientation and local imbrication, indicate movement and settling of the individual clasts within the flow. The matrix of some beds is typically graded: the base of the flow in which large clasts and slabs were transported consists of coarse quartz sand, whereas that of the top of the flow is clay. Hence these beds are turbidite rather than debris-flow deposits.

It is not clear whether the conglomerates of the Taconic Sequence were deposited as sheets or were confined to channels. Some of the conglomerates are associated with turbidites, which are generally considered to be confined to submarine canyons or to channels on a submarine fan. Thus, these conglomerates might have been similarly confined.

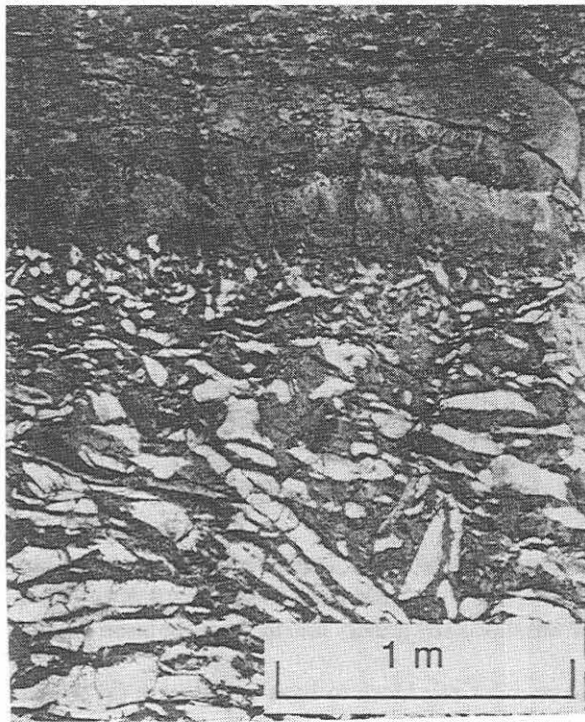


Figure 11. Graded layer of brecciola. The overlying strata took far longer to form than did the slump which may have accumulated in a few hours. Hatch Hill Formation (Cambrian) about 16 km south of Troy, New York, (Stop 7).

In places hummocky strata of brecciolas overlie regularly bedded microcrystalline or cryptocrystalline limestones (micrite)(Friedman and Sanders, 1995).

To break up beds of cemented slope deposits into slabs, some of which are a least 60 cm wide and up to several meters long, but most of which are only 20 to 40 cm long, takes a lot of force. These brecciolas are event deposits that record episodic erosion and deposition of coarse clasts by intense flows that must have involved substantial removal of cemented slope deposits. One mechanism may have been earthquake shocks which triggered tsunamis. The hummocky strata of brecciolas suggest a possible tsunami origin. Storm deposits, likely to be tsunamis, are widespread in Sauk deposits across the entire margin of the Sauk carbonate bank including China (Chuanmao et al., 1993) and Australia (Mount and Kidder, 1993), the Mid-Continent(Carozzi, 1989; Marsaglia and Klein, 1983), western Canada sedimentary basin (Brian Tuffs, personal communication), as well as more locally in Vermont, New York, and elsewhere in the Appalachians (Sepkoski, 1982; Whisonant, 1987; Pollock, 1989; Friedman, 1994), and along the western margin of the Sauk North American platform in Nevada (Cook and Taylor, 1977). Such a worldwide episodic event at the end of the Cambrian may even mark a meteorite impact. To break down such a tough, cemented rock at a water depth of up to several kilometers may require an energy level of a meteorite impact. Under the petrographic microscope I looked for microscopic lamellar deformation features in quartz that identify shock-generated events, such as meteorite impact (Alexopoulos et al., 1988). Almost all quartz particles were clear and devoid of such features, however two particles displayed well-defined and continuous planar features suggestive of shock metamorphism. However, such features may have been inherited from a previous geologic cycle, hence whether meteorite impact or earthquakes were responsible for the break-up of the bedded deposits into blocks, slabs, and other rubble is at this stage equivocal.

Olistoliths

Olistoliths are defined as *large blocks of rock in a debris-flow deposit* (Fig.12). One large block of orthoquartzite approximately 30 feet by 15 feet, settled in a deep-water shale and was exposed until recently on the grounds of the Troy YMCA (Friedman et al., 1982), formerly listed as Troy Jewish Community Center (Friedman, 1972). This block is now concealed beneath soil and vegetation. The exposed shale surrounding this erratic block showed that this block occurred singly. This block occurred along strike of the brecciolas and many more exposures of the brecciolas are present north of it in Frear Park. In a pit about 100 feet or so north of this block we exhumed from the shale a block of dark gray, fractured, and veined micritic dolomitic limestone.

The size and shape of the block of orthoquartzite suggests more than a steep slope. To detach a block of this dimension required considerable instability, such as severe shakes as occur during earthquake or meteorite impact. This block

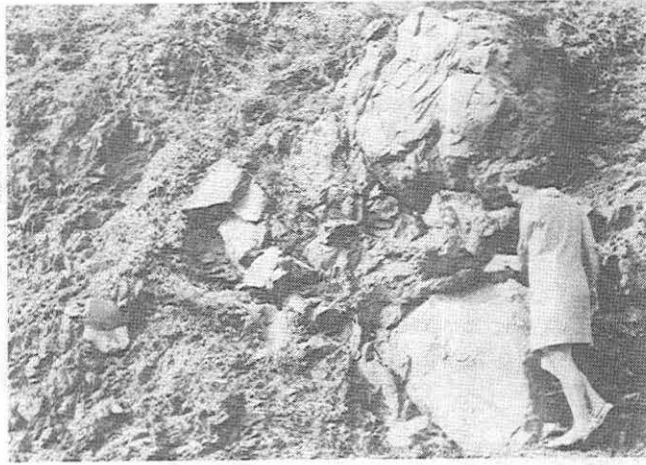


Figure 12. Two large blocks and various smaller blocks of limestone and sandstone set in a fine-grained dark-colored, deep-water shale. The blocks known as olistoliths are interpreted as having moved down a steep slope from near a shelf edge to a basin or basin margin. Note how well rounded the upper block is. Rensselaer Conglomerate, Hatch Hill Formation, (Middle Ordovician), Rysedorph Hill, Rensselaer, New York (Stop 4) (Friedman, Sanders, and Kopaska-Merkel, 1992, figure 17-1, p. 635)

of rock differs in lithology from the brecciolas. In contrast to the flat limestone boulders this huge block with its irregular outline suggests that it was forcibly detached from the shelf edge or basin slope. Could it be that this block was part of the wall of a submarine canyon which became detached and was moved by gravity into the basin or basin margin?

The alternative interpretation would be to consider this block to have been caught up in fault movement. Indeed slicken-sides are present on this block. However, the lower exposed contact with the shale is depositional and not faulted. Because the orthoquartzite block occurs along strike with brecciolas, and a limestone block has been found about 100 feet away, the evidence suggests that emplacement was by gravity rather than by faulting. This displaced orthoquartzite block may be as old as Precambrian. Large blocks of limestone likewise occur as olistoliths .

Massive Coarse Sandstone

These beds of massive sandstone show no bedding, lamination, or grading. The beds seem to fall into two groups, which are: (1) coarse-grained sandstone, and (2) thicker, coarse- to very coarse-grained sandstone. The beds are generally very coarse grained, with no internal features other than a few micrite pebbles. In places the beds contain either micrite pebbles, or wisps that stand out on the weathered surface (Keith and Friedman, 1977, 1978).

These sandstones are well sorted and porous and analogous to those forming deep-water reservoirs in California and elsewhere. In fact, they are as well sorted as beach sands and those not knowing the regional geology may mistake them for high-energy shallow water deposits. Yet these massive beds correspond to beds described extensively from turbidite sequences in the literature (Friedman et al., 1992).

A depositional mechanism that appears to fit these coarse-grained generally structureless sandstone beds is fluidized sediment flow. This mechanism works when a loosely packed sand is subjected to an initial shock, destroying its fabric, so that water is incorporated and the sand liquefies, that is, the grains are supported by excess pore pressure. Since the sand is not sealed, pore-fluid loss is rapid, and the flow short-lived. As the pore fluid escapes, the viscous properties of the mass disappear and the sediment comes to rest. Because the concentration of sediment relative to fluid is high, features associated with traction deposits, such as different types of lamination, cannot form (Keith and Friedman, 1977, 1978).

Generally, the beds of this lithofacies appear to fit a nebulous category of thick, coarse-grained massive sandstones "proximal" in nature (or possibly channel deposits). They were deposited by one or more processes, involving fluidization of the sediment (Keith and Friedman, 1977, 1978).

Graded Sandstones

The graded beds are found associated with beds of other lithofacies. These beds are prominent except south of Hudson, where they are only a minor constituent of the exposed section. Shales are interbedded with this lithofacies at all exposures, except for Judson Point, where sandstone beds are commonly in depositional contact with each other, or with only a very thin shale parting between them (Keith and Friedman, 1977, 1978).

The graded beds range in composition from pure sandstone to limestone, with little or no sand. There are some beds that are half sand and half carbonate. Generally, within one exposure the lithology will be fairly constant. At Judson Point, the beds of this lithofacies are essentially pure sandstone. South of Hudson the beds all contain nearly equal amounts of carbonate and sand. Carbonate is present as rounded intraclasts, individual grains, and as a matrix in the sandy beds. The rounded intraclasts are commonly found near the base of the bed. The intraclasts are composed of pelmicrite, pelsparite or micrite. One intraclast of oomicrite was seen. Sparite and pelmicrite occur as matrix for sandy carbonates (Keith and Friedman, 1977, 1978).

Beds of this lithofacies display many kinds of sedimentary structures. Graded beds, parallel lamination, and cross-lamination (commonly ripple lamination) are all common. Grading takes on several forms in the beds studied. Many beds at Judson Point show delayed grading (Dzulynski and Walton, 1965), where most of the bed is coarse- or medium-grained sand, uniformly distributed, up to the very top, where the bed quickly becomes argillaceous with essentially no intermediate grain sizes. The grading then takes place in a narrow zone at the top, rather than throughout the bed. Beds at the locality south of Hudson commonly show coarse bimodal sand at the base in a carbonate matrix, with the sand decreasing in amount upward, leaving only the carbonate at the top. This would be a type of discontinuous grading with no medium-grained portion (Keith and Friedman, 1977, 1978).

Parallel lamination is quite common. It appears to be especially well developed in the medium-grained sandstones. The laminae are generally less than 1 mm in scale. The coarse-grained sandstones, as seen at Judson Point, show only faint lamination, if any at all. Ripple lamination is quite well developed in some beds, but is not common. Not seen elsewhere was larger scale cross-lamination that could be considered cross-bedding in a bed south of Schodack Landing. Many examples of the various internal structures, alone or in combination with others, can be seen (Keith and Friedman, 1977, 1978).

Beds of sand-sized material, displaying grading and lamination in a systematic order (Bouma Sequence) and which are interbedded with basinal shales are turbidites. In general, these graded beds bear more resemblance to distal, rather than proximal, turbidites, but may be transitional (Keith and Friedman, 1977, 1978; Friedman et al., 1992)

Parallel-Laminated Sandstones

Beds identified as belonging to this lithofacies comprise a significant amount of the lithofacies at all of the major sections to be seen on this field trip. The beds of this lithofacies range from medium-grained, parallel-laminated sandstones, to medium-grained sandstones with parallel lamination and some cross-lamination, to coarse-grained sandstones. Most of the sandstone beds are composed of medium-grained quartz sand with a variable amount of carbonate matrix forming the laminae. Some of the sandstone beds will contain fossil fragments, and, in fact, nearly all the identifiable trilobite fauna recovered by Bird and Rasetti (1968) from Judson Point, and Nutten Hook, and used by them for dating, came from beds identified in the Keith and Friedman (1977) study as belonging to this lithofacies. All but one of the sandstone beds of this lithofacies show lamination of some sort. Commonly, only parallel lamination is present in the sandstones, but some sandstone beds show some cross-lamination.

The beds here probably represent channel-edge equivalents of the coarser, probable channel deposits represented by the conglomerates, massive sandstones and turbidites.

Beds of this lithofacies are intimately associated with turbidites and may even be types of turbidites themselves (Keith and Friedman, 1977, 1978; Friedman, 1979; Friedman et al., 1982).

FOR COMPARISON: GEOLOGIC SETTING OF DEPOSITS OF MIDDLE TO LATE ORDOVICIAN NORMANSKILL FORMATION.

On this field trip we shall compare flysch deposits of the Taconic Sequence with those of the post-Taconic Sequence, specifically with the Austin Glen Member of the Upper Normanskill Formation, which is Middle to Late Ordovician in age. Deposition of the Austin Glen Member followed the collision between the North American platform and the Taconic island arc (Fig. 3). In contrast to the rocks of the Taconic Sequence these Normanskill deep-water rocks were derived from a direction that is now west, but during the Early Paleozoic was north.

We shall see the deep-water facies of the Austin Glen Member at two stops and then look for Austin Glen lithology at a third stop.

OUTCROP GUIDE AND ITINERARY: FLYSCH OF THE TACONICS

Figure 13 is the road log.

Depart from Union College and drive to Troy via U.S. 7. From U.S. 7 on arrival in Troy head west to River Street, head south on River Street, past Castaway Restaurant, Super 8 Motel, and City Garage, cross Fulton Street, and between Fulton Street and Broadway on Third Street park across from the Rensselaer Center of Applied Geology on 15 Third Street. We shall stop at the Rensselaer Center of Applied Geology, headquarters of the Northeastern Science Foundation affiliated with Brooklyn College of the City University of New York, for a brief review of the geology.

Mileage Between Points	Cumulative
0.9	0.9

From Rensselaer Center of Applied Geology on Third Street proceed to corner of Broadway, one block (right) on Broadway to Second Street (Monument Square), turn right on River Street which becomes Fulton Street. Head east (uphill) on Fulton Street to Sixth Street. Make a left turn onto Sixth Street for one block to traffic light, turn right onto Federal Street (no street sign at corner) which becomes Sage Ave., hang right and drive to RPI '87 gymnasium.

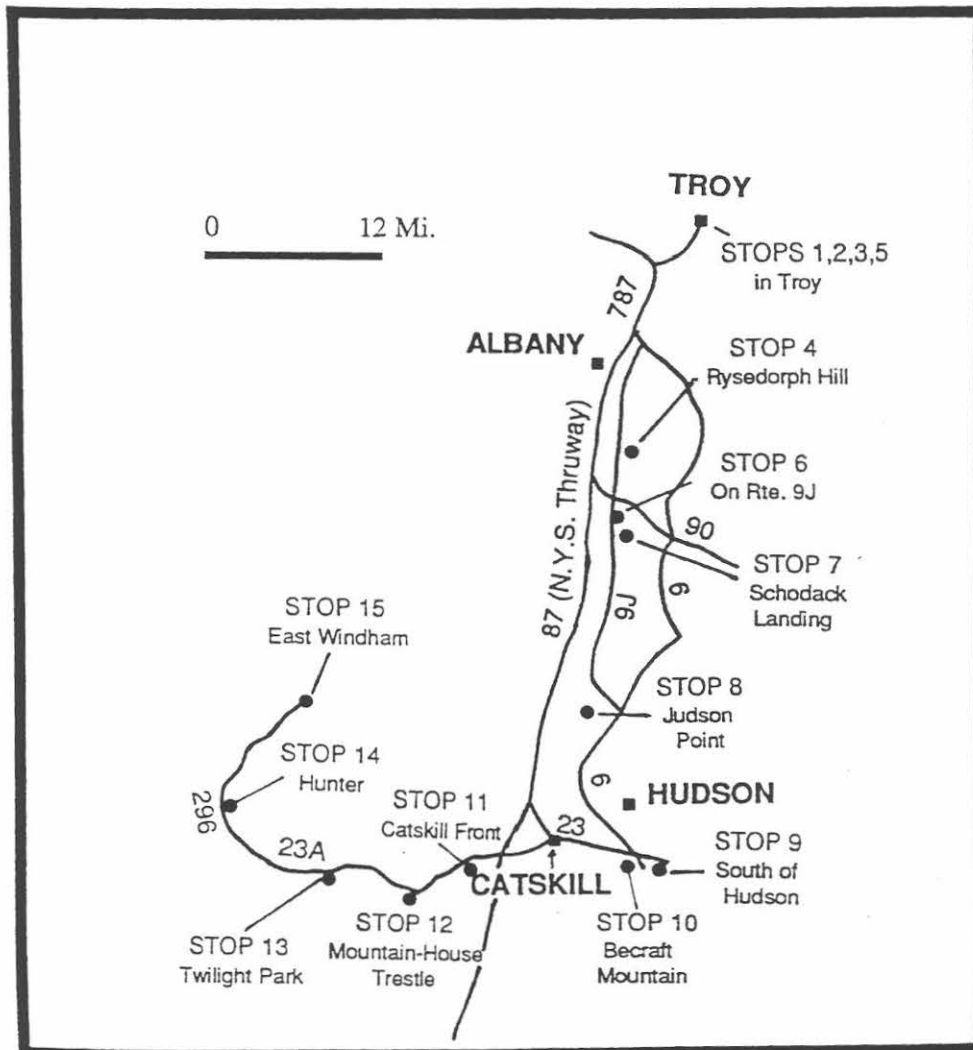


Figure 13. Road log with stops.

STOP 1 . RENSSELAER POLYTECHNIC INSTITUTE, '87 Gym

Exposure behind fence adjacent to gym.

Ruedemann (1930, p. 114; also Fig. 64) described and photographed this exposure as a good example of a “cliff of mylonite,” one of the “excellent exposures of a fault breccia” on the campus of Rensselaer Polytechnic Institute. According to Ruedemann and reconfirmed by Elam (1960) a thrust fault follows part of this street (Sage Ave.) and Ruedemann mistook this conglomerate for a fault breccia. Perhaps the presence of criss-crossing veins in this exposure led to his interpretation of a “cliff of mylonite.” Jack G. Elam (1960; unpublished Ph.D. thesis at Rensselaer Polytechnic Institute) assigned the rocks at this exposure to the Schodack lithofacies of Early Cambrian age. Cushing and Ruedemann (1914) had introduced the “Schodack Formation”. Zen (1964) has renamed this formation the West Castleton Formation. As already explained, this formation is now assigned to the Hatch Hill Formation (Fig. 1).

Lowman (1961) recognized that the boulders are a conglomerate and not a breccia, and following Kuenen and Migliorini (1950), he introduced the term brecciolas for these rocks.

The limestone, sandstone, and chert boulders which are embedded in shales at this exposure range from angular to rounded and show considerable variation in size (Fig. 9). Some boulders are coarse-grained fossiliferous limestone fragments with a micritic dolomite matrix. The rocks above the brecciolas are greenish-gray shale.

- 0.4 1.3 Continue on Sage Ave. due east past 15th Street, hang right and continue to Burdett Ave. and park in parking lot of Doyle Middle School across the street.

STOP 2. TROY HIGH SCHOOL QUARRY

Walk to the running track and on track proceed left (north) to exposure below RPI housing units.

The spectacular brecciolas at this exposure consist of three members with eleven sub-members (Lowman, 1961); some of the deposits that Lowman studied have since been destroyed. For details of the rocks, refer to Lowman’s descriptions (1961). The brecciolas are lithofacies of the Hatch Hill Formation, as at Stop 1.

A thin-section study shows the limestone clasts to consist of biomicrites, biointramicrites, and micrites with varying terrigenous quartz and clay-minerals. The intraclasts are of pelmicrite. Shell fragments have been selectively dolomitized. The carbonate sediments must have lithified before their displacement downslope.

The brecciola bed is steeply dipping (80°) and is approximately 3 m thick. It is interbedded within a fissile shale. The basal contact of the bed is at the top of the exposure, and the top of the bed is at the base of the exposure. The matrix between the clasts ranges from sand at the base of the bed (top of the exposure) to clay at top of the bed (base of the exposure). This change in the texture of the matrix is typical graded bedding.

Although some clasts are rounded, most of them are angular and slab-shaped (Fig. 10). Some slabs experienced multiple bending reflecting high-energy impact (Fig. 10). Sporadic slabs ½ m to 1 m long appear to have been in the process of breaking up, with clay wedged into the fractures of the separating clasts. The clasts are sub-parallel to the boundaries of the bed.

Mileage

Between Cumulative
Points

- | | | |
|-----|-----|---|
| 1.5 | 2.8 | From parking lot turn left onto Burdett Ave., continue to Tibbits Ave. (traffic light), turn right onto Tibbits Ave. for one block to Brunswick Ave., turn left onto Brunswick Ave. and drive downhill to Congress Street (traffic light). Turn left onto Congress Street and hang right at fork, follow NY 66 across Poestenkill Bridge to Linden Ave. Turn right on Linden Ave. and drive downhill (past Poestenkill Falls Park) to Spring Ave. Turn right onto Spring Ave. for one block to Canal Street. Turn left on Canal Street to South Troy Recreation Center at corner of 5 th Ave. Turn left on 5 th Ave. for one block and park at corner of 5 th Ave. and Madison Street. Walk old wagon road uphill to ruins of former buildings (now only the floor of the buildings is preserved), site of former abandoned quarry. Look at exposure in old quarry (Rushor’s Quarry) |
|-----|-----|---|

STOP 3. RUSHOR'S QUARRY, TROY

The rocks at this exposure are part of the Austin Glen Member of the Normanskill Formation of Middle Ordovician age. In contrast to the Cambrian (Hatch Hill Formation) deep-water rocks which were tectonically emplaced in the Troy area, the Normanskill deep-water suite formed *in situ* after deep submergence of the Cambrian-Early Ordovician carbonate shelf. The sediment composing the rocks at this site were derived from what is now east (Fig. 3), but was north during the Paleozoic. At this stop we see basin-margin sediments devoid of brecciolas. The clue to the presence of a paleoslope are sole marks on the undersides of sandstone beds (Fig. 14). These marks are infillings (molds) of depressions that formed in the soft bottom clays

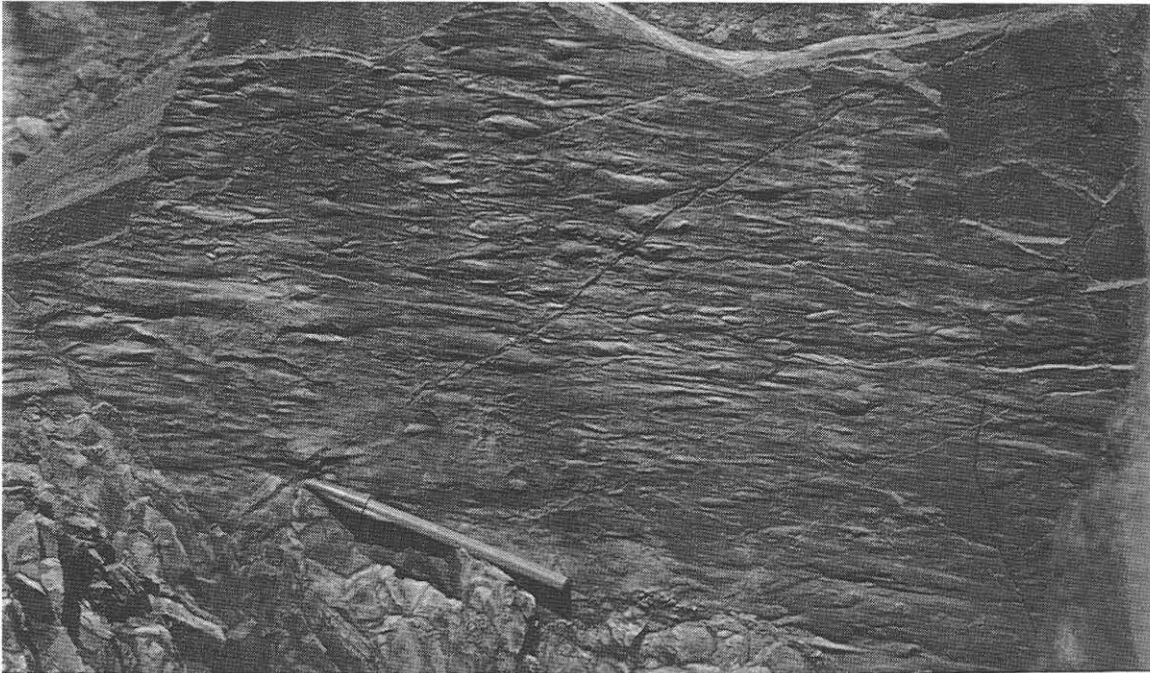


Figure 14. Sole marks made on cohesive mud substrate by erosive/reworking current and preserved as counterparts on the base of the overlying sandstone. These scour marks are known as flutes. Austin Glen Member of the Normanskill Formation. Rushor's Quarry, Troy, N.Y. (Stop 3).

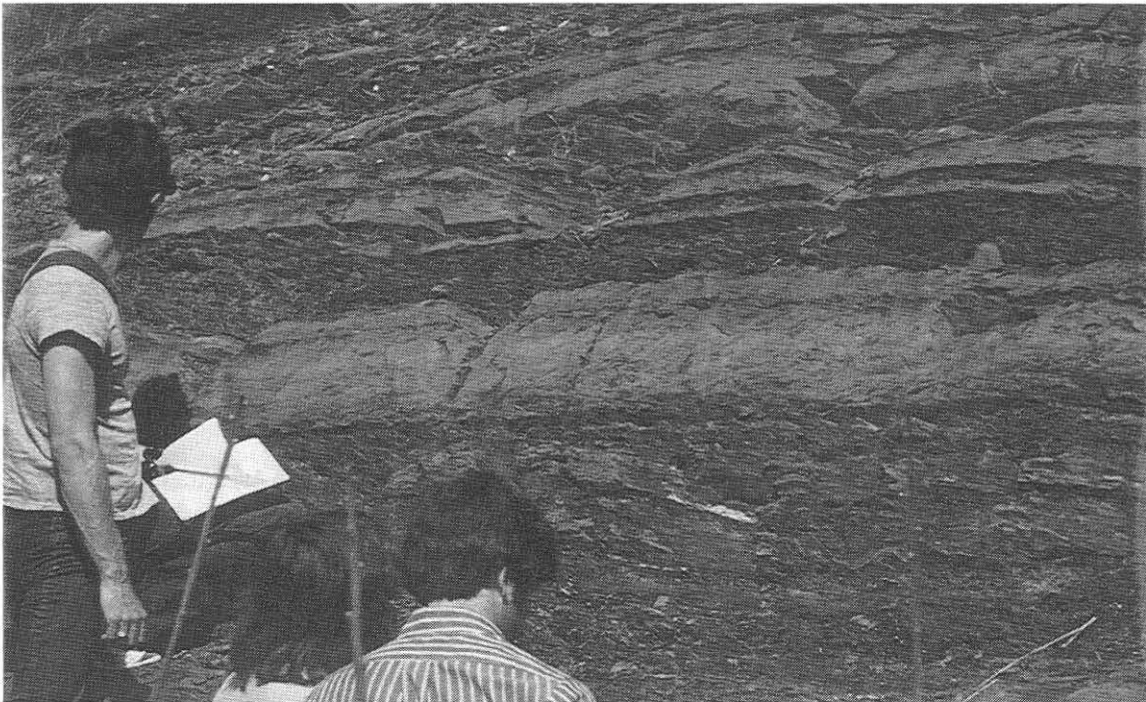


Figure 15. Alternating sandstones and shale of typical flysch facies, Austin Glen Member of the Normanskill Formation. Rushor's Quarry, Troy, N.Y. (Stop 3).

(now shales) as particles in turbidity current flows or sandfalls scoured or gouged the bottom. They are also known as impact marks or tool marks. The following sole marks can be seen (definitions modified from Pettijohn and Potter, 1964): flute molds - a raised sub-conical structure, the upcurrent end of which is rounded or bulbous, the other end flaring out and merging with the bedding plane; groove molds - rounded or sharp-crested rectilinear ridges produced by filling of grooves; brush marks - essentially a bounce cast with a crescentic depression on the down-current end; prod molds - a short ridge, parallel to the current, which unlike flute molds, rises down-current, and ends abruptly; frondescant marks - a type of load-flow structure that covers some soles with crowded lobate molds overlapping in the down-current direction.

The rocks in this quarry consist of interbedded sandstone and shale (Fig.15) which show the characteristics of distal turbidites (Walker, 1967): cross laminae, convoluted laminae, sole marks, graded beds, parallel sides and regular beds, thin beds, fine grain size; individual sandstone beds rarely amalgamate.

The exposure in this quarry is that of a typical flysch composed of alternating sandstones and shales (Fig. 15). Most sandstone beds are composed of two units: an underlying finely laminated sandstone and an overlying ripple cross-laminated sandstone (Figs. 16, 17). The ripple cross-laminated sandstone may scour into the underlying finely laminated sandstone. These two sandstone units correspond to the B and C units of a Bouma sequence; the shale would be a Bouma E unit. Please look carefully at the two sandstone units, keep a mental image of their appearance, and carry this image to the next stop where you will need it for interpreting the geological history.

10.1

12.9

Turn right on Madison Ave. to Third Street. Turn left onto Third Street which runs into Fourth Street and continue (south) on to Burden Ave. , keep hanging right on Burden Ave. to bridge over Hudson River. Bear left on bridge to Interstate 787 south (towards Albany). Exit at Rensselaer sign and cross Hudson River Bridge and take East Greenbush ramp to City of Rensselaer and make a left turn onto Washington Street. Drive down Washington Street to Third Street and make right turn (US 43 east) to US 151. Turn right on US 151; outcrops are located just north of corner with Eastern Ave.

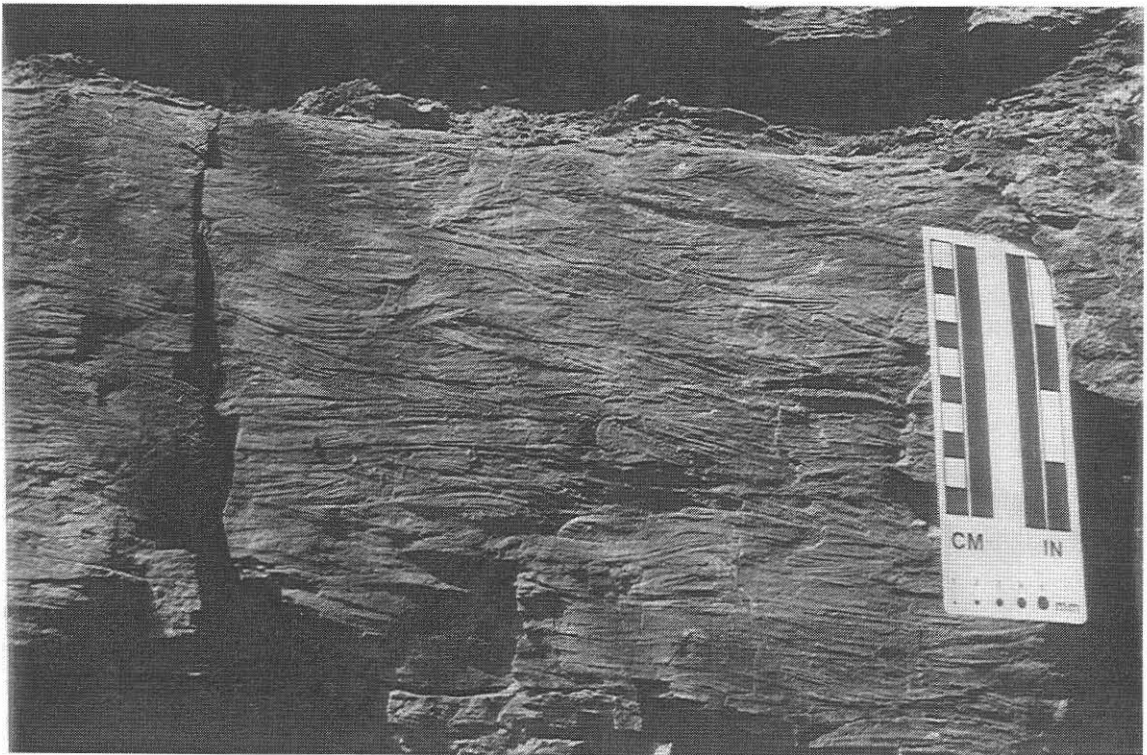


Figure 16. Ripple-crosslaminated sandstone unit (C unit of Bouma) which overlies finely laminated sandstone unit (B unit of Bouma). Austin Glen Member of the Normanskill Formation, Rushor's Quarry, Troy, N.Y. (Stop 3).



Figure 17. Flysch sandstone composed of two units: an underlying finely laminated sandstone and an overlying ripple crosslaminated sandstone. Geologist points at contact between these two units. Austin Glen Member of the Normanskill Formation, Rushor's Quarry, Troy, N.Y. (Stop 3).



Figure 19. Block of Ordovician Austin Glen / Normanskill Sandstone in shale matrix. Top of block is on left. Note fine laminae of Bouma B unit in sandstone and on their left ripple-cross laminae of C unit. To the left of the block is slickensided calcite-healed fracture. Rensselaer Conglomerate, Hatch Hill Formation, Rysedorph Hill, N.Y. (Stop 4).

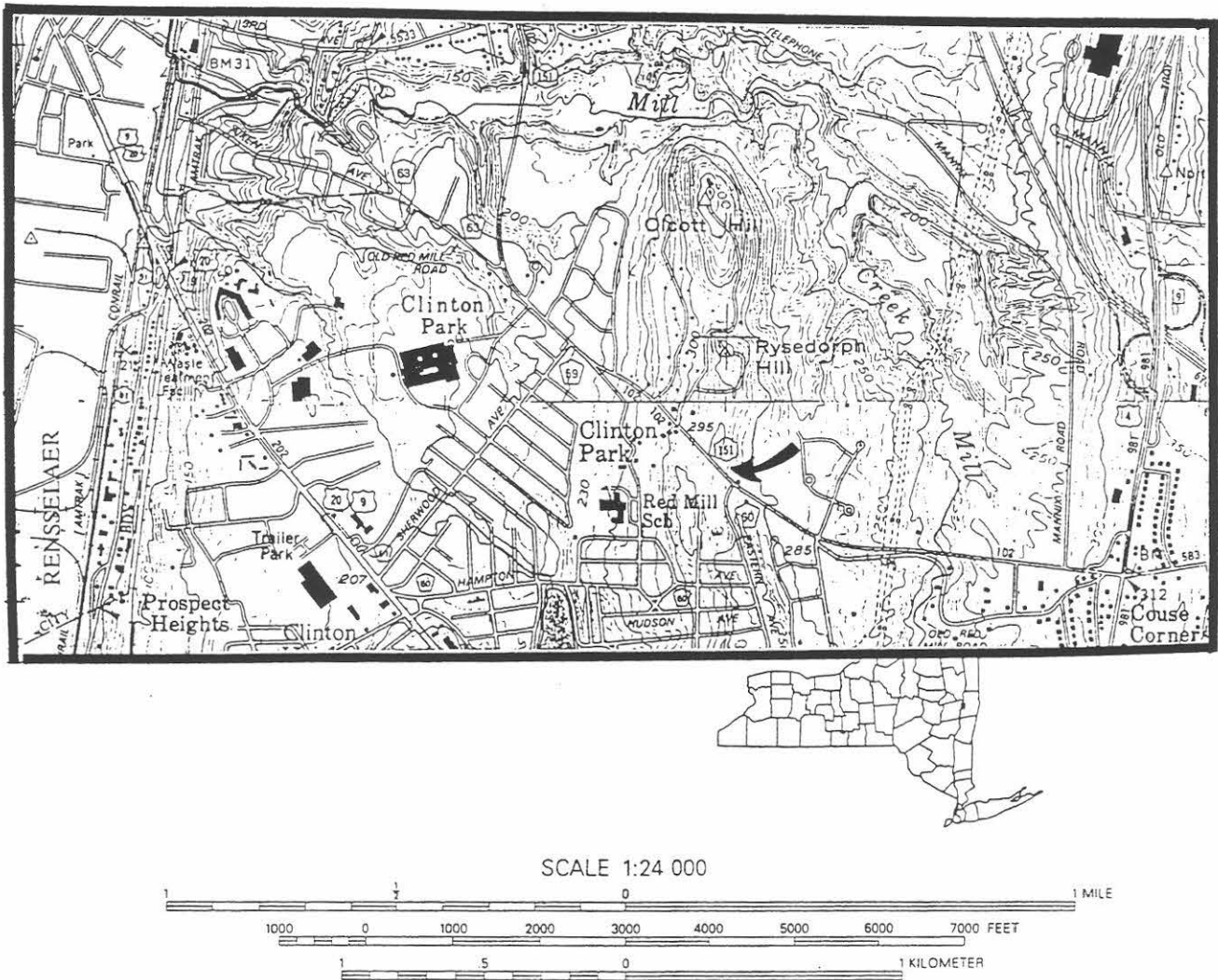


Figure 18. Location of map of Rysedorph Hill.

STOP 4. RYSEDORPH HILL

This roadcut is on the southern flanks of an unnamed peak of Rysedorph Hill which is topographically part of a ridge composed of two hills (Rysedorph Hill and Olcott Hill)(Fig. 18) . At and near the peak of Rysedorph Hill Ruedemann (1930) discovered spectacular conglomerates and faunas which are now obscured by vegetation; only small sporadic blocks or pebbles of limestone are now found dispersed in shale. This spectacular deposit is known as Rysedorph Hill Conglomerate and, as Ruedemann described, consisted for the most part of limestone conglomerate (see Sanders, 1995).

The rocks at this roadcut are quite different from those which Ruedemann described. Large and small blocks of sandstone and carbonate rock are set in a dark-colored deep-water shale. One of the blocks on the east side of the road is of particular interest. This block consists of Austin Glen/Normanskill sandstone lithology. In fact, it reveals the typical two kinds of sandstone units that we have seen at STOP 3 (Rushor's Quarry in Troy): an underlying finely laminated sandstone and overlying ripple cross-laminated sandstone which correspond to the B and C units of a Bouma sequence (Fig.19). One wonders how this one block traveled from Rushor's Quarry in Troy to this site!

The kind of lithology exposed at this stop has been designated **wildflysch**, a term applied to a *spectacular deposit consisting of small- to enormous blocks of sedimentary-, igneous, and metamorphic rocks set in a matrix of fine-grained, typically dark-colored marine shale, siltstone, or mudstone* (Fig.12). In modern terms, we would designate the wildflysch as one kind of diamictite (Friedman et al., 1992).

Study of sandstone blocks show them to be fine- to medium-grained and for the most part composed of quartz. Authigenic quartz overgrowth is in places so intense that the original texture is obscured. Plagioclase feldspar and microcline are common. They are commonly sericitized and calcitized. Fragments of sedimentary, igneous, and metamorphic rock are abundant, especially those of sedimentary rocks. Particles of carbonate rock include micrite, oolitic and pseudo-oolitic limestone, peloidal limestone, dolomitic limestone, and crinoidal limestone. These fragments of carbonate rock contain authigenic feldspars which are diagnostic of pre-Knox unconformity (Beekmantown or Sauk) deposits (Buyce and Friedman, 1975). Other kinds of rock fragments are composed of siltstones and shales, especially bituminous shales. Metamorphic rock fragments are those of sericite-chlorite schists and quartzites. Igneous rock fragments have porphyritic-ophitic texture. Finely-crystallized rock-fragment particles are those of recrystallized volcanic glasses or volcanic rocks, and contain chalcedony and chlorite.

Table 1 shows the petrographic composition of the sandstone blocks. They do not qualify to be called graywacke, despite such usage for these rocks in the literature (Potter, 1979). According to the classification of Friedman et al. (1992) the sandstones are classified as quartz-feldspar-rock fragment sandstones.

Carbonate blocks are for the most part composed of dolostone. The provenance of these blocks includes Cambro-Lower Ordovician platform carbonates (oolitic and peloidal carbonates and authigenic feldspar) and deep-water Austin Glen/Normanskill sandstone. Hence these conglomerates are younger than Normanskill which would make them Upper Ordovician Snake Hill Formation which is the only Ordovician unit younger than Normanskill present in New York (Fig. 1b)(Fisher, 1977). Because of age uncertainties I prefer to label these rocks Tippecanoe-correlative conglomerates and call them Rensselaer Conglomerate since they occur in the town and county of Rensselaer.

To recycle deep-water Austin Glen/Normanskill deposits an initial uplift is necessary. As Sanders (1995) points out "because eastern New York is part of the Appalachian orogenic belt... 'uplift' means thrust faults". This mechanism displaced deep-water Normanskill deposits into a setting of shallow water or emergence from which erosion detached the blocks and recycled them downslope, once again into deep water. This view is at variance with that of those who feel that these conglomerates resulted from tectonic emplacement (Potter, 1979). However, these deposits are incoherent subaqueous slumps (debris flow) of deep-water setting which were part of a slumped mass that moved by gravity downslope (Friedman et al., 1992, p.). Some of the boulders form a nearly perfect sphere (Fig.12) implying erosional rather than tectonic forces. The blocks are related to erosion on the front of the thrusts which contained rocks from each of the three sequences: Taconic, Sauk, and Tippecanoe (Normanskill) (see Fig. 1a) . The presence of Sauk material indicates that some thrusts broke loose inboard of the former shelf edge and transported the carbonate-sequence material over the foreland-basin shales of the Tippecanoe Sequence (see also Sanders, 1995; DeAngelis, 1995).

TABLE 1. QUANTITATIVE PETROGRAPHIC ANALYSES OF SANDSTONE BLOCKS FROM RYSEDORPH HILL

Minerals and Rocks	Sample Numbers											
	%	3	4	5	6	7	8	9	12	16	5a	6a
Quartz & Acces.Minerals	47.5	71.8	51.8	68.6	67.2	71.5	57.9	58.4	48.5	55.2	62.6	46.0
Feldspars	19.0	8.4	15.9	10.4	9.1	10.2	13.7	12.7	12.6	12.0	16.0	14.7
Sedimentary Rocks	4.5	3.1	3.5	5.2	1.0	3.7	2.5	4.7	2.2	3.6	5.3	5.3
Metamorphic Rocks	-	-	0.3	-	-	-	1.1	0.2	-	0.01	-	0.01
Igneous Rocks	0.01	0.01	0.01	-	-	-	0.01	-	-	0.01	-	0.01
Matrix & Cement	29.0	16.8	28.6	15.9	22.6	14.7	24.9	24.1	36.8	29.2	16.0	34.0
Total	100.01	100.11	100.11	100.1	99.9	100.1	100.11	100.1	100.1	100.02	99.9	100.02
Total No.of Points Counted	7.186	6.010	6.506	7.052	7.142	3.108	7.776	5.770	4.964	4.626	5.026	7.486

Analyst: Vincent Durovic

Mileage
Between Cumulative
Points

15.8 28.7 Turn around and return on US 151 to Washington Ave., now US 43 East, and turn right and drive to junction with US 4 North (but still US 43 East). Turn left at junction (traffic light) and continue to intersection, where US 43 East turns right. Turn right onto 43 and follow Sand Lake Road. Note Town of Sand Lake and junction with US 150. Turn left onto US 150 and follow to junction with US 66. Turn left onto US 66 (north) which becomes Pawling Ave. in Troy. Follow US 66 to Linden Ave. on left side of street. Turn left and drive downhill to parking lot on right.

STOP 5. POESTENKILL FALLS AND GORGE (SOUTH TROY QUADRANGLE, N.Y.)

We are back in Troy from where we initially set out; in fact we passed this site on our way between stops 2 and 3. You may ask why has a visit to this remarkable site been postponed till now, and why did we have to backtrack this circuitous route between Rysedorph Hill and Poestenkill Falls? The answer is that the experience of Rushor's Quarry (Stop 3) and Rysedorph Hill (Stop 4) is essential to an understanding of the geology of Poestenkill Falls and Gorge.

At this classical site Ruedemann (1930) shows "Logan's Line" (now known as Emmons' Line; Rodgers, 1970), the thrust plane which places Cambrian over Ordovician rocks surfacing at this site. Elam (1960) concurs and places Lower Cambrian rocks (his Poesten lithofacies, later West Castleton Formation; Zen, 1961) in contact with Middle Ordovician rocks (the Austin Glen Member of the Normanskill Formation). As already explained, the Cambrian strata , a typical flysch, have now been assigned to the Hatch Hill Formation (Fig. 1a). The Ordovician rocks are wildflysch, as at Rysedorph Hill, and hence I prefer to label them Tippecanoe-correlative strata and call them Rensselaer conglomerate (Hatch Hill Formation) as at Rysedorph Hill. As at Rysedorph Hill, the boulders and blocks consist of sandstones, shales, and limestones (Fig 20). Among the sandstones, blocks of Normanskill rocks can be identified. At this site Cambrian flysch has been thrust over Ordovician wildflysch. In places interbedded with the Ordovician wildflysch are strata of bedded flysch of Normanskill lithology which remind us of Rushor's quarry. Hence although consisting for the most part of debris-flow deposits the Ordovician strata are in part of turbidite origin.

This overthrust (Fig. 21) is interpreted as a segment that extends from Canada through Vermont, New York, and farther south. The fault line is well exposed in the south wall of Poestenkill Gorge. Cambrian shales occur above (east of) the fault, and Ordovician strata below (west of) the fault. The Ordovician strata have been described as a fault breccia or mylonite in which blocks of large size have been incorporated in shaly matrix. The matrix shows an anastomosing cleavage pattern that does not penetrate the boulders and blocks. Although described as a fault breccia or mylonite (Ruedemann, 1930;Elam, 1960)



Figure 20. Boulders and blocks of sandstone and other lithologies in shale matrix, a debris-flow deposit. Ordovician Rensselaer Conglomerate (Hatch Hill Formation), Poestenkill Falls and Gorge, Troy, N.Y. (Stop 5).

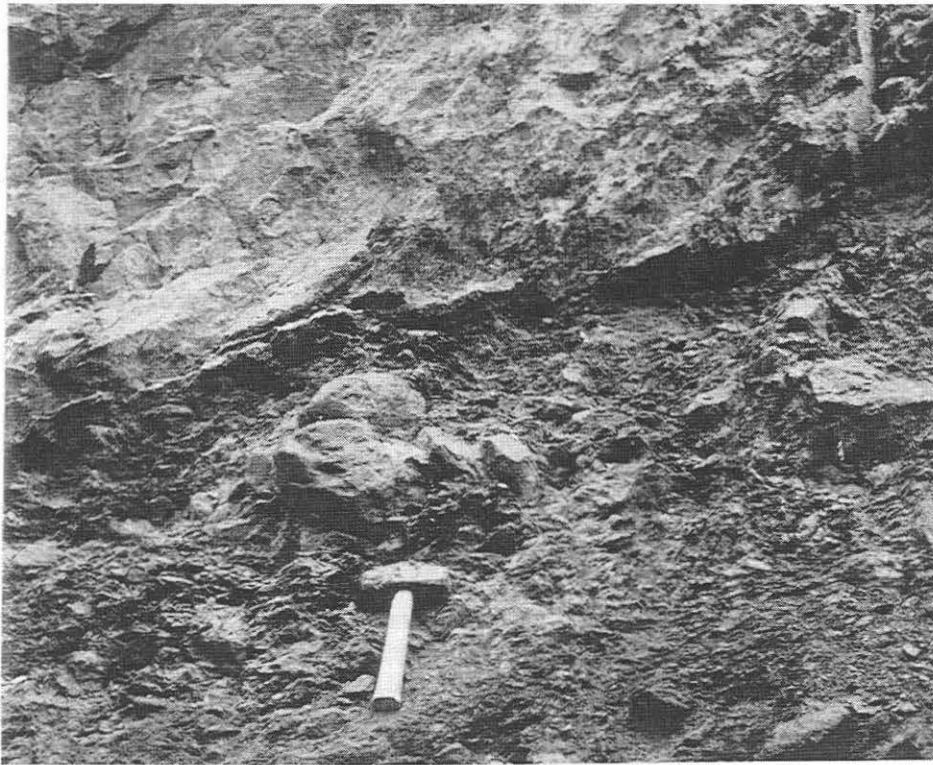


Figure 21. Thrust plane, known as part of “Logan’s Line” or Emmon’s Line, showing overlying Cambrian and underlying Ordovician deposits. Poestenkill Falls and Gorge, Troy, N.Y. (Stop 5).

this rock is a wildflysch. The blocks are interpreted as having moved down a steep slope from near a shelf edge to a basin or basin margin.

The Ordovician rocks are highly tectonized, especially towards the thrust. Antithetic post-thrust faulting is noted on the north side of the thrust sheet with an approximately 2 m displacement seen on the steep slope above the thrust. At and below the thrust sheet in the south wall of Poestenkill Gorge a dense black material may be *pseudotachylite*, a rock produced in the compression and shear related to intense fault movement, involving extreme mylonitization. My own professor Dr. S. James Shand introduced this term (Shand, 1916) for mylonites he encountered in the Orange Free State of South Africa. As Shand (1947) points out, this kind of “mylonites... proved most puzzling to field geologists meeting them for the first time”.

The Cambrian rocks contain abundant black particles which may be phosphatic.

During the Industrial Revolution numerous factories, clustered on the north slope of Poestenkill Gorge, were making cotton cloth, and curry combs, barbed wire, and buckwheat flour machines, and much more. The first mills were developed in the lower section of the gorge as early as 1791. Their full potential was realized when Benjamin Marshall constructed a brick cotton mill on the north side. Water power began its decline with the popularity of steam power at the turn of the century. Today obscure, moldering industrial ruins testify to this former busy activity. The last of the abandoned mill buildings tumbled into the stream in the fierce flood of 1938.

Mileage		
Between	Cumulative	
Points		
12.0	40.7	From parking lot take right turn onto Linden Ave. to Spring Ave. and make a right turn onto Spring Ave. Continue on Spring Ave. which becomes Hill Street and hang left to where Hill Street merges into Fourth Street. Continue on Fourth Street to Congress Street and make a left turn onto Congress Street. Cross Hudson River over Congress-Street Bridge and follow sign to 787 (right turn after crossing bridge on to Second Street). Make right turn at Twenty-Third Street and enter ramp to Interstate 787 south. Exit at sign for Rensselaer, cross Hudson River and exit on East Greenbush ramp. Continue on US 9 and 20 to junction with Route 9J. Head south on Route 9J past Port of Rensselaer through Town of East Greenbush. Stop at site of olistoliths on slope of left (east) side of highway, located 0.7 mi south of Rensselaer Town line. Note especially a large white block of limestone.



Figure 22. Olistoliths in shaley matrix. Note especially large, rounded block of limestone in approx. center of view. Snake Hill Formation, Ordovician, cut on east side of Route 9J, 0.7 mile south of Rensselaer Town Line Rensselaer County, N. Y. (Stop 6).

STOP 6 . OLISTOLITHS

At this site “exotic” boulders of various kinds are scattered through shale. Of these one large well-rounded boulder of white limestone demands particular attention (Fig. 22). Smaller angular blocks include those of sandstone and limestone. These olistoliths recall those of Rysedorph Hill and Poestenkill Falls and Gorge and are considered Tippecanoe correlative Snake Hill Formation which at the other two sites have been designated Rensselaer Conglomerate and deserve the same label here. Fisher (1977) applied the term Poughkeepsie Mélange to this deposit. However, for a deposit to be called a *mélange* it should be incorporated within a unit “that was moving as an overthrust or as a gravity-gliding mass”(Friedman et al., 1992). The deposit at this site is composed of olistoliths and is not a *mélange*, as Rowley and Kidd (1982) likewise observed.

- 11.4 52.1 Continue south on Route 9J through Town of Schodack, Village of Castleton on Hudson and Schodack Landing, past signs of Columbia County and Town of Stuyvesant, and pull into unpaved driveway on left at AT&T Stuyvesant facility (which is somewhat hidden behind vegetation).

STOP 7. EXPOSURES SOUTH OF SCHODACK LANDING

We shall first examine the road cut on the east side of the highway and then walk on dirt road across the railroad tracks to view more fine exposures.

Figure 23 describes and illustrates the section seen. Note exposures of bedded micrite. Micrite is overlain by distinctive lenticular brecciolas. Many of the clasts in these conglomerates/brecciolas evidently were derived from the digging up and local transport of nodular bodies of carbonate that may have become segregated as isolated “nodules” during early marine diagenesis of these deep-water carbonates. The conglomerates form convex-up lenses up to 2 m thick and at least 10 m across . Their bases are flat surfaces or fillings of local channels cut a decimeter or so into the underlying nodular zone. Siltstone strata lacking carbonates drape over the convex-up lenses (Friedman and Sanders, 1995, in press).

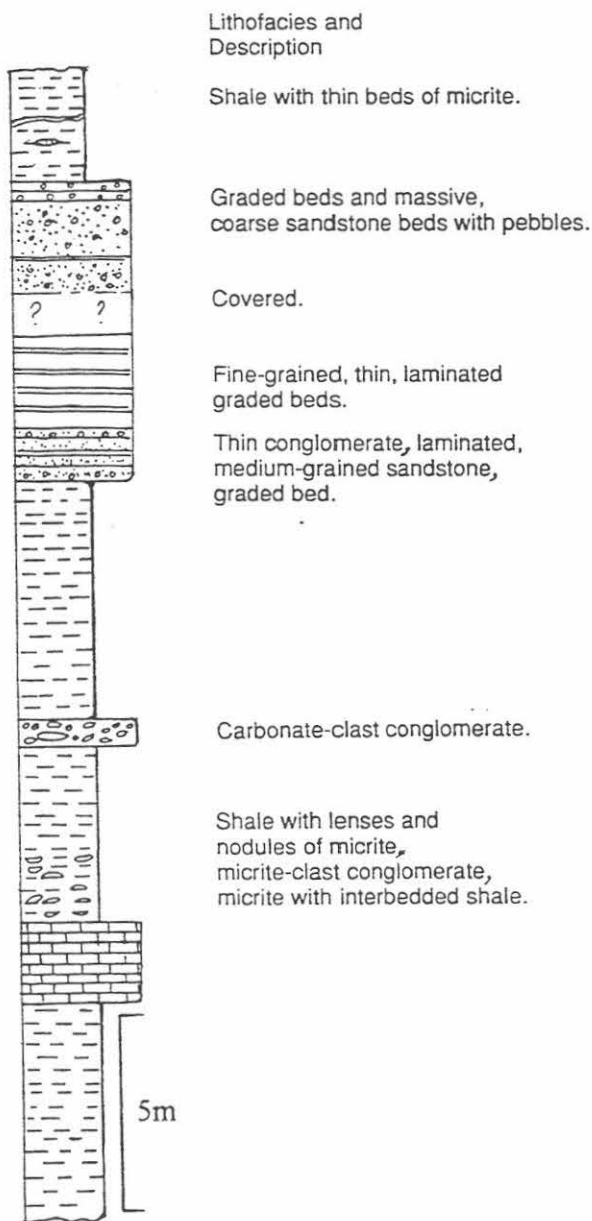
These lenses are analogous to hummocky strata, which have been ascribed to combined effects on shelf sediments of waves and currents. The waves involved in the origin of these deep-water, off-shelf brecciolas probably were tsunami (see previous discussion).

At the top of the slope on the east side of the railroad track an exposure shows a graded layer (2 m thick) of shingled clasts that may have been dumped in a matter of hours (Fig.11). The comparable thicknesses of the overlying strata may have accumulated over thousands of years.

Mileage
Between
Points
10.8

Cumulative
62.9

Continue south on Route 9J through Stuyvesant towards junction with US 9 North and South. After double signs for US 9 (North and South) and just before junction with Route 9 (100 ft.) take a right turn onto road designated Dead End. Bear right at fork. Park at railroad track.



Lithofacies and Description

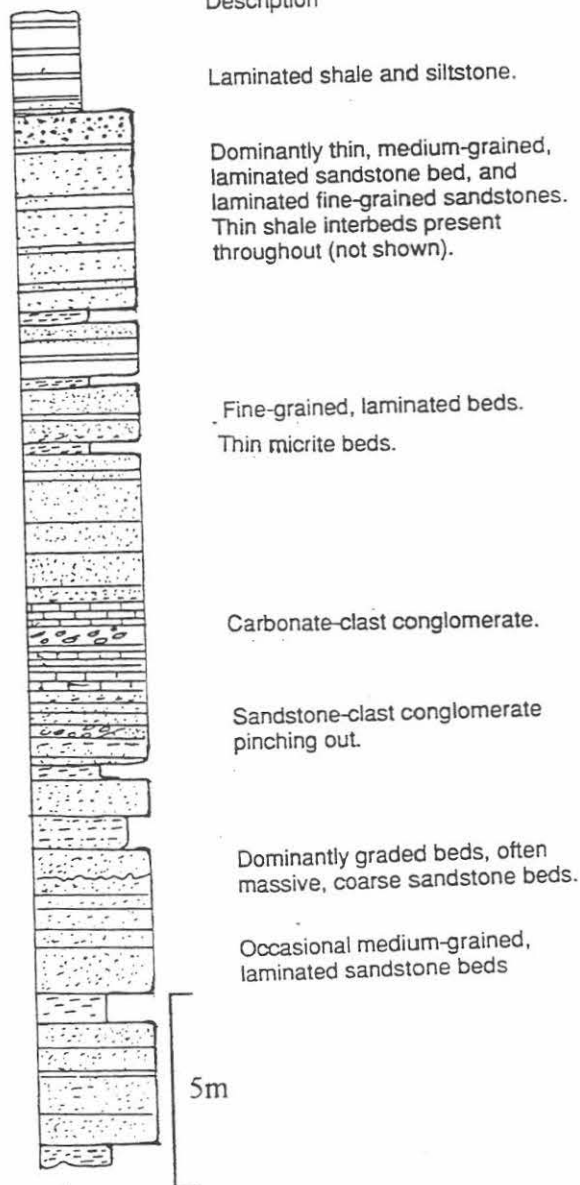


Figure 23. Section of Schodack Landing, N.Y. (Stop 7), (modified from Keith and Friedman, 1977, Fig. 20, p.1234).

Figure 24. Section at Judson Point, N.Y. (Stop 8), (modified from Keith and Friedman, 1977, Fig. 21, p. 1235).

STOP 8. JUDSON POINT

Figure 24 illustrates the section seen at this stop and describes lithofacies. The section is dominated by massive coarse sandstone beds of excellent reservoir quality which display high porosity and excellent sorting. They are analogous to sandstones forming deep-water reservoirs in California and elsewhere. Those not knowing the regional geology may mistake them for high-energy shallow-water deposits. In places the beds contain micrite pebbles.

11.7 74.6 Turn around and at Route 9 turn right and head south on Route 9 through Town of Hudson and Town of Greenport to junction with US 23. Turn left onto US 23 east which coincides with Route 9 south and continue to white house on right side (0.9 mile from junction of Route 9 and US 23).

STOP 9. EXPOSURE ON EAST SIDE OF ROUTE 9 (across from a white house)(SECTION SOUTH OF HUDSON)

Figure 25 shows and describes the lithofacies exposed at this stop and located across from white house. Note especially the interesting interbeds of fine-grained limestone (micrite) and dark shale (Fig. 7) and the carbonate-clast conglomerate (Fig. 8). The cycles consist of alternating thin-bedded micrite and calcareous shale (Fig. 7). The thickness of the micrite beds is about 1 to 5 cm and that of the calcareous shale varies between 3 and 20 cm. Table 2 and figure 26 provide data on the mineralogic and isotopic compositions of the alternating beds of micrite and calcareous shale. The micrite is pure calcite and encloses particles of quartz. The calcareous shale is composed of clay minerals and feldspar particles; its carbonate concentration is about 10% (calcite, dolomite, siderite) (Table 2). The stable carbon isotopic composition of the carbonate of the two interbedded lithologies is almost identical; the oxygen isotopic composition reflects deep-burial diagenesis (Table 2). The carbon isotopic composition of the brecciola overlying the micrite-calcareous shale cycles is enriched in ^{13}C . Although the range of carbon isotopic composition of total dissolved carbon in seawater is relatively narrow, a consistent decrease in $\delta^{13}\text{C}$ occurs with depth of water (around 2.6 per mil) (Berger and Vincent, 1986). Thin-section study of the brecciola confirms its shallow-water origin: radial ooids, echinoderm fragments, and intraclasts of oolitic facies. An approximately 2.6 per mil increase in $\delta^{13}\text{C}$ is shown by comparing the shallow-water brecciola facies with the deep-water micrite. The brecciola facies at this stop represents a shelf-edge facies that a tsunami broke loose and transported to great depth.

Lithofacies and Description

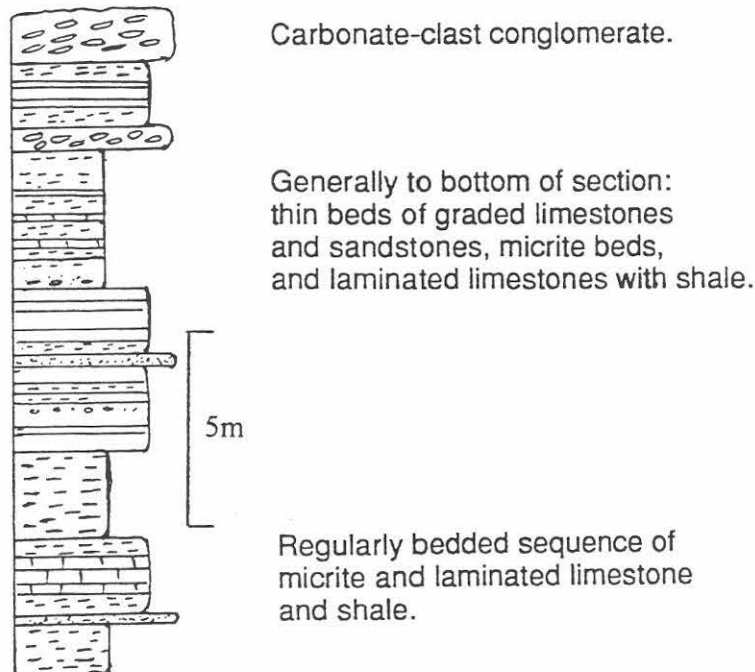


Figure 25. Section south of Hudson, N.Y. (Stop 9), (modified from Keith and Friedman, 1977, Fig. 23, p. 1237).

TABLE 2 TACONIC CARBONATE-SHALE CYCLES (SOUTH OF HUDSON, N.Y. STOP 9)

Sample No.	Lithology	Thickness above base of section (Meters)	$\delta^{13}\text{C}_{\text{PDB}}$	$\delta^{18}\text{O}_{\text{PDB}}$	$^{87}\text{Sr}/^{86}\text{Sr}$	Carbonate %**	% Clay Minerals ***	Quartz %	Feldspar %
1-1	Micrite	0.00	-1.0			92	0	5	3
1-2	Micrite	0.03	-0.5	-11.4		96	0	3	1
1-3	Micrite	0.08	-0.3	-11.4		97	0	3	0
1-4b	Calc.shale	0.22	-1.1	-11.5		20	27	35	18
1-4a	Micrite	0.35	-0.4	-11.6		97	0	3	0
1-5b	Calc.shale	0.48	-1.2	-11.5	0.715163(10)	10	36	36	18
1-5a	Micrite	0.63	-1.0	-11.3	0.711110(10)	96	0	4	0
1-6	Micrite	0.40	-1.2	-11.5		96	0	4	0
1-7b	Calc.shale	1.00	+0.3	-11.8		5	36	38	17
1-7a	Micrite	1.15	-0.9	-11.5		93	0	6	1
2-1	Micrite	2.0	-1.0; -1.1	-11.5	0.711353(10)	92	0	5	3
2-2	Calc.shale	2.2	-1.4	-11.5		11	33	44	10
2-3	Micrite	3.1	-2.0	-10.6		38	5	48	7
2-4	Micrite	3.4	-1.1	-11.3		60	0	39	0
3-0	Micrite	4.7	-1.3	-11.2		97	0	3	0
3-1	Micrite	4.8	-0.5	-11.0	0.710167(10)	91	0	8	0
3-2	Micrite	5.3	-0.6	-11.0		94	0	4	2
3-3	Micrite	5.6	-1.2	-9.2		78	1	19	2
3-4	Micrite	5.7	-1.1	-10.6		70	3	22	4
3-5	Micrite	5.8	-1.4	-10.7		48	2	42	4
3-6	Micrite	5.9	-1.6	-11.0		73	0	24	3
3-7	Calc.shale	6.0	-2.2	-11.3		1	52	37	10
4-1	Carbonate-clast congl.	6.5	+0.6	-10.7	0.710177(10)	94	0	5	1

*The Sr analyses are normalized to $^{86}\text{Sr}/^{88}\text{Sr} = 0.11940$.

Analyses of NBS 987 averaged 0.710241 (09) (n = 39) during the period of these analyses.

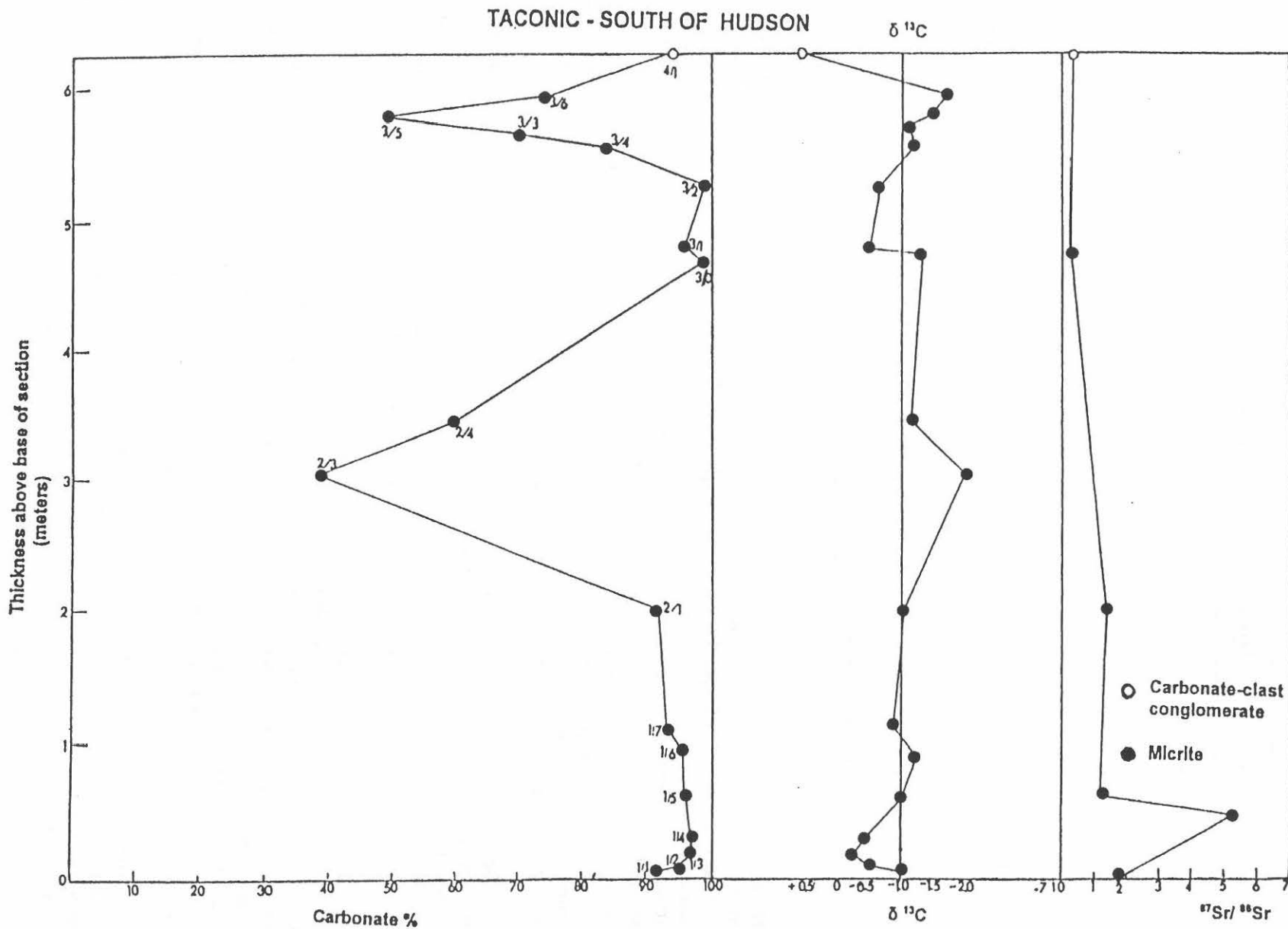
Errors on $^{87}\text{Sr}/^{86}\text{Sr}$ are given as 2 sigma (95%) in the last two digits.

**The carbonate minerals calcite, dolomite and siderite occur in calcareous shale.

***The clay minerals are illite and chlorite.

The strontium isotopic composition requires discussion (Table 2). The principal sources of marine Sr with distinctive $^{87}\text{Sr}/^{86}\text{Sr}$ ratios are (1) old granitic basement rocks of the continental crust (high Rb/Sr, high $^{87}\text{Sr}/^{86}\text{Sr}$); (2) young volcanic rocks (low Rb/Sr; low $^{87}\text{Sr}/^{86}\text{Sr}$); and (3) marine carbonate rocks on the continents (low Rb/Sr, intermediate $^{87}\text{Sr}/^{86}\text{Sr}$) (Faure, 1991, p.359). Because young volcanic rocks are not relevant in a discussion of the Lower Paleozoic, only granitic basement rocks and marine platform carbonates determine strontium isotopic composition. The $^{87}\text{Sr}/^{86}\text{Sr}$ compositions of the carbonate in micrite and calcareous shale (0.711110 and 0.715163) are higher than in Lower Paleozoic seawater (0.7090-0.7095) (Burke et al., 1982) (Fig. 27) and consistent with precipitation from waters containing significant proportions of continent-derived fluids enriched in ^{87}Sr as a result of basement weathering or percolation through soils or sediment on the continental platform.

Figure 26. Isotopic composition of carbonate in micrite and carbonate-clast conglomerate in cycles of section south of Hudson, N.Y. Stop 9.



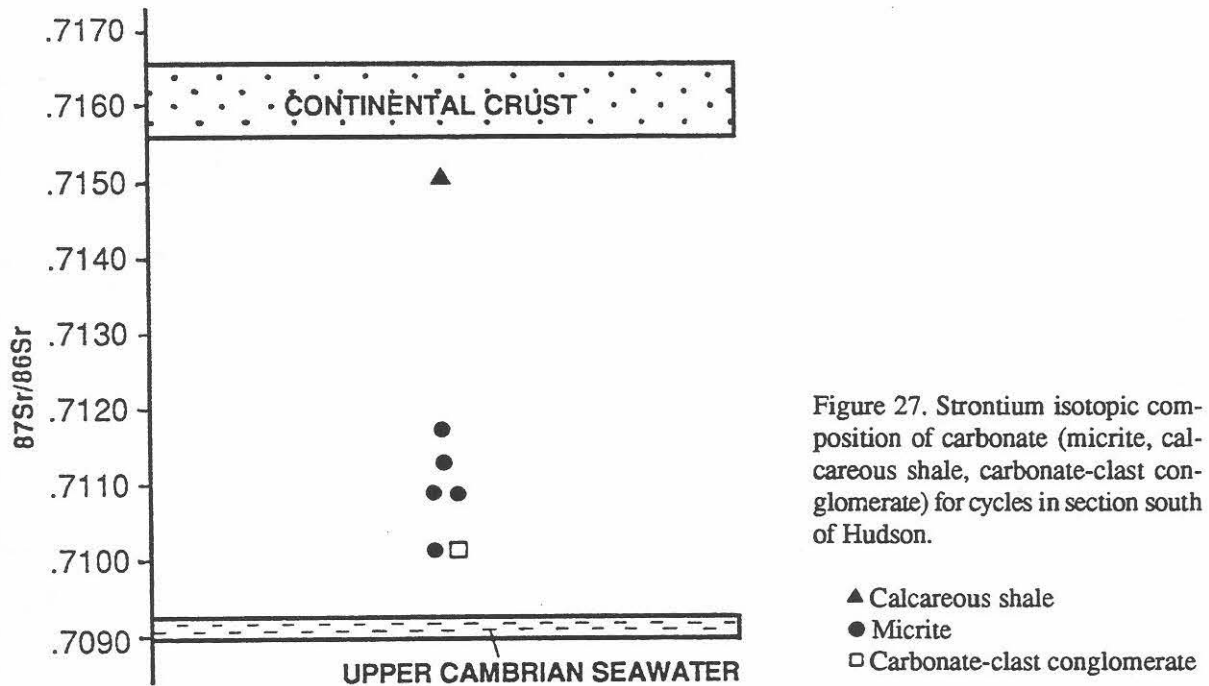


Figure 27. Strontium isotopic composition of carbonate (micrite, calcareous shale, carbonate-clast conglomerate) for cycles in section south of Hudson.

- ▲ Calcareous shale
- Micrite
- ◻ Carbonate-clast conglomerate

Weathering of Precambrian terrain causes enrichment of radiogenic ^{87}Sr ($^{87}\text{Sr}/^{86}\text{Sr}$ 0.716 \pm 0.004) within river and lake waters (Faure et al., 1963). Waters originating from a granitic continental source have a high $^{87}\text{Sr}/^{86}\text{Sr}$ ratio relative to those derived from a platform of marine carbonates (Fig. 27). Even though the $^{87}\text{Sr}/^{86}\text{Sr}$ composition of the carbonate in both the micrite and calcareous shale is higher than in Paleozoic sea water, the difference between the two compositions is remarkable. The strontium isotopic composition of the carbonate in the calcareous shale almost overlaps that of the granitic basement crust, whereas that of the micrite is much lower and closer to that of Paleozoic sea water (Fig. 27). What determines these drastic changes in isotopic composition of such closely interbedded units? The highly radiogenic isotopic ratios of the carbonate of the calcareous shale reflect supply of radiogenic strontium from the decomposition of the detrital particles, such as feldspar and clay minerals which are present in this rock. By contrast, because the calcite of the micrite is essentially devoid of such detrital particles, its strontium isotopic composition is closer to that of Lower Paleozoic seawater.

Because strontium isotopes are not subject to significant mass fractionation during precipitation (Faure, 1986), the waters responsible for precipitating the carbonate of the calcareous shale must have derived from a granitic continental source. The only such source would be the Precambrian shield which must have been exposed during Early Cambrian times. The $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of the micrite, though much lower, still indicates enrichment of radiogenic strontium from the granitic continental terrain, but its signature is considerably closer to that of marine platform carbonates.

The strontium isotopic evidence indicates that the calcareous shales were derived from exposed granitic basement. Sea level must have been low so that weathering and erosion of the granitic continental terrain generated the fine-textured particles. Rivers then distributed the debris across the emergent continental platform and down the slope, probably through submarine canyons. As pointed out earlier, a large block of quartzite in this formation has been tentatively identified as derived from part of the wall of a submarine canyon (Friedman, 1972; Friedman Sanders and Martini, 1982). By contrast, the strontium isotopic evidence for the micrite suggests mixing of waters originating from granitic continental terrain with waters from the epeiric marine carbonate platform under conditions of high stand of sea level. Thus the calcareous shale beds represent low-stand sea-level facies tracts, whereas the micrites are high-stand sea-level facies tracts.

The alternating thin-bedded micrite and calcareous shale beds reflect rapid changes in sea level. To produce such closely interbedded lithologies, the sea level must have moved up and down like a piston. Such cycles have been explained as astronomical rhythms (de Boer, 1991). For low latitudes, the precession of the Earth's axis has been assumed the cause of such rhythmicities. Precession cycles last about 19,000 and 23,000 years. At this time no dates are available to determine the rate of sea-level changes, but 19,000 to 23,000 years are possible time frames for micrite-calcareous shale couplets to accumulate.

Mileage Between Points	Cumulative	
0.8	75.4	

Turn around and return on Route 9 and US 23 to intersection of Route 9 north and US 23 west. Note exposure on southwest corner of Becraft Mountain.

STOP 10. SOUTHWEST CORNER OF BECRAFT MOUNTAIN (Fig. 28)

At the base of this exposure rocks of the Mount Merino Formation of the Taconic Sequence (Fig. 1) consist of highly

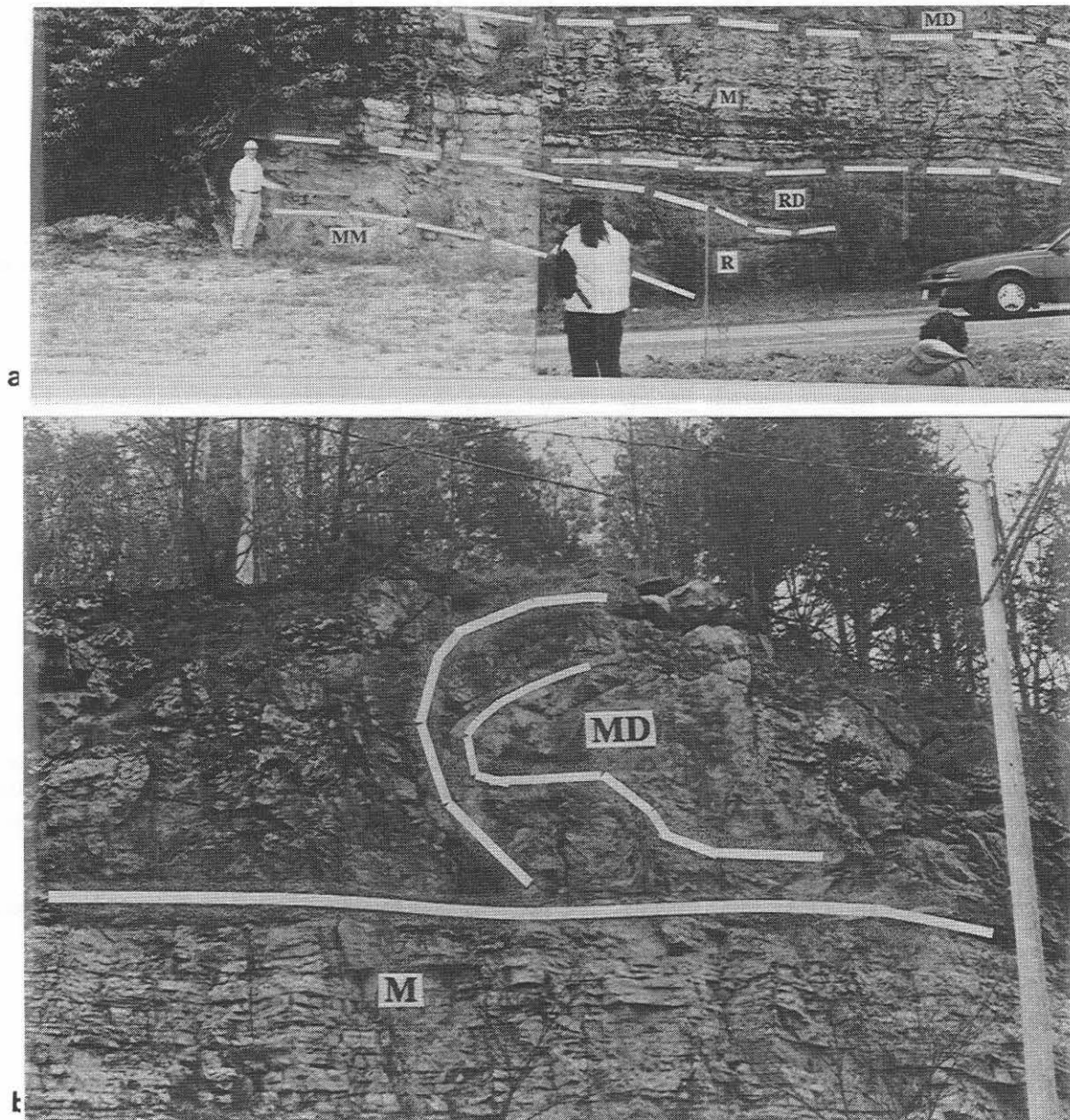


Figure 28. View of exposure at southwest corner of Becraft Mountain. (Stop 10).

(a) MM: Mount Merino Formation of Taconic Sequence.

R: Rondout Formation (Silurian) unconformably overlying Mount Merino Formation.

RD: Deformed Rondout Formation: note thrust between undeformed and deformed Rondout Formation.

M: Manlius Formation.

MD: Deformed Manlius Formation.

Geologist at left points stick just above unconformity with underlying Ordovician Mount Merino Formation.

(b) Note intense folding in deformed Manlius Formation as a result of thrusting.

M: Manlius Formation. MD: Deformed Manlius Formation.

deformed cherty dolomite. A low-angle, irregular unconformity separates this Ordovician formation from the overlying Silurian Rondout Formation, a silty dolomite. The lower part of the Rondout Formation is undeformed, however, a bedding thrust through this formation has deformed the Rondout Formation above this thrust (Fig. 28). Undeformed, laminated, whitish limestones of the Manlius Formation overlie the Rondout Formation. Within the Manlius Formation a bedding thrust folded the upper part of this formation (Fig. 28).

X-ray study of the rock of the bedding thrust within the Manlius Formation shows it to consist of calcite (96%) and quartz (4%). Its isotopic signatures are $\delta^{13}\text{C}_{\text{PDB}} +2.4$, $\delta^{18}\text{O}_{\text{PDB}} -9.9$, and $^{87}\text{Sr}/^{86}\text{Sr}$ 0.715096 (10) (± 2 S.D.) (The Sr analyses are normalized to $^{86}\text{Sr}/^{88}\text{Sr} = 0.11940$. Analyses of NBS 987 averaged 0.710241 (09) ($n = 39$) during the period of these analyses. Errors on $^{87}\text{Sr}/^{86}\text{Sr}$ are given as 2 sigma (95 %) in the last two digits). The carbon and oxygen isotopes tag a hot heavy-carbon enriched fluid and the strontium isotopes a radiogenic strontium of composition close to that of the continental crust.

The thrusts at this exposure are post-Taconic and resulted from Acadian or Alleghanian orogenesis.

OUTCROP GUIDE AND ITINERARY: MOLASSE OF CATSKILLS

The term **molasse** designates a *tectono-stratigraphic unit consisting of a wedge-shaped body of extrabasinal sediments typified by patterned successions of shallow-marine- and nonmarine strata, whose particles were derived from erosion of the older rocks, including flysch, that composed the rising mountain chain* (Friedman, Sanders, and Kopaska-Merkel (1992). The Devonian Acadian orogeny generated the Catskill deposits. The Devonian succession in the Catskill region has been, named a *tectonic fan-delta complex* (Friedman, 1988a).

14.9 90.3 Drive on US 23 west to Rip Van Winkle Bridge, note Catskill front from distance as you approach bridge. Cross bridge over Hudson River, get off US 23 at sign Jefferson Heights and Leeds, and make left turn towards Jefferson Heights on to Green County 23B through Catskill. Take right turn at junction with US 23A towards Hunter.

Stop 0.1 mile east of junction of US 23A with US 32.

VIEW OF CATSKILL FRONT

STOP 11 . BRAIDED AND MEANDERING STREAM FACIES SEEN FROM A DISTANCE

From the vantage point of this stop we see to the west the classical fluvial sequence of the Middle and Upper Devonian rocks of the Catskill front. Note continuous ledges near the top of the front which reflect the presence of laterally continuous, coarse-grained sandstones and conglomerates deposited in braided streams (Butner, 1968). Below the continuous ledges are discontinuous cliffs which reflect laterally interfingering channel sandstones and overbank shales deposited in meandering streams.

Infrared photography of the Catskill front from this stop brings out amazing details (Mutch, Head, and Saunders, 1968).

2.3 Continue west on NY 23A to Palenville. Turn right (north) on Boggart Road.
1.1 93.7 Turn left at fork on to dirt road and drive <0.1 mile to tresle to Mountain House.

STOP 12. TRESTLE TO MOUNTAIN HOUSE

Walk approximately 300 feet uphill to first sandstone body.

Meandering-Stream Facies

This grayish green, medium-grained graywacke of the Hamilton Group of Middle Devonian age displays abundant truncations resulting from lateral cutting; the sandstones within the channels are crossbedded and display abundant reactivation surfaces. This channel complex is about 15 feet thick and is sandwiched between overbank shale which is well exposed in the hangover at the lower sandstone contact. Note sporadic pebbles, wood fragments, and wood casts as well as onlap and

toplap. Looking uphill along the clearing of the old trestle to Mountain House between cliffs of similar sandstone bodies the slopes mark the sites of interbedded shale. The interbedded sandstones and shales represent interfingering channel and overbank deposits of meandering- stream facies. The lateral relationship between channel and overbank will be apparent at stop 15. The green color of the graywacke reflects the abundance of chlorite in the rocks. This chlorite was derived from the source terrain to the east which consisted of metamorphic rocks of the greenschist facies.

A similar sandstone body occurs about 300 feet below which is underlain by gray, red, and green shales.

- | | | |
|-----|------|---|
| 1.1 | 94.8 | Return to NY 23A. Turn right (west) on NY 23A; cross Kaaterskill Clove three times ascending Catskill Front. Note red beds in gorge: interbedded channel sandstones and overbank shales. At sign on right designated STATE LANDS 1885-1935 pull over to view interbedded red sandstones and shales. |
| 3.8 | 98.6 | Turn left at Twilight Park entrance and park at bridge of Kaaterskill Clove. A “clove” is a deep ravine. |

STOP 13. TWILIGHT PARK

Twilight Park is a private residential section and permission is needed to park here which may be obtained from the Superintendent Hillard Hommel or Justine L. Hommel, Box 129. Haines Falls, NY 12436 (phone 518-589-6191).

The site at this bridge was one of the preferred views of the Hudson River School of Landscape Paintings. Asher Durand (1786-1886) in his painting *Kindred Spirits* illustrated this site, including the crossbedding developed within the pointbars of the meandering stream facies visible below this bridge (Jordan, 1995).

- | | | |
|-----|------|--|
| 0.1 | 98.7 | Return to NY 23A and turn left (west). Drive through Haines Falls, Tannersville and Hunter on NY 23A. At junction of NY 23A and NY 296 is the next stop. |
|-----|------|--|

STOP 14. BRAIDED-STREAM FACIES, HUNTER

At this site polymictic conglomerate of braided-stream facies overlies sandstone of meandering-stream facies (Friedman, Sanders, and Kopaska-Merkel, Fig. 4-17). Small channels of conglomerate also cut through sandstone. The pebbles in this conglomerate range in diameter between 3 and 10 cm. The pebbles and sand particles are composed of quartz, including vein quartz, and various sedimentary and metamorphic rock fragments.

Streams of high energy must have been at work to transport these large pebbles. As we have seen from stop 11 these streams were laterally continuous channel complexes devoid of intervening overbank deposits. The coarse particle size of the conglomerates and the absence of overbank shales or siltstones suggest a braided-stream deposit. This complex stream system developed on a slope of steep gradient, most probably in association with a series of coalescing alluvial fans that spread westward from the high, tectonically active source terrain to the east.

- | | | |
|-----|-------|--|
| 6.9 | 105.6 | Continue north on NY 296. Drive through Hensonville. |
| 7.0 | 112.6 | Turn right (east) on NY 23; drive beyond Point Lookout. |
| 5.8 | 118.4 | STOP 14. EAST WINDHAM. Stop before curve at yellow road sign showing bent arrow (< 0.1 mile before parking lot on left). |

STOP 15: MEANDERING-STREAM POINT-BAR FACIES: CHANNELS, OVERBANK, AND SWAMP

This exposure shows a point-bar sequence in Tully-clastic correlative strata of the Gilboa Formation (Upper Givetian, Uppermost Middle Devonian). Two channels are exposed in this road cut, a lower channel and an upper channel (Fig. 29). The lower channel truncates overbank siltstone, whereas the upper channel truncates the lower channel and laterally adjacent overbank siltstone. A shale-pebble conglomerate, as a lag concentrate, overlies the truncation surface of the lower channel; in part this conglomerate is now hidden by fallen debris. This point-bar sequence at this site represents an abandoned meander which became an oxbow lake.

In rocks of the overbank facies, dark gray to black interbeds and, lenses containing abundant coarse plant remains,

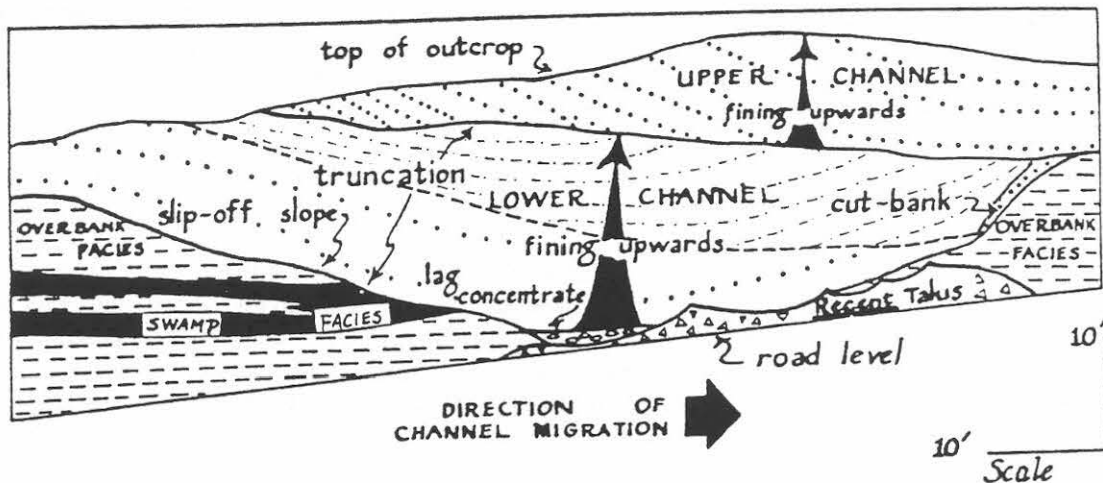


Figure 29. Compound channel in meandering stream point-bar facies. East Windham, N.Y. (modified from Johnson and Friedman, 1969) (Stop 15).

consisting of what appear to be stems and branches primarily, and locally developed coaly material, represent a back-swamp environment located on the slip-off slope side of the lower channel. The back-swamp facies appears to overlie crevasse-splay deposits. These beds, which occur in red overbank siltstone and contain very abundant plant material, some of which has been altered to a coaly substance, are some 5 feet thick and measure well over 60 feet in lateral extent.

The woody cells of the plant debris were in part converted to anthracite and in part replaced by pyrite. Replacement occurred at an early stage as evidenced by the lack of deformation of the replaced woody cells. Most of the pyrite has now been oxidized to hematite and limonite, probably by surface weathering. Even the anthracite, which usually resists oxidation, shows evidence of undergoing oxidation in this material. Surface oxidation was probably intensified by sulfate-bearing water derived from pyrite. The vitrinite reflectance of samples of this plant debris is 2.2% to 2.3%. The mean reflectivity of the vitrinites in oil using light with wavelength of 546 m is 2.5%. This indicates that the vitrinite contains 93% to 94% fixed carbon, about 3.5% hydrogen, and other volatile matter. Locally the coalified plant chambers contain anhedral pyrite and galena: the paragenetic sequence is pyrite and then galena. The vitrinite (coalified woody tissue) contains micron-sized pyrite cubes. Outside the plant tissues, small amounts of chalcopryrite and possibly sphalerite are present (Friedman, 1987a,b; Friedman and Sanders, 1982). The deposits observed at this site were buried to a depth of 6.5 km for a duration of about 200 million years (Friedman and Sanders, 1982).

The steep cut-bank on the right of the lower channel may initially have extended upward for many more tens of feet and tapped sand from an earlier upper channel. Sand, now sandstone, moved down the cut-bank as a liquefied cohesionless particle flow or grain flow and is now preserved as an inclined sand body paralleling the cut bank.

Bivalves of the Species *Archanodon* (= *Amnigenia*, Hall 1874) *catskillensis* (Heteroconchia, Archanodontacea) inhabited the back-swamp facies on the slip-off slope margin of the lower channel. These bivalves (up to 22 cm x 6 cm) are among the largest known fossil fresh-water pelecypods in geologic history. Unfettered, opportunistic growth in this swamp environment marked by rich organic matter and devoid of competition with other pelecypods, and possibly other benthic fauna of any kind, resulted in the unusually large sizes of these clams (Friedman and Chamberlain, 1995).

To the left of the point-bar sequence note onlap of stream facies on to an interpreted soil horizon.

Johnson and Friedman (1969, p. 463-468, including Figs. 7-17) provide a detailed description of this stop (their section 53); Figure 29 is a modified version of Figure 7 of Johnson and Friedman.

Return to Union College via New York Thruway (Interstate 787).

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TRAVERSE ACROSS THE TACONIAN OROGEN

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INTRODUCTION

In Early Ordovician time the rifted and submerged eastern edge of the North American continent (current coordinates) was located approximately at the present location of the Connecticut River in western New England (Figures 1A and 2). To the east, across a narrow ocean basin, lay an island arc. Subduction of oceanic crust beneath the arc allowed the arc to advance toward North America, colliding in Middle to Late Ordovician time (Figures 1B to 1F). The remnants of the island arc itself are present in western New England as a sequence of Ordovician sediments (e.g., Bradley and Kidd, 1991) and bimodal volcanic and plutonic rocks (e.g., Leo et al., 1984; Leo, 1985, 1991; Schumacher, 1988; Hollocher, 1993, 1994). The remnants of the island arc are found in four different forms: 1) unmetamorphosed sediments derived from the arc and forearc regions that were deposited west of the arc on North American crust; 2) variously metamorphosed continental shelf, continental slope, trench, and forearc rocks that became part of the arc's accretionary wedge and were thrust westward onto North America during subduction and collision; 3) interbedded sediments and volcanics associated with arc, and possibly forearc and back arc volcanics; and 4) the plutonic roots to the arc. This trip examines all four of these.

THE EARLY PALEOZOIC NORTH AMERICAN CONTINENTAL MARGIN

In the Late Precambrian and Early Cambrian much of eastern North America was emergent from the oceans, with Late Precambrian and Early Cambrian sediments accumulating on the continental shelf. Deeply eroded Precambrian metamorphic rocks, including those of Grenville age, were exposed over much of the surface of the continent. As the continental margin gradually subsided, the first unit deposited on the unconformity in many areas was quartz sandstone and quartz pebble conglomerate. This deposit is the Cheshire Quartzite in the Green Mountains and Berkshires of New England, and the Potsdam Sandstone in the area surrounding Precambrian rocks in the Adirondacks of New York State (Stanley and Ratcliff, 1985, Ratcliff et al., 1988). The unmetamorphosed Potsdam Formation is quite complex and includes fluvial, beach, dune, and shallow water offshore facies (Selleck, 1993). The basal Paleozoic sandstones are overlain by a variety of Cambrian and Lower Ordovician clastic and carbonate rocks that were deposited in shallow seas in near shore environments on the North American continental margin.

IGNEOUS ROCKS ASSOCIATED WITH THE TACONIAN ISLAND ARC

By Middle Ordovician time an island arc (the Taconic arc) began approaching North America because of subduction of the intervening oceanic crust beneath the arc. Large volumes of magma were emplaced into the arc. The remnants of the arc plutonic complex were metamorphosed during the Devonian Acadian orogeny, and are presently exposed as gneisses in eroded Acadian structural domes in the Bronson Hill anticlinorium (Figures 2, 3) and in a few outlying domes to the west. These Taconian arc plutonic rocks are predominantly felsic, metaluminous, and calc-alkaline, and range in composition from tonalite through granite (Leo et al., 1984; Leo, 1991; Hollocher, 1994). Small quantities of mafic, ultramafic, and anorthositic rocks also occur.

The compositions of the Taconian plutonic rocks in the Bronson Hill anticlinorium vary geographically (Figure 3). In the northern section of the Bronson Hill anticlinorium, Taconian felsic gneisses are predominantly granitic, rich in K_2O and Ba, poor in Sr, and have generally high Y concentrations. These rocks are generally strongly LREE-enriched, have flat MREE's and HREE's, and have moderate negative Eu anomalies. The rocks in the west-central part of the Bronson Hill anticlinorium are dominantly tonalitic. These rocks have less K_2O and Ba than those in the northern section, and tend to be somewhat less LREE-enriched. Rocks in the southeastern section are also dominantly tonalitic, but occur in two lithologically similar but geochemically distinct groups. The first group is identical to rocks of the west-central section, being largely tonalitic, having flat MREE and HREE patterns, and having

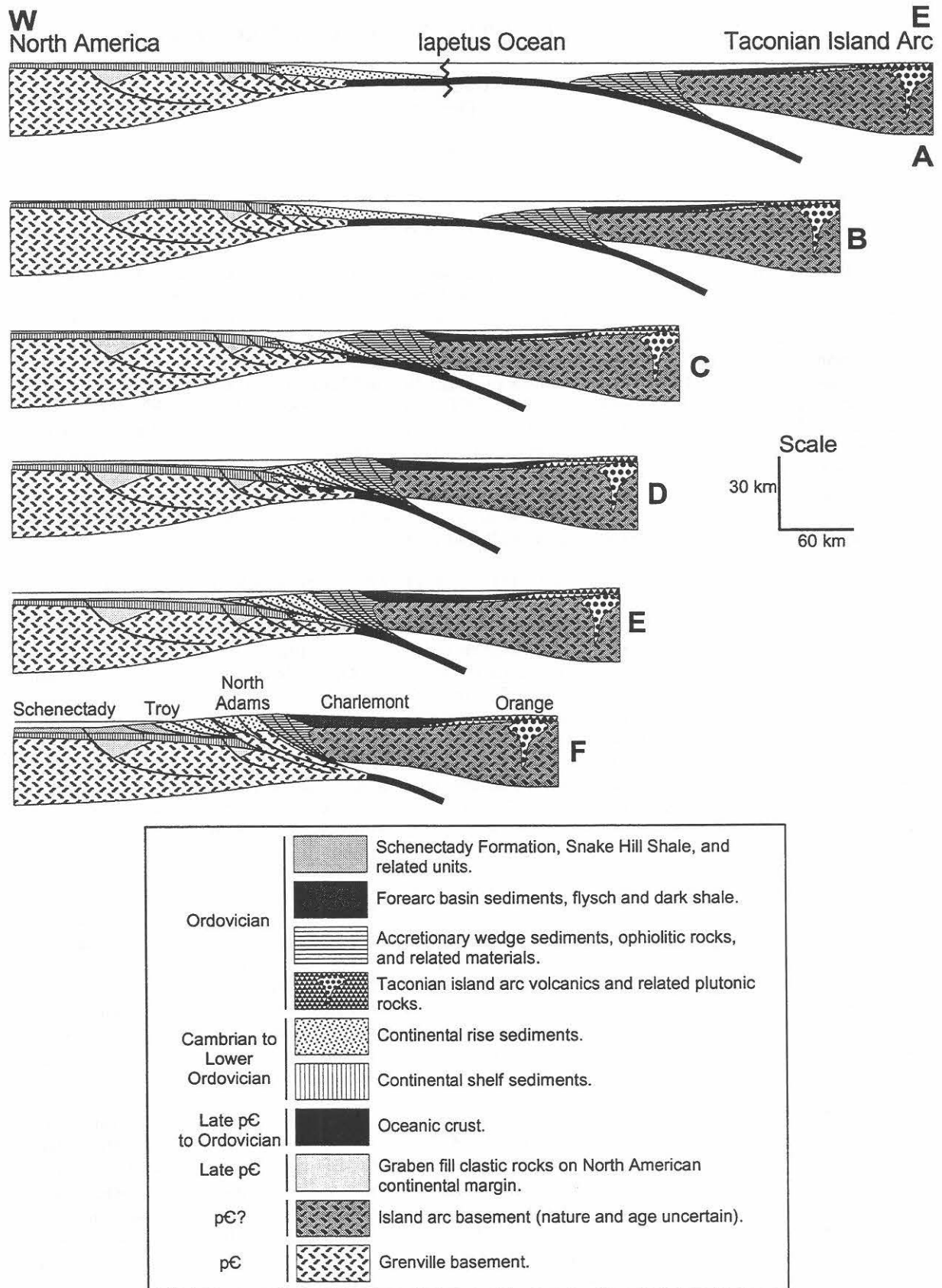


Figure 1. Schematic cross sections showing the tectonic development of the Taconic Orogen from Early Middle Ordovician (A) to Early Upper Ordovician (B). Adapted from Rowley and Kidd (1981).

marked negative Eu anomalies, low Sr, and generally high Y. In contrast, the second group has strongly depleted HREE's, no significant Eu anomaly, high Sr, and low Y.

The dominantly granitic Taconian arc plutonic rocks of the northern section are likely to have been derived from the melting of relatively K-rich intermediate or felsic igneous rocks (Roberts and Clemens, 1993) in the lower crust of the Taconian island arc. The preponderance of granitic rocks in the northern section of the Bronson Hill anticlinorium suggests that the basement to the Taconian arc in this area was composed of continental crust. The generally tonalitic Sr-poor rocks in the west-central and southeastern sections were probably derived from the melting of relatively K-poor mafic igneous rocks at moderate pressure in the stability field of plagioclase. Residual plagioclase during melting yielded the observed negative Eu anomalies and low Sr concentrations in the resulting liquids, because of the strong partitioning of these elements into plagioclase. The generally tonalitic high-Sr, low-Y, HREE-depleted rocks in the southeastern section were also probably derived from the melting of relatively K-poor mafic igneous rocks, but at a higher pressure in which garnet was abundant in the restite after melting rather than plagioclase. Garnet in the source area explains the observed lack of negative Eu anomalies, the strong HREE-depletion, the high Sr contents, and low Y contents of these rocks. The change across the strike of the Taconian arc from rocks having flat MREE and HREE element patterns to rocks having depleted HREE's is similar to variations in composition of plutonic rocks observed across strike in more recent subduction zone environments (e.g., the Peninsular Ranges batholith; Gromet and Silver, 1987).

The volcanic rocks that are associated with the Taconian island arc system are rather complex and somewhat problematic. The large volumes of volcanics that are usually associated with island arcs, and expected considering the large volumes of Ordovician plutonic rocks that are exposed, do not exist in the Bronson Hill anticlinorium, the presumed axis of the arc. The Taconian plutonic rocks in the Bronson Hill anticlinorium in southern New England are overlain by the Ordovician Ammonoosuc Volcanics and the Partridge Formation, which also contains metamorphosed volcanics. These rocks are relatively thin (generally only few hundred meters thick), and are in part mineralogically and chemically distinct from the Taconian plutonic rocks (Hollocher and Lent, 1987; Hollocher 1994). These and related volcanics have been interpreted, at least in part, to belong to a back-arc basin environment rather than to the arc itself (Hollocher, 1993; Stoll and Karabinos, 1993). It seems likely that large volumes of volcanics on the arc itself were eroded away during and after arc development in the Late Ordovician or Silurian. A variety of other metamorphosed volcanic and shallow intrusive rocks occur in units of probable Ordovician age between the Bronson Hill anticlinorium and the Green Mountain and Berkshire massifs. Some of these volcanics and related intrusives in the Hawley Formation (Kim and Jacobi, 1994, 1995 and in review; in part Stop 8) have recently been interpreted as an early set of basalts having variously island arc, mid-ocean ridge, and forearc boninitic affinities, and a later set having back-arc basin affinities. These suggest a complex evolution of the Taconian island arc, and perhaps even more than one arc.

EFFECT OF THE TACONIAN COLLISION ON NORTH AMERICA

Sediments that were scraped off the subducting plate accumulated on the trench side of the advancing island arc to form an accretionary wedge or thrust complex (Figure 1). Each package of accreted sediment was incorporated in the accretionary wedge along thrust faults. These faults stepped westward during collision of the arc with North America, forming a series of allochthonous, thrust-bounded packages of rock that now make up the Taconic Mountains and underlie much of the Hudson River Valley (Figures 1, 2, 4).

Just prior to subduction, the downgoing plate tends to bulge upward a little bit due to the strength of the subducting lithosphere. This strength transmits flexural stress oceanward from the downward-bending plate as it descends into the trench. As the Taconian island arc began to advance on the North American continent, this flexural stress caused up-bowing and emergence of parts of the North American margin. Emergence in the Middle Ordovician resulted in erosion of shallow marine sediments over much of New York State, forming the extensive Upper Middle Ordovician unconformity in this area (Bradley and Kidd, 1991; Isachsen et al., 1991). Flexural stresses during collision also caused extensive normal faulting due to downloading of the North American margin, some of which involved reactivation of older faults (Bradley and Kidd, 1991).

As the arc continued to advance, continued loading of the eastern margin of North America by the arc and the overriding accretionary wedge caused the region to submerge again, resulting in a deep-marine basin into which gray and black shales of the Snake Hill Shale accumulated. Gradually, sediments were shed westward off of the advancing arc and forearc region and were deposited into this basin as flysch of the Schenectady Formation. The Schenectady Formation, and its tectonized equivalents such as the Austin Glen Formation in the Hudson River Valley and the Pawlet Formation in the Taconics, consist of

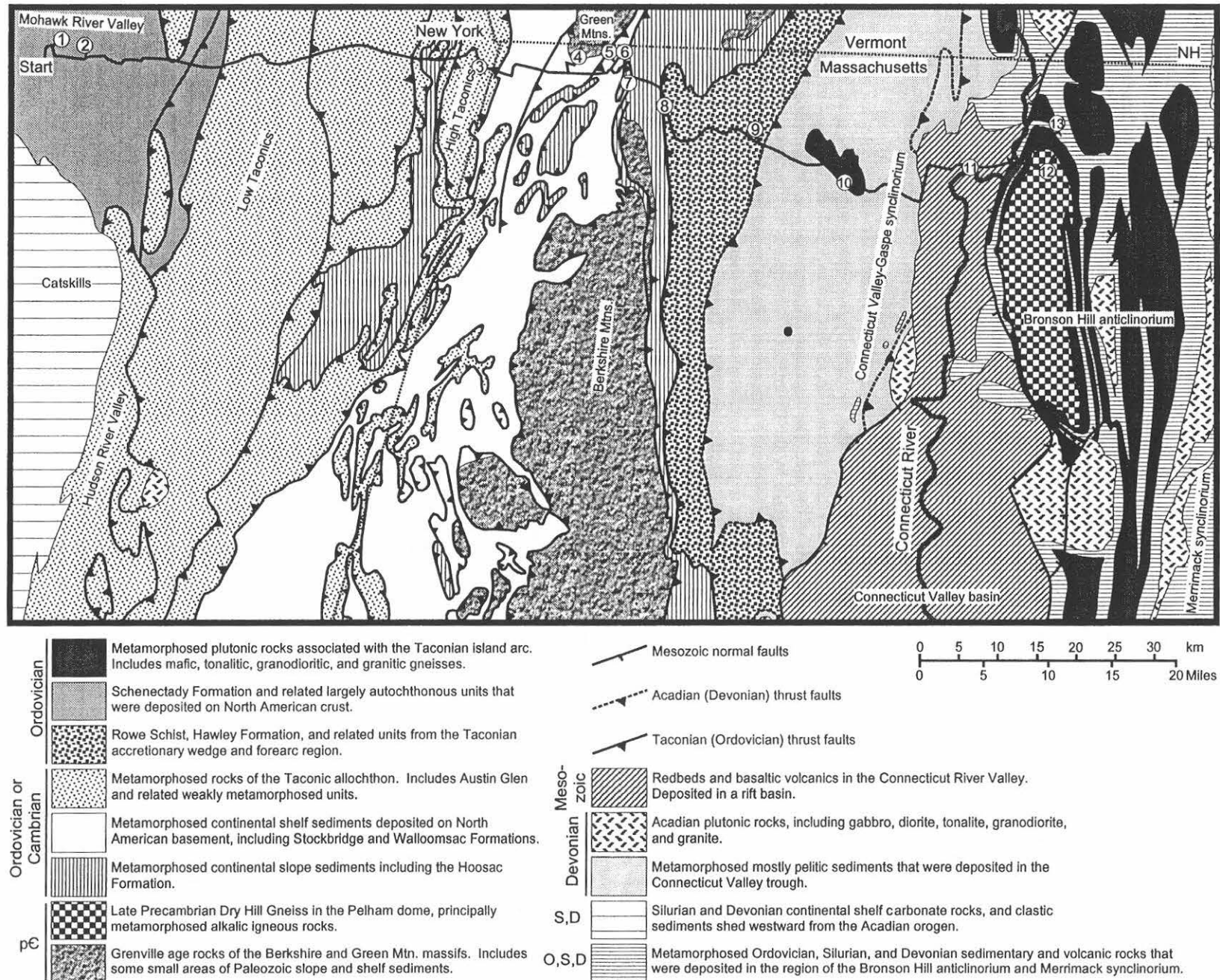


Figure 2. Generalized geologic map of eastern New York, western Massachusetts, and parts of adjoining states, showing the field trip route and trip stop locations (adapted from Stanley and Ratcliff, 1985; Zen, 1983; Rogers, 1990).

alternating gray shale and gray turbidite sandstones. The sandstones range in thickness from a few millimeters to over a meter, and in places have good internal grading, internal sedimentation features, and sole marks including superb flute and groove casts.

The collision of the arc with North America not only involved westward thrust transport of accretionary wedge and forearc sediments, but also involved transport Precambrian North American basement rock and overlying Lower Paleozoic sediments. These west-transported allochthonous blocks of metamorphic basement and cover rocks are resistant to erosion and hold up the Green Mountains and Berkshires of western New England (Figures 2 and 4).

METAMORPHISM

Figure 5 shows the realms of metamorphism in the area of this field trip. The trip starts in the Schenectady Formation, which is clearly unfoliated sedimentary rock (Stops 1 and 2). In the Schenectady area this rock contains coarse illite, chlorite, and kaolinite that, with other evidence, suggests diagenetic temperatures approaching 200°C (supported by work on units nearby, e.g., Friedman and Sanders, 1982; Conrad et al., 1983; Johnsson, 1986). These high diagenetic temperatures may have been the result of deep burial in the Devonian or Late Paleozoic. Rocks in the Taconic thrust slices, in and to the east of Troy, NY, are generally well-foliated and are of chlorite grade and higher. Ordovician metamorphic grade increases to chloritoid grade in the vicinity of the Taconic crest (Stop 3), to biotite grade in the Green Mountains-Mt. Graylock area (Stop 4), and to garnet grade in the eastern North Adams and the north-western Berkshires (Stops 5, 6, and 7). Ordovician metamorphic grades up to sillimanite grade are found in the central and southern Berkshires, south of the route of this trip. To the east of Stop 7, the effects of Ordovician metamorphism either never existed or have been overprinted by Acadian (Devonian) metamorphism, which ranges from the chlorite grade to the lower granulite facies (Figure 5). For an excellent discussion of Taconian and Acadian metamorphism, see Robinson (1986).

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Figure 3. Generalized lithological and geochemical domains of Ordovician (Taconian) felsic gneisses that are exposed in the cores of structural domes in the Bronson Hill anticlinorium. There are three domains that are distinguished by generally different lithology, higher vs. lower concentrations of K_2O and Ba, higher vs. lower Sr and Y, and different REE patterns. All graphs of each type are at the same scale. Data are from Hodgkins (1985), Leo (1991), Leo et al. (1984), Webster and Wintsch (1987), and references therein, and unpublished data of Kurt Hollocher and Jon Bull. Ordovician rocks are shown in black, and unrelated Late Precambrian rocks in the Pelham dome are shown in a vertical line pattern.

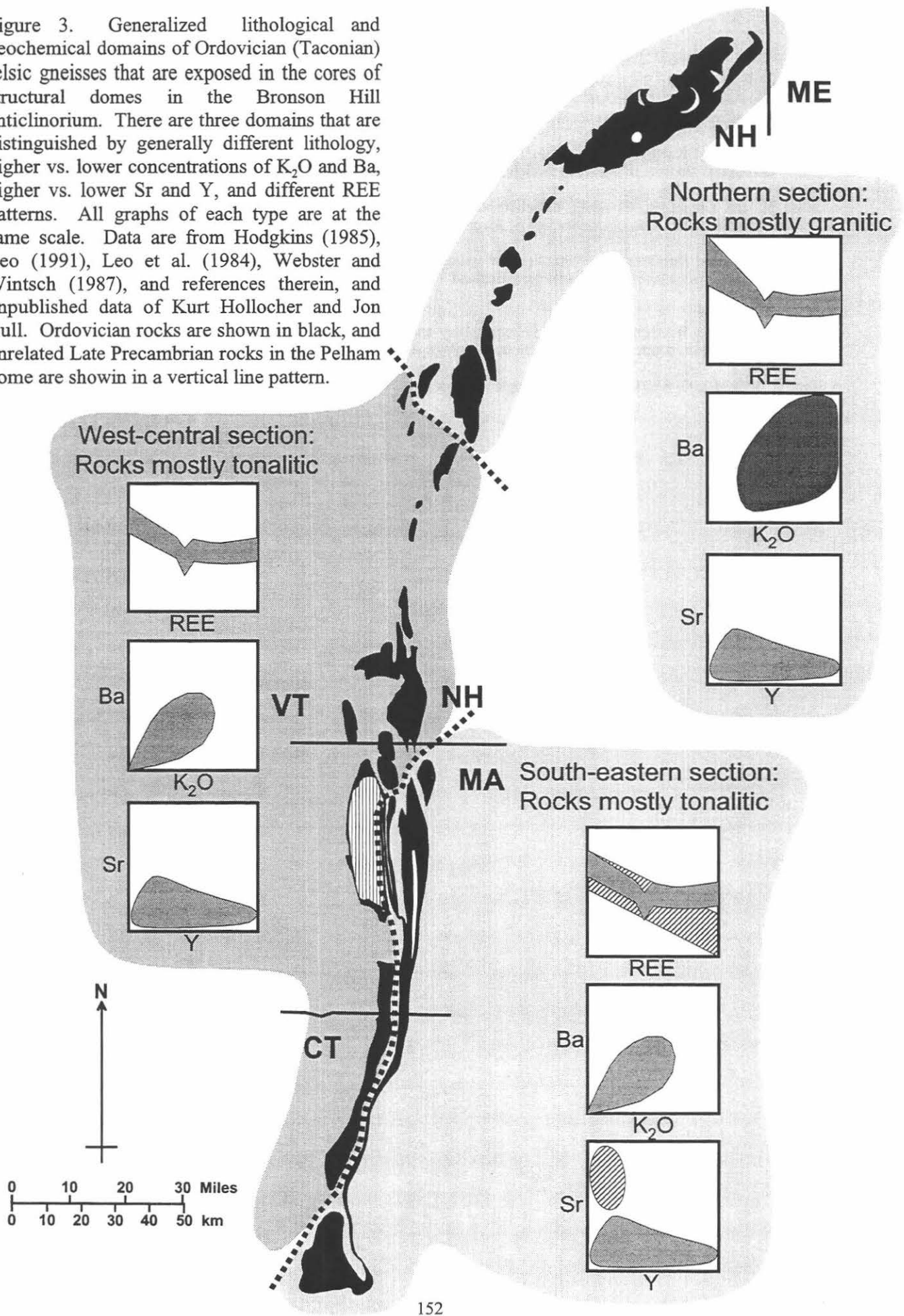
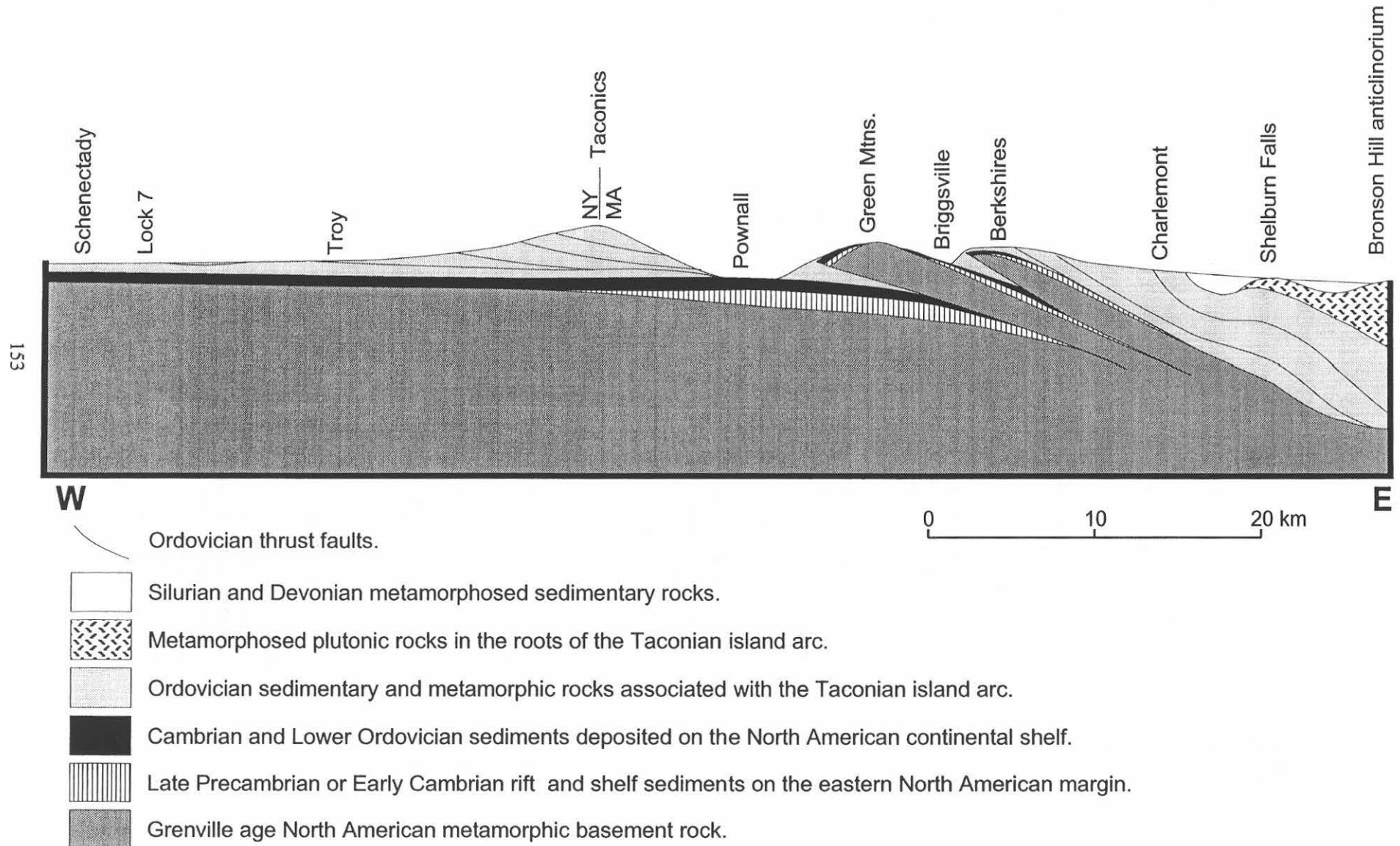


Figure 4. Schematic cross section from Schenectady, NY, across the Taconics, the southern Green Mountains, the Berkshires, and the remnants of the Taconic island arc. High-angle Taconian faults and post-Ordovician deformation are not shown. Greatly modified after Stanley and Ratcliff (1985) and Zen (1983).

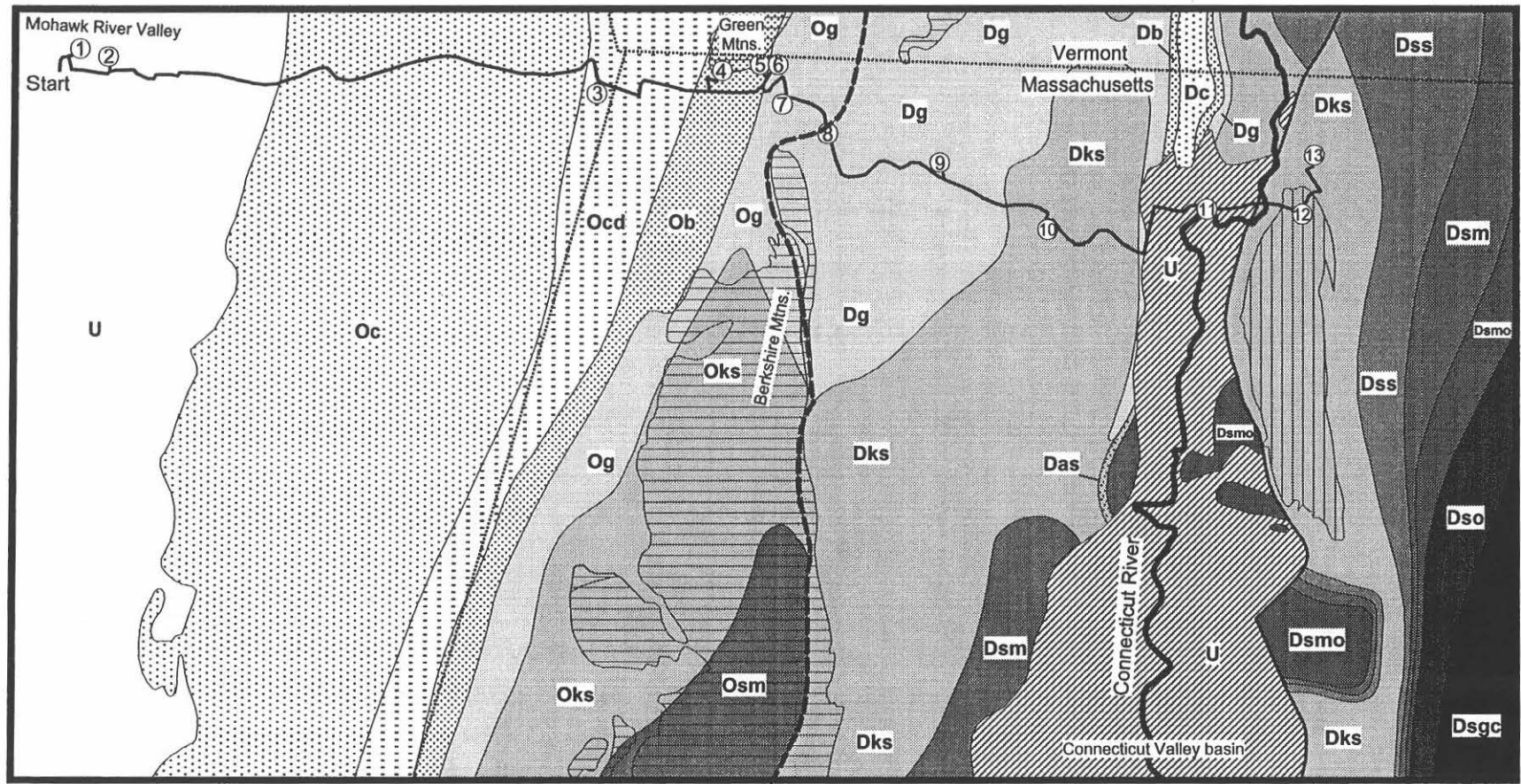


TRAVERSE ACROSS THE TACONIAN OROGEN

ROAD LOG

Miles from Start	Miles from last point	point
0.0		Start at the parking lot near the power house, Union College, near the Geology Department.
0.1	0.1	Leave the Union College gate. Cross the intersection at Nott St. straight ahead (north) onto Van Vranken Ave. Van Vranken eventually turns into Aqueduct Rd.
2.9	2.8	Road crosses bicycle path and curves sharply to the right.
3.2	0.3	Cross Balltown Rd. (Rt. 146) and enter Williams St. Immediately turn left into small parking lot by bridge.
		<p>Stop 1. Rexford Bridge. No hammers! On the cliff across the Mohawk River you can see an excellent exposure of undeformed autochthonous Schenectady Formation, which is composed of interbedded gray shale and gray sandstone (graywacke). The sandstones are turbidite beds that are made of sediment derived from erosion of the Taconian island arc, and possibly associated emergent accretionary wedge and forearc material (see Kidd et al., this volume). Flute casts and other current indicators in eastern New York State indicate a generally northward turbidity current transport direction, suggesting deposition in a north-south trending, northward-deepening sedimentary basin. Garver et al. (in press) has shown that there is a dramatic increase in Cr and Ni concentrations part way up the stratigraphic section in these rocks in New York State and similar rocks in Quebec and Newfoundland. They interpret this increase as resulting from erosional unroofing of mafic and ultramafic ophiolitic rocks that were exposed in the forearc region during the arc-continent collision. Ophiolites are still exposed in western Newfoundland and eastern Quebec, but, except for small ultramafic pods, have been eroded away in western New England. Unstable mineral grains and rock fragments in these gray sandstones have decomposed to form illite, chlorite, and other diagenetic minerals. The mineralogy of these rocks is dominated by quartz, illite, chlorite, and kaolinite. Leave the parking lot and return to the traffic lights at Balltown Rd, and turn left.</p>
4.7	1.5	At the next set of lights turn left (east) onto River Rd. On the skyline to the east you can see the high Taconics.
5.4	0.7	Enter the traffic circle near General Electric, go halfway around the circle and continue east on River Rd.
8.0	2.6	At T-intersection turn left onto Rosendale Rd.
8.4	0.4	As the road makes a sharp right at the bottom of the hill, turn left onto Waterplant Rd., with signs toward Lock 7. Watch out for traffic from the right!
9.0	0.6	Cross bike path at stop sign, continue to the left.
9.2	0.2	Pull off to right after small roadcuts on right.

Stop 2. Lock 7 access road. This small roadcut has allochthonous "Schenectady Formation" that has been deformed. The thrust zone along which these rocks have been transported extends approximately 500 m to the west. Notice that these are still sedimentary rocks and have not been metamorphosed significantly. Graded beds and sole marks in some sandstones indicate topping direction. Note that this outcrop is about 6 km west of a klippe of Austin Glen Formation, the westernmost mapped klippe in the Schenectady area (Figure 1). This and surrounding outcrops probably represent a poorly exposed thrust slice that was transported a relatively short distance. Continue straight ahead.



U Unmetamorphosed Mesozoic rocks of the Connecticut Valley basin

U Unmetamorphosed Paleozoic rocks

U Late pC rocks in the Pelham dome

U Middle Proterozoic rocks of Grenville age

Ordovician metamorphic zones

Oc Chlorite zone

Ocd Chloritoid zone

Ob Biotite zone

Og Garnet zone

Oks Kyanite-staurolite zone

Osm Sillimanite-muscovite zone

Ordovician metamorphism dominant ←→ Devonian metamorphism dominant

Devonian metamorphic zones

Dc Chlorite zone

Db Biotite zone

Dg Garnet zone

Das Andalusite-staurolite zone

Dks Kyanite-staurolite zone

Dss Sillimanite-staurolite zone

Dsm Sillimanite-muscovite zone

Dsmo Sillimanite-muscovite-orthoclase zone

Dso Sillimanite-orthoclase zone

Dsgc Sillimanite-garnet-cordierite zone

0 5 10 15 20 25 30 km
0 5 10 15 20 Miles

Figure 5. Generalized metamorphic map of eastern New York, western Massachusetts, and parts of adjoining states, showing the field trip route and trip stop locations (adapted from Zen, 1983; Ratcliffe et al., 1988). The metamorphic zones are based on common mineral assemblages in pelitic rocks. Sillimanite-muscovite grade rocks west of the Connecticut Valley Basin may include rocks of sillimanite-staurolite and sillimanite-muscovite-orthoclase grades. Ordovician metamorphic grade increases from west to east, up to the sillimanite-muscovite zone. Farther to the east, Devonian metamorphism partially or wholly overprints the earlier mineralogy and fabric. Devonian metamorphism reaches the lower granulite facies in the sillimanite-garnet-cordierite zone.

- 9.4 0.2 Bear left at the 'Y' intersection and turn around in the parking lot. Backtrack to Rosendale Rd.
- 10.3 0.9 Stop sign at Rosendale Rd. Turn left to continue east on Rosendale Rd.
- 12.1 1.8 Turn left onto Old River Rd. Watch out for cars from the right! Rosendale Rd. continues right up a steep hill. If you encounter Rt. 7, backtrack down the hill and take the correct turn.
- 14.3 2.2 Old River Rd. meets a T-intersection. Turn right (southeast) onto Forts Ferry Rd.
- 15.3 1.0 Turn left (north) onto Sparrowbush Rd. Watch out for approaching traffic! Immediately bear right (east) at the 'Y' intersection to continue on Sparrowbush Rd.
- 16.4 1.1 Cross the I-87 overpass.
- 16.7 0.3 Traffic lights at intersection with Rt. 9. Go straight across the intersection to take Rt. 7 east toward Troy.
- 17.4 0.7 Outcrops to left and right are Ordovician Austin Glen Formation. These weakly metamorphosed rocks have a strong east-dipping foliation, contain numerous small thrust faults, and is quite inhomogeneous, giving the outcrop a mélange-like appearance.
- 18.0 0.6 Low Taconics visible in the distance ahead.
- 20.7 2.7 Cross the Hudson River. Get in the left-most lane. After the traffic lights enter the middle lane and continue straight up the hill on Rt. 7 east, Hoosic St.
- 24.8 4.1 Junction of Rt. 142 with Rt. 7. Continue straight on Rt. 7 east.
- 25.8 1.0 Turn right (southeast) off Rt. 7 onto Rt. 278, toward Rt. 2.
- 27.4 1.6 Rt. 278 ends at a T-intersection, junction with Rt. 2. Turn left (east) onto Rt. 2.
- 35.1 7.7 Town center of Grafton, NY.
- 37.3 2.2 High Taconics visible on the skyline ahead.
- 41.2 3.9 Town center of Petersburg, NY.
- 46.5 5.3 Taconic crest, Petersburg Pass. Turn right into the dirt parking lot.

Stop 3. Taconic crest. These highly deformed rocks of the Nassau Formation had a shaley deep water facies protolith. Now they are chloritoid-grade phyllites with abundant deformed quartz veins (I have not found chloritoid in this outcrop). The quartz veins occur in several generations, as can be discerned by different degrees of deformation and crosscutting relationships. The phyllites contain the mineral assemblage quartz-white mica (muscovite and paragonite)-chlorite-carbonate (probably ankerite). The large rusty cavities in the quartz veins are weathered carbonate. Although two or three fold generations are apparent in the outcrop, the general dip of the schistosity is east, typical of the entire Taconian accretionary wedge system. Shear sense indicators suggest a top-to-the-west transport direction. Notice that the dominant foliation has been folded by a steeply east-dipping crenulation cleavage. Leave the parking lot and turn right, continuing east on Rt. 2 into Massachusetts.

- 50.7 4.2 At bottom of the valley you encounter a T-intersection with Massachusetts State Rt. 7. Turn left (north) to continue on Rt. 2.

- 52.0 1.3 Enter Williamstown. Follow signs around a small square to continue on Rt. 2 east into town.
- 53.7 1.7 At the first set of traffic lights since the Taconic crest turn left (north) onto Cole Rd.
- 54.4 0.7 Cross the Hoosic River.
- 54.5 0.1 At the T-intersection, turn right (east) onto Bridge Rd.
- 54.9 0.4 Turn left onto Pine Cobble Rd.
- 55.1 0.2 Turn left into little the parking lot and park. This hike will be to Pine Cobble on the Long Trail. The hike is 3.2 miles round trip and 800' of climbing. It will take about 2.5 hours. Bring water, a snack, and appropriate clothing.

Stop 4. Pine Cobble. **No hammers!** This outcrop of Cheshire Quartzite is located on the southern tip of the Green Mountains, and is equivalent to the Potsdam Formation in New York State. This is the basal Cambrian clastic unit that overlies older rocks, including Grenville age metamorphic basement rocks, of the Green Mountains and Berkshires. The quartzite has a weak foliation defined by muscovite that dips gently east. A set of steeply east-dipping joints may be a fracture cleavage, because the joints have an orientation similar to the crenulation cleavage evident in the more ductile rocks at Stops 3 and 7, and are approximately parallel to the antiformal axial surface of the Green Mountains. The Taconic Mountains are visible to the west, Mt. Graylock to the south, and the Berkshires to the east. The mountains are held up by allochthonous metamorphic rocks that were thrust-transported westward in the Taconian accretionary wedge. The valley is floored by Cambrian and Lower Ordovician carbonate and clastic continental shelf rocks that stratigraphically overlie the Cheshire Quartzite of this stop. The Cheshire and underlying rocks, which is part of a broad anticlinal structure that makes up the Green Mountains, plunges southward beneath the valley. The rocks in the valley floor are also allochthonous, having been transported piggyback-style on the Precambrian-based block of the Green Mountains. Hike back down, leave the parking lot and backtrack down Pine Cobble Rd.

- 55.2 0.1 At the stop sign turn right (west) onto Bridge Rd.
- 55.6 0.4 Turn left (south) back onto Cole Rd.
- 56.4 0.8 At the traffic lights, turn left to continue on Rt. 2 east. Continue through Williamstown and North Adams on Rt. 2. Highlands to the left (north) are the southern end of the Green Mtns.
- 62.2 5.8 Rt. 2 goes up a hill while sharply turning right. On the turn take the left onto Rt. 8 north toward Clarkesville and Stanford Vt.
- 62.7 0.5 Turn left (west) onto McCaully Rd. to Natural Bridge State Park.
- 62.8 0.1 Cross stream and take the dirt road to the right. Park in small parking lot on left before the gate, if possible. Otherwise continue up road to park and pay the \$2.00 parking fee.

Stop 5. Natural Bridge State Park. **No hammers!** These are rocks of the Cambrian to Ordovician Stockbridge Formation. This unit represents North American Cambrian or Lower Ordovician continental shelf carbonate rocks that were deformed and transported westward along Taconian thrusts. These have themselves been overridden by Taconian thrust sheets, including the Berkshire block immediately to the east. Backtrack to Rt. 8.

- 63.1 0.3 At the intersection with Rt. 8, turn left (north), continuing on Rt. 8.

63.3 0.2 Pull off the road to right after roadcut.

Stop 6. Road cut on Rt. 8. This rock is composed of relatively low-grade metamorphosed calcareous sandstone in the Ordovician Walloomsack Formation. This unit was also deposited on the North American continental margin. Although the rock is folded, sedimentary structures such as cross bedding can still be found in some places. The folds are best outlined by carbonate-rich layers in which the carbonate weathers out to form rusty pits. Turn vehicles around and backtrack south on Rt. 8 toward Rt. 2.

64.1 0.8 Intersection of Rt. 8 with Rt. 2. Turn left to continue on Rt. 2 east.

64.7 0.6 Climb up the Berkshire front.

67.0 2.3 Hairpin turn. Continue up the hill.

67.9 0.9 At the top of the hill turn right into the Wigwam and Western Summit Gift Shop parking lot. Excellent views of the Williamstown-North Adams valley with the Taconics, Mt. Graylock massif, Green Mtns., and the carbonate-floored valley.

Stop 7. Berkshires "western summit". This outcrop of Hoosac Formation contains biotite grade metamorphosed clastic sediments similar in composition to those seen at the Taconic crest. At these higher grades, biotite is the dominant mafic mineral rather than chlorite. Gray albite porphyroblasts up to 3 mm across are very common, and have grown to replace paragonite as the principal Na-bearing phase. Although the foliation generally dips east, there are numerous folds visible. Most are late, broad, open, folds having a prominent steeply east-dipping crenulation cleavage similar to that at Stop 3. More rare are isoclinal folds that indicate a top-to-the-west shear sense in the plane of the dominant foliation. This rock overlies Precambrian rocks in the Berkshire massif, and so it is part of a block of North American basement that has been transported westward over the shelf sediments seen at Stops 5 and 6. Leave parking lot and turn right (east) to continue on Rt. 2.

70.5 2.6 Pass by the Witcomb Summit Resort Inn. Rt. 2 starts descending the eastern slope of the Berkshires.

71.7 1.2 Turn left into the Eastern Summit Gift Shop parking lot. Excellent view of the eastern slope of the Berkshires, and the Bear Swamp pumped storage hydroelectric reservoir.

Stop 8. Eastern Summit Gift Shop. **No hammers!** This glacially striated pavement outcrop of Rowe Formation contains garnet grade phyllite (but no garnets) that has a composition richer in Fe and Mg and poorer in K, Na, and Si than the schists and phyllites of Stops 3 and 6. This more mafic composition is probably caused by a large volcanoclastic component in the sediments, presumably derived from erosion of volcanics from the Taconian island arc. The dominant minerals are chlorite, white mica, quartz, and octahedral crystals of magnetite. The patchy appearance of the outcrop is caused by mechanical mixing of a variety of slightly different sedimentary lithologies, probably when the rock was a soft sediment. On a large scale, the Rowe Formation has been characterized as a *mélange*, and includes basaltic and ultramafic rocks. Leave the parking lot and turn left (south) down the hill on Rt. 2 east. Units passed on the steep, winding descent of the Berkshires include the Rowe and Hoosac Formations.

79.3 7.6 Cross bridge over railroad tracks.

80.0 0.7 Cross bridge over the Deerfield River.

80.1 0.1 Take left immediately after the bridge, following the sign toward Rowe and Monroe.

80.3 0.2 Outcrops on right are Hawley Formation.

- 80.4 0.1 Turn vehicles around in small pulloff on the right. Backtrack toward Rt. 2 about 100 m.
- 80.5 0.1 Pull off onto the shoulder to the right.

Stop 9. Hawley Formation. The Hawley Formation contains large quantities of volcanic and intrusive rocks along with the volcanoclastic sediments. The igneous rocks include mafic and felsic varieties. The mafic rocks have compositions similar to basalts, and seem to include mid-ocean ridge, back-arc basin, and boninitic, and calc-alkaline varieties. These metamorphosed igneous rocks are related to the Taconian island arc, but their diverse lithology, compositions, and ages indicate a complex history for the island arc system in which they were emplaced. The rocks at this outcrop are amphibolite grade carbonate- and hornblende-bearing chlorite-muscovite schists. They contain abundant acicular hornblende up to 20 cm long in some areas, although they are typically 1 to 3 cm long in this outcrop. Euhedral carbonate rhombs (ankerite?) can be seen weathering out on outcrop surfaces to form rhombohedral pits. On the steep slope on the river side of the road there are some nice exposures of coticule (fine-grained pink garnet quartzite). This unusual rock type is thought to represent metamorphosed Mn- or Fe-rich chert that formed from silica derived from the weathering or hydrothermal alteration of volcanic rocks. Continue down the hill toward Rt. 2.

- 80.8 0.3 Stop sign at the intersection with Rt. 2. Turn left (east) onto Rt. 2 east toward Charlemont.
- 82.8 2.0 Town center of Charlemont, MA.
- 85.7 2.9 Good outcrops of garnet-rich Devonian Goshen Formation schist on the north shore of the Deerfield River to the right (south).
- 89.6 3.9 Outcrops are calc-alkaline felsic gneisses and related amphibolites in core rocks of the Shelburn Falls dome (roots of the Taconian island arc).
- 90.4 0.8 Turn right (south) onto Rt. 2A toward Shelburn Falls.
- 91.2 0.8 In the center of Shelburn Falls turn left (east) onto the steel bridge across the Deerfield River.
- 91.3 0.1 Small access road to the right (south) soon after the bridge. Do not drive down this road, but find a place to park after the bridge. On foot, follow signs on the small access road to "glacial potholes". Find the steps going down to the river behind the gift shop. Do not go down to the rocks if the gate is locked, or if water is flowing strongly over the rocks.

Stop 10. Shelburn Falls. **No hammers!** This exposure is an excellent example of the core gneisses of the Shelburn Falls dome. This is an Acadian (Devonian) structural dome that exposes a variety of plutonic-looking calc-alkaline gneisses and amphibolites. These rocks have generally been interpreted to be part of the plutonic roots to the Taconian island arc. Igneous features that can be seen in this outcrop include crosscutting relationships, dikes, xenoliths, intrusion breccia, and late-stage quartz veins and faults. These rocks are similar to, but less deformed than, the Ordovician gneisses exposed in the cores of domes in the Bronson Hill anticlinorium, farther to the east. The precise relationship of the Shelburn Falls rocks to those in the Bronson Hill anticlinorium is somewhat problematic. Although the two are lithologically and chemically similar, a preliminary age date on the Shelburn Falls rocks is substantially older (Paul Karabinos, oral communication, 1993) than the 443-453 MY age of the calc-alkaline gneisses in the Bronson Hill anticlinorium (Tucker and Robinson, 1990). Other features that can be seen here include wonderful large potholes that were formed by the Deerfield River when it was swollen with glacial melt water. Continue in the same direction away from the bridge (east on Rt. 2A). Do not backtrack across the bridge.

- 91.7 0.4 Turn right (southeast) following sign for Rt. 2A toward Greenfield and Boston.
- 91.9 0.2 Intersection with Rt. 2. Turn right (east) onto Rt. 2.
- 94.3 2.4 Road climbs out of the topographic depression of the Shelburn Falls dome. Hills bounding the dome here are kyanite-staurolite grade Goshen Formation schists that are more resistant to erosion than the calc-alkaline gneisses in the dome interior.
- 99.4 5.1 Excellent view of the Deerfield basin, the northernmost sizable portion of the Mesozoic rocks of the Connecticut River Valley. The N-S trending ridge in the middle of the basin is held up by the Deerfield Diabase, probably correlative with the Holyoke Basalt farther south. The large mountain to the southeast is Mt. Toby.
- 100.5 1.1 Enter the traffic rotary in Greenfield, which circles underneath I-91. Continue 3/4 of the way around the rotary.
- 100.7 0.2 Enter the on ramp for I-91 north, which is also Rt. 2 east.
- 102.5 1.8 Excellent exposures of fluvial facies of the Mesozoic red beds of the Connecticut Valley basin.
- 103.1 0.6 Take Exit 27 (the first exit) to continue on Rt. 2 east.
- 105.3 2.2 At the next traffic lights, bear left to continue on Rt. 2 east.
- 106.0 0.7 At the bottom of the hill, cross the Falls River. Immediately after the river turn right into the parking lot.

Stop 11. Turner's Falls. Although it has little to do with the Taconian orogen, this wonderful set of outcrops has excellent exposures of Mesozoic basaltic lava and terrestrial sediments that formed in a rift valley. The outcrops near the parking lot are of tholeiitic Deerfield basalt. The flow base, in contact with red beds, is accessible on the south side of the road to the west of the parking lot. The flow base has pillow lavas that formed as the lava flowed into a saline rift valley lake. The flow top and overlying red beds are visible in the roadcut across the road from the eastern end of the parking lot. Several faint, coarse-grained layers sub-parallel to the flow top are visible in the central part of the roadcut. These are sill-like segregations of residual magmatic liquid that separated from the crystal mush of the compacting, solidifying flow (described for a different locality by McHone and Philpotts, 1995; Philpotts et al., 1995). The outcrops near the river, accessible by foot path from the parking lot, have excellent exposures of the vesicular lava flow top, and one of the coarse-grained sill-like layers. The flow is overlain by fluvial conglomerate, arkose, siltstone, and shale. The fluvial sediments have excellent depositional and soft sediment features including bed forms, mud cracks, trace fossils, raindrop imprints, shale rip up clasts, and filled channels. Near the dam the sediments grade into finer-grained gray and black lake sediments in which fish and plant fossils can be found. Leave the parking lot and turn right to continue on Rt. 2 east. For excellent introductions to the geology of the Connecticut Valley basin, see Hubert et al. (1978) and Horne et al. (1995).

- 108.5 2.5 Small road cuts on both sides of the road are composed of very coarse-grained matrix-supported conglomerate. This outcrop represents the alluvial fan facies of the Mesozoic red beds of the Connecticut River Valley. The alluvial fans were principally built on the east side of the basin, below the mountains on the east side of the border fault along which the basin subsided.
- 109.4 0.9 Cross the French King Bridge over the Connecticut River. This is the approximate location of the normal fault that forms the eastern margin to the Connecticut River Valley rift basin.

111.6 2.2 Large road cuts on either side of the road. Pull off road to right and stop.

Stop 12. Millers Falls. This large roadcut is in the biotite member of the Dry Hill Gneiss, a Late Precambrian (Tucker and Robinson, 1990) unit exposed in the core of the Pelham dome, one of the structural domes in the Bronson Hill anticlinorium. This unit is interpreted to be a thick series of metamorphosed alkalic volcanics that probably erupted in a continental rifting environment (Hodgkins, 1985). The relationship of these rocks to surrounding rocks is unclear, but they are unrelated to the Taconian island arc except to the extent that they are exposed in this dome in the Bronson Hill anticlinorium. The Dry Hill Gneiss is overlain, perhaps along a fault, by felsic Fourmile Gneiss, which is part of the Taconian igneous suite. The Dry Hill Gneiss is a moderately well foliated, layered granitic gneiss that contains the assemblage quartz-plagioclase-microcline-biotite±amphibole (hornblende or hastingsite), with accessory sphene, magnetite, apatite, zircon, and allanite. The allanite crystals are large and are visible as dull tan-colored crystals in hand specimen. The large K-feldspar megacrysts in the gneiss were probably derived from severely deformed and dismembered pegmatites. This texture indicates the great strength of feldspar compared to the quartz- and biotite-rich matrix during extreme solid-state deformation. Continue east on Rt. 2.

116.1 4.5 Turn left onto Mountain Rd., in Erving, at the cemetery on the side of the hill to the left.

119.0 2.9 Intersection with South Mountain Rd. Turn right up the hill.

119.3 0.3 Pull off to the right at the top of the hill and park. Follow white blazes along the private driveway. At the double blaze, turn right into the woods. Follow the trail to the quartzite ledges after about 1/4 mile.

Stop 13. Crag Mountain. **No hammers!** This prominence is composed of thickened (by folding) Silurian Clough Quartzite. In this locality it is a quartz pebble and cobble conglomerate metamorphosed to kyanite-staurolite grade. Pelitic schists in the area can contain the assemblage quartz-plagioclase-garnet-biotite-muscovite-staurolite-kyanite. The Clough Quartzite was deposited on an extensive Silurian erosion surface in western New England, and it rests on a variety of Ordovician and older rocks. The quartzite has been deformed so that quartz cobbles and pebbles have been flattened and extended to different degrees, visible in outcrop. The milky quartz is full of fluid inclusions, suggesting that it was originally metamorphic vein quartz. Silurian erosion of Taconian low- and medium-grade metamorphic rocks in the Taconics, Green Mountains, and Berkshires may have been the source for the necessary large quantities of high-purity vein quartz. End of trip!

TO RETURN

Turn around and head down the hill (west) down South Mountain Rd..

Turn left (east) onto Mountain Rd.

Stop sign at the intersection of Mountain Rd. with Rt. 2. Turn right (west) onto Rt. 2.

Follow Rt. 2 back into New York State. Continue on Rt. 2 through Troy and into Latham. Don't take Rt. 278 and 7 into Troy.

Cross the Hudson River. Continue straight on Rt. 2 west.

In Latham, NY, Rt. 2 enters a traffic circle where the road crosses over Rt. 9. Continue halfway around the circle to continue west on Rt. 2 toward Schenectady.

Rt. 2 turns into Rt. 7. Continue straight on Rt. 7 (west) toward Schenectady.

In Niskayuna, Rt. 7 bears left. Bear right (west) onto Union St. toward downtown Schenectady and Union College.

Traffic lights at Union Ave. Union College visible on the right.

Traffic lights at Nott St. Stay in the right hand lane.

Traffic lights at Seward Place. Turn right (north) onto Seward Place.

Traffic lights at Nott St. Turn right (east) onto Nott St.

Entrance right to main (Van Vranken Ave.) Union College parking lot.

DEVONIAN CARBONATES AND ECONOMIC RESOURCES: THE BLUE CIRCLE AND CALLANAN QUARRIES

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CRUSHED STONE AND CEMENT PRODUCTION - GENERAL CONSIDERATIONS

The production of crushed stone in New York was 33,500,000 tons 1994 with a value of \$196 million. Most of the crushed stone produced in New York is limestone and dolostone with traprock (diabase) and various metamorphic rocks making up the bulk of the remainder. Cement manufacturing amounted to 2,925,000 tons valued at slightly over \$149 million. Taken together, crushed stone and cement account for 40% of the value of mineral produced in New York. There is an order of magnitude difference in the price of these two commodities with crushed stone trading for an average of \$5.85/ton and cement valued at an average of \$50.94/ton. This price difference leads directly to the contrast in mining operations that will be examined. On this field trip, we will visit two quarries operating in the rocks of the Helderbergs. One of these produces crushed stone for construction aggregate and the other produces cement. Although both companies are working in the same stratigraphic section with similar equipment and mining plans, the end use of the rock drives the companies to use very different geologic units within the overall stratigraphic framework. The following brief discussion outlines the processes and constraints on the construction aggregate and cement industries.

Aggregates

Construction aggregates are hard, inert materials suitable for being formed into a stable mass by the addition of cementing materials to produce concrete (portland or bituminous), or by compaction or natural weight to produce a road base or foundation fill. Conditions necessary for a rock deposit to be developed for construction crushed stone and much of the following discussion are given by Herrick (1994) to include:

- Quality - passes specifications for strength and durability;
- Quantity - adequate volume of rock is present to support a production life of 10-20 years;
- Market - must have an adequate market to sustain the costs of a new operation;
- Transportation - costs must be competitive for the intended market;
- Environmental - impacts of mining and attendant operations must be within acceptable limits;
- Permitability - all operations associated with the mine must be permitted by one or more governmental agencies.

Specific chemical and physical properties control the use of a particular rock for crushed stone. Chemically, the rock should be inert and should not change chemically in use. However some rocks contain minerals that are reactive in portland cement, bituminous concrete and other environments. Rocks containing silica in the form of glass, chalcedony, opal, chert, and finely divided quartz may react with high alkali cement to form a gel which, due to the increased volume of the gel, caused deterioration of the concrete. Moderate to high clay content of dolomitic limestones may also react with high-alkali cement. Expansion of the cement appears to be caused by microfracturing throughout the cement and aggregate as well as in reaction rims around the carbonate rock particles (Rogers, 1979). Sulfides such as pyrite, marcasite, and pyrrhotite react with water to form iron hydroxides and sulfates leading to discoloration and weakening of concrete.

In Garver, J.I., and Smith, J.A. (editors), Field Trips for the 67th annual meeting of the New York State Geological Association, Union College, Schenectady NY, 1995, p. 163-171.

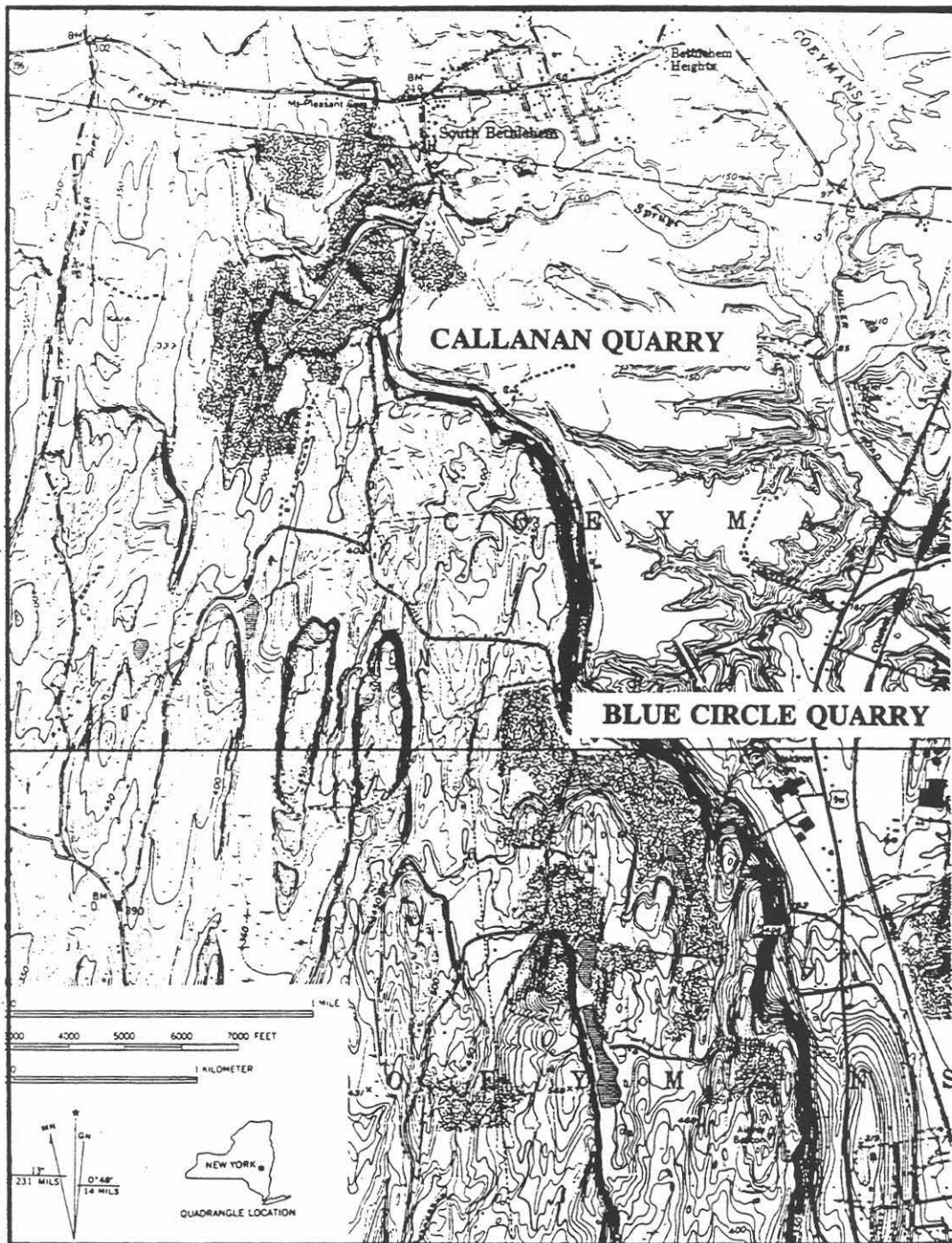


Figure 1. Location of Callanan Industries and Blue Circle Industries quarries.

In bituminous mixes, certain rocks present a problem when the bituminous film separates (strips) from the aggregate. Stripping is related to the electrical charge on the surface of the rock particle because surfaces with a negative charge may attract water and thus promote lack of adhesion of the bituminous film. Rocks with a high quartz content, such as quartzites and some granites, gneisses and schists, may present stripping problems.

Physical properties critical for use of a given rock as construction aggregate are strength and durability, porosity and pore size, and volume integrity - that is, maintaining constant volume when subjected to variable moisture or freeze/thaw conditions. Another important property is the tendency of the rock to break into relatively equant fragments. Large amounts of platy or flat fragments, such as are derived from slate, shale and some schists are not acceptable. Coarsely crystalline igneous rocks, quartzite, or marble may be excessively brittle and too easily shattered. The presence of certain minerals such as olivine or metamorphic amphiboles may create a weak, brittle rock.

Cement

The discussion that follows is adapted from (Ames and Cutcliff, 1983). Although there is much confusion between the terms cement and concrete, cement manufacture is the processing of selected and prepared raw materials into a synthetic mixture, called clinker, that can be ground into a powder having a specific chemical composition and the physical properties of cement. Cement manufacture is accomplished in a number of steps which may vary from plant to plant but it is characterized by a key pyroprocessing (burning) step which brings about the necessary changes in the raw materials. This burning step results in a chemical change. Concrete is a combination of cement, aggregate, and water. Cement, then, is one of the raw materials of concrete.

The primary requirement for making cement is a source of lime (CaO) which is generally available as calcium carbonate in limestone or some close relative thereof. However, other sources of lime have been used including shell, aragonitic sand, slag, anhydrite and feldspar. Secondary raw materials are needed to supply the other chemical components of cement, specifically silica, alumina, and iron. Although the sources of these components may be diverse, the ratios of the chemical constituents must be carefully controlled. Typical sources of these components are sand, silt and clay and their corresponding rocks types. Manufacturing wastes and ash may also be used. In addition to the raw materials listed above, a source of SO₃ is required to control setting times for the concrete made with the cement. Generally, the addition of gypsum is mandatory. The considerable flexibility in selection of raw materials to be blended into acceptable kiln feed lies in the fact that cement making is basically a chemical process. Table 1 lists the raw material used for cement in the United States.

Raw materials and fuels in the cement kiln introduce components other than those sought. Within specified limits, these components are tolerable, even beneficial. However, if present above certain defined levels, these may constitute deleterious impurities. Magnesium compounds are the most familiar and common of these. At low levels, generally under four percent, MgO acts as a fluxing agent and is regarded as innocuous. But at higher levels, magnesium compounds that form cause expansion and ultimately results in disruption of the concrete.

The production of cement involves a large amount of mining activity. Generally, open pit methods are used to produce raw cement materials. All activities associated with crushed stone mining are employed by cement plants including stripping, drilling, blasting and breaking, loading, transportation, and reclamation. The raw materials are then milled and blended. The objective of milling is to prepare sizes and mixtures of raw materials for proper kiln feed. Grinding can be done wet or dry. At Blue Circle, wet milling produces a slurry of ground kiln feed in which water content has been kept as close as possible to the minimum (35-40%) that can be pumped and handled.

Table 1 . Raw material used in making cement clinker in the United States.

<u>Calcium carbonate sources</u>	<u>Alumina sources</u>
limestone, including	shale
lithified limestone	clay and mud
chalk	loess
marble	slag
marl	fly ash

oyster shell reef
coquina
aragonite sand
slag

bauxite
alumina processing waste
staurolite
mill fines (granite)

Silica sources

sand, sandstone, quartzite
clay and claystone
shale
loess
mill fines
fly ash

Iron sources

iron ore
blast furnace flue dust
pyrite cinder
mill scale
fly ash

Pyroprocessing (burning) is the key process in cement manufacture. Burning at high temperature causes the raw materials to react and combine to produce *clinker* - a balanced mixture of synthetic "minerals" and glasses that can be ground into cement. The process is carried out in a rotary kiln.

Rotary kilns are steel cylinders about 25 feet in diameter and up to 760 feet long. They are lined with refractory material, usually brick, which are designed to develop a coating of raw materials so that the finished materials are processed over similar material. Kilns are inclined slightly so that their rotation (50-90 rph) moves the materials from the feed end to the discharge end at the desired rate. Retention time in the kiln is several hours although the burning zone occupies only about 15% of the kiln and burning temperature is much more important than the travel time of the raw material.

The kiln is fueled (with powdered coal, oil or gas) under pressure through a burner pipe positioned at the discharge end and the flame extends well up into the kiln. As the materials move from one end of kiln to the other, they pass through a series of stages. These stages include heating to lose water, calcination of the carbonates (i.e. driving off CO₂) and the fusion, melting and reacting to form clinker. The burning fuel is regulated so that the hottest part of the kiln is 2600° to 3000° F. The kiln zone in which the chemical combination of clinker compounds (Table 2) occurs begins 40 to 50 feet from the discharge end. Clinker leaves the kiln as sand-sized to golf ball-sized rounded particles that are cooled in a variety of ways on their way to finish grinding.

The clinker is relatively unreactive at this point but clinker that is finish-ground hot or old clinker that has become hydrated each produce poorer quality cement. At this point in the manufacturing process, the SO₃ content of the cement must be adjusted. This is critical in providing the desired setting time for the concrete made with the cement. Gypsum (3-6%) is traditionally interground with the clinker to provide the SO₃. Additions of air-entraining agents and other ingredients are metered in at this time. The cement is now ready for bagging or shipment in bulk via rail, truck or, in the case of Blue Circle, by water on self-unloading barges.

Table 2. Cement compounds (adapted from Clausen, 1960)

<u>Compound</u>	<u>Composition</u>	<u>Percent content in Type I cement</u>
Tricalcium silicate	Ca ₃ SiO ₅	45
Dicalcium silicate	Ca ₂ SiO ₄	27
Tricalcium aluminate	Ca ₃ Al ₂ O ₆	11
Tetracalcium- aluminoferrite*	Ca ₄ Al ₂ Fe ₂ O ₁₀	8

* The composition of the Fe-bearing phase is an approximation and may range from Ca₂Fe₂O₅ to Ca₆Al₄Fe₂O₁₅.

REGIONAL GEOLOGIC SETTING

The two quarries to be visited on this trip lie within the Helderberg plateau, which rises on the west side of the Hudson Valley. Westward, the Catskill Mountains rise above the plateau to an elevation of roughly 4000 feet. East of the Plateau are the Hudson lowlands and the Taconic uplands. The Hudson River flows southward, at tidal level, a few miles east of the quarries.

The rock units in the mid-Hudson Valley region range in age from Ordovician (≈ 450 my) to Middle Devonian (≈ 380 my). The rocks are mainly carbonates and shales with two thin clastic units. Within this sequence is the Lower Devonian Helderberg Group which, in this part of New York is represented by five of the seven Helderberg units. These are, from oldest to youngest, the Manlius, Coeymans, Kalkberg, New Scotland, and Becraft. The youngest units, the Alsen and Port Ewen, are not present here. Figure 2 is a generalized stratigraphic column of the Helderberg rocks and Figure 3 is a diagrammatic cross section of Lower Devonian formations along the outcrop belt. Regionally, the Helderberg rocks are exposed from Cayuga Lake in central New York eastward to the so-called Helderberg Escarpment southwest of Albany and thence southward and southwestward through the Hudson Valley, culminating in the vicinity of Port Jervis.

The rocks in which the quarries are developed lie between the highly deformed rocks of the Taconics and the much less disturbed rocks of the Catskill highlands. Structurally, the rocks in the quarries have been broadly folded with wavelengths of 900-1000 feet and amplitudes of 50 feet. Fold axes generally plunge southward although an anticline in the southernmost face of the Callanan Quarry is nearly horizontal. The folds are broken by several thrust faults that dip 20-30 degrees to the east. Offset on these faults varies from 50 to 200 feet.

FORMATION DESCRIPTIONS

Much of the following discussion is paraphrased from Banino and Brown (1978).

Manlius Formation

The Manlius Formation has a mixed carbonate lithology, ranging from a dark-gray, thin-bedded, shaly limestone to argillaceous, silty, laminated dolomite. The formation has been subdivided into the four members in the Mohawk Valley (Thatcher, Olney, Elmwood and Clark Reservation) but only the Thatcher is present in the Hudson Valley (Rickard, 1962). The Manlius consists of interbedded dolomitic "ribbon" or "paper" limestones and pure massive to biostromal limestones. Fossils, particularly *Tentaculites gyracanthus*, are concentrated in the limestone beds. Primarily for the purposes of mining control in the Hudson Valley cement quarries, this unit has been divided into six units designated (youngest to oldest) M6-M1.

M6 is a transitional unit between the overlying Coeymans and the much darker gray and finer grained Manlius limestone. Typical M6 is a dark-gray, dense, fine-grained material. Bedding is generally two to 12 inches with local shaly partings.

M5 has been called the "paper" or "ribbon" bed of the Manlius. It is a dolomitic limestone that characteristically weathers to a light gray to tarnished green color.

M4 is a dark-gray, finely crystalline, massive limestone. Bedding is typically one to two feet. Near the bottom of M4, algal reef structures have been reported.

M3 is another dark gray, dense limestone. The unit has lamellar partings 1/8-1/4 inch thick containing black shale. Sparry calcite is common in the more massive crystalline beds.

M2 is a unit similar to M5.

M1, the lowest unit, is a dense, dark-gray limestone with interbedded shaly partings. The more crystalline beds increase in thickness towards the bottom of the unit. The lower beds contain cephalopods. The lower contact with the Roundout Formation is recognized by a color change from dark-gray to tan and an increase in Mg-carbonate in the Roundout.

Coeymans Formation

The Coeymans is a pure, bluish-gray, medium- to coarse-crystalline limestone that forms prominent ledges along the Helderberg Escarpment. Generally massive-bedded, individual beds are difficult to recognize. Crinoid stems, locally silicified, are common. The brachiopod *Gypidula coeymanensis* is a very common fossil in this formation and is used to identify the Manlius-Coeymans contact. The clean limestone of the Coeymans is thought to represent more open marine conditions relative to the depositional environment of the Manlius.

Helderberg Group

Formation	Rock Types, Grain Size, Sedimentary Structures	Fossils	Environments
Port Ewen	see New Scotland, below	see New Scotland, below	see New Scotland, below
Alsen	see Kalkberg, below	see Kalkberg, below	see Kalkberg, below
Becraft	see Coeymans, below	see Coeymans, below	see Coeymans, below
New Scotland	fine- to medium-grained limestone that contains clay shale that contains calcium carbonate thin to medium layers of uniform thickness	high number & variety of sea bottom dwellers	deepest water of the Helderberg Sea; below motion of fair-weather waves; bottom agitated by storm waves
Kalkberg	medium-grained limestone rich in clay & silica chert thin to medium layers	high number & variety: bryozoans brachiopods crinoids corals trilobites mollusks ostracodes	deeper water at or near lowest point reached by fair-weather waves bottom occasionally agitated
Coeymans	clean medium- to coarse-grained limestone scattered small coral and stromatoporoid reefs uneven, medium to thick layers cross-bedding	moderate number pelmatozoans corals brachiopods mollusks trilobites ostracodes	shallow water shelf vigorous wave motion well-agitated bottom
Upper Manlius	fine- to medium-grained limestone slightly uneven, medium to thick layers scour & fill, birdseye, ripple marks, cross-bedding	low to moderate number & variety stromatoporoids brachiopods mollusks ostracodes trilobites	shallow water near the shore & near low tide moderate wave motion-- protected by a barrier
Lower Manlius	fine-grained limestone & dolostone medium to thin layers; some laminations alternating layers of shale rich in carbonate sediments scour & fill, birdseye, desiccation cracks	low number & variety stromatolites oncolites ostracodes brachiopods gastropods tentaculites	between high & low tides and shallow water below low tide

Figure 2. Stratigraphy of the Helderberg group. Modified from Isachsen et al. (1991)

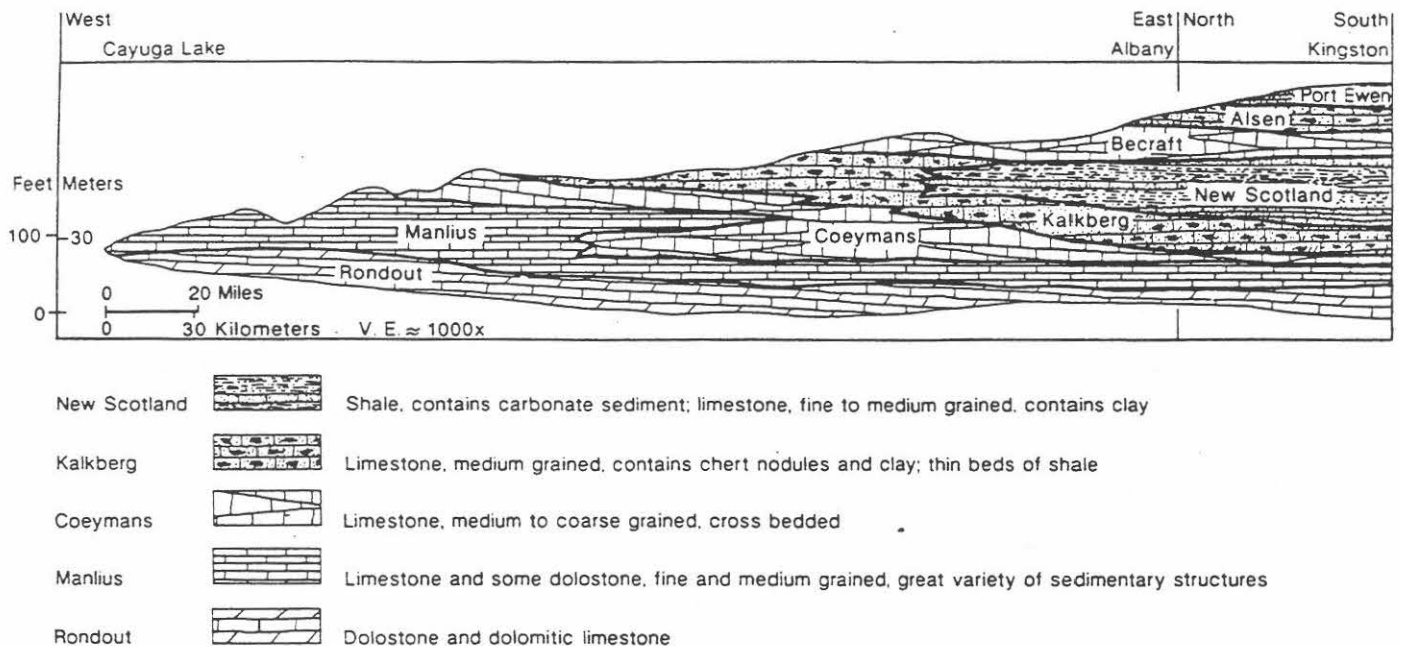


Figure 3. Diagrammatic cross section of Lower Devonian formations, east to west along the outcrop belt from central New York to the mid-Hudson Valley. Figure modified from Isachsen et al. (1991).

Kalkberg Formation

The Kalkberg ranges from a bluish-gray, chert-rich limestone near the base to gray, fine-grained, argillaceous limestone near the top. Lithologically, the formation has been divided into four members from oldest to youngest: the Lower and Upper Hannacroix and the Lower and Upper Broncks Lake. The environment of deposition of the Kalkberg is interpreted to be transitional between a shallow, wave-agitated sea and a deeper, quieter setting represented by the overlying New Scotland Formation.

The Lower Hannacroix is the dominant bluff-forming unit of the Helderberg Escarpment. It is recognized by the prominent black chert nodules and layers spaced about one foot apart, each with a thickness of about four inches. The rock is massive-bedded and is finer grained and darker than the Coeymans. The Upper Hannacroix is a fine-grained, fairly massive, gray limestone with anastomosing argillaceous partings which give a net-like appearance to weathered surfaces. This unit does not contain layers of chert but has numerous small nodules of black and dark gray chert. Except for the presence of chert, this unit is similar to the Coeymans although finer grained and less fossiliferous. The top of the unit is marked by the first appearance of a dark-gray, euxinic shale bed about two feet thick containing pyrite nodules and small brachiopods.

The Lower Broncks Lake is a fine-grained, bluish gray limestone with beds one to three feet thick interbedded with one to two inch calcareous shale beds. Fossils are abundant in this unit and encrusting bryozoans are common. The Upper Broncks Lake is fine-grained, bluish-gray limestone with beds from three to more than 12 inches and fewer shale layers than the underlying unit. One, two or three thin, black to dark gray chert layers occur near the base.

New Scotland Formation

The New Scotland Formation is composed of alternating medium-gray, very-fine-grained, impure limestone and dark-gray calcareous mudstone and siltstone with variable quantities of chert and pyrite. The mudstone and siltstone at the base grade upward into argillaceous and silty limestone. (The unit has, at times, been described as a calcareous shale.) Though a greater influx of mud occurred during the deposition of the New Scotland than during the deposition of older units, a richer fauna was present including sponges, corals, bryozoans, brachiopods, pelecypods, gastropods, and trilobites.

Becraft Formation

The Becraft Formation is a light-gray to pink, coarsely crystalline, biofragmental limestone. It is massive and bedding planes are often difficult to distinguish. Locally, gray chert occurs near the base and top. Informally two units are sometimes observed in the Becraft. The lower unit is shalier and has a lower lime and higher silica content than the upper. Abundant fossils occur in the Becraft, dominantly crinoids and brachiopods. The environment of deposition is interpreted to be a clean, clear sea.

SITE GEOLOGY

The Callanan Quarry is located primarily in the Town of Coeymans, south of NY Route 396 and the village of South Bethlehem (Fig. 1). The location can be found on the Delmar 7.5' quadrangle. The quarry is owned and operated by Callanan Industries, Inc. All correspondence concerning access should be directed to Mr. Charles A. Stokes, Senior Vice-President, Callanan Industries, One South Street, South Bethlehem, NY 12161. The Blue Circle Quarry is located in the town of Coeymans, east of NY Route 9W (Fig. 1). The location can be found on the Ravena 7.5' quadrangle. Correspondence concerning quarry access should be directed to Mr. Kevin Riley, Quarry Superintendent, Blue Circle Industries, Ravena, NY 12143.

The two quarries described herein are separated from each other by nine miles. Consequently the geology at the two sites is quite similar. Both quarries are developed in the lower and middle units of the Helderberg Group rocks. Here, the Manlius is 54 feet thick and subdivided as described above. Units M1, M3, and M6 are dark-gray, fine-grained to sublithographic, fine to medium bedded, subtidal limestone. Units M2 and M5 are light-gray, fine-grained, thin-bedded, supertidal dolomitic limestone. Unit M4 is medium to dark-gray, massive, intertidal, stromatoporoid limestone. The Coeymans is 27 feet thick, light-gray, medium-grained, massive, non-argillaceous, fossiliferous limestone. It is homogeneous and lacks marker beds.

The Kalkberg is 66 feet in total thickness. The Lower Hannacroix is 11-18 feet thick, medium-gray, medium-bedded limestone with interbedded dark gray to black chert nodules and chert beds one to four inches thick. The Upper Hannacroix is 10 feet thick, fine- to medium-grained limestone. This unit is similar to the Coeymans except that it is less fossiliferous and the fossils are somewhat smaller. The Lower and Upper Broncks Lake total 45 feet in thickness. They are fine- to medium-grained, medium-bedded, argillaceous limestones.

The New Scotland is 100 feet thick and is composed of medium to thick-bedded, argillaceous limestone similar in appearance to the Kalkberg Formation. The uppermost part of the New Scotland is transitional into the overlying Becraft. The Becraft is 40 feet thick, tan to grayish-white with greenish-gray shaly partings. It is a coarse-grained and highly fossiliferous limestone.

Several marker beds and stratigraphic horizons are visible in the quarries and can be used for stratigraphic location. The Coeymans-Kalkberg contact is a key horizon. This contact can be identified by the following characteristics: (1) on a weathered surface, the Coeymans is light-gray to buff and the Kalkberg is medium-gray and is less resistant to weathering; (2) a black, shaly layer, two to six inches thick, commonly occurs between the formations; (3) the Lower Hannacroix of the Kalkberg contains discontinuous bedded black chert.

The contact between the Kalkberg and New Scotland is marked by an increase in siliceous material in the New Scotland giving it a shaly appearance relative to the more massive Upper Broncks Lake Member of the Kalkberg.

The top of the New Scotland contains a transitional zone into the overlying Becraft Formation characterized by alternating beds of crystalline (Becraft lithology) and shaly beds (New Scotland lithology). Also notable at the contact are green shale bands of the Becraft as opposed to the black shale of the New Scotland.

PRODUCTION

The majority of the aggregate produced from the Callanan Quarry in South Bethlehem is used for New York State Department of Transportation (NYSDOT) road construction. Therefore, NYSDOT specifications are most crucial to Callanan's mining. High-friction aggregate, material with greater than 20% non-carbonate content, is used in top-coarse paving material in New York State. High-friction material commands the highest price, and, for this reason, is the most valuable to producers. Of the five formations mined at Callanan's quarry, only the New Scotland and Kalkberg are approved for high-friction use. As Callanan excavates the New Scotland and Kalkberg, most of the Manlius, Coeymans and Becraft is left behind. These formations are, however, mined for non-friction products, which include commercial uses.

A few miles south, the situation is startlingly different. Blue Circle is discarding in waste piles approximately 85% of the Kalkberg and nearly 100% of the New Scotland that is mined. Callanan's low-valued Manlius, Coeymans and Becraft formations command high value at Blue Circle's quarry. These three formations

have a high carbonate content which is extremely important in cement production. So much so that Blue Circle strips up to 160 feet of waste rock (rock which is valuable to Callanan) to get to the Manlius and Coeymans formations. A comparison of the cost per ton of aggregate versus cement in New York, cited above, illustrates the difference in value and explains Blue Circle's mining strategy. The much higher cost (and value) of cement is attributed to a number of factors including processing costs, cost of raw materials (limestone, iron, silica, alumina, gypsum) and market demand.

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ROAD LOG

NOTICE: Specific permission to enter the properties of Callanan Industries Inc. and Blue Circle Industries PLC has been granted to the New York State Geological Association by those companies. DO NOT attempt to revisit the stops on this trip without contacting these companies for permission. This road log begins at the intersection of Interstate Routes 87 and 90 near the Exit 24 intersection on the New York Thruway.

Start	Milage	Cumulative miles
Rt. I-87/I-90 intersection. Travel east on I-90	0	0
Turn south on I-787.	6	6
At the end of I-787, turn right on Rt. 9W (south) also called McCarty Avenue.	3.7	9.7
At third traffic light, Rt. 32 joins Rt. 9W, go straight.	2.9	12.6
Traffic light on Rts. 9W & 32, go straight (Rt. 32 diverges to right, fork to left, remain on Rt. 9W).	0.1	12.7
Traffic light at Feura Bush/Glenmont Rds., go straight on 9W.	1.5	14.2
Traffic light at Wemple Rd., go straight on 9W.	1.5	15.7
Blinking light at Rt. 55, go straight on 9W.	1.7	17.4
Turn right onto Rt. 396 at nest traffic light.	0.7	18.1
Enter town of South Bethlehem.	1.6	19.7
Turn left onto Rt. 101 (South Street).	1.0	20.7
Enter Callanan Industries mine on right.	1.1	21.8
Retrace route to Rt. 9W, turn right (south) on 9W.	4.1	25.9
Turn right onto Fuller Rd.	4.3	30.2
Turn left at "T" intersection.	0.2	30.4
Turn right, sharply uphill, into Blue Circle mine, to "Dead End" sign.	0.7	31.1

THE PALEOFLUVIAL RECORD OF GLACIAL LAKE IROQUOIS IN THE EASTERN MOHAWK VALLEY, NEW YORK

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ABSTRACT

During the final retreat of Laurentide ice from New York State, the Mohawk Valley played a major role in the routing of glacial meltwater. The Mohawk Valley contained the Iromohawk River, which drained Glacial Lake Iroquois, in the Ontario Basin, into Hudson Valley glacial lakes Albany II, Quaker Springs, Coveville and Fort Ann. Iromohawk drainage occurred for a period of approximately 500 years while the St. Lawrence Lowland remained glaciated and the Mohawk Valley was ice free.

Iromohawk drainage developed a network of distributary channels across the Hudson-Mohawk Lowland, as well as carving the modern Mohawk channel between Schenectady and Cohoes. Progradation of these channels occurred as the Iromohawk drained toward lowering Hudson Valley glacial lake levels. Depositional and erosional surfaces associated with distributary and modern channels have been used to establish a chronology of channel development and usage relative to Hudson Valley glacial lakes. This chronology constrains the timing and duration of later Hudson Valley lake phases between ~12,500 and 12,000 years BP.

Sedimentologic evidence in the Scotia Gravel at Scotia, one of the principle Iromohawk deposits in the valley, indicates Iromohawk drainage was a long-term, high-discharge event with cyclic (probably seasonal) variation in flow.

INTRODUCTION AND HISTORICAL REVIEW

The Mohawk Valley has long been recognized as the outlet for Glacial Lake Iroquois, the largest Late Pleistocene glacial lake to occupy the Lake Ontario Basin. Much attention has been given to Late Pleistocene water bodies in the basin (MacClintock and Stewart, 1965; Muller and Prest, 1985; Clark and Karrow, 1984; Pair and Rodrigues, 1993), but very little work has addressed the effect of Iroquois drainage on the Mohawk Valley and Hudson-Mohawk Lowland. As will be discussed, the duration of this event is short in geologic terms, but drainage during this period was the primary sculptor of modern Mohawk Valley and Hudson-Mohawk Lowland morphology.

Eastern Mohawk Valley fluvial cobble gravel deposits (Scotia Gravel) have been the subject of debate concerning both their emplacement mechanism and implications for Hudson-Mohawk Lowland morphologic development. Newly observed sedimentary structures in the Scotia Gravel indicate the drainage associated with their emplacement was a single, long-term, high-discharge event. The first author has correlated their emplacement with Lake Iroquois outflow and developed a theory regarding how Lake Iroquois outflow formed the present morphology in the Hudson-Mohawk Lowland.

Scarcity of quality datable material in the Hudson Valley has forced a relative dating scheme for eastern New York State deglacial events. Only a poor sense exists of how long most events lasted and where the set of events fits chronologically. Absolute timing constraints from the Lake Ontario Basin provide an opportunity to use eastern Mohawk Valley and Hudson-Mohawk Lowland evidence of Lake Iroquois drainage to chronologically link deglacial events in the Ontario Basin and Hudson Valley.

The area in the Mohawk Valley and Hudson-Mohawk Lowland influenced by the Iromohawk River is expansive, and as such, this trip will cover a lot of ground. As the terrain and associated geology changes very quickly, we've made the road log fairly detailed so follow the mileage carefully. The geologic interpretation of this area has been debated since the turn of the century, we invite you to continue this tradition.

Much of the following text is from the first authors' dissertation (Wall, 1995). The text is divided into three sections. The first section, "Introduction and Historical Review", gives a fairly detailed summary of background information and the evolution of thought regarding the field area. The second section, "Scotia Gravel", presents a description and detailed discussion on the principle Iromohawk River deposit in the Mohawk Valley. The third section presents our current understanding regarding the evolution of Hudson-Mohawk Lowland morphology.

Great (glacial) Lakes Drainage

Two periods of drainage from the Lake Ontario Basin to the Hudson Valley occurred during the late Pleistocene. Morner and Dremanis (1973) postulated eastward drainage of Lake Leverett from the Lake Erie basin during the Erie Interstade. They had no direct evidence of the event in New York, only that drainage toward the Ontario Basin and through the Mohawk and Hudson Valleys was the likely lake outlet. Ridge (1991) identified the

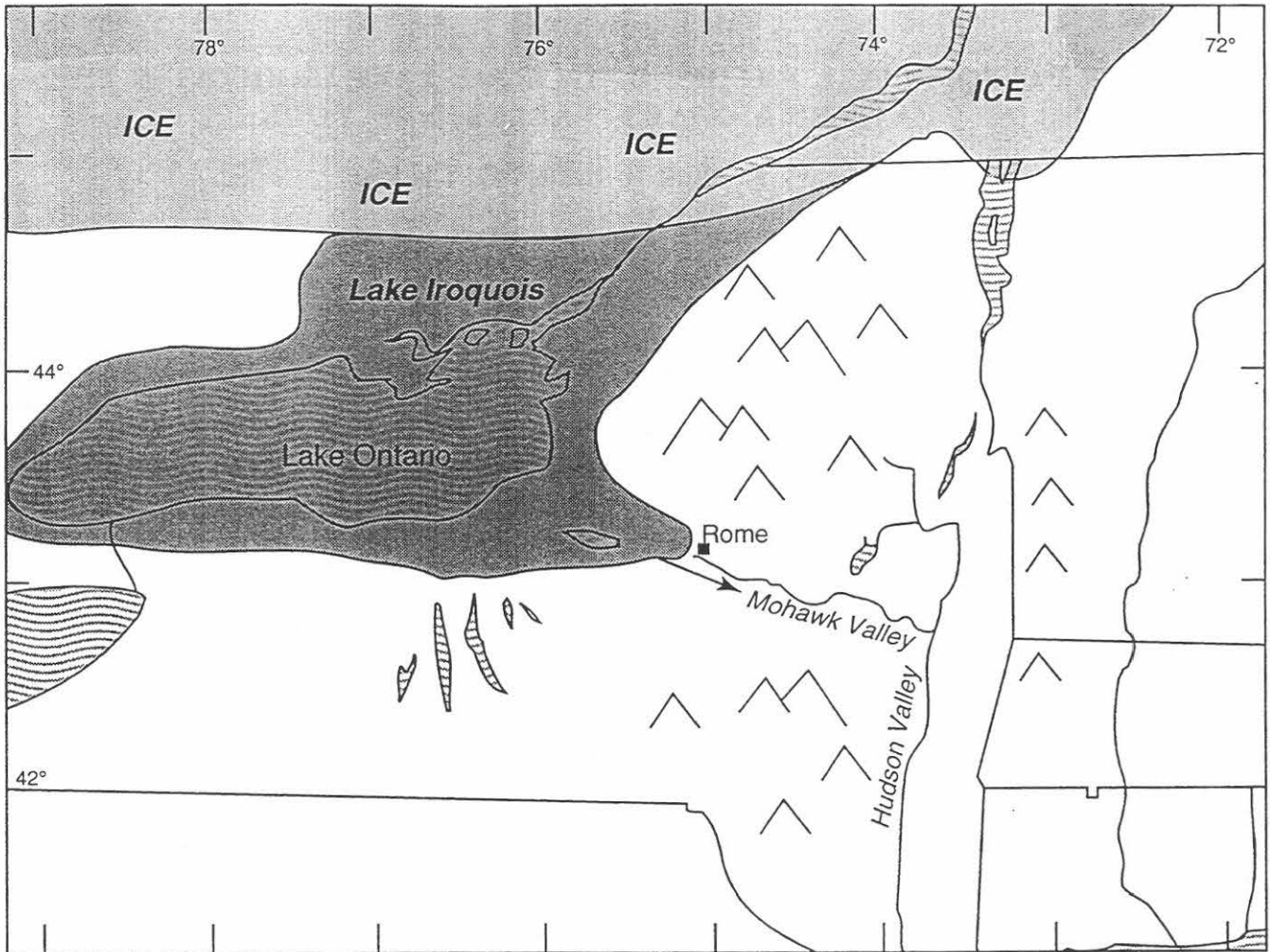


Figure 1: Cartoon map of Glacial Lake Iroquois.

Shed Brook Discontinuity and Little Falls Gravel (LFG) in the western Mohawk Valley and postulated an Erie Interstade age. Dineen and Hanson (1992) note a similar discontinuity and gravel in the Hudson Valley and postulated a correlation with Ridge's Mohawk Valley observations.

Lake Iroquois was the second Late Wisconsin water body to drain through the Mohawk Valley (Figure 1). Lake Iroquois beaches were first recognized by Thomas Roy in 1837, and named "Iroquois" by Spencer in 1890 "in memory of the aborigines who trailed over its gravel ridges" (Chute, 1979). Muller and Prest (1985) adapt the convention of defining Ontario Basin glacial lakes by their outlet:

"Drainage of the Ontario Basin by the Rome outlet to the Mohawk River is considered to define a single lake, that is, Lake Iroquois, in spite of changing extent and complexity of strandlines. Glacial Lake Iroquois ceased to exist, however, when outflow from the Ontario Basin shifted north of the Adirondack Mountains."

Fairchild (1909) refers to lakes in the western Ontario Basin, which ultimately found their outlet at Rome but were impounded by ice west of the Genesee Valley, as "Hyper-Iroquois". Chute (1979) details several stillstands of Lake Iroquois in the Syracuse area. Clark and Karrow (1984) attribute their Level I and II strandlines at Covey Hill (northern Adirondacks) to two levels of Lake Iroquois. Pair and Rodrigues (1993) identify two Iroquois waterplanes: 1) Iroquois - Watertown phase, which extended into the lowland as far north as Watertown, and 2) Iroquois - Main phase, which extended through the Lowland to Covey Hill.

Rome is commonly referred to as the outlet for Lake Iroquois (Prest, 1970; Denny, 1974; Muller and Prest, 1985). However, the level of the Rome outlet may have been controlled by downcutting of the bedrock channel at Little Falls assuming no differential uplift between Little Falls and Rome (Fullerton, 1980). Pair and Rodrigues (1993) suggest the Little Falls sill controlled initial water levels of Lake Iroquois.

Lake Iroquois Drainage and Iroquois (Scotia) Gravel

The morphology of the gravel deposit at Scotia, NY was recognized by Brigham in 1898. He mistakenly thought the deposit to be composed of sand and built as a bar or shoal into ponded waters. Woodsworth (1905), Stoller (1911), and Fairchild (1917) attributed gravels at Scotia to Iroquois outflow and consider their deposition to be contemporaneous with the Schenectady Delta. Brigham (1929) related the locations of gravel at Scotia and Yosts (Randall) to valley expansions. He considered the deposits to be made in an extension of Lake Albany (the largest Hudson Valley glacial lake) up the Mohawk Valley to Little Falls, evidently contending that the momentum of Iroquois outflow was enough to transport coarse gravel over 70 miles in standing water. The gravels at Scotia were studied by Winslow (1965) from a hydrogeologic perspective, but little attempt was made at understanding the deposit from a geologic point of view. LaFleur (1979) first coined the name Scotia Gravel for the gravels at Scotia, and advocated a series of at least three post-Lake Albany, high-discharge, glacial lake outbursts to erode and transport gravel downvalley (LaFleur, 1975, 1979, 1983). He proposed that an initial outburst eroded gravel east of Little Falls (later named Little Falls Gravel (Ridge 1991)), and redeposited it at Randall. A second outburst eroded the Randall deposit and redeposited it as valley fill at Scotia. The third outburst eroded portions of the Scotia Gravel, leaving it with two distinct terrace surfaces as it appears today. LaFleur (1979, 1983) suggested that the last lake outburst correlates with the draining of Lake Iroquois.

Timing of Late Wisconsin Glacial/Deglacial Events

Radiocarbon dates (fossil wood) for Fairchild's "Hyper-Iroquois" at Lewiston are $12,660 \pm 400$ BP and $12,080 \pm 300$ BP (Muller and Prest, 1985). Lake Iroquois east of this impoundment was already draining through the Rome outlet (Muller and Prest, 1985). Prest (1970) dates the maximum development of Lake Iroquois between 12,500 and 12,400 BP. Calkin (1982) suggests formation of Lake Iroquois prior to 12,200 BP. Clark and Karrow (1984) use four radiocarbon dates on wood to obtain an average age of 12,100 BP. Pair and Rodrigues (1993) give a minimum age for deglaciation of the northwestern Adirondack flank and incursion by Lake Iroquois of $12,500 \pm 140$ BP (radiocarbon date of kettle lake organics). Anderson and Lewis (1985) note the curvature of a well dated (11,400 BP) early Lake Ontario water plane that closely paralleled that of Lake Iroquois, and suggest only a few hundred years had elapsed between lakes. Anderson and Lewis (1985) bracket the existence of Lake Iroquois between ~12,500 and 12,000 years BP. Denny (1974) suggests a range between 12,700 and 12,000 years BP for deglaciation of the northern Adirondack flank. Muller and Calkin (1993) suggest Lake Iroquois only existed for a "few centuries" and came into existence between 12,500 and 12,100 BP. Prest (1970) suggests 12,200 BP for a stable strand of Lake Frontenac, which succeeded Lake Iroquois.

Hudson-Mohawk Lowland

Woodsworth (1905) was the first to comprehensively review Hudson Valley post-glacial deposits, and presented evidence for a water body that followed the retreating ice north. He named the water body "Lake Albany" after the Albany clay in the valley which was named by Ebenezer Emmons in 1846. Fairchild (1917) suggested that Lake Albany was not a lake, but rather a marine strait between New York City and the St. Lawrence Lowland. Stoller (1919) cited topographical evidence at Mechanicville as conclusive evidence against a Lake Albany estuary. Stoller was the first to interpret and name stages in Lake Albany based on deposits in the Schenectady and Saratoga 15" quadrangles (Stoller, 1922). Woodsworth (1905) recognized the Quaker Springs phase of Lake Vermont (Champlain Valley) on the Schuylerville 15" quadrangle, noting the erosive work of a "powerful stream of water" at Quaker Springs as evidence of a lake outlet there. Chadwick (1928) recognized a lowering Lake Albany. He named the Coveville phase of Lake Vermont as he considered its outlet to be at "The Cove" in Coveville. He made no mention of Woodsworth's Quaker Springs phase. LaFleur (1965) rejected the interpretation of Quaker Springs and Coveville outlets. He suggested early phases of Lake Vermont were controlled by later phases of Lake Albany and proposed the renaming of lower Lake Albany phases. Connally and Sirkin (1969) proposed the names "Quaker Springs" and "Coveville" for lower Albany phases. DeSimone (1985) recognized three phases of Lake Fort Ann in the Hudson Valley. DeSimone (1985) and DeSimone and LaFleur (1985) postulated a slightly lower Albany phase, between Lakes Albany and Quaker Springs. The table below shows the current interpretation of Hudson Valley glacial lake level elevations in the study area.

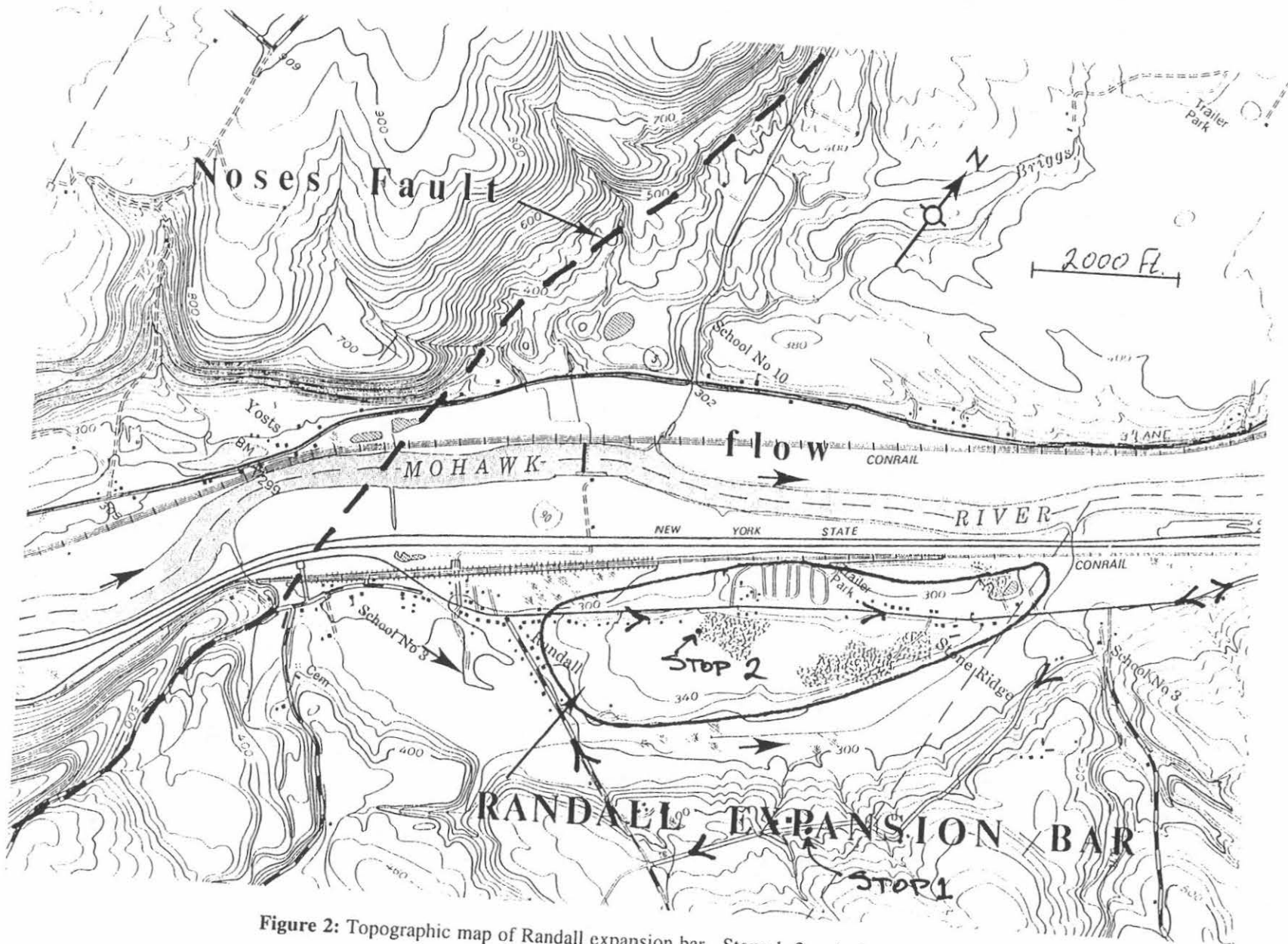


Figure 2: Topographic map of Randall expansion bar. Stops 1, 2 and trip route shown.

HUDSON VALLEY GLACIAL LAKE	SCHUYLERVILLE/ SARATOGA SPRINGS	MECHANICVILLE/ EAST LINE	TROY/ SCHENECTADY
Albany	370 feet	350 feet	340 feet
Albany II	340	330 (1)	~310
Quaker Springs	310	300	~290
Coveville	270-240	240-220	~220
Fort Ann I	200	190	~180
Fort Ann II	160	150	~140
Fort Ann III	130	120	~110

Modified from DeSimone (1985)

(1) This study

Hudson-Mohawk Lowland Paleochannel Development

Stoller's surficial mapping in the Schenectady (Stoller, 1911), Saratoga (Stoller, 1916), and Cohoes (Stoller, 1920) 15' quadrangles was the first detailed examination of the area. Schock (1963), DeSimone (1977) and Dahl (1978) mapped the Troy North, Schuylerville, and north half of the Mechanicville 7 1/2' quadrangles, respectively. Hanson (1977) mapped the Round Lake, Niskayuna and southern half of the Mechanicville 7 1/2' quadrangles.

Fairchild recognized the contribution of Great Lakes drainage to the Hudson Valley via the Mohawk Valley (Fairchild, 1904). He coined the terms "Glaciomohawk" and "Iromohawk" referring to Mohawk Valley drainage before and after the initiation of Lake Iroquois. Stoller (1911) considered the development of, what he terms, "distributary channels north of Schenectady Delta" as due to a lowering of Lake Albany. Fairchild (1917) considered "land uplift ... by a wave movement" as the mechanism responsible for diversion of Iroquois waters into the channels. Stoller (1911) considered distributary channel occupancy to be controlled by undercutting of other distributary channels from overtopping and headward erosion. Remnant ice may have played a part in channel diversion (LaFleur, 1979, 1983) and at least three catastrophic lake outbursts in the western Mohawk Valley and/or Ontario Basin may have been responsible for channel overtopping and initiation of distributary channel incision (LaFleur, 1975, 1979, 1983, and Hanson, 1977). Hanson (1977) rejected simple headward erosion as a mechanism for modern channel occupancy.

Stoller (1911, 1922) advocated fluvial action as the mechanism for removal of sediments from Ballston, Round and Saratoga Lake basins. Fairchild (1917), Cook (1930), Hanson (1977), and LaFleur (1979, 1983) all considered the basins to be kettles. Cushing and Ruedemann (1914) considered Lonely Lake basin a kettle, but Saratoga Lake Basin to be a "less deeply filled" portion of preglacial drainage. Hanson (1977) identified a relatively small kettle at Elnora and called it the Elnora ice block. Woodsworth (1905) and Cook (1909) consider Ballston Channel and the location of Saratoga and Round Lakes as preglacial fluvial channels. Dineen and Hanson (1983) mapped the extent of preglacial drainage in the area, and in particular the Colonie Channel which Woodsworth and Cook first recognized. Cook (1909) traced the preglacial Mohawk gorge between Schenectady and just north of Coeymans (where it joins the Hudson), concluding that the modern river channel east of Rexford is post-glacial.

SCOTIA GRAVEL

The Scotia Gravel is a disjointed fluvial deposit in the eastern Mohawk River Valley at Scotia, Randall, Fort Hunter, and Fultonville(?). The type locale for the deposit is in Scotia, this segment being by far the largest. The Scotia Gravel at Scotia (Figure 3) extends along the valley trough a distance of 8.3 miles in an expanding downvalley "wedge", from Hoffmans to Collins Lake, Scotia. The deposit is about 1 mile across at its widest point where well logs indicate it to be over 100 feet thick (Winslow et al., 1965). The highest part of the wedge is approximately 70 feet above the Mohawk River, with a maximum elevation of 310+ feet near Hoffmans and 290+ feet in Scotia.

In general, the Scotia deposit fines downvalley. Gravel near Hoffmans is typically well-sorted, clean, and clast supported, with individual clasts ranging up to softball size. Farther downvalley, the deposit is sandier and more matrix supported. At Collins Lake, the deposit has no gravel; it is strictly a very well-sorted, coarse-grained sand. Gravel lithology is dominated by quartzite, gneiss, and red quartzitic sandstone of Adirondack provenance along with carbonates minor shale presumably with a western Mohawk provenance.

Much of the deposit, in apparently random locales, is bound by a weak carbonate cement. The origin and precipitational environment of the cement is enigmatic, because it apparently shows signs of both vadose and phreatic conditions (S. Gaffey, personal communication, 1994). The carbonate could come from dissolution of carbonate clasts in the deposit or from carbonate rich Iromohawk waters. The Mohawk Valley contains many potential carbonate sources, making the latter possibility quite reasonable. Carbonate precipitation could have been

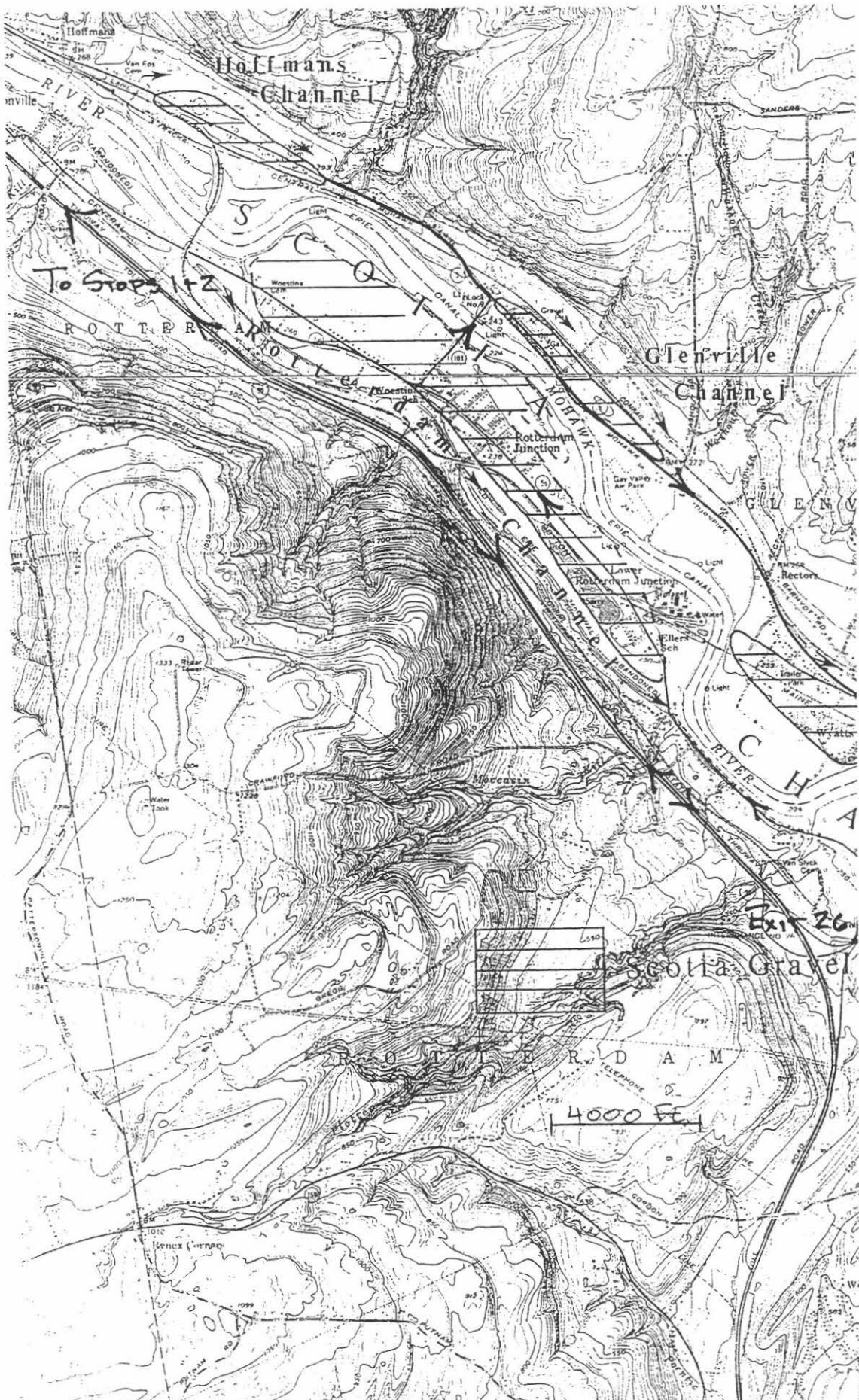


Figure 3: Topographic map of Scotia Gravel at Scotia. Relevant morphologic features, stops 3 and 4, and trip route shown.

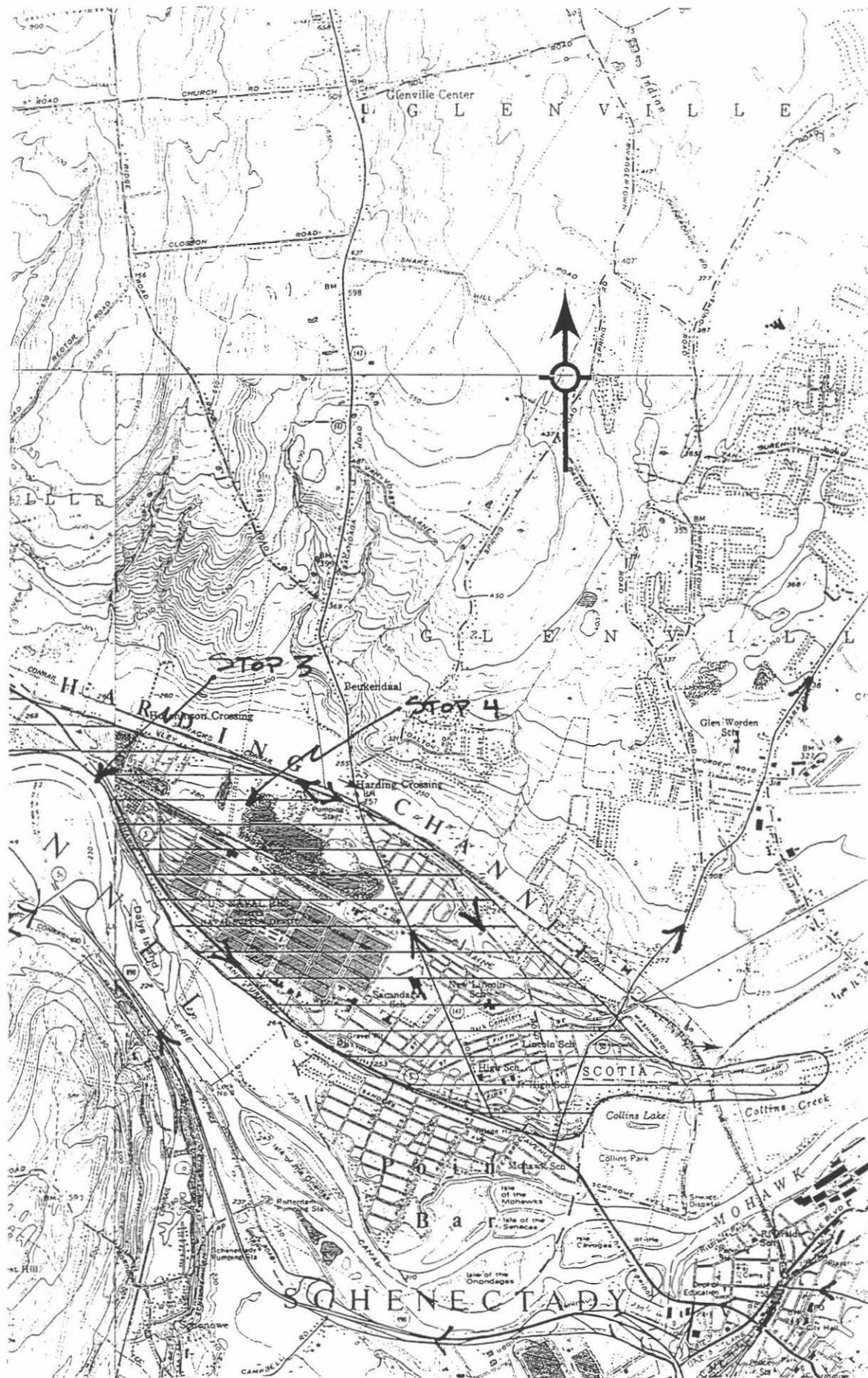


Figure 3 (continued): Topographic map of Scotia Gravel at Scotia.

caused by changing water levels associated with Iromohawk discharge leaving gravel clasts alternately wet and dry, thereby changing carbonate equilibrium conditions enough to cause precipitation. More detailed study of the cement is needed to test this hypothesis.

Exposures of the Scotia deposit reveal little in the way of bedforms, appearing quite massive in many locations. Clast imbrication is evident, but not pervasive in these areas. The only observed exception to the general massive nature of the deposit is east of Wyatts along a precarious exposure bordering the Mohawk River. This exposure is marked by numerous east-dipping graded gravel foreset beds. Dozens of these foresets are visible across the exposure, each grading from cobble to coarse sand and pebble gravel, and measuring approximately 3 feet in thickness (Figure 4). The cyclic nature of the graded foresets is more typical of a long-term rather than catastrophic depositional event; furthermore, the cyclically suggests a seasonal variation in transport energy - essentially gravel varves. If a gravel varve hypothesis is correct, the Scotia deposit records many tens if not hundreds of years of deposition. Baker (1973) describes similar graded features in catastrophic outburst pendant bars in the channeled scablands of Washington State. However, he attributes the graded nature of the deposit to eddies developing behind the channel protuberance associated with the bar. The Scotia deposit is clearly not a pendant bar as no such protuberance is evident.

The gradient and discharge of the modern Mohawk River may be too low to account for the thickness and grain size of Scotia Gravel (LaFleur, 1983). The stratigraphic position of the gravels requires it to be either the last depositional (ignoring modern alluvium) event in the valley or the remnant of an exhumed topography. The exhumed scenario would require the gravel to be at latest Erie Interstade in age (i.e., equivalent to Little Falls Gravel (LFG)). If this is the case, the gravels would have been overridden by glacial ice at least once during the Port Bruce Stade. Exhumation would then need to remove all glacial material because none is found stratigraphically above the gravel. It seems unreasonable that no material would remain, especially erratic boulders that would have been associated with the deposit and unlikely to have been moved during exhumation. Portions of the LFG have been exhumed near Little Falls, but by far the majority of the remaining deposit is buried in glacial and lacustrine sediments. A second problem with exhumation is the erosion of LFG at Little Falls. Clearly, a huge amount of gravel has been eroded from a deposit that likely spanned the entire width of the valley east of Little Falls. If the gravels at Scotia are considered equivalent to LFG, where are the eroded LFGs? Lastly, projecting the modern valley slope from the top of gravel at Scotia, it intersects fairly well with the top of the gravel bar at Randall. Farther projection of this line westward intersects the LFG well below its peak. Explanations of differential uplift or eroded nickpoints could explain this last observation, but taken as a whole it seems probable that the Scotia Gravel is the product of a second fluvial event in the valley, that being the drainage of Lake Iroquois. Lake Iroquois drainage eroded LFG between Little Falls and St. Johnsville and transported it downvalley, redepositing it in Randall, Fultonville(?), Fort Hunter and Scotia.

The locations of Scotia Gravel are directly related to valley morphology. The three deposits of Scotia Gravel are found in valley expansions. The Randall and Scotia deposits are located downvalley of the Noses and Hoffmans faults, respectively. Brigham was apparently the first to make the connection between the faults and the locations of deposits, however he considered the depositional environment as lacustrine rather than fluvial (Brigham, 1929). The locations of the Fultonville and Fort Hunter deposits are related to valley expansions associated with fairly dramatic valley bends.

The width of the modern Mohawk floodplain between Hoffmans and Barge Canal Lock 8 is fairly consistent at about 2,000 feet. Throughout this reach, the Mohawk floodplain north of the river is bounded by a bank of gravel 50 to 70 feet high, peaking in elevation between 290 and 310 feet. The slope of this bank appears too steep to be depositional in origin; however, with the exception of active bank erosion just east of Wyatts, the Mohawk River has apparently done nothing to modify the bank. Observed bedrock(?) just below ground surface, south of the river and east of Wyatts exposure, may have forced the modern Mohawk into the bank. South of the Mohawk River at Rotterdam Junction, the 2,000 foot wide swath is bounded by a low ridge of gravel that parallels the valley trough and crests at 260+ feet elevation. Downvalley of Rotterdam Junction, the Mohawk or its floodplain are in contact with the southern valley wall. Both valley walls throughout the reach are composed of shale covered by a thin veneer of glacial till or diamicton. The floodplain is flooded in gravel and covered by a patchy, fairly thin (<10 feet) layer of alluvium.

East of Lock 8 the valley rapidly expands into Schenectady Basin (Figure 3). The southern bedrock wall at Lock 8 swings to the south, following the course of the preglacial Mohawk channel (Winslow et al., 1965), where bedrock gives way to glacial and glaciolacustrine sediments that rim the southern basin wall. Well logs (Winslow et al., 1965) indicate the alluvium covered floor of Schenectady Basin is underlain by a 20 to 50 foot layer of sand and gravel over till. This sand and gravel layer is pervasive across the basin, but absent from stratigraphy in the southern basin wall, indicating that sand and gravel in the basin was deposited following or during erosion of the southern basin wall.

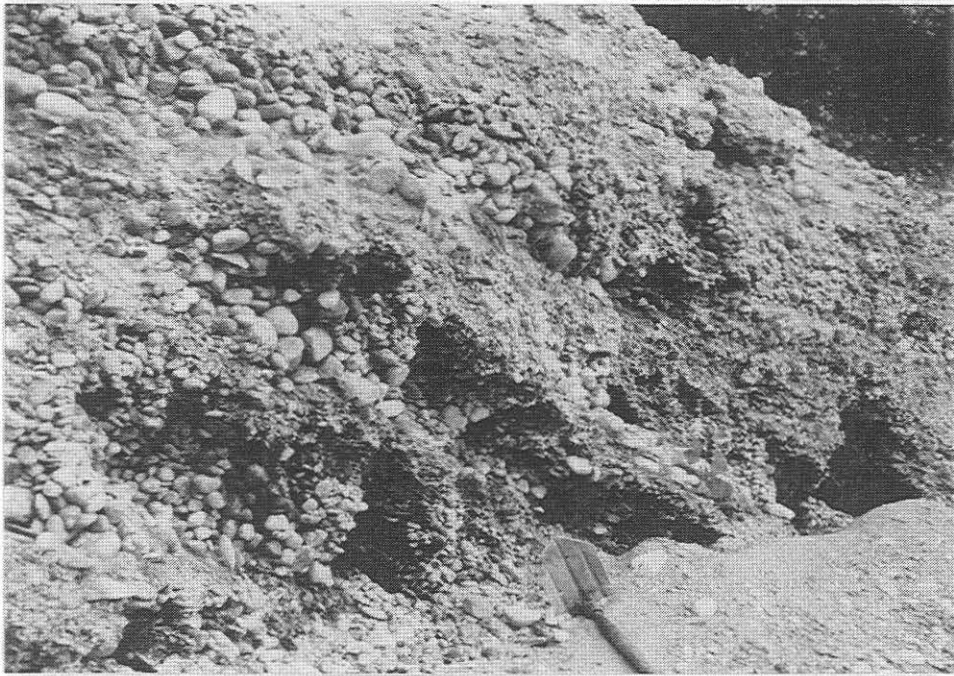


Figure 4: Exposure of graded Scotia Gravel foreset beds near Wyatts.



Figure 5: Internal structure of Randall expansion bar. Downvalley is to the left, frontend loader is approximately 15 feet high.

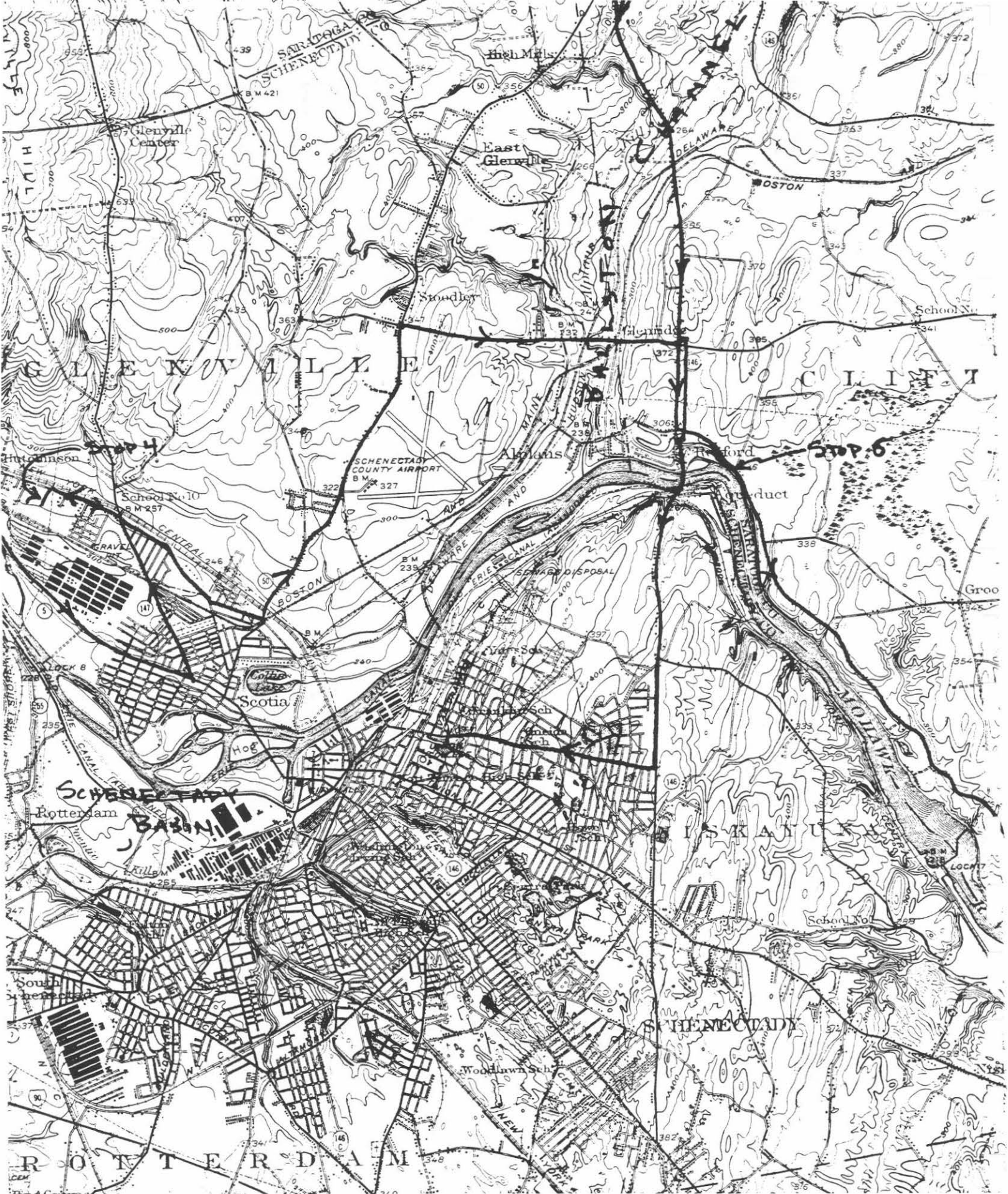


Figure 6a: Topographic map of southern half of Hudson/Mohawk Lowland. Relevant morphologic features and trip route shown. Scale: 1"= apx. 1 mile.

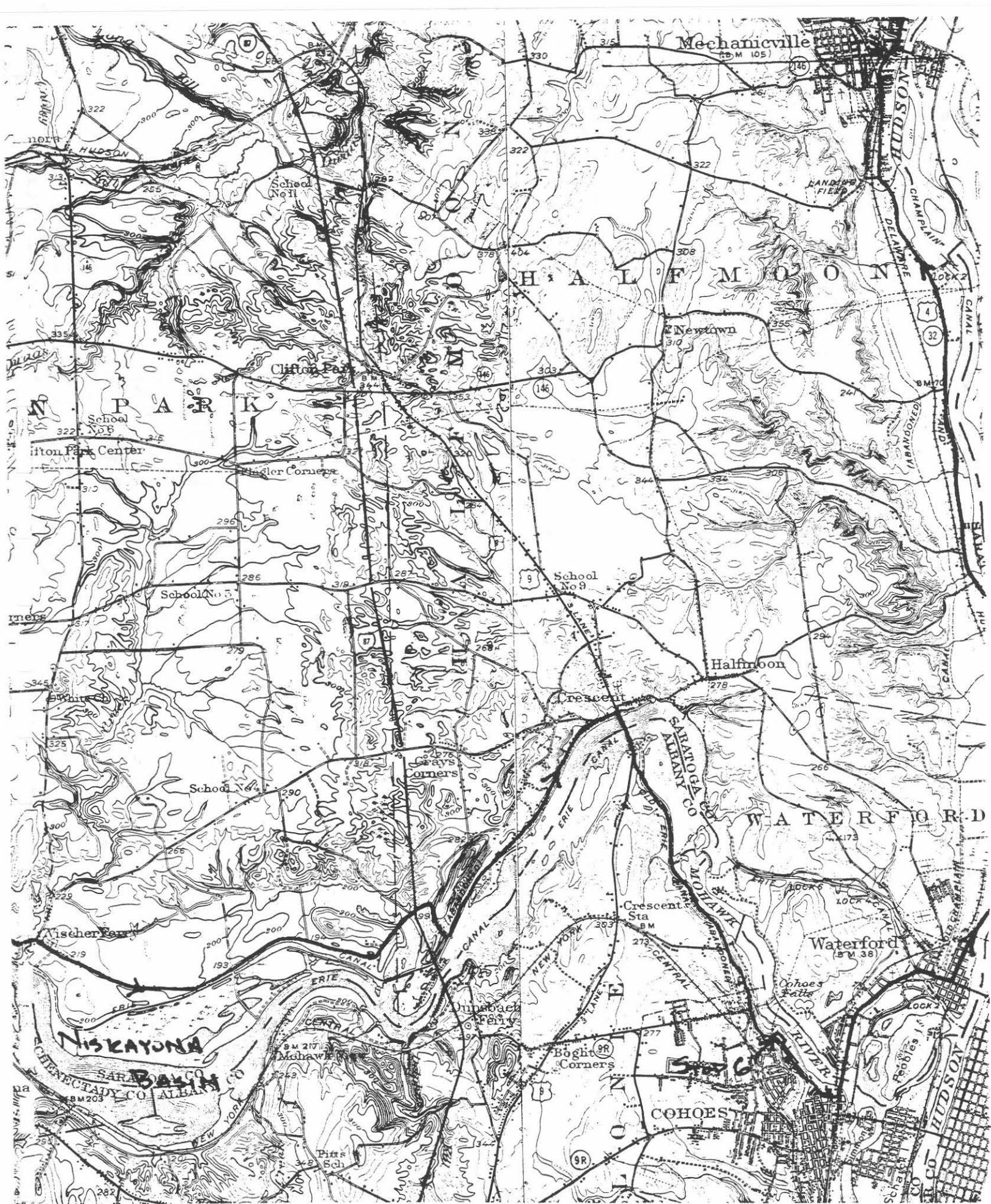


Figure 6a (continued):

Figure 6b: Topographic map of northern half of Hudson/Mohawk Lowland. Relevant morphologic features and trip route shown. Scale: 1"= apx. 1 mile.



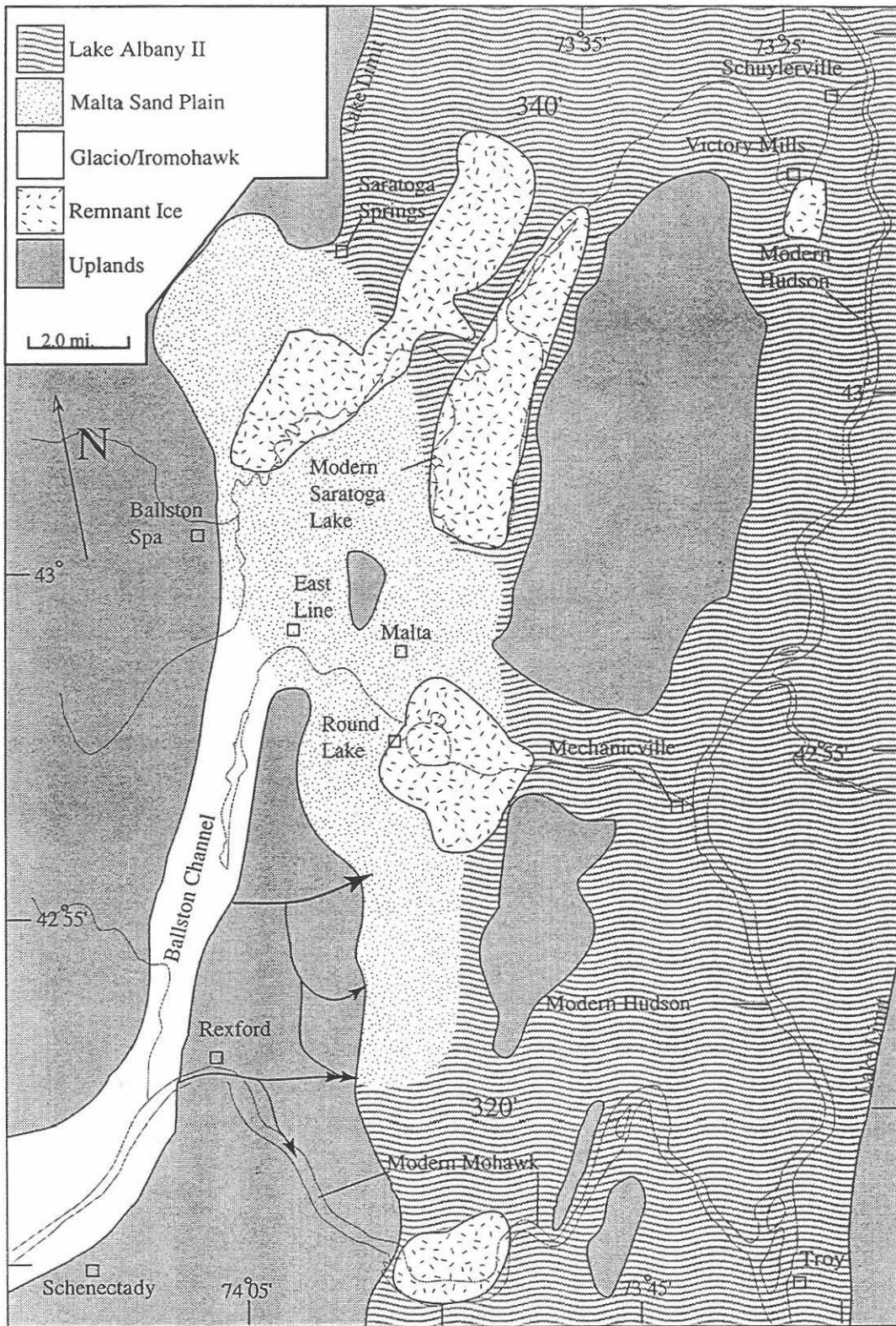


Figure 7: Map of Hudson/Mohawk Lowland during Lake Albany II.

The morphology of the Scotia deposit in Schenectady Basin changes quite dramatically from that immediately upvalley. The abrupt wall of gravel bordering the Mohawk floodplain near lock 8 is modified to a step in Scotia. This step drops from the gravel terrace top of 290+ feet to ~250 feet along Route 5 in Scotia. From Route 5, the gravel gradually slopes to the south toward the Mohawk River at 210 feet. It is reasonable to think this gravel is continuous across the Mohawk where it is overlain by alluvium and visible at ~210 feet in Winslow et al.'s well logs.

The deposit is inset by four minor channels (Figure 3), three of which hang well above the modern floodplain. Hoffmans Channel has a bottom elevation of 280+ feet, approximately 30 feet above the Mohawk floodplain, with channel dimensions of approximately 5,000 feet in length and 400 feet in width. The channel is bordered to the north by the valley wall and to the south by a ridge of apparently undisturbed gravel peaking at 310+ feet. The downvalley end of the channel is filled by reworked Verf Kill sediment draining from the north valley wall. Some 4,000 feet downvalley, Glenville Channel flanks the northern valley wall in a fashion similar to Hoffmans Channel. The southern channel wall consists of gravel with a peak surface elevation of 300+ feet. The channel is 8,000 feet long and about 500 feet wide with a bottom surface elevation of 280+ feet. The downvalley end of the channel is met by deposits of Washout Creek, with two distinct terrace surfaces of 290+ and 310+ feet.

Harding channel is the largest of the four minor channels. This channel, with dimensions of 20,000 feet in length and approximately 1000 feet in width, flanks the northern valley wall. The upvalley end of the channel is poorly defined, but apparently begins as a shallow trough northwest of Wyatts. The downvalley end of the channel is met by a spit of sand between the channel and Mohawk floodplain northeast of Collins Pond. The spit of sediment extends the channel length an additional 3,000 feet to the east. Channel bottom elevation grades from 250+ near Wyatts to 230-240 feet near Collins Lake. The fourth channel, Rotterdam channel runs along the southern valley wall in Rotterdam Junction. The northern side of the channel is bounded by the previously mentioned 260+ foot elevation ridge of gravel in Rotterdam Junction. The bottom of Rotterdam channel is marginally above the modern floodplain, grading from ~250 feet in the upvalley end at Woestina Cemetery to 230+ feet at Lower Rotterdam Junction. The channel is 12,000 feet long and 600 feet wide.

The modern morphology of the Scotia deposit looks very similar to that of a braided fluvial system. However, the internal structure of the deposit is very regular, similar to a deltaic front rather than a braided bar. Similarly, the deposits' position in a valley expansion suggests a valley delta or expansion bar. The thickness of the deposit (100+ feet) also suggests the deposit was laid down in deep water, also consistent with a delta or expansion bar than a braided fluvial system interpretation.

The most convincing evidence against a braided fluvial system comes from an examination of the Schenectady Basin outlet. As will be illustrated, Iromohawk waters drained through Ballston Channel to Lake Quaker Springs near Saratoga Springs. Adjusting for isostatic rebound, the 310 foot level of Lake Quaker Springs at Saratoga Springs is ~10 feet above the 280 foot elevation of the Ballston Channel divide (the divide is presently in a narrow bedrock "sub-channel" at 270 feet, but erosion to this level did not occur until a lower Hudson Valley base level). Therefore, water level at the modern divide in Ballston Channel was at least 290 feet. Considering the size of Ballston Channel and the volume likely exiting through it (see next section), water depth at the divide must have exceeded 10 feet. Considering the distal end of the Scotia deposit is at 290+ feet, together with the excessive deposit thickness (>100 feet), the Scotia deposit must have been deposited in a water-filled basin. The main body of Scotia gravel continued to be built until the modern channel at Rexford eroded below 290 feet. The size of the deposit and the aforementioned time likely associated with its deposition suggests a considerable amount of time was required for lowering of Rexford sill.

The uneven, channeled, and braided appearance of the deposit between Hoffmans and Schenectady Basin is explained by a transition of the Iromohawk from depositional to erosional flow. This transition could have occurred in one of two ways, or in a combination. First, base level could have lowered, causing initial and secondary outlets from Schenectady Basin to lower. Significant erosion of either outlet below the top of the Scotia deposit (~290 feet) would have initiated erosional flow. Second, the transition could be due to a wane of Iromohawk discharge. According to D. Franzi and D. Pair (personal communication, 1994) the Iroquois transition from Mohawk to St. Lawrence outlets could have been gradual. A waning discharge would tend to channelize across its previous depositional surface, thereby eroding that surface. The second scenario cannot be eliminated, but a response to lowering base level seems reasonable considering Mohawk Gorge (the modern channel/outlet) eroded well below the level of Ballston channel divide and the Scotia deposit.

Following Mohawk Gorge lowering, Harding, Glenville and Hoffmans channels became active, with the majority of flow contained in the 2,000-foot-wide swath referred to as the Scotia Channel (LaFleur and Wall, 1992) (Figure 3). The former bottom of the Scotia Channel exists as the modern floodplain between Hoffmans and Schenectady Basin. Lowering of water level is also recorded in the 20 foot drop of Washout Creek terrace deposits. At Rotterdam Junction, a bend in Scotia channel migrated away from the southern valley wall, leaving either a former channel bottom at 260 feet elevation, or depositing a point bar. Very limited exposures are available,

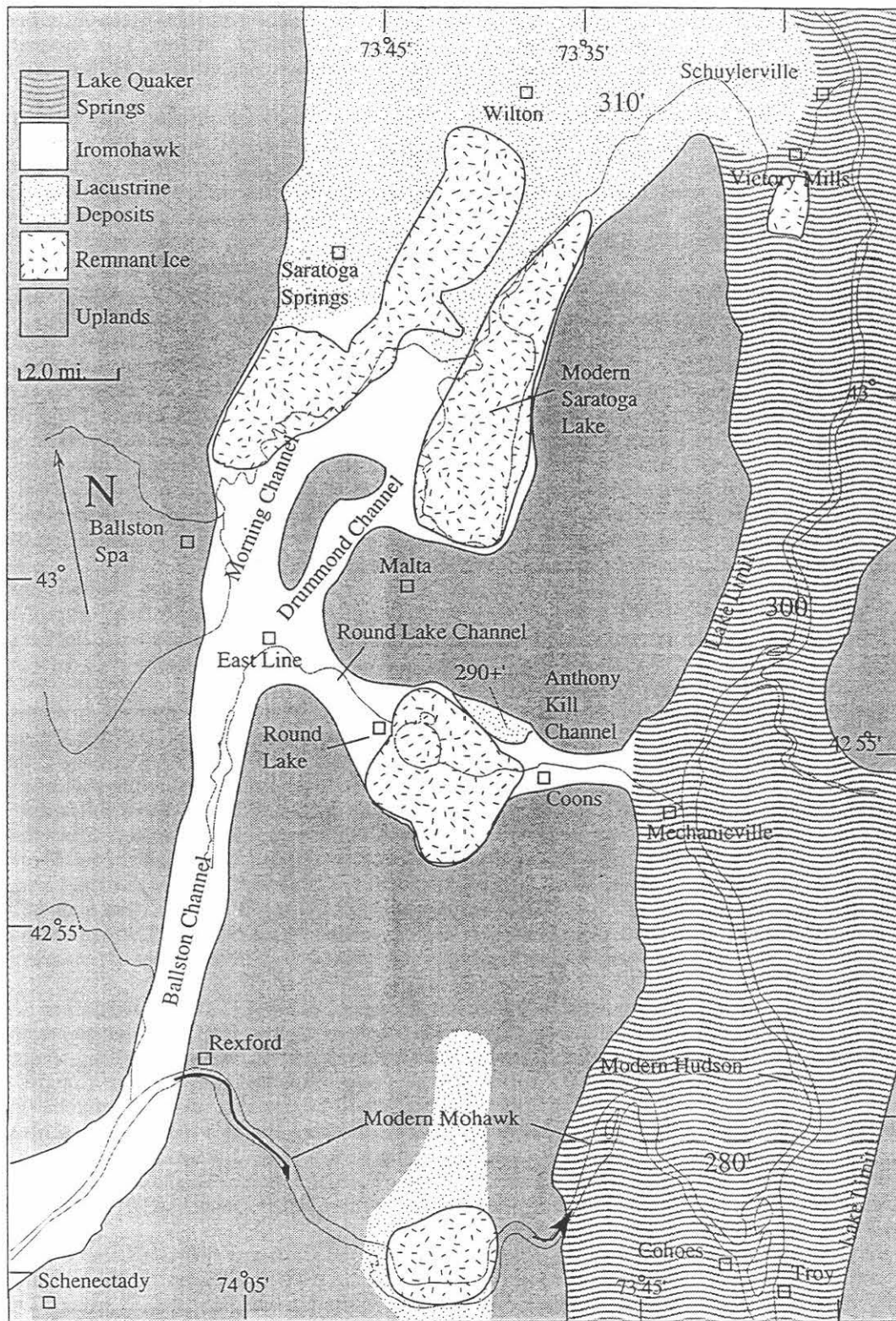


Figure 8: Map of Hudson/Mohawk Lowland during Lake Quaker Springs.

however, to test either hypothesis. At some point, Scotia Channel lowered to where it deactivated Glenville and Hoffmans channels, as they hang well above the modern valley floor. Based on its elevation, Rotterdam channel was the last minor channel to develop, incising the back of the Rotterdam Junction point bar(?) much in the fashion of a chute.

Gravel incision by Scotia Channel was accompanied by channel migration south of Scotia into Schenectady Basin. This migration eroded the southern wall of Schenectady Basin and deposited a point bar against the main body of the deposit south of Route 5 in Scotia. Route 5 in Scotia is built on the top of this point bar and roughly marks a lower limit to the maximum Schenectady Basin water line associated with point bar development. Most, if not all, of the gravel south of Route 5 is reworked from development of the Scotia Channel and redeposited in Schenectady Basin.

The floor of the downvalley end of Harding Channel joins the modern floodplain, suggesting it was active throughout the existence of Scotia Channel, perhaps acting as a chute during periods of high flow. The spit of sand at its downvalley confluence with Scotia channel may be the result of converging Scotia and Harding Channel flows. The morphology of the distal end of the gravel at Collins Lake is enigmatic. It may be either a headward erosional feature developed during the initial transition from depositional to erosional flow regimes, or perhaps an eddy feature developed during the same period.

Based on location, morphology and sedimentary structures, the Randall segment of Scotia Gravel is a classic giant expansion bar (Figure 2), similar to deposits in the channeled scabland of Washington State. The bar measures 7,000 feet in length and 1,800 feet at its maximum width. The 3.8:1 L/W ratio falls within the typical 3 to 4 range of large-scale fluvial landforms (Komar, 1983, 1984). The interior of the deposit is exposed by an active sand and gravel quarry. Large-scale east-dipping (downvalley) cross-bedding is visible throughout the exposed pit (Figure 5), confirming the eastward flow direction inferred from the deposits east-pointing tapered tail. Exposures in contact with the deposit surface show bedding that mimics external morphology. This observation is consistent with an original depositional surface as opposed to an erosional remnant as LaFleur (1979, 1983) infers.

Overall, the Randall deposit is very poorly sorted with individual clasts ranging from small boulders to coarse sand. Thickness of the deposit is approximately 60 feet. One well log from the deposit indicates the bottom of the bar is approximately at the elevation of the modern floodplain (U.S. Geological Survey, Open File). The height of the bar above the floodplain therefore represents a minimum water depth (60 feet) for bar deposition.

The Fort Hunter deposit is barely perceptible at land surface. The deposit appears as a fluvial ridge on the Tribes Hill 71/2' quadrangle. No exposures were found in the deposit, but the characteristic Scotia Gravel lithology is evident at the surface. Similar fluvial ridges in Fultonville are observable on the Randall 71/2' quadrangle, but urban development has made them inaccessible.

PALEOCHANNEL DEVELOPMENT AND USAGE CHRONOLOGY

Introduction

A major problem with previous interpretations of Hudson/Mohawk Lowland distributary channel development and occupation has been the assumption that only one or possibly two channels could have been active at once. Based on this assumption, previous authors (see Historical Review) invoked mechanisms by which flow could be transferred from one channel to the next, in essence shutting off one channel and turning on another. Mechanisms include: an "uplift wave", isostatic rebound, headward erosion, overtopping and overtopping due to catastrophic lake outbursts. Central to the assumption of a progression of channel usage, is a gross underestimation of the quantity of water that developed and occupied these channels. Even the more conservative paleodischarge estimates (Wall, 1995), indicate potentially a tremendous volume of water drained through the Mohawk Valley during this period. It is the author's contention that the Mourning, Drummond, and Round Lake distributary channels (Figures 6a and 6b), as well as the modern channel east of Rexford, *all* accommodated Lake Iroquois outflow throughout the duration of Lake Iroquois. The quantity of water occupying each channel gradually changed, primarily as a function of modern channel incision, but all channels remained active.

The Hudson-Mohawk Lowland north of the Mohawk River is scarred by basins and misfit channels, but perhaps most striking is the amount and extent of lacustrine sand between Albany and Gansevoort. Connally and Sirkin (1972) refer to this body of sand as an erg, a sea of sand. Beyond attempting to attribute portions of the erg to different Hudson Valley lacustrine phases, little effort has been made at understanding the source and route by which sediment in these locations was deposited. This section relates deposits and source with misfit channels. In so doing, a relative chronology for the development of deposits and channels emerges.

Lake Iroquois Outflow and Lake Albany II

Clear evidence of high-volume discharge in the Mohawk Valley is the transport and deposition of Scotia Gravel. The deposit, as described previously, was made in a relatively low-energy, water-filled basin, which had a surface elevation at Scotia of ~300 feet. The Lake Albany Schenectady Delta stands at 340+ feet elevation just to

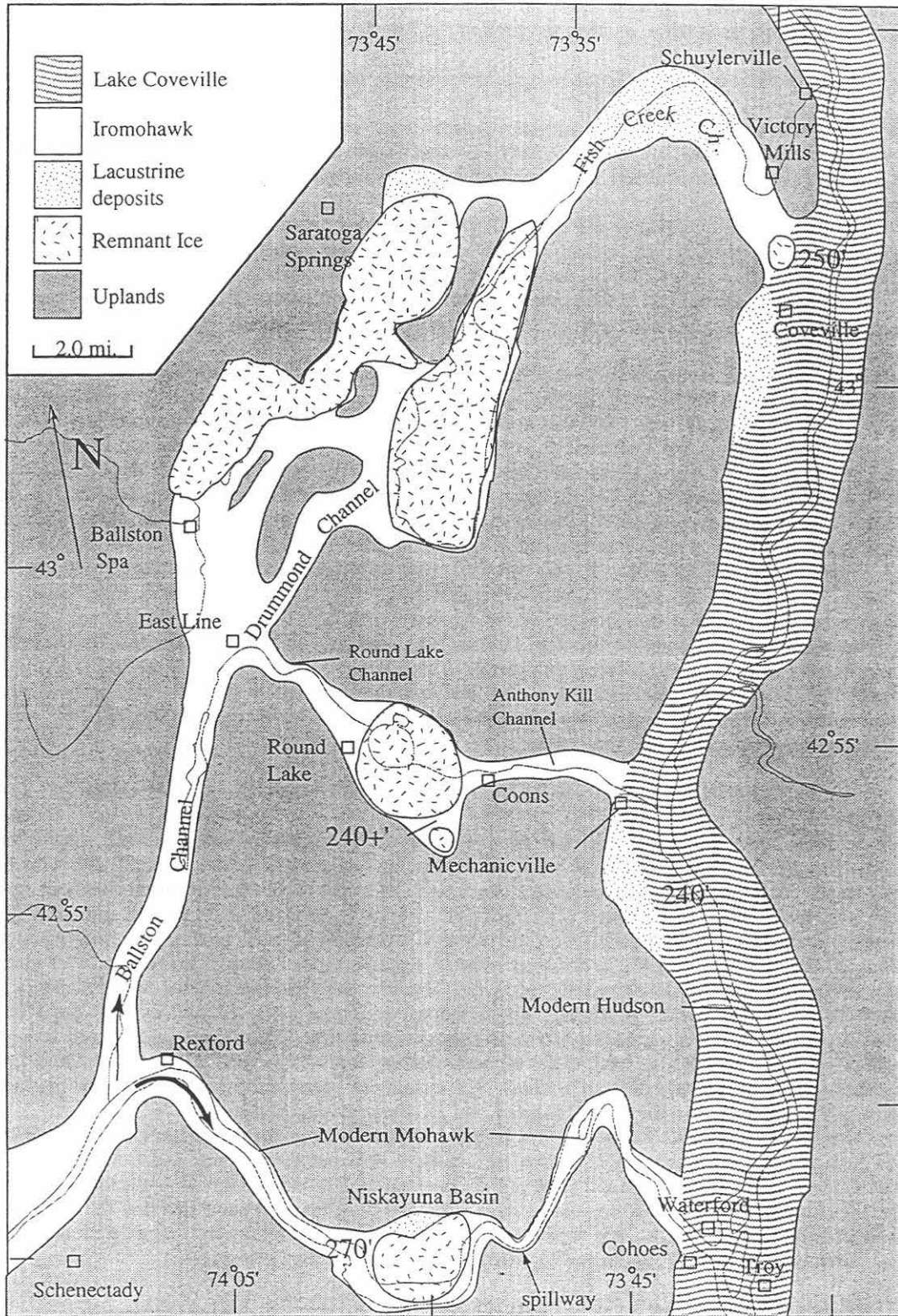


Figure 9: Map of Hudson/Mohawk Lowland during Lake Coveville.

the south of Scotia. Based on the difference in elevation between depositional surfaces and the water levels they represent, the Scotia Gravel was deposited sometime after Lake Albany. This alone does not preclude Lake Albany from receiving high volume discharge; however, it's difficult to envision high-volume discharge making the dramatic ~90° bend at what is now Schenectady Basin unless Schenectady Delta was already inset and cut off.

Cutting off a delta or delta lobe can occur by one or a combination of two processes. First, the feeder channel to the delta becomes choked with sediment, forcing the channel to find an easier path to base level, and second, a change in base level occurs. The presence of the Malta Sand Plain at 330+ feet elevation presents the possibility that a change in base level caused the Mohawk to cut off Schenectady Delta.

The Malta Sand Plain is an extensive sand body requiring a significant sediment source. Considering Ballston Channel provides a direct link between the Mohawk Valley and the sand plain, the Mohawk has to be considered the sediment source. Correcting for isostatic adjustment (DeSimone, 1985), the Malta Sand Plain lies nearly 40 feet below Schenectady Delta, and 30 feet above Lake Quaker Springs deposits both having experienced approximately equal uplift at the Hoosic River mouth. Although no topset-foreset contacts were found in the Malta Sand Plain, it seems unlikely that it was deposited in 40 feet of water 11 miles north of Schenectady Delta. Therefore, an intermediate lake level between Albany and Quaker Springs, as first suggested by DeSimone (1985) and DeSimone and LaFleur (1985), seems likely (Figure 7). Although DeSimone (1985) uses only a small number of features to develop a "lowered Albany" water plane curve, the Malta Sand Plain falls on his curve quite well. The outlet of this lake is debatable, so a naming of this level based on an outlet is not advisable. A temporary and uninspiring name of Albany II, therefore is assigned to this phase. Lake Albany II is equivalent to Stoller's (1922) Malta Delta Stage of Lake Albany.

Considering the extreme ~90° channel bend at Schenectady Basin and initial flow through Ballston Channel was to a base level other than Lake Albany, high discharge is excluded from entering Lake Albany. This implies that at least initially, Lake Albany II did not receive high discharge.

Distributary Channels

A lack of Albany II deltas or sand plains associated with Mourning, Drummond, and Round Lake channels and a massive Quaker Springs sand plain associated with Drummond and Mourning Channels, indicates channel development was due to a lowering lake level and not channel infilling and progradation. Initial development, therefore, of Mourning, Drummond, and Round Lake channels occurred following Lake Albany II. All three channels have erosional surfaces below and adjacent to the Malta Sand Plain. Channel erosional surfaces are between 300 and 310 feet elevation and characterized by a surface of washed till in Round Lake Channel and an exhumed clay surface northeast of East Line between Drummond and Mourning Channels. Drummond channel exhibits an exposure of coarse sand and gravel at its bend southwest of Rt. 9. This sand and gravel is at the top of the stratigraphic section between 290 and 300 feet. Its stratigraphic position and location along the inner channel bend suggest it is a remnant point bar, now hanging 40-50 feet above the channel floor. The development of equal-elevation erosional surfaces in each channel indicates their development was contemporaneous and their nature truly distributary, rather than a series of channels used in some sequential manner. Distributary channels form when a main channel can no longer accommodate flow volume. Considering: (1) the size of these channels; (2) their distributary nature; (3) the coarse-grained point bar in Drummond Channel; and (4) the gigantic Saratoga-Wilton sand plain adjacent to Mourning and Drummond channels, high volume discharge (i.e., Lake Iroquois outflow) must have occurred no earlier than the Lake Albany II - Quaker Springs transition. Such a dramatic flood pulse into the Hudson Valley could have triggered the lake outlet failure (LaFleur, 1979; Stanford and Harper, 1991). DeSimone (1985) considered the Albany-lowered Albany (Albany II) transition to have been gradual. If this was the case, the lake outlet may have been weakened and perhaps all the more susceptible to failing from Iroquois outflow. If it is shown that some other mechanism is responsible for the Albany II-Quaker Springs transition, Iroquois outflow must have occurred some time during Lake Quaker Springs time.

Ice Block Basins

As previous authors have proposed (see Historical Review), the Lake Lonely, Saratoga, Victory Mills, Kayaderosseras, Round Lake, Elnora and Niskayuna basins (Figures 6a and 6b) are considered locations of remnant glacial ice. The argument presented by these authors, and supported here, is that basin morphology is such that fluvial erosion of material is simply not a reasonable explanation of these features. From a sediment budget perspective, the amount of sediment that would have to be removed from the basins by fluvial action is not easily accounted for in the Hudson Valley. Further evidence for the Saratoga ice block can be inferred by ice marginal channels north of Malta Ridge.

Ballston Channel has also been considered by many (see Historical Review) as the former location of remnant ice. It has been argued that the extreme depth at the southern end of Ballston Lake requires an ice block to prohibit infilling by flow through Ballston Channel. This explanation is not considered satisfactory for the

following reasons. The southern end of Ballston Lake is at the base of a fairly steep drop in the Ballston Channel bed past its modern divide. Considering this gradient, the discharge, and the size of the channel, flow velocities through this reach must have been substantial. Therefore, the possibility of erosion in this deep part of the lake cannot be easily dismissed. Furthermore, the coarse bedload that might have filled this channel depression would have been deposited along with Scotia Gravel in Schenectady Basin. Sand-sized material could have been easily transported through this reach if it were ice free, considering it would have been transported through a much larger basin at Schenectady. The only available coarse bedload between Schenectady Basin and Ballston Lake is in the southern portion of Ballston Channel. The 4+ foot thick deposit along the western side of the lake of shale rich sand and gravel likely has its provenance in this channel reach. It is not known if this deposit is present on the floor of Ballston Lake, but the availability of such material is limited and perhaps not sufficient to fill the depression. If this scenario is realistic, the northward shallowing of Ballston Lake is a reflection of an Iromohawk competence-loss associated with the observed channel widening midway along its length.

Quaker Springs Base Level

Lake Iroquois discharge into Lake Quaker Springs deposited the Saratoga-Wilton sand plain (Figure 8). Chadwick (1928) was the only person to suggest this sand plain was an Iromohawk deposit (in Lake Albany). He left room for the possibility, however, that the sand plain was a Snook Kill delta, because the Mourning and Drummond Channels did not clearly intersect it. Chadwick failed to recognize the role of remnant ice between the channels and sand plain. Iromohawk drainage through Mourning and Drummond channels drained across, between, and around the Saratoga, Kayaderosseras, and Lake Lonely ice blocks, reaching the southern (proximal) end of the sand plain at, and northeast of, Saratoga Springs. The Albany II sand cap (320-330 feet elevation) on Malta Ridge stands as an erosional remnant to flow through Drummond and Mourning channels. This sand cap likely extended to the west and originally bordered Kayaderosseras Basin remnant ice.

Sand plains west, north, and south of Niskayuna Basin are at an elevation correlative with Lake Quaker Springs. The Elnora Channel was not likely active during Lake Quaker Springs because deltaform features bordering Colonie Channel at the end of Dwass Kill and Grooms Corners Channel are not incised by drainage to this base level. Therefore, the source of sands infilling Colonie Channel to the level of Lake Quaker Springs must have been along the course of the modern Mohawk River. Quaker Springs sands just south of Albany County International Airport are below and adjacent to Lake Albany Schenectady Delta sands. As Quaker Springs age deposits are absent from Niskayuna Basin itself, remnant ice must have still occupied the basin. The highest bedrock terrace along the southern wall of Mohawk Gorge began to develop as Iroquois waters drained across the Rexford sill and toward Niskayuna.

Coveville Base Level

Erosional surfaces at 260-270 feet elevation in Mourning, Drummond, and Round Lake channels indicate the channels continued to serve as distributary channels after Lake Quaker Springs (Figure 9). Clear erosional scarps in Mourning and Round Lake channels between this and the 300-310 Quaker Springs erosional surface distinguish this surface from Quaker Springs drainage.

In Mourning Channel, the 260-270 foot elevation erosional surface expands north of East Line, bifurcating around the fluvial rise northwest of Malta Ridge. Because there is nothing bounding this surface to the west and no headward-erosional features along the Kayaderosseras basin wall, flow to the west of the rise would have been bounded by Kayaderosseras Basin remnant ice. The apparent eastward redirection of the surface north of Malta Ridge possibly reflects flow around and not over remnant ice, but the flat-topped drumlinoid features in Kayaderosseras Basin at roughly an equivalent elevation might have been planed by flow across decaying ice. Upon encountering Saratoga Lake Basin, the surface was again redirected by remnant ice to develop the ice marginal channels north of Malta Ridge. *All of these channels join Saratoga Basin without any headward erosional features along the basin wall, again indicating water drained across ice.*

Although these Mourning Channel erosional surfaces are at approximately the same elevation, they decrease substantially in width to the north. Furthermore, the erosional surface west of Malta Ridge and bordering Kayaderosseras Basin truncates the slightly higher erosional surface bordering Malta Ridge. These features are interpreted as recording diminished flow through Mourning Channel during Lake Coveville time.

The sand capped ridge, at 270+ feet, between Saratoga and Lake Lonely basins was deposited as water drained across and between Saratoga and Lake Lonely ice blocks from erosional surfaces north of Malta Ridge. Sandy terraces along the south side of Fish Creek channel and the western side of the Hudson Valley southwest of Coveville at 270+ feet elevation record the high stand of Lake Coveville. Sands at 260+ feet north of Lonely Lake Basin and partially occupied today by Meadow Brook can be attributed to this Lake Coveville level as well. The Meadow Brook area was occupied by Lake Lonely ice during Lake Quaker Springs, and thereby prevented deposition

of Iromohawk-Quaker Springs sand. Meltback of Lake Lonely ice to the modern Lake Lonely Basin margin occurred following Lake Quaker Springs and during Lake Coveville.

A lowering of Lake Coveville to its stable 240 foot level (at Schuylerville) backed the lake out of Fish Creek channel at least as far as Grangerville. Clay flow shadows preserved downchannel of drumlins in the eastern part of Fish Creek Channel (DeSimone, 1977) developed during this lower Coveville phase. Sands at 240 feet elevation southwest of Coveville likely found their source from Fish Creek Channel and record continued use of the channel throughout Lake Coveville. The Victory Mills ice block must have remained throughout Lake Coveville, blocking deposition.

Drummond Channel records Coveville-Iromohawk channel bottoms at 270 foot elevation northeast of East Line, and at a 260+ foot elevation bedrock terrace at the outer Drummond Channel bend. Beyond the junction of Drummond Channel with Saratoga Basin, there is limited evidence of the course of Drummond Channel waters. Sand and gravel capping the largest of three mounds along the western margin of Saratoga Lake is similar to the club-shaped ridge between Lake Lonely and Saratoga Basins. This cap possibly represents deposition in a hole or crack in Saratoga Basin ice. It seems likely flow from Drummond Channel joined Mourning channel waters at the club-shaped ridge, but direct evidence of this has melted away with the Saratoga ice block.

In Round Lake Channel, Iromohawk drainage to Lake Coveville is recorded in two erosional terraces at 270+ feet and 260 feet elevation. The down-channel edge of these terraces has been truncated by drainage to lower base levels, so it is not clear how far the channel extended to the east. A lack of erosional features along the basin margin, indicates drainage was likely across Round Lake Basin ice. The south-central portion of the basin is a likely candidate for the position of an 'ice channel'. This position is suggested by the mound in the central portion of the basin, possibly the product of glacial debris filling a ice hole or crack.

The 290+ foot elevation sand plain in the northeastern portion of the basin may have been active as late as Lake Coveville, recording a slightly higher base level in Round Lake Basin during Lake Coveville than that in the Hudson Valley at Mechanicville. It seems possible that this sand plain was active no later than the high stand of Lake Coveville, as the gradient from a 290+ foot Round Lake Basin base level to 220 foot Mechanicville level seems too steep considering the discharge and composition of Anthony Kill Channel. In any case, the sand plain does not appear disturbed or affected by drainage through the basin, furthering the idea that flow was through the southern portion of the basin.

Sands at ~240 feet elevation between Round Lake Basin and Elnora ice possibly record the lower Coveville phase in the basin, as it is significantly lower than the northern 290+ foot elevation sand plain. Infilling to this level between ice blocks may indicate the blocks were originally connected, which prevented deposition of Albany II sands continuous with the Malta Sand Plain.

A thin layer of coarse sand and gravel east of Coons, hanging ~60 feet above the floor of Anthony Kill Channel, records drainage across Round Lake ice to Anthony Kill Channel. The coarse-grained poorly-sorted nature of the deposit suggests it may have been channel lag and therefore represents the Anthony Kill channel bottom during Lake Coveville time. Anthony Kill Channel developed in response to lowering Hudson Valley base levels and connected water adjacent to Round Lake Basin ice to the Hudson Valley. Sands at 240+ feet elevation south of Mechanicville record usage of Anthony Kill Channel during Lake Coveville time.

East of Niskayuna Basin, the modern channel began to develop toward Coveville base level. The elevation of the bedrock ridge at the Twin Bridges (I-87, the Northway) prior to downcutting by the modern Mohawk is not known. However, the approximate top of rock on the southern valley wall and peculiar break in slope along the bedrock ridge comprising the northern valley wall, both at 270 feet elevation, suggests this ridge of bedrock extended across the Niskayuna Basin outlet. The ridge would have dammed basin water and formed a waterfall between Niskayuna Basin and Lake Coveville. The cutback nature of the southern valley wall at Dunsbach Ferry resembles that of a former plunge basin and supports this position. This waterfall was perhaps 90 feet above the modern Mohawk (dammed) water level and 30 feet above Lake Coveville base level. This spillway/waterfall at the eastern edge of Niskayuna Basin would have allowed continued deposition of sands north and south of Niskayuna Basin above the Hudson Valley Coveville base level. Initial erosion of glacial and lacustrine sediment between Niskayuna Basin and Waterford was initiated during Lake Coveville time.

Modern Channel Development

A dramatic contrast exists between Iromohawk erosional and depositional features associated with Coveville and Quaker Springs base levels. The extent of deposits associated with Drummond and Mourning channels during Lake Quaker Springs time greatly outweighs those of Lake Coveville. Coveville erosional surfaces appear quite substantial in the channels themselves, but the diversion of flow by remnant ice in Kayaderosseras and Saratoga Basins and the closeness in elevations between Coveville erosional and depositional surfaces in Mourning and Drummond channels suggest flow was fairly slow and shallow. Conversely, erosional features in Mohawk Gorge

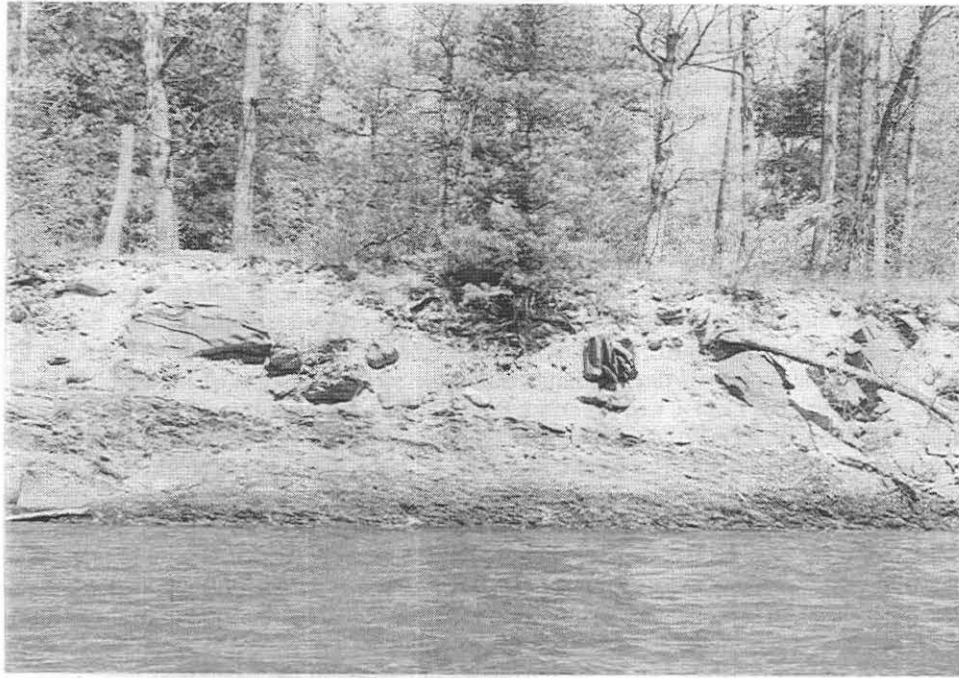


Figure 10: Iromohawk deposit in Mohawk Gorge overlying Schenectady Formation

appear to increase in energy from Albany II to Lake Coveville time. This shift in energy from distributary to modern channels is directly related to the lowering of the bedrock sill at Rexford.

Goat Island, at the downchannel end of Mohawk Gorge, and a 220+ foot elevation gravel-boulder terrace along the northern side of the gorge, are clear evidence that high discharge developed Mohawk Gorge. A bedrock island in a bedrock channel does not form under normal flow conditions. Such islands require simultaneous incision on both sides, normal flow would tend to develop one side preferentially. The position of the gravel-boulder terrace in the gorge requires it to be of fluvial origin. The deposit (Figure 10) lies on a bedrock surface that was exposed by postglacial erosion and is far enough away from the gorge walls to eliminate the incorporation of boulder clasts by mass wasting. The only agent capable and available to move boulder size material found in the deposit is fluvial. It is doubtful that the modern Mohawk is capable of transporting this size material, and furthermore, the modern Mohawk is erosive in this portion of the gorge today, indicating some change in flow conditions since the deposit was made. It seems reasonable that this change is the result of the modern Mohawk flowing across and incising the Iromohawk channel bottom, leaving it as a terrace. If this is the case, the modern Mohawk has cut through ~8 feet of Iromohawk channel lag and 20-25 feet of bedrock.

The 260+ foot elevation bedrock bench at Rexford is slightly lower than the isostatically corrected (DeSimone, 1985) Ballston Channel divide. Because the bench is lower, it can be inferred that the majority of Iromohawk water, following bench formation, drained through Mohawk Gorge. Bench development is interpreted as an indication of a fairly stable base level of significant duration. The Niskayuna Basin spillway maintained a basin base level very near that of Lake Quaker Springs during Lake Coveville time. This interpretation taken with the observed *Quaker Springs-Coveville transition of distributary channels features, indicates the majority of Iromohawk flow was contained in the modern channel during Lake Coveville time.*

Lowering of Mohawk Gorge to the 230+ foot elevation Rexford bench occurred only after the Niskayuna Basin spillway eroded, thereby lowering Niskayuna Basin base level and reactivating Mohawk Gorge erosion. Continued lowering of Mohawk Gorge allowed the transition from depositional to erosional environments in Schenectady Basin. The Scotia Gravel was incised, developing Scotia Channel and associated minor channels.

The symmetry of the small delta along the western flank of Ballston Lake indicates its development was unaffected by flow in the channel. The flow through Ballston Channel necessary to develop distributary channel features correlated with Lake Quaker Springs, suggests a post-Lake Quaker Springs age for delta formation. Its elevation above the modern channel divide indicates it pre-dates modern drainage. Furthermore, if the divide 'subchannel' formed because of waning flow through Ballston Channel during Lake Fort Ann time, and the 'subchannel' contained all of this flow, the delta elevation above the top of the 'subchannel' indicates its age can be

further constrained to Lake Coveville. This last point is purely conjecture, but internally consistent with the idea that Iromohawk discharge through Ballston Channel to Lake Coveville was clearly diminished relative to Lake Quaker Springs.

Fort Ann Base Level

Flow through Drummond Channel to Lake Fort Ann abandoned the bedrock terrace at the major channel bend as an erosional remnant. Based on the course of Drummond Creek today, it appears that remnant ice continued to force flow north along the western Saratoga ice margin (Figures 11 and 12). This northward flow direction might have been compounded by sediment deposition at the mouth of Drummond channel and adjacent to Saratoga Lake Basin ice.

Base level between East Line and Schuylerville was controlled by a bedrock spillway at ~210 feet elevation west of Coveville. Correcting for isostatic adjustment (DeSimone, 1985), this spillway would have caused an apparent base level at about the present day Saratoga Lake level. Drainage across the spillway developed "The Cove" at Coveville. The size of this feature has caused speculation that Iromohawk flow was necessary to develop it; however, sedimentation at ~210 feet in Fish Creek Channel and the poorly defined nature of the channel indicates this was not the case. Drainage of Saratoga Lake Basin waters across the spillway may have continued well into the Holocene before the lower Schuylerville outlet opened.

Headward erosional features in the Coveville channel bottom join the bottom of Mourning Channel at 220+ feet elevation and evidence drainage to a post-Coveville Kayaderosseras Basin base level. The narrow subchannel, now occupied by the Mourning Kill, developed as a result of this base level. The Mourning Kill is clearly misfit in this subchannel suggesting that it became misfit as Iromohawk flow diminished.

Two post-Coveville water levels are recorded in Round Lake Basin at 200 and 170-180 feet elevation. This level is further recorded around nearly the entire basin margin in the form of small deltas and sand plains. Significant basin infilling north of Little Round Lake and along the southern basin margin occurred during this base level, indicating significant remnant ice meltback during or prior to this base level. This 200 foot Round Lake Basin base level is correlated with the 180 foot elevation level of Lake Fort Ann at Mechanicville.

The lowest erosional surface in Round Lake Channel terminates in a small delta at the modern lake margin. This delta is composed almost entirely of shale clasts derived from Round Lake Channel. Creation of the 170-180 foot elevation sand plains in the southern and eastern portions of Round Lake Basin indicate further and perhaps accelerated meltback of Round Lake Basin ice during later Lake Fort Ann levels.

The extent of erosional and depositional features between East Line and Mechanicville, together with the diminished features north of East Line, suggest Round Lake Channel was the dominant distributary channel during Lake Fort Ann time. The shorter route (and steeper gradient) to Hudson Valley base levels allowed Round Lake Channel to undercut and pirate Drummond and Mourning Channels.

Drainage through Anthony Kill Channel lowered the channel floor below the Coons-Coveville age channel bottom. Flow through Anthony Kill Channel deposited a delta south and west of Mechanicville in the 180 foot Lake Fort Ann base level. The sedimentology of the deposit (Wall, 1995) suggests that material was locally derived and likely reworked from tills in the developing Anthony Kill Channel. The graded nature of the deposit suggests Iromohawk discharge was long-term and cyclic, not catastrophic.

Approximately 80 feet of downcutting in Anthony Kill Channel occurred following Mechanicville delta formation. The cutback feature in the northern channel wall at Willow Glen is a likely eddy feature associated with channel widening downchannel of the narrowest channel reach. The Anthony Kill Channel bottom is at the same elevation as an erosional surface just west of Reynolds on the eastern side of the Hudson River. The valley wall adjacent to this surface is concave suggesting modification during Anthony Kill Channel outflow. This, plus the misfit nature of the modern Anthony Kill, indicate a significant amount of drainage occupied the channel during later stages of Lake Fort Ann. Contrary to previous investigations, it is not necessary to flush the entire volume of Iromohawk water through the Anthony Kill Channel to develop these features. Because the channel bottom gradient today is 0.003, it seems quite reasonable that discharge on the order of the modern Mohawk could have developed these features. That is to say, only a portion of Iromohawk flow would be necessary.

The floor of the Elnora Basin is between 210 and 220 feet. This level is very close to the Fort Ann water level in Round Lake Basin, suggesting Elnora Basin may have filled with sediment from surrounding tributaries during Lake Fort Ann time. If this inference is correct, it indicates Elnora ice melted prior to or during Lake Fort Ann time.

Sometime following or in the latter stages of Lake Coveville, the Niskayuna Basin spillway lowered. This lowering allowed a reactivation of erosion in Mohawk Gorge and Schenectady Basin, the development of the 230+ foot elevation Rexford bench, erosion of Goat Island, and deposition of boulder gravel channel lag in Mohawk Gorge, which includes reworked Scotia Gravel.

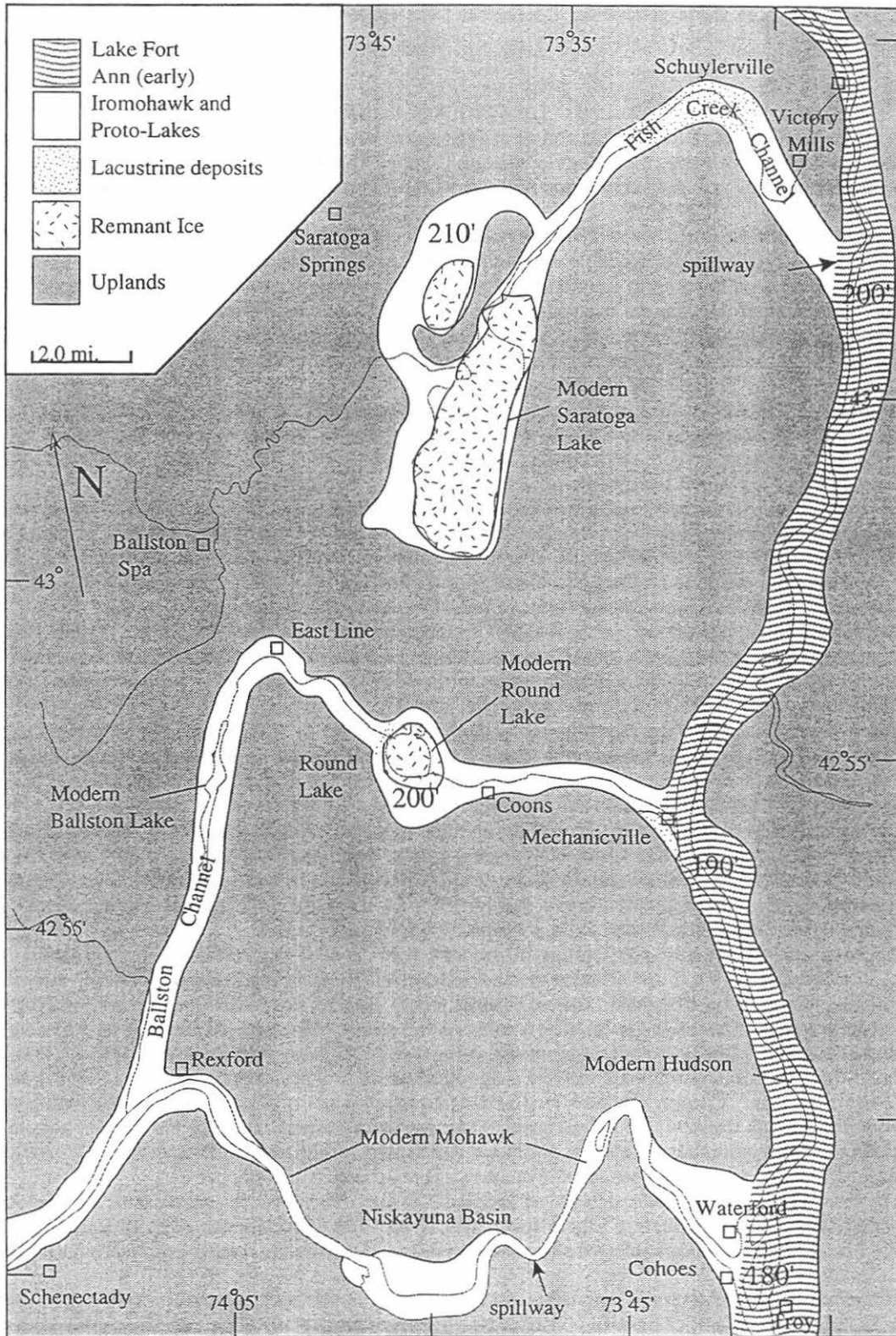


Figure 11: Map of Hudson/Mohawk Lowland during early stages of Lake Fort Ann.

The position of remnant ice in Niskayuna Basin during this period is not clear, but if it lingered during Lake Fort Ann time, it likely resided in the southern portion of the basin where the river flows today. Vischers Ferry sands at 200-210 feet elevation, are slightly higher than modern basin water level. It is not clear if this level is correlative with Lake Fort Ann or a localized base level associated with a lowered Niskayuna Basin spillway. It is also not clear if Stony Creek, the Mohawk or both are the source of these terrace sands, as ample sediment was likely available from both sources. The cutback feature west of Niska Isle is likely due to eddying flow emerging from Mohawk Gorge. The scale of the feature suggests it was developed by high discharge.

Iromohawk drainage stripped glacial and lacustrine material from the Taconic-age thrust slice west of Waterford (see Kidd et al., this volume). It is not clear if the north-south trending ridges and troughs in this area were gullied by fluvial action or reflect the structural geology, but in either case, these troughs likely focused flow to the south, developing the trench the modern river flows through. The pond in Waterford adjacent to the eastern edge of the slice is an artifact of a former plunge basin. The large headward erosional feature occupied by the 'Waterford flight' formed as water drained along the thrust slice margin, eroding the thick sequence of lacustrine sediments directly to the east.

The same argument for high discharge needed to develop Goat Island applies for islands at the modern Mohawk and Hudson River confluence. As some inter-island channels are ~30 feet deep, however, and the highest island is some 20 feet below the lowest level of Lake Fort Ann at Troy (~110'; DeSimone, 1985), it is difficult to envision bedrock erosion in ~50 feet of relatively quietstanding water as far as a mile away from where this discharge would enter Lake Fort Ann III. This observation raises the possibility that a high-discharge Mohawk drained to a Hudson Valley base level closely resembling that of the modern Hudson and clearly below the level of Lake Fort Ann III.

If evidence of high-flow discharge observed in the Anthony Kill and modern channels is attributable to the Iromohawk, and the Fort Ann water plane curves of DeSimone (1985) are accurate, then all Hudson Valley Fort Ann phases must have occurred prior to the Iroquois-Frontenac transition and are therefore correlative with Denny's (1974) Lake Fort Ann I in the Champlain Lowland. Clark and Karrow (1984) correlate a post-Frontenac Ontario Basin level with Lake Fort Ann II in the Champlain Lowland; therefore, if their correlation is correct, Iromohawk drainage had ceased well before Lake Fort Ann II in the Champlain Lowland. Because Lake Fort Ann II (Champlain Lowland) drained through the Hudson Valley, and the Iromohawk drained to a level below Lake Fort Ann III but chronologically before Lake Fort Ann II (Champlain Lowland) outflow, at least one base level is suggested between Lake Fort Ann III and the modern Hudson.

Alternatively, significant inputs from melting ice in the Catskills and Adirondacks could have increased post-Iromohawk discharge to a post-Fort Ann III base level, thereby developing the observed features at the Hudson-Mohawk confluence. Although this alternative is less attractive, both scenarios raise the possibility that Cohoes Falls underwent significant headward erosion prior to modern Mohawk drainage.

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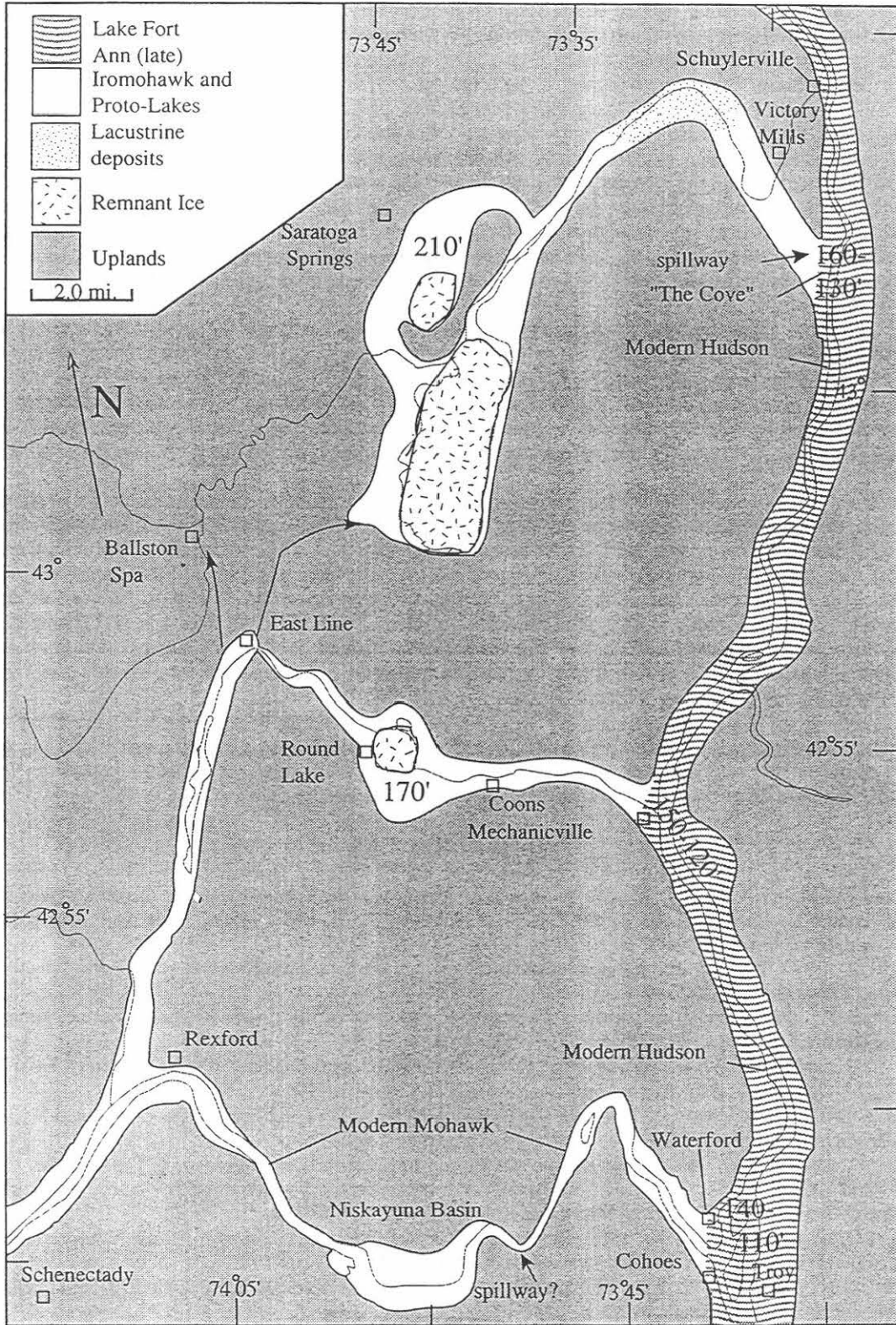


Figure 12: Map of Hudson/Mohawk Lowland during late stages of Lake Fort Ann.

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THE PALEOFLUVIAL RECORD OF GLACIAL LAKE IROQUOIS IN THE EASTERN MOHAWK VALLEY, NEW YORK

ROADLOG

<u>Miles</u>	<u>Log</u>
	From Butterfield Hall parking lot, turn right onto Union Ave.
0.0	At 1st traffic light set trip odometer and bear right onto Union St.
0.5	Left at 1st traffic light past railroad underpass.
0.7	Right at 2nd traffic light onto Rt. 5 west.
1.0	Left at 2nd traffic light to 890 west.
4.0	Mohawk River and Barge Canal Lock 8 on right.
4.7	Right on I-90 onramp. Past toll plaza, bear right onto I-90 west.
7.6	Mohawk Valley and Rotterdam Jct. on left.
21.2	<i>Schoharie Bridge - The original Thruway crossing of Schoharie Creek catastrophically failed in 1987 due to Schoharie Creek high flow and scour of bridge supports.</i>
23.4	<i>Mohawk River floodplain and approximate Iromohawk River channel bottom. Iromohawk depth in this location was approximately 50 feet.</i>
25.3	Leave Thruway - Exit 28, Fultonville Fonda - Toll \$0.65. Left turn after toll plaza. Mohawk River and Barge Canal on right.
26.3	Left at 1st traffic light onto Rt. 30A south.
26.6	Right on 5S west.
27.2	<i>Middle Ordovician Canajoharie Shale exposed on left.</i>
29.6	Left on Borden Rd.
29.7	Right on Dillenbeck Rd. and bear right past red house.
30.4	Stop 1 - Randall Expansion Bar overview. <i>Uplifted land associated with the Noses fault (Figure RL1) is visible to the west. At the base of the hill is the Iromohawk channel bottom, and beyond that the Randall expansion bar. The bar measures 7,000 feet in length and 1,800 feet at its maximum width. Thickness of the deposit is approximately 60 feet. One well log from the deposit indicates the bottom of the bar is approximately at the elevation of the modern floodplain (U.S. Geological Survey, Open File). The height of the bar above the floodplain therefore represents a minimum water depth (60 feet) for bar deposition.</i>
	Continue on road.
31.2	Right at stop sign. <i>Road descends to Iromohawk channel bottom. Randall bar is straight ahead and to the right.</i>
31.9	Right at stop sign onto 5S west. <i>Road climbs along Randall bar long axis.</i>
32.5	Stop 2 - Randall Expansion Bar. Pit owned by Santos Construction Company (SCC) - our access to the pit is under consideration by SCC. Contact: Dave Santos (owner) (518) 842-6201 <i>Large-scale east--dipping (downvalley) cross-bedding is visible throughout the exposed pit, confirming the eastward flow direction inferred from the deposits east-pointing tapered tail. Exposures in contact with the deposit surface show bedding that mimics external morphology. Overall, the Randall deposit is very poorly sorted with individual clasts ranging from small boulders to coarse sand.</i>
	Continue on Rt. 5S east. <i>Road descends along eastern tail of Randall bar.</i>
36.5	Left on 30A north.
36.8	Right at 1st traffic light.
37.3	Right at Thruway (I-90) entrance. I-90 east after toll plaza.
57.4	Leave Thruway - Exit 26, Rts. I-890 and 5S - Toll \$0.65.
58.3	Stay left past toll plaza - Rt. 5S Pattersonville.
61.0	<i>Road crosses top of Rotterdam Jct. portion of Scotia Gravel</i>
62.1	Right on Rt. 103
62.4	Cross Mohawk River and Barge Canal Lock 9

- 62.6 Right on Rt. 5 east
62.8 *Glenville Channel visible on left - one of four minor channels to incise the Scotia deposit of Scotia Gravel.*
- 63.1 *Partially cemented blocks of Scotia Gravel visible in old pit on left.*
65.5 **Stop 3** - Scotia Gravel and Mohawk River overlook.
If water level is low and interest high, hike from here to exposure of graded gravel foresets.
This exposure is marked by numerous east dipping graded gravel foreset beds. Dozens of these foresets are visible across the exposure, each grading from cobble to coarse sand and pebble gravel, and measuring approximately 3 feet in thickness (Figures 3 and 3). The cyclic nature of the graded foresets is more typical of a long term rather than catastrophic depositional event; furthermore, the cyclically suggests a seasonal variation transport energy, essentially gravel varves.
- in*
65.6 *Top of Scotia Gravel (~300' asl). Deposit continues to the northern (left) valley wall (3000+ feet)*
67.3 *This stretch of Rt. 5 runs along the top of a giant point bar extending to the south (right) into Schenectady Basin.*
- 67.9 Left at light onto Rt. 147 - Road climbs to top of Scotia Gravel.
69.3 Right on Vley Rd. and left at stop sign.
69.9 Left on Viele Rd.
70.1 **Stop 4** - Scotia Gravel overlook and paleodischarge discussion.
Another Scotia Gravel overlook, this one into a semi-active gravel pit.
Turn around and turn right back onto main road. Road parallels Harding Channel on left.
- 71.7 Bear left at yellow flashing light <<CAUTION - dangerous intersection>>
72.1 Left at light onto N. Ballston Ave. (Rt. 50 north)
73.3 Schenectady County Airport on right.
The airport sits atop a sand plain deposited in Lake Albany by the Glaciomohawk (pre-Iromohawk) and/or Alplaus Kill.
- 74.9 Right on Glenridge Rd.
75.5 *Road descends into Ballston Channel.*
76.4 Straight at light under railroad bridge. *Bottom of Ballston Channel.*
76.8 *Road climbs out of channel.*
77.2 Right on Rt. 146 south.
77.9 Left at light onto Riverview Rd.
78.5 First left onto Nott Rd. **Stop 5** - Mohawk Gorge
Directly across the gorge, a 1,200 foot wide bedrock bench stands at ~270 feet elevation. This bench descends and narrows to 260+ feet elevation some 6,000 feet downvalley where it pinches out at the narrowest gorge reach. A lower bedrock bench stands at 220+ feet.
Turn around and continue on Riverview Rd.
- 79.2 Stay right at fork.
83.5 Bear right at fork (stay on Riverview Rd.). *Village of Vischers Ferry is built on the floor of Niskayuna Basin, the southernmost of three giant kettles in the Hudson Mohawk Lowland. Niskayuna Basin is roughly 4 miles (E-W) by 2 miles (N-S). All three basins lie along the N-S trending preglacial Colonie Channel.*
- 85.0 *Original Erie Canal visible on right.*
87.2 Cross Northway (I-87).
87.3 Right turn at fork. *Road leaves Niskayuna Basin*
87.6 Left at end of road.
88.5 *Mohawk River on right.*
89.9 Right at stop sign.
90.0 Right at light onto Rt. 9 south.
90.2 Left onto Cohoes-Crescent Rd.
93.3 Left onto Front St. (just past old row houses).
93.4 Left onto Cataract St.
93.5 **Stop 6** - Cohoes Falls
The falls are dry throughout most of the year due to a diversion of flow for the Barge Canal and the hydroelectric plant immediately to the west.

Cohoes Falls has eroded back some 2000+ feet in post-glacial time. Inferences from bedrock islands at the Hudson -Mohawk confluence immediately downvalley suggest much of this erosion may have occurred during Iromohawk drainage. Iromohawk flow exposed and eroded the top of bedrock on both sides of the gorge.

- Continue on Cataract St. and turn left on Cohoes-Crescent Rd.
- 94.3 Left at light onto Rt. 32 north.
- 94.5 *Cross Mohawk River - Bedrock islands visible to right.*
- 95.7 *Barge Canal locks 2 and 3 visible to right and left respectively.*
- 96.0 Left on Rt. 4 north.
- 104.9 Left at light onto Rt. 67 west.
- 105.5 *Enter Anthony Kill Channel - this channel developed as Round Lake Basin waters drained toward falling Hudson Valley glacial lake levels.*
- 106.8 *Willow Glen channel expansion (on right).
Anthony Kill Channel dramatically expands to the north. This feature was likely carved out by eddying flow through Anthony Kill Channel, eroding the less resistant Willow Glen kame delta (the majority of the channel walls are composed of till and bedrock).*
- 108.3 Turn left on unnamed road
- 108.5 Stop just past bridge to pull off on right. **Stop 7** - Anthony Kill Channel.
A classic example of a misfit stream.
- Turn around and continue on Rt. 67 west.
- 109.0 *Enter Round Lake Basin, the middle of three giant kettles aligned with the preglacial Colonie Channel. Round Lake Basin is nearly circular with a diameter of apx. 2.5 miles*
- 111.2 Left onto County Rt. 80
- 111.5 *Road crosses top of small delta at mouth of Round Lake Channel.*
- 111.9 Cross Rt. 9 at light.
- 112.7 Right onto Interstate 87 north.
- 113.1 *Cross Round Lake Channel. Flow was from left to right into Round Lake Basin.*
- 114.2 Exit 12 off I-87, left at light at end of off-ramp.
- 115.4 Left on Rhule Rd. No.
- 116.0 At end of road pull off to right. **Stop 8** - Round Lake Channel
Visible in this portion of Round Lake Channel are two bedrock erosional surfaces which grade to Round Lake Basin water levels correlative with lakes Coveville and Fort Ann in the Hudson Valley. The upper Coveville surface is distinguished from the lower Fort Ann by the bedrock channel wall flanking Rhule Rd. No. The opposite channel wall is visible below the power lines crossing the channel.
- Turn around and double back on Rt. 67 east.
- 117.6 Crossover I-87.
- 118.1 Cross Rt. 9. - *Road runs across the Malta Sand Plain, a Glaciomohawk deposit into Lake Albany II.*
- 119.9 *Road descends into Saratoga Lake Basin, the northernmost of three giant kettles aligned with the preglacial Colonie Channel. Saratoga Lake Basin has approximate dimensions of 2 miles (E-W) by 6 miles (N-S).*
- 120.4 Left at 'T' onto Rt. 9P. *Saratoga Lake is visible straight ahead.*
- 120.9 Ascend back onto Malta Sand Plain.
- 122.3 Cross Rt. 9
- 122.9 *Road descends into Ballston Channel. Ballston Channel is the middle of three distributary channels branching from the northern end of Ballston Channel at East Line. Drummond Channel connects with Saratoga Lake Basin to the northeast (right).*
- 123.1 Cross over I-87.
- 123.4 *Road ascends out of channel onto Malta Ridge, an erosional remnant between distributary channels (Drummond and Mourning) equivalent with the Malta Sand Plain.*
- 123.6 Straight at 4-way stop.
- 124.3 Left onto East Line Rd. *Road parallels Mourning Channel (to right).*
- 126.3 Straight at traffic light. *East Line - Point at which Ballston Channel splits into the distributary Mourning, Drummond, and Round Lake Channels.*
- 126.8 Right onto Lake Rd. *Road descends into Ballston Channel.*
- 127.9 Right onto Outlet Rd.
- 128.2 **Stop 9** - Northern end of Ballston Lake in Ballston Channel.

GEOCHRONOLOGY AND PETROGENESIS OF ADIRONDACK IGNEOUS AND METAMORPHIC ROCKS

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INTRODUCTION AND GEOCHRONOLOGY

The Adirondacks form a southwestern extension of the Grenville Province (figure 1) and are physiographically divided into the Adirondack Highlands (granulite facies) and Lowlands (amphibolite facies) by a broad zone of high strain referred to as the Carthage-Colton Mylonite Zone (figs. 2,3) which is continuous with the Chibougamau-Gatineau line, or Labelle shear zone, in Canada (AB on figure 1). Together these two zones separate the Grenville Province into two major blocks with the Central Granulite Terrane (CGT) lying east of AB and the Central Metasedimentary Belt (CMB) and Central Gneiss Belt (CGB) lying to the west. Within the southwestern portion of the Grenville Province further subdivisions exist and are shown in figure 3.

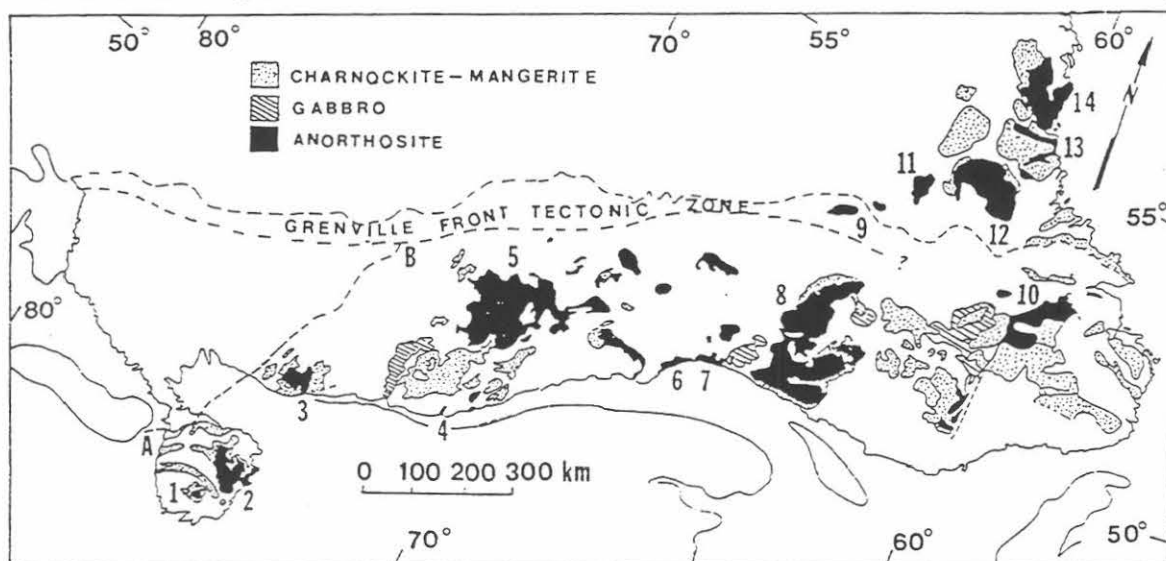


Figure 1. Generalized map of anorthositic massifs within the Grenville Province and adjacent Labrador. The dashed line, AB, separates terranes with anorthosite massifs on the east from ones lacking them on the west and corresponds to the Carthage-Colton-Gatineau-Chibougamau Line. 1-Snowy Mt. and Oregon domes (ca. 1130 Ma); 2-Marcy massif (ca. 1135 Ma); 3-Morin anorthosite and Lac Croche complex (1160 ± 7 Ma); 4-St. Urbain anorthosite (ca. 1070 Ma); 5-Lac St. Jean complex (1148 ± 4 Ma); 6-Sept Isles (1646 ± 2 Ma); 7-8-Havre St. Pierre complex (1126 ± 7 Ma) including the Pentecote (1365 ± 7 Ma) anorthosite; 9-Shabagamo intrusives; 10-Mealy Mts. anorthosite (1646 ± 2 Ma); 11-12-Harp Lake anorthosite (ca. 1450 Ma); 13-Flowers River complex (ca. 1260 Ma); 14-Nain complex (1295 Ma) including Kiglapait intrusive (1305 ± 5 Ma). From McLelland (1989).

As demonstrated by recent U-Pb zircon and Sm-Nd geochronology summarized (table 1) by McLelland and Chiarenzelli (1990) and McLelland et al. (1993), the Adirondack-CMB sector of the Grenville Province contains large volumes of metaigneous rocks that represent recent (i.e., ca. 1400-1200 Ma) additions of juvenile continental crust. These results (figure 4) indicate that the Adirondack-CMB region experienced wide-spread calcalkaline magmatism ca. 1300-1220 Ma. Associated high grade metamorphism has been fixed at 1226 ± 10 Ma by Aleinikoff (pers. comm.) who dated dust air abraded from metamorphic rims on 1300 Ma zircons. Identical rocks, with identical ages, have been described from the Green Mts. of Vermont by Ratcliffe et al. (1991), in northern Ireland by Menuge and Daly (1991), in the Llano uplift of Texas (Walker, 1993) and in the Texas-Mexico belt of Grenville rocks by Patchett and Ruiz (1990). It appears, therefore, that a major collisional-magmatic belt was operative along the present southern flank of the Grenville Province during the interval 1300-1220 Ma and may have been related to the assembly of a supercontinent. More locally, this magmatism and associated metamorphism, represents the Elzevirian Orogeny of the Grenville Orogenic Cycle, as defined by Moore and Thompson (1980). Within the Adirondacks, Elzevirian rocks are represented by 1300-1220 Ma tonalites and alaskites whose distribution is shown in figure 5. The apparent absence of this suite from the central Highlands is believed to be the combined result of later magmatic intrusion and recent doming along a NNE axis. Within the Frontenac-

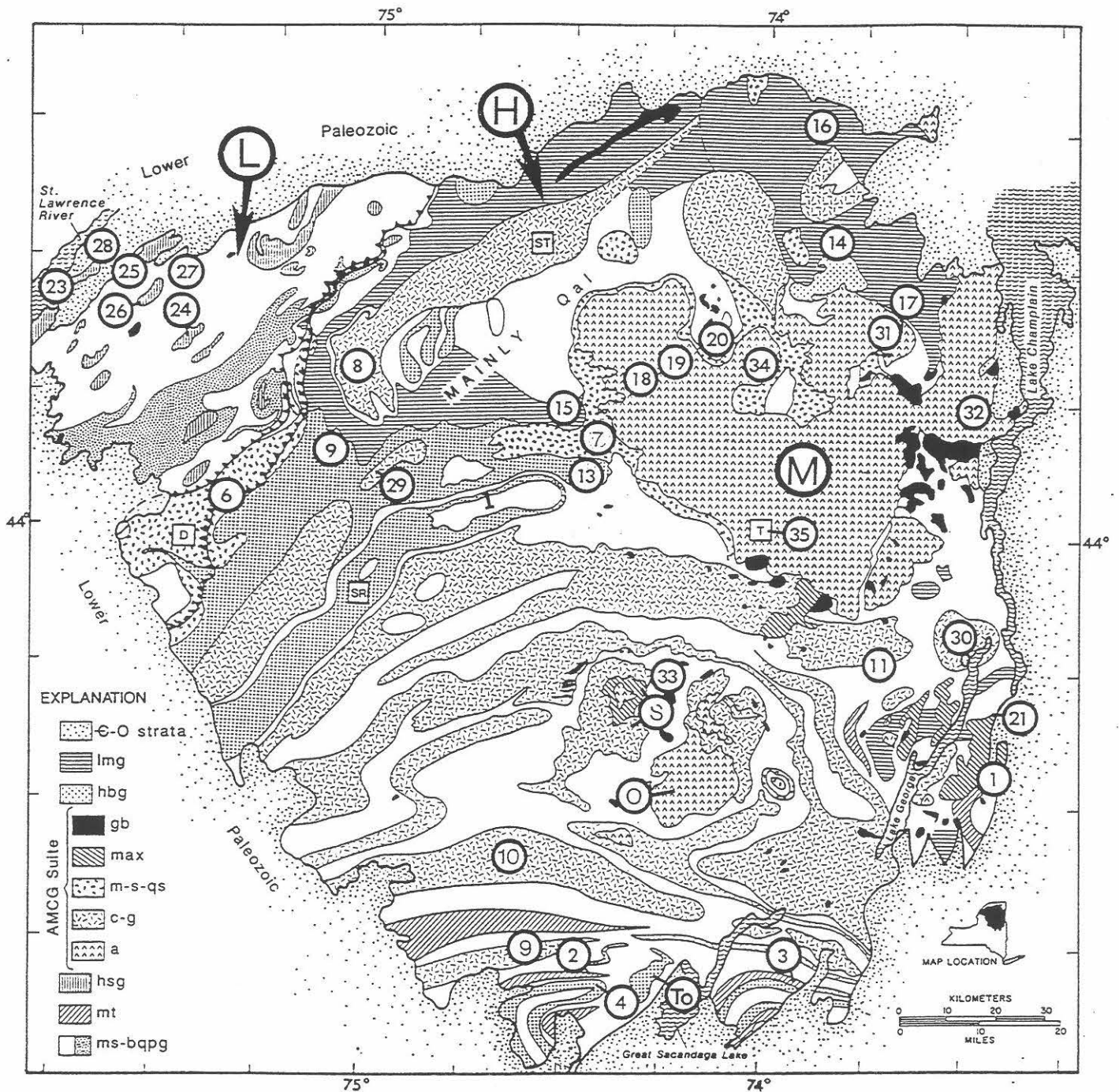


Fig. 2. Generalized geologic map of the Adirondack Highlands (H) and Lowlands (L). The Carthage-Colton Mylonite Zone (CCMZ) is shown with sawteeth indicating directions of dip. Numbers refer to samples listed in Tables 1 and 2. Map symbols: lmg=Lyon Mt. Gneiss, hbg=hornblende-biotite granitic gneiss, gb=olivine metagabbro, max=mangerite with andesine xenocrysts, a=metanorthosite, m-s-qs=mangeritic-syenitic-quartz-syenitic gneiss, ms=metasediments, bqpg=biotite-quartz-plagioclase gneiss, hsg=Hyde School Gneiss, mt=metatolalitic gneiss. Locality symbols: A=Arab Mt. anticline, C=Carthage anorthosite, D=Diana complex, O=Oregon dome, S=Snowy Mt. dome, ST=Stark complex, SR=Stillwater Reservoir, T=Tahawus, To=Tomantown pluton. From McLelland and Chiarenzelli (1990) and Daly and McLelland (1991).

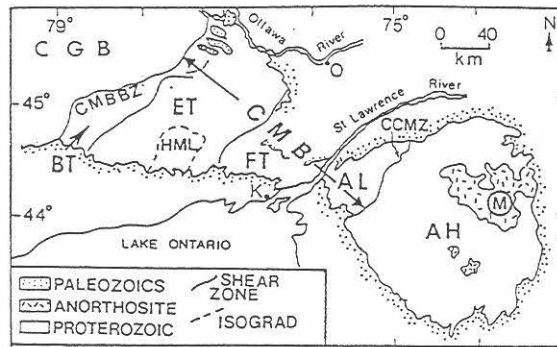


Fig. 3. Southwestern Grenville Province. CMB=Central Metasedimentary Belt, CGB=Central Gneiss Belt, BT=Bancroft Terrane, ET=Elzevir Terrane, FT=Frontenac Terrane, AL=Adirondack Lowlands, HL=Adirondack Highlands, HML=Hastings metamorphic low, K=Kingston, O=Ottawa, CCMZ=Carthage-Colton Mylonite Zone, M=Marcy massif.

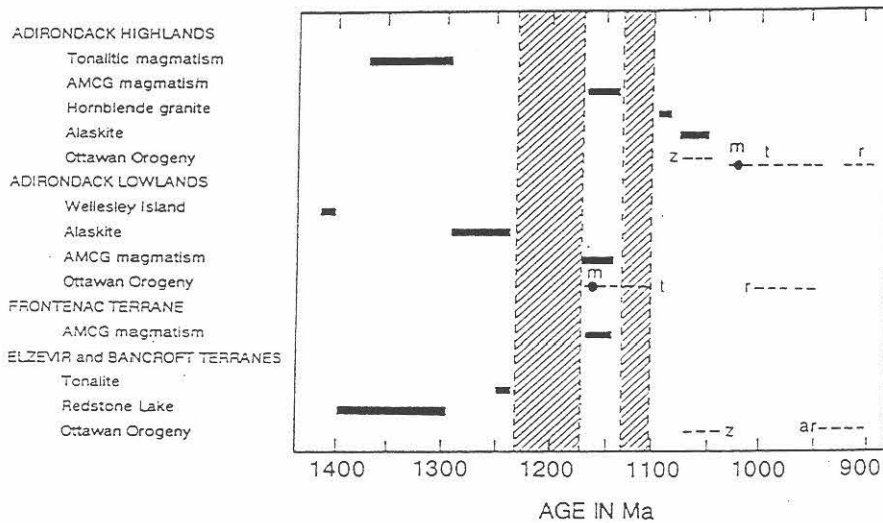


Fig. 4. Chronology of major geological events in the southwestern Grenville Province. z=zircon, t=titanite, m=monazite, r=rutile, ar=Ar/Ar. Diagonal ruling=quiescence. From McLelland and Chiarenzelli (1991).

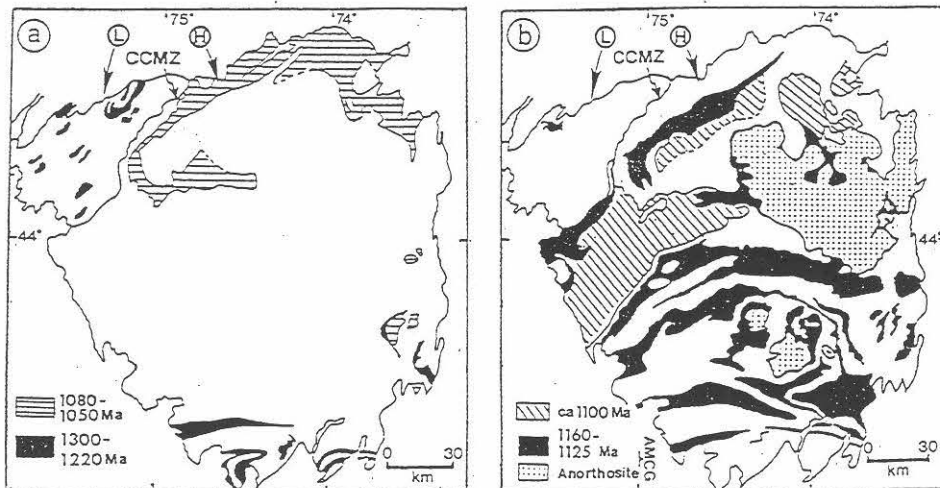


Fig. 5. Chronological designation of Adirondack units. L=Adirondack Lowlands, H=Adirondack Highlands, CCMZ=Carthage-Colton Mylonite Zone. From Chiarenzelli and McLelland (1991).

TABLE 1
U-PB ZIRCON AGES FOR META-IGNEOUS ROCKS
OF THE ADIRONDACK MOUNTAINS

No.	Age (Ma)	Location	Sample No.
HIGHLANDS			
Tonalitic gneiss and older charnockite			
1	1329 ±37	South Bay	AM-87-12
2	1301*	Canada Lake	AM-86-12
3	1336*	Lake Desolation	LDT
4	1233*	Canada Lake	AM-87-13
Mangeritic and charnockitic gneiss			
5	1155 ±4	Diana complex	
6	1147 ±10	Stark complex	AM-86-15
7	1134 ±4	Tupper Lake	AC-85-6
8	1125 ±10	Schroon Lake	9-23-85-7
Older hornblende granitic gneiss			
9	1156 ±3	Rooster Hill	AM-86-17
10	1150 ±5	Piseco dome	AM-86-9
11	1146 ±5	Oswegatchie	AC-85-2
Younger hornblende granitic gneiss			
12	1100 ±12	Carry Falls	AM-86-3
13	1098 ±4	Tupper Lake	AM-86-6
14	1093 ±11	Hawkeye	AM-86-13
Alaskitic gneiss			
15	1075 ±17	Tupper Lake	AM-86-4
16	1073 ±5	Dannemora	AM-86-10
17	1057 ±10	Ausable Forks	AM-86-14
Anorthosite and metagabbro			
18	1054 ±20	Saranac Lake	AC-85-8 [§]
19	1050 ±20	Saranac Lake	AC-85-7 [§]
20	996 ±5	Saranac Lake	AC-85-9
Xenolith-bearing olivine metagabbro			
21	1144 ±7	Dresden Station	AM-87-11
22	1057	North Hudson	CGAB**
LOWLANDS			
Leucogranitic gneiss			
23	1415 ±5	Wellesley Island	AM-86-16
Alaskitic gneiss			
24	1284 ±7	Gouverneur dome	AC-85-4
25	1236 ±5	Fish Creek	AM-87-4
26	1230 ±33	Hyde School	AC-85-5
Granitic and syenitic gneiss			
27	1160 ±4	Edwardsville	AM-87-5
28	1155 ±15	North Hammond	AM-87-3
HIGHLAND SAMPLES OF SILVER (1969)^{§§}			
29	1113 ±10	Fayalite granite, Wanakena	
30	1109 ±11	Charnockite, Ticonderoga	
31	1084 ±15	Undeformed syenite dike, Jay	
32	1074 ±10	Anorthosite pegmatite, Elizabethtown	
33	1064 ±10	Metanorite, Snowy Mountain dome	
34	1054 ±20	Sheared anorthosite pegmatite, Jay	
35	1009 ±10	Magnetite-ilmenite ore, Tahawus ^{##}	

Note: Errors at two sigma.

*Minimum Pb/Pb age.

§Data from Grant et al (1986).

§Contains zircon cores >1113 Ma, air abraded.

§Baddeleyite age of >1086 ±5 Ma from this sample.

**Contains baddeleyite >1109 Ma.

Monazite age of 1137 ±1 Ma.

§§Decay constants of Steiger and Jager (1977).

##Location same as Sanford Lake (SL) in Figure 1.

Table 2.: Sm-Nd data (sample numbers as in Table 1)

sample	L	Zircon age ¹	t _{DM} ²
ADIRONDACK HIGHLANDS			
Tonalites			
1 :AM87-12	t	1329 ± 36	1403
2 :AM86-12	t	1307 ± 2	1366
3 :LDT	t	>1366	1380
AMCG granitoids			
5 :DIA	s	1155 ± 4	1430
6 :AM86-15	r	1147 ± 10	1495
7 :AC85-6	m	1134 ± 4	1345
9 :AM86-17	e	1156 ± 8	1436
10 :AM86-9	g	1150 ± 5	1346
Younger granitoids			
13 :AM86-6	gd	1098 ± 4	1314
15 :AM86-4	a	1075 ± 17	1576
(repeat)			
:SK2A	tr	c.1060	1330
(repeat)			1373
Metasediment			
:JMCL-1	p	>c.1330	2075
Gabbro			
21:Ali-1	g	1144 ± 7	1331
ADIRONDACK LOWLANDS			
Wellesely Island			
23:AM86-16	l	1415 ± 6	1440
Fish Creek			
25:AM87-4	a	1236 ± 6	1210
:5/90-5	t		
Hyde School			
26:AC85-5	a	1230 ± 33	1351
:HS3	t	1230 ± 33	1397
:HS4	t	1230 ± 33	1350
Gouverneur			
24:AC85-5	a	1284 ± 7	1525
ELZEVIR TERRANE			
Northbrook			
9/88-9	t	1250	1245
Elzevir			
9/88-10	t	1275	1397

1: U-Pb zircon ages in Ma from McLelland and Chiarenzelli (1990a,b) and Grant et al. (1986); 2: Sm-Nd model ages in Ma (DePaolo 1981) from Daly and McLelland (1991) for the Highlands and McLelland, Daly and Perham (1991) for the Lowlands; L: lithologies, a=alaskite, e=enderbite, g=granite, gd=granodiorite, m=mangerite, p=pelite, s=syenite, t=tonalite, tr=trondhjemite, l=leucogranite, initial digits of sample numbers refer to localities in Fig. 2.

Adirondack region, the Elzevirian Orogeny was followed by 40-50 Ma of quiescence terminated at 1170-1130 Ma by voluminous anorogenic (figure 4) magmatism referred to as the anorthosite-mangerite-charnockite-granite (AMCG) suite. The older ages are characteristic of AMCG magmatism in the Frontenac Terrane (including the Lowlands) while the Highlands commonly exhibit ages of 1150-1130 Ma (figure 5). The large Marcy anorthosite massif (figure 2) and its associated granitoid envelope were emplaced at ca. 1135 Ma (McLelland and Chiarenzelli, 1990). These ages are similar to those determined (Emslie and Hunt, 1990) for the Morin, Lac St. Jean, and several other large massifs farther northeast in the Grenville Province (figure 1). Rocks of similar age and chemistry (i.e., Storm King Granite) have been described within the Hudson Highlands (Grauch and Aleinikoff, 1985). The extremely large dimensions of the AMCG magmatic terrane emphasize its global-scale nature corresponding, perhaps, to supercontinent rifting with the rifting axis located farther to the east. Valley et al. (1990) and McLelland et al. (1991) have provided evidence that contact, and perhaps also regional, metamorphism accompanied emplacement of hot (1100°C), hypersolvus AMCG magmas. Wollastonite and monticellite occurrences related to thermal pulses from AMCG intrusions occur in proximity to AMCG intrusions (Valley et al., 1990). In the Lowlands and the Canadian sector of the Frontenac Terrane, monazite (table 1., no. 28), titanite, and garnet ages (Mezger et al., 1992) all indicate high temperatures (600-800°C) at ca. 1150 Ma. Rutile ages and Rb/Sr whole rock isochron ages document temperatures not exceeding 400 °C at ca. 1050-1000 Ma.

Following approximately 30 Ma of quiescence (figure 4), the Adirondacks, along with the entire Grenville Province, experienced the onset of the Ottawa Orogeny of the Grenville Orogenic Cycle. Initially the Ottawa Orogeny was represented by 1090-1100 Ma hornblende granites in the northwest Highlands. These rather sparse granites were followed by deformation, high grade metamorphism, and the emplacement of trondhjemitic to alaskitic magnetite-rich rocks (Lyon Mt. Gneiss of Whitney and Olmsted, 1988) in the northern and eastern Adirondacks. The zircon ages of these rocks fall into an interval of 1050-1080 Ma (table 1) which corresponds to the peak of granulite facies metamorphism when crust, currently at the surface, was at 25 km. Accordingly, the alaskitic to trondhjemitic rocks are interpreted as synorogenic to late-orogenic intrusives. They were followed by the emplacement of small bodies of fayalite granite (ca. 1050 Ma) at Wanakena and Ausable Forks (figure 2).

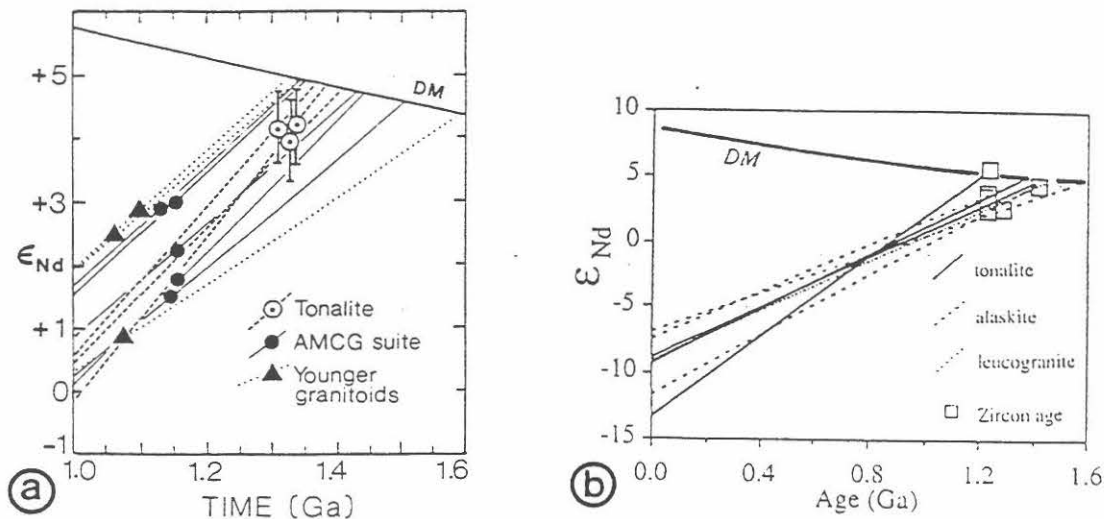


Figure 6. Epsilon-Nd diagrams for orthogneisses of the Adirondack Highlands (A) and Lowlands (B). Symbols fixed by zircon ages.

Sm-Nd analysis (Daly and McLelland, 1991) demonstrates that the emplacement ages of the ca. 1300 Ma tonalitic rocks of the Highlands correspond closely to their neodymium model ages (table 1 and figure 6) indicating that these represent juvenile crustal additions. As seen in figure 6, ϵ_{Nd} evolution curves for AMCG and younger granite suites pass within error of the tonalitic rocks and suggest that the tonalites, together with coeval granitoids, served as source rocks for succeeding magmatic pulses. Remarkably, none of these igneous suites gives evidence for any pre-1600 Ma crust in the Adirondack region and the entire terrain appears to have come into existence in the Middle to Late Proterozoic. Significantly, Sm-Nd

TABLE 3

<u>EARLY CALCALKALINE ROCKS</u>				<u>OLDER ANOROGENIC PLUTONIC ROCKS</u>							
	AM-87-13	TOE	CL-6	AM-86-17	AM-86-1	AC-85-1	AM-86-9	AM-86-15	AC-85-2		
SiO ₂	65.00	65.68	74.63	68.90	71.88	73.72	69.14	67.47	75.17		
TiO ₂	0.75	1.16	0.37	0.59	0.36	0.04	0.89	0.72	0.20		
Al ₂ O ₃	15.10	14.97	14.22	14.50	14.82	13.54	13.78	15.12	12.63		
FeO	Nd	Nd	Nd	2.16	1.27	0.87	2.83	3.34	1.11		
Fe ₂ O ₃	6.03	6.79	1.53	1.1	0.96	0.11	1.82	1.59	1.07		
MnO	0.10	0.08	0.04	0.02	0.03	0.01	0.04	0.10	0.02		
MgO	0.46	1.15	0.55	0.84	0.43	0.20	0.45	0.51	0.19		
CaO	2.71	2.56	1.66	2.3	1.87	0.85	2.26	2.57	0.88		
Na ₂ O	4.10	2.80	3.56	3.06	3.93	5.71	3.07	3.41	2.99		
K ₂ O	5.13	4.52	4.26	4.18	3.99	4.42	4.91	5.18	5.49		
P ₂ O ₅	0.10	0.51	0.10	0.24	0.09	0.01	0.23	0.19	0.04		
LOI	<u>0.39</u>	<u>0.74</u>		<u>0.40</u>	<u>0.27</u>	<u>0.17</u>	<u>0.19</u>	<u>0.39</u>	<u>0.17</u>		
Σ	99.87	100.96	99.42	99.63	99.65	99.61	100.59	99.96			
Ba(ppm)	1230	710	510	680	660	1100	736	810	442		
Rb (ppm)	100	97	170	160	200	406	81	128	230		
Sr (ppm)	260	230	260	200	130	26	211	215	99		
Y (ppm)	70	70	37	40	110	321	60	71	77		
Nb (ppm)	20	19	17	30	30	15	19	21	15		
Zr (ppm)	790	345	160	270	670	118	538	546	284		
<u>YOUNGER ANOROGENIC PLUTONIC ROCKS</u>											
	AC-85-6	AC-85-10	AM-86-7	WPG	SLC	AM-87-9	AM-86-8	AM-87-10			
SiO ₂	62.15	54.94	58.50	69.20	60.64	61.05	60.94	62.14			
TiO ₂	0.88	1.55	0.68	0.51	1.14	0.78	1.39	0.36			
Al ₂ O ₃	16.40	14.87	20.32	13.90	15.27	15.98	15.91	12.35			
FeO	3.96	10.25	2.81	3.1	9.28	4.60	6.51	10.32			
Fe ₂ O ₃	1.49	2.80	0.43	1.34	1.77	2.10	1.02	1.7			
MnO	0.09	0.24	0.01	0.05	0.19	0.05	0.14	0.01			
MgO	1.06	0.96	1.47	0.52	0.74	1.64	1.70	0.83			
CaO	3.27	5.52	6.16	2.03	3.97	3.63	4.53	3.65			
Na ₂ O	4.81	3.45	5.02	3.02	3.34	3.41	3.55	6.05			
K ₂ O	5.13	3.83	3.35	5.48	3.70	4.76	3.86	1.26			
P ₂ O ₅	0.30	0.65	0.32	0.11	0.31	0.42	0.46	0.09			
LOI	<u>0.41</u>	<u>0.37</u>	<u>0.43</u>	<u>0.39</u>	<u>0.01</u>	<u>0.91</u>	<u>0.37</u>	<u>0.67</u>			
Σ	99.95	99.43	99.50	99.65	100.46	99.63	100.38	99.24			
Ba(ppm)	850	625	Nd	1279	823	Nd	1100	Nd			
Rb (ppm)	106	47	Nd	124	87	Nd	83	29			
Sr (ppm)	335	367	Nd	184	215	Nd	410	180			
Y (ppm)	60	55	Nd	48	121	Nd	110	63			
Nb (ppm)	21	14	Nd	14	38	Nd	25	79			
Zr (ppm)	464	431	Nd	382	647	Nd	1200	309			
<u>YOUNGER GRANITIC ROCKS</u>						<u>LATE LEUCOGRANITIC ROCKS</u>					
	AM-86-3	AM-86-6	NO-Fo1	AM-86-13	AM-87-6	GHA	AM-86-11	AM-87-7	AM-86-4	AM-86-10	AM-86-14
SiO ₂	68.62	68.05	71.75	76.30	69.00	73.2	69.98	67.80	70.01	69.05	72.39
TiO ₂	0.48	0.55	0.41	0.18	1.43	0.35	0.46	0.42	0.69	0.57	0.38
Al ₂ O ₃	14.37	14.67	13.49	11.64	12.12	13.1	12.37	15.76	12.43	13.06	12.63
FeO	3.01	3.51	2.39	1.13	4.94	0.84	5.13	1.2	4.23	3.50	1.6
Fe ₂ O ₃	0.93	1.18	1.12	0.61	1.16	2.1	1.11	2.8	1.5	1.42	4.13
MnO	0.06	0.07	0.05	0.01	0.02	0.02	0.14	0.03	0.03	0.01	0.03
MgO	0.49	0.45	0.11	0.01	0.67	0.33	0.08	0.56	0.01	0.17	0.29
CaO	1.99	1.81	1.43	0.45	0.75	1.55	1.25	2.35	0.94	0.35	1.07
Na ₂ O	3.71	3.81	2.99	3.32	2.79	4.16	3.99	3.71	1.92	1.08	6.63
K ₂ O	5.67	5.61	5.79	5.22	6.50	4.01	4.91	4.91	8.34	9.64	0.52
P ₂ O ₅	0.13	0.12	0.07	0.03	0.17	0.07	0.04	0.13	0.17	0.12	0.70
LOI	<u>0.40</u>	<u>0.58</u>	<u>0.5</u>	<u>0.30</u>	<u>0.25</u>	<u>0.39</u>	<u>0.21</u>	<u>0.31</u>	<u>0.23</u>	<u>0.41</u>	<u>0.10</u>
Σ	99.86	100.41	100.00	99.20	99.80	100.3	99.67	99.98	100.50	99.38	100.47
Ba(ppm)	861	715	692	Nd	1014	1249	160	Nd	840	290	98
Rb (ppm)	161	148	182	188	214	178	190	Nd	315	330	16
Sr (ppm)	209	240	115	132	303	211	20	Nd	73	60	42
Y (ppm)	65	62	72	157	35	86	120	Nd	66	75	117
Nb (ppm)	20	20	25	29	17	19	30	Nd	18	23	27
Zr (ppm)	394	542	595	392	507	338	1230	Nd	414	600	786

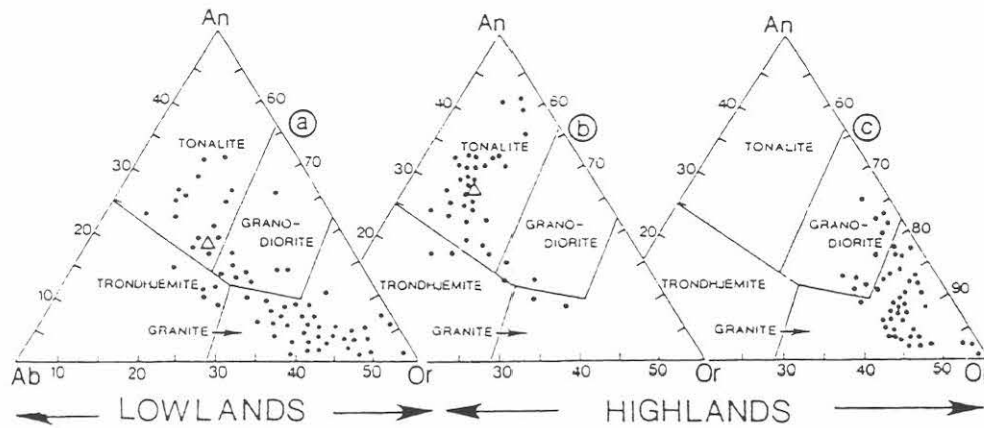


Figure 7. Plots of normative albite (Ab)-anorthite (An)-orthoclase (Or) for (a) Hyde School Gneiss, (b) Highlands tonalites, and (c) Tomantown pluton. Open triangles give average values for tonalitic samples. Definition of fields due to Barker (1979).

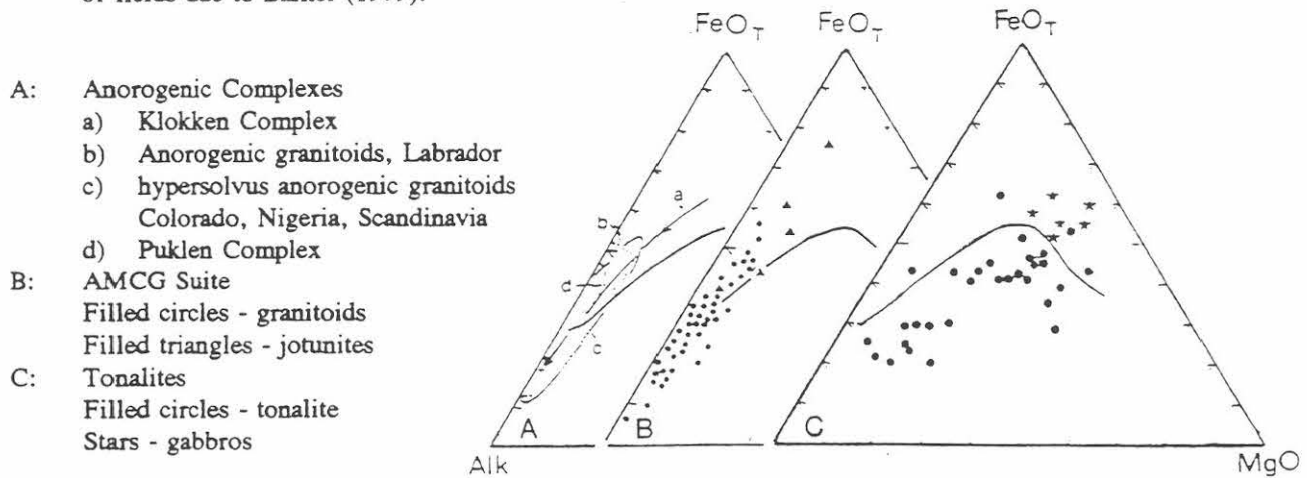


Figure 8. AFM variation diagrams for A) anorogenic complexes, B) AMCG suite, C) Highlands tonalites and associated gabbro (see McLelland 1991 for sources).

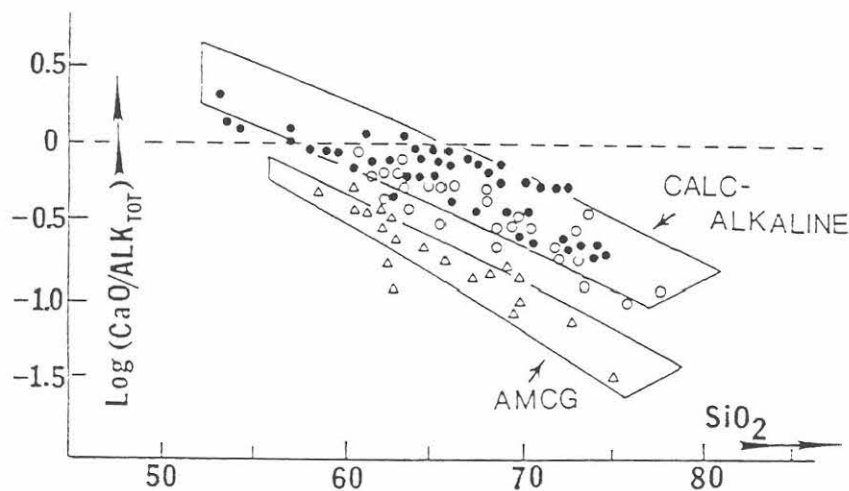


Figure 9. Calcalkali ratio vs. weight percent for AMCG granitoids (triangles), tonalites (closed circles) and Tomantown pluton granitoids (open circles).

analysis for the ca. 1230-1300 Ma tonalitic to alaskitic Hyde School Gneiss at the Lowlands (table 1, figure 6) demonstrates that it has model neodymium ages and ϵ_{Nd} values similar to Highland tonalites (McLelland et al., 1993). These results are interpreted to reflect the contiguity of the Highlands and Lowlands at ca. 1300 Ma. Given this, the Carthage-Colton Mylonite Zone is interpreted as a west-dipping extensional normal fault that formed during the Ottawa Orogeny in response to crustal thickening by thrust stacking (McLelland et al., 1993). East dipping extensional faults of this sort and age have been described by van der Pluijm and Carlson (1989) in the Central Metasedimentary Belt. Motion of this sort along the Carthage-Colton Mylonite Zone would help to explain the juxtaposition of amphibolite and granulite facies assemblages across the zone. A downward displacement of 3-4 km would satisfactorily explain the somewhat lower grade of the Lowlands terrane.

PETROLOGIC CHARACTERISTICS OF THE PRINCIPAL ROCK TYPES IN THE ADIRONDACKS

The following discussion is divided into igneous and metasedimentary sections. Whole rock analyses for granitoids are given in table 2 while those for anorthositic and gabbroic rock appear in table 4.

Igneous Rocks

Tonalites and related granitoids. Typical whole rock chemistries for these rocks are shown in figures 7-9. Figure 8 shows the normative anorthite (An)-albite (Ab)-orthoclase (Or) data for these rocks and compares them to similar rocks in the Lowlands. AFM plots are given in figure 8 and calc-alkali index versus silica plots in figure 9; both figures illustrate the strongly calcalkaline nature of the Highland tonalite to granitoid suite. The tonalitic rocks, which will be visited at Stop 1, outcrop in several belts within the southern and eastern Adirondacks. In the field they can be distinguished from, otherwise similar, charnockitic rocks by the white alteration of their weathered surfaces and the bluish grey on fresh surfaces. A distinctive characteristic is the almost ubiquitous presence of discontinuous mafic sheets. These have been interpreted as disrupted mafic dikes coeval with emplacement of the tonalites. Associated with the tonalitic rocks are granodioritic to granitic rocks containing variable concentrations of orthopyroxene.

AMCG Suite. Within the Adirondack Highlands AMCG rocks are widely developed and abundantly represented in the Marcy massif as well as the Oregon and Snowy Mt. Domes. The chemistry of granitoid (mangeritic to charnockitic) facies of these rocks is given in Table 2 and figures 9 and 10, both for the older as well as the younger anorogenic plutonic rocks. As shown in figure 9, the AMCG rocks have calcalkali-silica trends that are distinctly different than those shown by the tonalitic suites. McLelland and Whitney (1991) have shown that the AMCG rocks exhibit anorogenic geochemical characteristics and constitute bimodal magmatic complexes in which anorthositic to gabbroic cores are coeval with, but not related via fractional crystallization to, the mangeritic-charnockitic envelopes of the AMCG massifs (i.e., Marcy massif, figure 2). Bimodality is best demonstrated by the divergent differentiation trends (figure 11) of the granitoid members on the one hand and the anorthositic-gabbroic rocks on the other (Buddington, 1972). Eiler and Valley (pers. comm.) report that $\delta^{18}O$ values for AMCG granitoids are magmatic in origin and demonstrate that these rocks are related by fractional crystallization and were metamorphosed under vapor-absent conditions. Anorthositic members of the AMCG suite have distinctly different $\delta^{18}O$ values. The extreme low-SiO₂, high-iron end members of the anorthosite-gabbro family will be seen at Stop 10 and are believed to represent late liquids developed by plagioclase fractionation under conditions of low oxygen fugacities (i.e., dry, Fenner-type trends). Associated with these are large magnetite-ilmenite deposits which will be visited at Sanford Lake.

Younger Hornblende Granitic Rocks. The distribution of these rocks is shown in figure 5a, and their ages are given in table 1. An example of these rocks will not be visited. In the field these rocks consist of medium grained, pink, streaky granitic rocks containing hornblende and minor biotite. They are difficult to distinguish from the granitic facies of the AMCG suite. As pointed out by Chiarenzelli and McLelland (1991), their restriction to the northwestern Highlands is intriguing but not yet understood.

Alaskitic and Leucogranitic Rocks. The distribution of these distinctive rocks is shown in figure 5a, and their geochronology is summarized in table 1. An example of these rocks will not be visited. They consist principally of pink quartz-mesoperthite gneiss commonly with magnetite as the only dark phase. A less voluminous, but important, trondhjemitic facies is also common and is commonly associated with

low-Ti magnetite deposits in the unit. Granitic facies also occur within this group which, together, constitutes the Lyon Mt. Gneiss (Whitney and Olmsted, 1989). U-Pb zircon ages of 1080-1050 Ma for the alaskites are interpreted as dating emplacement, and, since this time interval corresponds to granulite facies metamorphism at 25 km, the Lyon Mt. Gneiss is interpreted as intrusive (Chiarenzelli and McLelland, 1991). This is in contrast to Whitney and Olmsted (1989) who have interpreted the Lyon Mt. Gneiss as a metamorphosed series of altered acidic volcanics. This issue is discussed in detail in the text for Stop 12.

Olivine Metagabbro. Numerous bodies of tholeiitic metagabbros are scattered throughout the Adirondacks and are especially abundant near the anorthosite massifs with which they are coeval. Most of these bodies are coronites whose evolution has been discussed by McLelland and Whitney (1980) but which will not be visited on this trip.

Metasedimentary Rocks. Within the southern and eastern Adirondacks the metasedimentary sequence is dominated by quartzites and metapelites with marbles being virtually absent. The major quartzite of the southern region is exceptionally pure and comprises an 1000 m-thick unit. Of even greater extent, as well as thickness, are garnet-biotite-quartz-oligoclase \pm sillimanite gneisses (referred to as kinzigite) together with sheets, pods, and stringers of white, garnetiferous anatectite. The common presence of hercynitic spinel supports an anatectic origin for the leucosomes. Kinzigitic rocks grade into khondalites as sillimanite and garnet increase at the expense of biotite.

In contrast to the southern and eastern Adirondacks, the central and northern Adirondacks contain only sparse kinzigite, and metasediments are principally represented by synclinal keels of marble and calcisilicate with minor quartzite (Stop 15). It is possible that the change from carbonate to pelitic metasediments corresponds to an original shelf to deep water transition, now largely removed by intrusion, doming, and erosion.

STRUCTURAL GEOLOGY

The Adirondacks region is one of intense ductile strain, essentially all of which must postdate the ca. 1150 Ma AMCG suite which is involved in each of the major phases of deformation, i.e., the regional strain is associated with the Ottawa Orogeny. Earlier Elzevirian fabric may be present and rotated into parallelism. Here we describe the structure of the southern Adirondacks which is best known and representative of the entire region. A complete set of references is given in McLelland and Isachsen (1986).

As shown in figures 2 and 11, the southern Adirondacks are underlain by very large folds. Four major phases of folding can be identified and their intersections produce the characteristic fold interference outcrop patterns of the region (figure 11). The earliest recognizable map-scale folds (F_1) are exceptionally large isoclinal recumbent structures characterized by the Canada Lake, Little Moose Mt., and Wakely Mt. isoclines, whose axes trend E-W and plunge 10-15° about the horizontal. The Little Moose Mt. isocline is synformal and the other two are antiformal, and suspected to be anticlinal, but the lack of any facing criteria precludes any age assignments. All of these structures fold an earlier tectonic foliation consisting of flattened mineral grains of unknown age and origin. An axial planar cleavage is well developed in the Canada Lake isocline, particularly in the metapelitic rocks.

F_2 -folds of exceptionally large dimensions trend E-W across the region and have upright axial planes (figure 11). They are coaxial with the F_1 folds suggesting that the earlier fold axes have been rotated into parallelism with F_2 , and that the current configurations of both fold sets may be the result of a common set of forces. An intense ribbon lineation defined by quartz and feldspar rods parallels the F_2 -axes along the Piseco anticline, Gloversville syncline, and Glens Falls syncline and documents the high temperatures, ductile deformation and mylonitization that accompanied the formation of these folds. Large NNE trending upright folds (F_3) define the Snowy Mt. and Oregon domes (figure 11). Where the F_3 folds intersect F_2 axes structural domes (i.e., Piseco dome) and intervening saddles result. A late NW-trending fold set results in a few F_4 folds between Canada Lake and Sacandaga Reservoir (figure 11).

Kinematic indicators (mostly feldspar tails) in the area suggest that the dominant displacement in the region involved motion in which the east side moved up and to the west (McLelland, 1984). In most instances this implies thrusting motion, however, displacement in the opposite sense has also been

Fig. 12. Block diagram showing how interference between F_1 , F_2 , and F_4 fold sets produce the outcrop pattern of the F_1 Canada Lake isocline. the axial plane of the F_1 fold is stippled and its fold axis plunges 10-15° to the southeast. The city of Gloversville is shown.

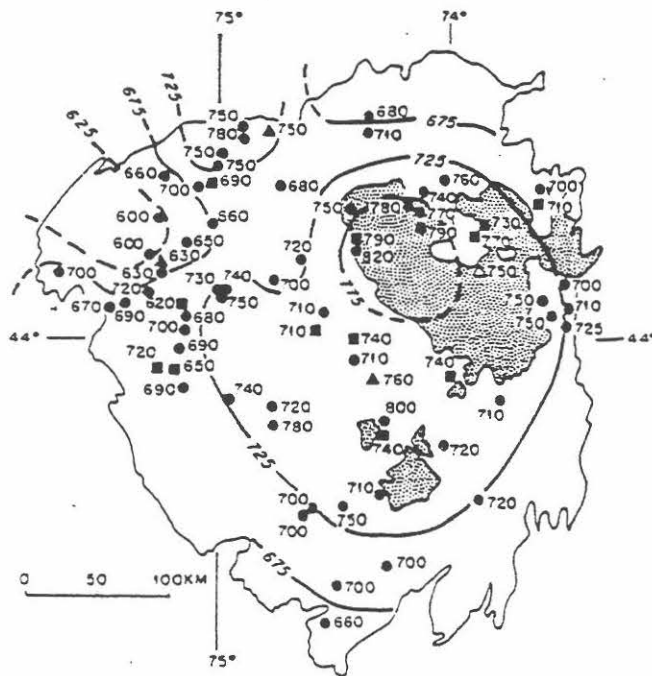
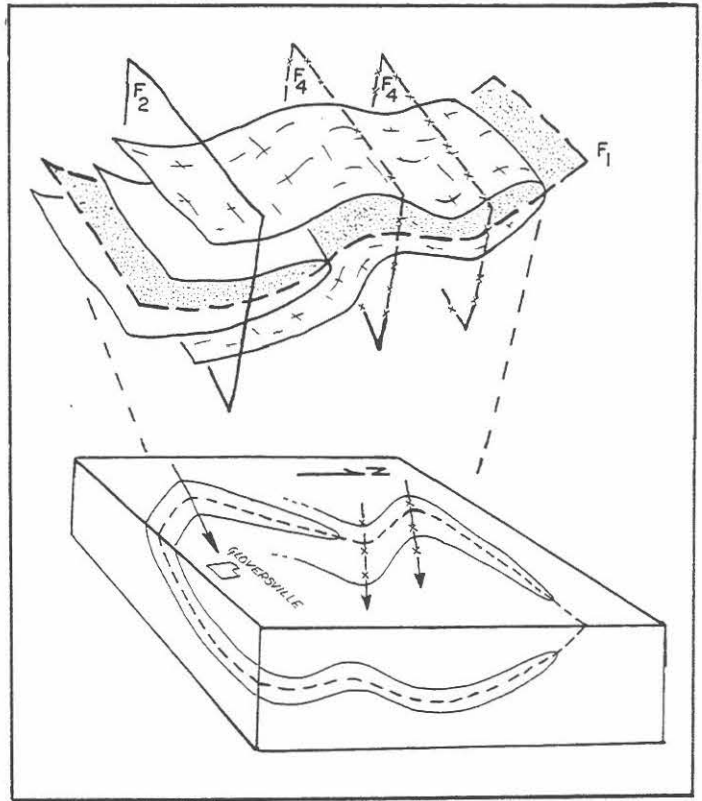


Fig. 13. Metamorphic temperatures, in °C, after Bohlen and others (1985). Temperatures from coexisting feldspars (filled circles); magnetite-ilmenite (squares); calcite-dolomite (filled triangles); and akermanite (open triangle). Stippled area: anorthosite.

documented. This suggests that relative displacement may have taken place in both senses during formation of the indicators.

METAMORPHISM

Figure 12 shows the well known pattern of paleoisotherms reported by Bohlen et al. (1985). paleotemperatures have been established largely on the basis of two-feldspar geothermometry but (Fe, Ti)-oxide methods have also been used and, locally, temperature-restrictive mineral assemblages have been employed (Valley et al., 1990). The bull's eye pattern of paleoisotherms, centered on the Marcy massif, is believed to result from late doming of the massif. Paleopressures show a similar bull's eye configuration with pressures of 7-8 kbar decreasing outward to 6-7 kbar away from the massif and reaching 5-6 kbar in the Lowlands (Bohlen et al., 1985). The P,T pattern of figure 12 is interpreted as reflecting peak metamorphic conditions, although microtextures suggest that some retrogression exists. Generally, the P,T conditions of the Adirondack are those of granulite facies metamorphism, and most commonly correspond to the hornblende-clinopyroxene-almandine subfacies of the high-pressure range of the granulite facies. These conditions must have been imposed during the Ottawa Orogeny since they affect rocks as young as 1050 Ma. The identification of ca. 1050-1060 Ma metamorphic zircons by McLelland and Chiarenzelli (1990) fixes the time of peak metamorphic conditions and corresponds well with garnet and titanite U-Pb ages of ca. 1050-1000 Ma in the Highlands (Mezger et al., 1992). Rb-Sr whole rock isochron ages of ca. 1100-1000 Ma also reflect Ottawa temperatures and fluids. Despite the high-grade, regional character of the Ottawa Orogeny, the preservation of foliated garnet-sillimanite xenoliths in an 1147±4 Ma metagabbro (McLelland et al., 1987), and the report of several 1150 Ma U-Pb garnet ages (Mezger et al., 1992), reveals that earlier assemblages from the Elzevirian and AMCG metamorphic pulses managed to survive locally. The dehydrating effects of these high temperature events, as well as the anhydrous nature of the AMCG rocks themselves, are thought to be responsible for creating a water-poor terrane throughout the Adirondack Highlands prior to the Ottawa Orogeny which appears to have proceeded under generally vapor-absent conditions (Valley et al., 1990).

The present depth to the Moho beneath the Adirondack Highlands is 35 km (Katz, 1955; Hughes and Luetgert, 1992). Since metamorphic pressures of 7-8 kbar correspond to 20-25 km depth of burial, it follows that during Ottawa metamorphism the Adirondack region consisted of a double thickness of continental crust, and this portion of the Grenville orogen may have corresponded to a Himalayan-type collisional margin at 1050-1080 Ma. Bohlen et al. (1985) proposed a counterclockwise P-T-t path for the Ottawa Orogeny, including an almost isobaric cooling history. If this is correct, the necessary magmatic component of heat may have been derived from 1130-1150 Ma gabbroic magmas that were ponded at the base of the lithosphere during the AMCG magmatism. Upward transfer of this heat by conduction would require 80 Ma to reach the surface (Emslie and Hunt, 1990) and would, therefore, have been present in the crust during the height of the Ottawa Orogeny. Granitic rocks of the ca. 1050 Ma Lyon Mt. Gneiss may have helped to transport this thermal energy.

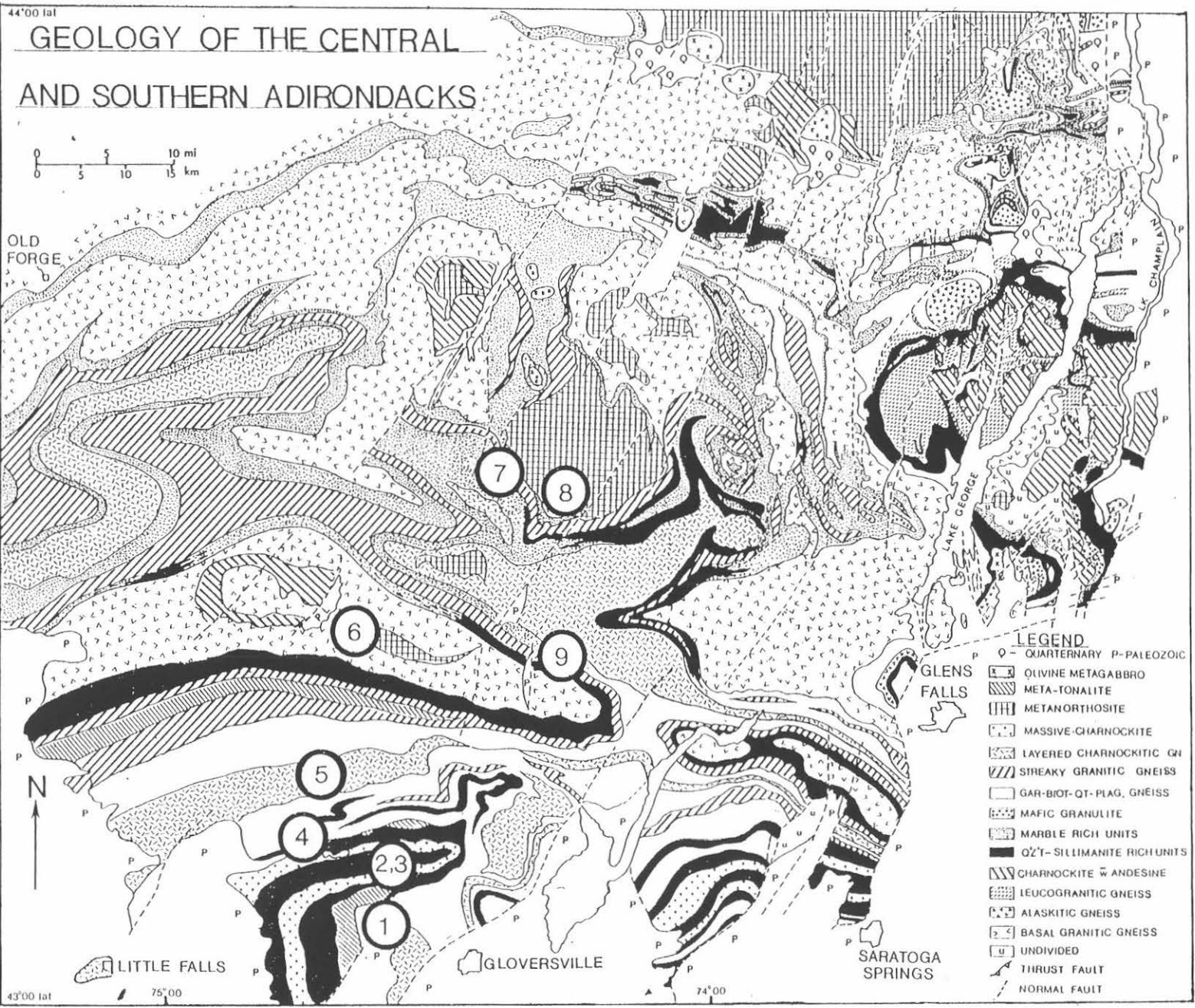


Fig. 13. Geologic map of the southern and central Adirondacks with field trip stops 1-9 indicated (McLelland and Isachsen 1986).

ROAD LOG

(See fig. 18 for stop locations)

CUMULATIVE MILES FROM MILEAGE	LAST POINT	ROUTE DESCRIPTION
	0	Junction of Willie Road, Peck Hill Road, and NY Rt. 29A
1.3	1.3	Mud Lake to northeast of NY Rt. 29A
1.8	1.5	Peck Lake to Northeast of NY Rt. 29A
3.6	1.8	STOP 1. Peck Lake Fm.

STOP 1.

This exposure along Rt. 29A just north of Peck Lake is the type locality of the sillimanite-garnet-biotite-quartz-feldspar gneisses (kinzigites) of the Peck Lake Fm. in addition, there are exposed excellent minor folds of several generations. Note that the F_1 folds rotate an earlier foliation. The white quartzo-feldspathic layers in the kinzigites consist of quartz, two feldspars, and garnet and are believed to be anatectic and have been folded by F_1 indicating pre- F_1 metamorphic events. Typical whole rock compositions are shown below. Spinel has been found enclosed in garnets at this outcrop. The similarity of the Peck Lake Fm. to the Major Paragneiss of the Lowlands suggests that the Adirondacks were contiguous at the time of deposition of the rocks.

	COMPOSITIONS OF REPRESENTATIVE LEUCOSOME AND HOST ROCK				SELECTED CLASTICS		
	Leucosome		Host Rock		Average Greywacke ^a ($\Sigma = 23$)	Average PC Slate ^b ($\Sigma = 33$)	Average Slate ^c ($\Sigma = 36$)
	LL1	9-17-2A	10-29-1B	9-11-4B			
SiO ₂	75.61	74.60	68.04	64.24	64.70	56.30	60.64
Al ₂ O ₃	13.75	13.49	13.93	16.16	14.80	17.24	17.32
TiO ₂	.02	.19	.86	.90	.50	.77	.73
Fe ₂ O ₃	.51	1.47	6.08	7.44	4.10	7.22	4.81
MgO	.11	.54	1.45	1.57	2.20	2.54	2.60
CaO	.36	1.64	1.65	3.41	3.10	1.00	1.20
Na ₂ O	2.19	3.25	2.84	3.20	3.10	1.23	1.20
K ₂ O	6.82	4.69	3.27	2.92	1.90	3.79	3.69
MnO	.02	.04	.06	.09	.10	.10	...
P ₂ O ₅	.09	.08	.18	.17	.20	.14	...
LOI	.31	.25	.66	.66	2.40	3.70	4.10
TOTAL	99.78	100.24	99.80	99.76	101.00	98.70	98.00

6.1	2.5	Junction NY Rt. 29A and NY Rt. 10
8.0	1.9	Nick Stoner's Inn on west side of NY Rt. 29A-10
8.6	.6	STOP 2. Irving Pond Fm., .5 mile north of Nick Stoner's Inn, Canada Lake. Very near hinge line of F_1 Canada Lake isocline.

STOP 2.

The outer portion of the Irving Pond Fm. is exposed in low cuts along the east side of Rt. 29A just prior to the crest in the road heading north.

At the southern end of the cut typical, massive quartzites of the Irving Pond are seen. Proceeding north the quartzites become "dirtier" until they develop sillimanite-garnet-biotite-feldspar (kinzigites) layers along with quartzite.

At the northern end of the cut, and approximately on the Irving Pond/Canada Lake Fm. contact there occurs an excellent set of F_1 minor folds. Polished slabs and thin sections demonstrate that these fold an earlier foliation defined by biotite flakes and flattened quartz grains.

At the southern end of the outcrop dark, fine grained metadiabase sheets crosscut the quartzite. Near the telephone pole erosional remnants of diabase appear to truncate approximately horizontal foliation in the quartzite suggesting that the diabase was emplaced after an early metamorphism. At the north end of the cut a diabase sheet of variable thickness is folded in the F_1 fold. The folding is interpreted as Ottawan, the diabase as AMCG in origin, and the early foliation as Elzevirian. This is consistent with the presence of quartzite xenoliths in the ca. 1300 Ma tonalites.

The Irving Pond Fm. is the uppermost unit in the lithotectonic sequence of the southern Adirondacks. Its present thickness is close to 1000 meters, and it is exposed across strike for approximately 4000 meters. Throughout this section massive quartzites dominate.

8.8 .2 STOP 3. Canada Lake Charnockite (>1233 Ma, table 1, sample AM-87-13. Now fixed at 1251±43)

STOP 3.

Large roadcuts expose the type section of the Canada Lake charnockite. Lithologically the charnockite consists of 20-30% quartz, 40-50% mesoperthite, 20-30% oligoclase, and 5-10% mafics. The occurrence of orthopyroxene is sporadic. These exposures exhibit the olive-drab coloration that is typical of charnockites. Note the strong foliation in the rock. Farther north along the highway there are exposed pink leucogranitic variants of this unit. The chemical composition of these is given in table 3 (ab-6). The whole rock chemistry of the charnockitic phase is similar to AM-86-17 in table 3. The lateral continuity of the Canada Lake is striking (fig. 2) but the presence of xenoliths reveals an intrusive origin.

10 1.2 STOP 4. Royal Mt. Tonalite (>1301 Ma, table 1, sample AM-86-12, now fixed at 1307±2 Ma).

STOP 4.

Steep roadcuts, exposed across from the Canada Lake Store, expose typical examples of the early tonalitic rocks that occur within the southern and eastern Adirondacks and that manifest the presence, throughout the region, of collisional magmatic arcs of calcalkaline chemistry that existed along the eastern margin of Laurentia from ca. 1400-1200 Ma. Amalgamation of these arcs culminated in the Elzevirian Orogeny at ca. 1250-1220 Ma.

The whole rock chemistry of the tonalitic rocks is given in table 3 and important chemical trends are portrayed in figures , , and . Figure 7 shows the ϵ_{Nd} characteristics of these rocks and emphasizes their petrologically juvenile character, i.e., they are not derived from any crustal rocks with long-term crustal residence but are essentially mantle derived (including derivation from melting of basaltic rock derived from the mantle at ca. 1300-1400 Ma). The ϵ_{Nd} characteristics are compared with those from Lowland tonalites and granitoids of similar age, and the similarity suggests that they are essentially the same, strongly suggesting contiguity across the entire Adirondacks at that time (~1300 Ma).

A disrupted layer of amphibolitic material runs down the outcrop to road level at the east end of the outcrop. This, and other mafic sheets in the outcrop, are interpreted as dikes and sheets coeval with the tonalite. In the eastern Adirondacks it has been possible to document mutually crosscutting relationships between these rock types. Also documented there are xenoliths of kinzigitic rock in the tonalites. Within the southern Adirondacks xenoliths of quartzite similar to the Irving Pond Fm. have been found in the tonalite.

11.8	1.8	Pine Lake, Junction NY Rt. 29A and NY Rt. 10. Proceed north on NY Rt. 10.
17.5	5.7	STOP 5. Rooster Hill megacrystic gneiss at the north end of Stoner Lake (1156±8 Ma, table 1, sample AM-86-17).

STOP 5.

This distinctive unit belongs to the AMCG suite and is widespread in the southern Adirondacks. Here the unit consists of a monotonous series of unlayered to poorly layered gneisses characterized by large (1-4") megacrysts of perthite and microcline perthite. For the most part these megacrysts have been flattened in the plane of foliation, however, a few megacrysts are situated at high angles to the foliation and show tails. The groundmass consists of quartz, oligoclase, biotite, hornblende, garnet, and occasional orthopyroxene. An igneous rock analogue would be monzonite to quartz-monzonite (see table 3 for chemical composition) and the presence of orthopyroxene makes the rock mangeritic to charnockitic.

The contacts of the Rooster Hill megacrystic gneiss are everywhere conformable, but the presence of xenoliths of kinzigite indicate its intrusive nature. Rocks such as the Rooster Hill are interpreted as derived from melting of ca. 1300 Ma tonalitic and lower crustal granitoid rocks with heat derived from large AMCG gabbroic intrusions that would ultimately differentiate to anorthosite. This suggestion is consistent with the ϵ_{Nd} trends of AMCG and tonalitic rocks in figure 6a and with the REE distributions shown in figure 14, where it appears that melting of tonalite so as to leave a plagioclase-rich residue can give the AMCG REE-trends.

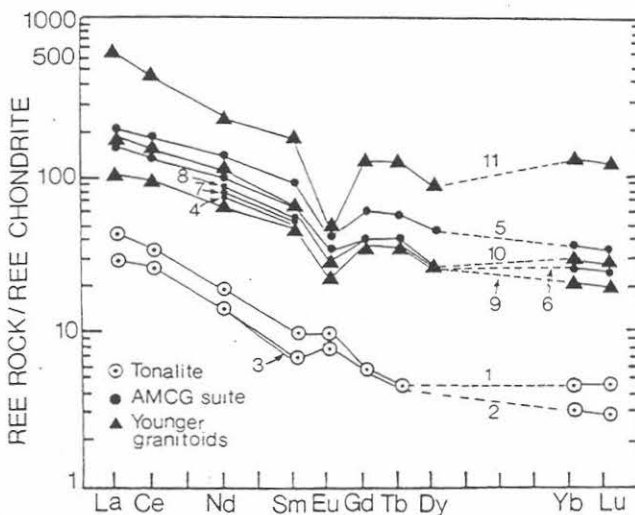


Fig. 14 . Chondrite normalized REE concentrations for the Adirondack highlands. Numbers refer to samples in table 1 of Daly and McLelland (1991).

20.0	2.5	Low roadcut in kinzigites.
21.4	1.4	Avery's Hotel on west side of NY Rt. 10
22.5	1.1	Long roadcuts of pink quartzofeldspathic gneisses and metasediments of intruded metagabbro and anorthosite metagabbro. The igneous rocks are believed to belong to the AMCG suite.
23.6	1.1	Roadcut of anorthositic metagabbro and metanorite of AMCG suite.
23.9	.3	Roadcut on west side of highway shows excellent examples of anorthositic gabbros intrusive into layered pink and light green quartzofeldspathic gneisses.
24.0	.1	Pink granitic gneiss of AMCG suite intruded by anorthositic AMCG gabbros and gabbroic anorthosites. Large boudin of calcsilicate in granite.
24.3	.3	Roadcuts of quartzites and other metasediments of the Sacandaga Fm. Mezger (1990) obtained a U-Pb garnet age of ca. 1154 from these rocks.
31.0	5.7	Red-stained AMCG quartzofeldspathic gneisses that have been faulted along NNE fractures.
31.5	.5	Junction of NY Rt. 10 and NY. Rt. 8. End Rt. 10. Turn east on NY Rt. 8.
32.0	.5	STOP 6. Core rocks of the Piseco anticline (1150±5 Ma, table 1, sample AM-86-9).

STOP 6.

This stop lies along the hinge line of the F_2 Piseco anticline near its domical culmination at Piseco Lake. The rocks here are typical of the granitic facies of quartzofeldspathic gneisses such as occur in the Piseco anticline and in other large anticlinal structures, for example Snowy Mt. dome, Oregon dome.

The pink "granitic" gneisses of the Piseco anticline do not exhibit marked lithologic variation. Locally grain size is variable and in places megacrysts seem to have been largely grain size reduced and only a few small remnants of cores are seen. The open folds at this locality are minor folds of the F_2 event. Their axes trend N70W and plunge 10-15° SE parallel to the axis of the Piseco anticline.

The most striking aspect of the gneisses in the Piseco anticline is their well-developed lineation. This is expressed by rod, or pencil-like, structures which are clearly the result of ductile extension of quartz and feldspar grains in a granitic protolith. The high temperature, grain size reduction that has occurred results in a mylonite. Where recognizable, early F_1 isoclinal fold axes parallel the lineation.

These rocks are similar in age and chemistry to other AMCG granites and are considered to be part of that suite.

Smooth outcrops of Piseco Core rocks showing exceptionally strong mylonitic ribbon lineations.

43.5	11.5	Junction NY Rt. 8 and NY Rt. 30 in Speculator. Head southeast on NY Rt. 8-30.
47	3.5	STOP 7. Northern intersection of old Rt. NY 30 and new Rt. NY 30, 3.3 miles east of Speculator, New York.

STOP 7.

Typical Adirondack marble is exposed in roadcuts on both sides of the highway. These exposures show examples of the extreme ductility of the carbonate-rich units. The south wall of the roadcut is particularly striking, for here relatively brittle layers of garnetiferous amphibolite have been intensely boudinaged and broken. The marbles, on the other hand, have yielded plastically and flowed extensively during the deformation. As a result, the marble-amphibolite and marble-charnockite relationships are similar to those that would be expected between magma and country rock. Numerous rotated, angular blocks of amphibolite and charnockite are scattered throughout the marble in the fashion of xenoliths in igneous intrusions. At the eastern end of the outcrop tight isoclinal folds of amphibolite and metapelitic gneisses have been broken apart and rotated. The isolated fold noses that remain "floating" in the marble have been aptly termed "tectonic fish". The early, isoclinal folds rotate on earlier foliation. The garnetiferous amphibolites have typical igneous compositions and are interpreted as flows or sills.

Near the west end of the outcrop a boudin of charnockite is well exposed. McLelland and others (1987) have presented evidence that boudin represents a local example of charnockitization by carbonic metamorphism. However, granites of similar composition outside the marble do not develop orthopyroxene, demonstrating the local nature of the process and the limited permeation of the fluid phase.

Exposed at several places in the roadcut are crosscutting veins of tourmaline and quartz displaying a symplectic type of intergrowth. Other veins include hornblende- and sphene-bearing pegmatites.

Almost certainly these marbles are of inorganic origin. No calcium carbonate secreting organisms appear to have existed during the time in which these carbonates were deposited (>1200 Ma ago). Presumably the graphite represents remains of stromatolite-like binding algae that operated in shallow water, intertidal zones. This is consistent with the presence of evaporitic minerals, such as gypsum, in Lowland marbles.

At the eastern end of the outcrop coarse diopside and tremolite are developed in almost monomineralic layers. Valley et al. (1983) showed that the breakdown of almost Mg-pure tremolite to enstatite, diopside, and quartz in these rocks requires low water activity at the regional P,T conditions. Similarly, the local presence of wollastine requires lowering of CO₂ activity, presumably by H₂O. These contrasts demonstrate the highly variable composition of the fluid phase and are consistent with a channelized fluid phase within a largely fluid-absent region.

47.5	.5	Extensive roadcuts in lower part of marble. Quartzites, kinzigites, and leucogneisses dominate. Minor marble and calcsilicate rock is present.
47.9	2.5	Large roadcuts in well-layered, pink quartzofeldspathic gneisses with subordinate amphibolite and calcsilicate rock. The layering here is believed to be tectonic in origin, and the granitic layers represent an intensely deformed granite. The calcsilicate layers may be deformed xenoliths.

STOP 8. One half mile south of southern intersection of old Rt. 30 and with new Rt. 30. Anorthositic rocks on the SW margin of the Oregon Dome.

STOP 8.

On the west side of the highway a small roadcut exposes typical Adirondack anorthosite and related phases.

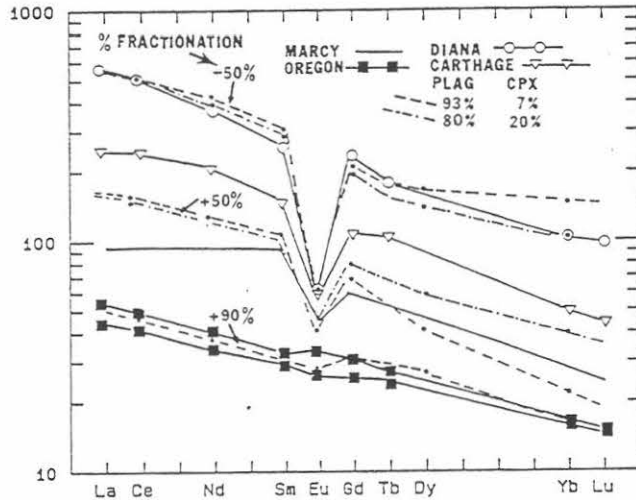


Fig. 15. Chondrite normalized REE concentrations for several Adirondack ferrogabbroic occurrences. Percentage fractionation of plagioclase and clinopyroxene are shown for a starting composition given by Carthage ferrogabbro (triangles). The Diana occurrence corresponds to sheets of breccia-bearing mafic material referred to by Buddington (1939) as shonkinite. The breccia consists of K-feldspar fragments from the host pyroxene syenite of the Diana complex.

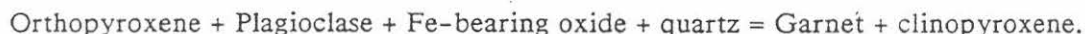
The glacially smoothed upper surface of the roadcut reveals the presence of three major igneous phases: 1) a dark, pyroxene-rich dike that crosscuts the anorthosite, contains anorthosite xenoliths, and contains a large irregular, disrupted mass of sulfidic calcisilicate; 2) a coarse grained, Marcy-type anorthosite facies with andesine crystals 6-8" across; and 3) a fine grained anorthositic phase. Some of the coarse grained facies has been crushed and these portions bear some resemblance to the finer grained phase (note, for example, those places where fractures cross large andesine grains and produce finer grained material). However, close inspection of the finer grained material reveals the presence of ophitic texture with pyroxenes of approximately the same size as the plagioclase, and this texture and association are much better explained as igneous in origin. Therefore, the texture of the fine grained phase is interpreted as igneous in origin and may be due to chilling near the contact of the Oregon Dome massif. By contrast, large (3-4 cm) rafts of coarse grained, ophitic gabbroic anorthosite seem to be "flat" in the fine grained phase. Analyses of typical anorthositic rocks are shown in Table 5.

The pyroxene-rich ferrogabbro dike shows "soft" contacts with the anorthosite and is interpreted as essentially coeval. Zircons from it give a minimum age of 1087 Ma and, by comparison with other Adirondack anorthosites, its emplacement age is set at ca. 1135 Ma. The composition of the ferrogabbro is shown in table 5 where it is seen to be rich in TiO_2 and P_2O_5 . Similar rocks occur together with other Adirondack anorthosites and are interpreted as late, Fe-enriched differentiates of a Fenner-type fractionation trend (see fig. 11). It is suggested that further differentiation within these rocks can result in liquid immiscibility and the production of magnetite-ilmenite liquids.

TABLE 5

No. of Samples	Jotunite	Anorthositic Gabbro Green Mt.	Anorthosite Green Mt.	Anorthosite Owis Head	Mangerte Tupper Lake	Keene Gn. Hulls Falls	Marcy-type Anorthosite ^a	Whiteface Anorthosite ^a
	1	1	1	1	3	2	4	7
SiO ₂	47.16	55.88	56.89	53.65	62.12	51.63	54.54	53.54
TiO ₂	2.20	1.8	0.47	0.52	0.87	3.1	0.67	0.72
Al ₂ O ₃	17.23	18.23	23.82	24.96	16.48	14.23	25.61	22.50
Fe ₂ O ₃	2.75	2.4	1.21	0.41	1.49	2.1	1.00	1.26
FeO	9.24	6.57	1.3	0.70	3.96	13.5	1.26	4.14
MnO	0.15	0.09	0.02	0.02	0.09	0.16	0.02	0.07
MgO	2.71	2.08	0.65	1.45	1.06	2.63	1.03	2.21
CaO	9.04	4.87	8.19	12.21	3.27	6.5	9.92	10.12
Na ₂ O	6.61	4.26	5.38	3.92	4.81	2.67	4.53	3.70
K ₂ O	2.27	2.76	1.13	1.20	5.13	2.41	1.01	1.19
P ₂ O ₅	0.59	0.48	0.09	0.09	0.30	0.57	0.09	0.13
H ₂ O	<u>0</u>	<u>0.09</u>	<u>0.42</u>	<u>0.04</u>	<u>0.32</u>	<u>0.07</u>	<u>0.55</u>	<u>0.12</u>
Total	99.70	99.94	99.57	99.17	99.90	99.57	100.17	100.00

The upper, weathered surface of the outcrop affords the best vantage point for studying the textures and mineralogy of the anorthositic rocks. In several places there can be seen excellent examples of garnet coronas of the type that are common throughout Adirondack anorthosites. These coronas are characterized by garnet rims developed around iron-titanium oxides and pyroxenes. Recently McLelland and Whitney (1977) have succeeded in describing the development of these coronas according to the following generalized reaction:



This reaction is similar to one proposed by de Waard (1965) but includes Fe-oxide and quartz as necessary reactant phases. The products are typomorphic of the garnet-clinopyroxene subfacies of the granulite facies (de Waard 1965). The application of various geothermometers to the phases present suggests that the P,T conditions of metamorphism were approximately 8 kb and 700±50°C respectively.

51.0	2.0	Minor marble, amphibolite, and calcsilicate rock. Predominantly very light colored sillimanite-garnet-quartz-feldspar leucogneisses interpreted as minimum-melt granitic due to anatexis of kinzigite near Oregon dome anorthosite. Enclaves of spinel- and sillimanite-bearing metapelite are present.
52.0	1.0	Junction to NY Rt. 8 and NY Rt. 30. Continue south on NY Rt. 30. To the west of the intersection are roadcuts in garnetiferous metasediment. A large NNE normal fault passes through here and fault breccias may be found in the roadcut and the woods beyond.
52.5	.5	Entering granitic-charnockitic gneiss on northern limb of the Glens Falls syncline. Note that dips of foliation are to the south.
54.8	2.3	Entering town of Wells which is situated on a downdropped block of lower Paleozoic sediments. The minimum displacement along the NNE border faults has been

		determined to be at least 1000 meters.
58.3	3.5	Silver bells ski area to the east. The slopes of the ski hill are underlain by coarse anorthositic gabbro that continues to the west and forms the large sheet just south of Speculator.
60.3	2.0	Entrance to Sacandaga public campsite. On the north side of NY Rt. 30 are quartzo-feldspathic gneisses and calcsilicates. An F ₁ recumbent fold trends sub-parallel to the outcrop and along its hinge line dips become vertical.
60.8	.5	Gabbro and anorthositic gabbro.
62.0	1.2	STOP 9. Pumpkin Hollow.

STOP 9.

Large roadcuts on the east side of Rt. 30 expose excellent samples of the Sacandaga Fm. At the northern end of the outcrop typical two pyroxene-plagioclase granulites can be seen. The central part of the outcrop contains good light-colored garnet-microcline-quartz gneisses (leucogneisses). Although the weathered surfaces of these rocks are often dark due to staining, fresh samples display the typical white vs. grey color of the Sacandaga Fm. The characteristic and excellent layering of the Sacandaga Fm. is clearly developed. Note the strong flattening parallel to layering. Towards the southern end of the outcrop calc-silicates and marbles make their entrance into the section. At one fresh surface a thin layer of diopsidic marble is exposed.

At the far southern end of the roadcut there exists an exposure of the contact between the quartzo-feldspathic gneisses of the Piseco anticline and the overlying Sacandaga Fm. The hills to the south are composed of homogenous quartzo-feldspathic gneisses coring the Piseco anticline (note how ruggedly this massive unit weathers). The Sacandaga Fm. here has a northerly dip off the northern flank of the Piseco anticline and begins its descent into the southern limb of the Glens Falls syncline.

The pronounced flaggy layering in the Sacandaga Fm. is not of primary sedimentary or volcanic origin. Instead it is tectonic layering within a "straight" gneiss. Hand specimen and microscopic inspection of the light layers, particularly, reveals the existence of extreme grain size reduction and ductile flow. Long quartz rods consist of rectangular compartments of recovered quartz and annealed feldspar grains occur throughout. The rock is clearly a mylonite with its mylonitic fabric parallel to compositional layering.

The chemistry of the light colored layers in the Sacandaga Fm. indicates that they are minimum melt granites. As one proceeds away from the core of the Piseco anticline, these granitic layers can be traced into less deformed sheets and veins of coarse granite and pegmatite. In the most illustrative cases the granitic material forms anastomosing sheets that get grain size reduced and drawn into parallelism as high strain zones are approached. The Sacandaga Fm. is interpreted as an end result of this process and represents a mylonitized migmatite envelope developed in metapelites where they were intruded by AMCG granites at ca. 1150 Ma and then intensely strained during the Ottawa Orogeny at ca. 1050 Ma. This interpretation is consistent with field relationships, the presence of spinel and sillimanite restites in the leucosomes, and with the fact that similar metapelitic rocks are crosscut by ca. 1300 Ma tonalites. The latter observation makes the Sacandaga Fm. protoliths older than the ca. 1150 Ma granitic rocks in the Piseco anticline and makes an intrusive relationship obligatory despite the conformable contact at the south end of the roadcut.

62.5-67.0	.5-4.5	All exposures are within the basal quartzo-feldspathic gneisses at the core of the Piseco anticline.
67.0	4.5	Re-enter the Sacandaga Fm. Dips are now southerly.
68.0	1.0	In long roadcuts of southerly dipping pink, quartzo-feldspathic gneisses with tectonic layering. The coarse grain size of the gneissic precursors can be seen in many layers.
70.4	2.4	Cross bridge over Sacandaga River.
74.4	4.0	Bridge crossing east corner of Sacandaga Reservoir into Northville, N.Y.

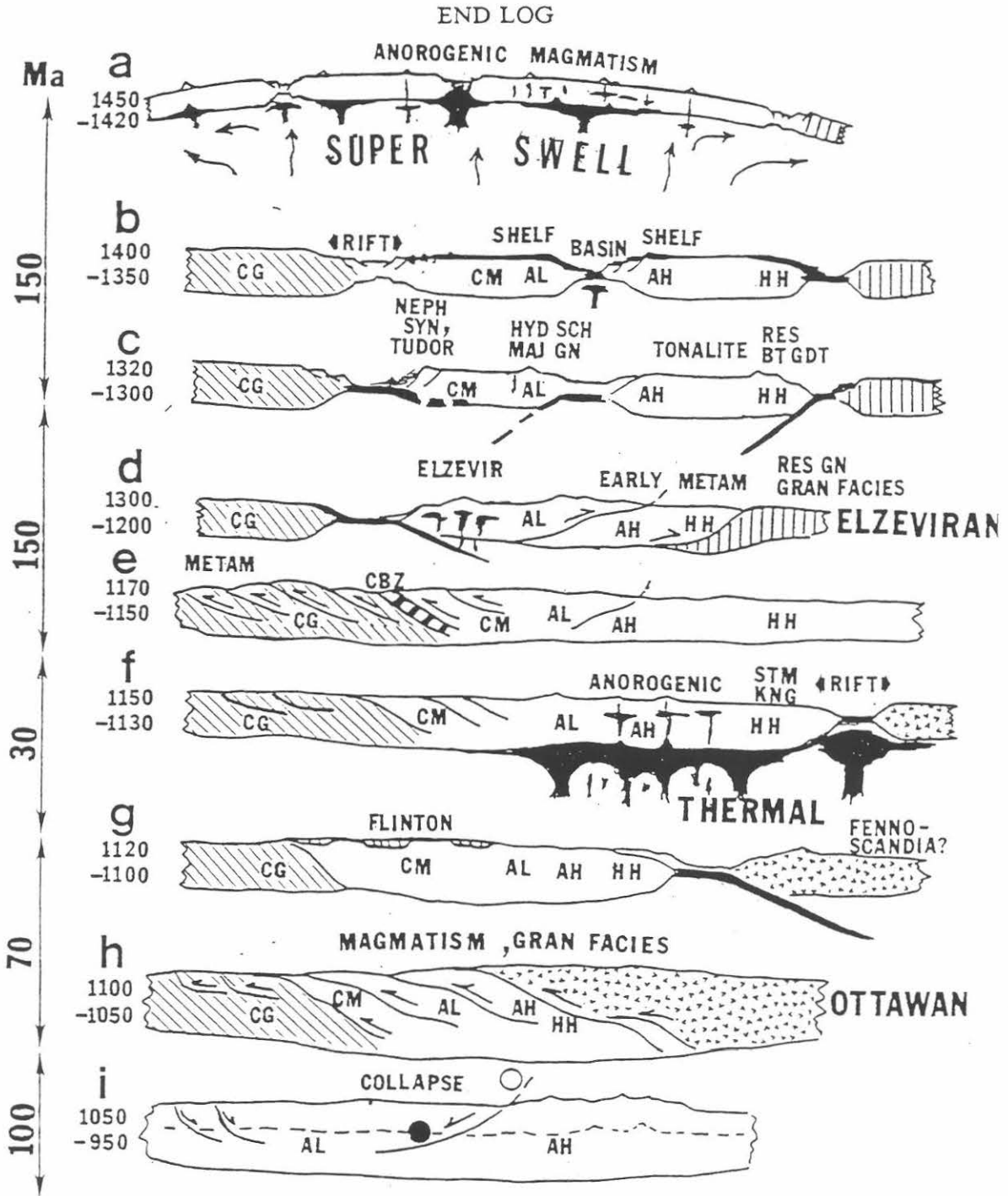


Fig. 16 Hypothetical plate tectonic scenarios for the southwestern Grenville Province.

MIDDLE DEVONIAN TEMPERATE WATER BIOHERMS OF WESTERN NEW YORK STATE (EDGECLIFF MEMBER, ONONDAGA FORMATION)

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INTRODUCTION

The bioherms of the Edgecliff member of the Onondaga Formation are a well known feature of the geology of New York State and the Niagara Peninsula in Ontario, Canada. Over the past thirty to forty years, the Onondaga and its bioherms have been the subject of a number of studies which have greatly increased our understanding of this unit (see Brett and Ver Straeten (1994) for references).

As is the case with "reefy" limestones in general, the Edgecliff has been assumed to represent a warm, tropical, shallow water environment. However stromatoporoids and calcareous algae are rare in the Edgecliff which is unusual for Devonian reefal limestones. The absence of these organisms lead Kissling and his students (Kissling and Coughlin, 1979; Cassa and Kissling, 1982; Kissling, 1987) to interpret these bioherms as deep water structures. The absence of an obvious peritidal facies in the Onondaga was also cited as supportive of a deep water interpretation, suggesting that the shallow water facies which rimmed the basin have since been removed by erosion. More recently Wolosz (1990, 1991) has argued that shallow water facies are present, but have gone un-noticed since they do not fit the standard model for tropical carbonate peritidal facies. However, over the past fifteen years there has been increasing paleontological evidence that the "reefy" Edgecliff Member of the Onondaga Formation represents an example of Devonian temperate water deposition (Koch and Boucot, 1982; Blodgett, et al., 1988). Stromatoporoid biogeographic data and isotopic analyses which further support the temperate water hypothesis are included in this field trip guide.

The field trip will examine a series of Edgecliff bioherms, which are interpreted as representing an on-shore to off-shore trend in biohermal development for this temperate water Devonian limestone.

STRATIGRAPHY AND REGIONAL SETTING

The stratigraphy of the Onondaga Formation has been extensively studied by Oliver (for complete list of references see Oliver (1976)), and recently interpreted in light of both sequence stratigraphic and depositional facies models by Brett and ver Straeten (1994). Only a brief summary of Oliver's Onondaga stratigraphy will be presented here.

In the easternmost part of New York (Fieldtrip Stops 1 thru 4) the Onondaga ranges up to approximately 34 meters in thickness, but with the exception of the basal 2 meter "C1" micrite, it is lithologically a cherty crinoidal packstone/grainstone which is only divisible into members biostratigraphically. In central New York (vicinity of Mt. Tom Reef), the formation thins to roughly 21 meters, but is easily divided on lithologic grounds into Oliver's four members, with the Edgecliff a massive, biostromal, very coarsely

In Garver, J.I., and Smith, J.A. (editors), *Field Trips for the 67th annual meeting of the New York State Geological Association, Union College, Schenectady NY, 1995, p. 227-250.*

crystalline limestone from about 2.5 to 7.5 meters thick; the Nedrow a thin bedded, very fine grained shaley limestone; the Moorehouse a very fine grained limestone with chert and shaley partings; and the Seneca similar to the Moorehouse lithologically, but with a different fauna. Near Buffalo the formation reaches a thickness of 43 meters with only a very thin Edgecliff grainstone/packstone unit (about 1.5 meters), overlain by a sparsely fossiliferous, fine grained, chert-rich limestone which Ozol (1963) named the Clarence member and interpreted as equivalent to the Nedrow. Brett and Ver Straeten (1994) however, have recently identified the Clarence as a facies within the Edgecliff.

The basal contact of the Onondaga is marked by a widespread unconformity (Rickard, 1975). In the east, the contact with the underlying Schoharie Formation has alternatively been interpreted as gradational (Goldring and Flower, 1942) or disconformable, with the presence of a glauconitic sand bed cited as evidence of a period of nondeposition (Chadwick, 1944). In the central part of the state, the base of the Onondaga is marked by the "Springvale Sand" which overlies either the patchily distributed Lower Devonian Oriskany Sandstone, or the older Helderberg limestones. The underlying units continue to be variable to the west, where the Onondaga rests upon either the Lower Middle Devonian Bois Blanc Formation or Silurian dolomites.

Both Lindholm (1967) and Mesolella (1978) identified the central New York Onondaga as representing the most basinal facies exposed at the surface, and located the topographic axis of the basin through that area.

EVIDENCE FOR A TEMPERATE WATER EDGECLIFF SEA

The observations which led Kissling and his students to interpret the Edgecliff as a deep water limestone have also resulted in an alternative hypothesis - deposition in a warm temperate (as opposed to tropical) environment. Koch and Boucot (1982) made this suggestion on the basis of brachiopod community analysis; while Blodget, et al. (1988) noted that the low diversity gastropod fauna along with the absence of stromatoporoids and algae would suggest temperate water conditions. Recently collected isotope data and an analysis of the distribution pattern of the rare stromatoporoids in the Edgecliff lend strong support to the interpretation of this unit as a warm temperate water limestone.

Isotope Data

Brachiopods collected from the Edgecliff for isotopic analysis comprise a geographically and taxonomically broad-based sample (Table 1), which includes specimens of five different Edgecliff genera collected across New York State and Ontario, Canada (brachiopod identifications are based on Feldman, 1985, and Boucot and Johnson, 1968). Non-Edgecliff brachiopods were collected from the Oriskany Sandstone in the Seneca Stone Quarry at Seneca Falls, N.Y. and the Clarence member of the Onondaga near Buffalo, New York. A total of sixteen brachiopods were sampled - fourteen from the Edgecliff - with one brachiopod from the Clarence Member and one from the Oriskany Sandstone. Thirty-three isotopic analyses were performed, twenty nine on the Edgecliff samples, and two each on the Clarence and Oriskany samples. To check for the possible effects of diagenesis, sixty-one samples of dull luminescent, pore-filling cements were analyzed.

Isotopic analyses yielded δO^{18} values ranging from -1.81 to -7.10 ‰ with 24 of 29 Edgecliff analyses falling into the -1.81 to -3.74 ‰ range (Table 1, Figure 1). Data from the Clarence member sample was in the same range (-1.98 ‰, -2.42 ‰); as was the Oriskany Sandstone sample (-2.38 ‰, -2.41 ‰). δC^{13} values ranged from 1.35 to 3.18 ‰, with the Clarence sample again in agreement (2.40 ‰, 2.25 ‰); but with the Oriskany sample at the heavy end of the range (3.26 ‰, 3.65 ‰). Analyses of the dull luminescent cements yielded δO^{18} values ranging from -11.85 to -4.32 ‰ and δC^{13} from -0.27 to 3.98 ‰ (Figure 2).

**CARBON AND OXYGEN ISOTOPE DATA
FROM NONLUMINESCENT PRISMATIC BRACHIOPODS**

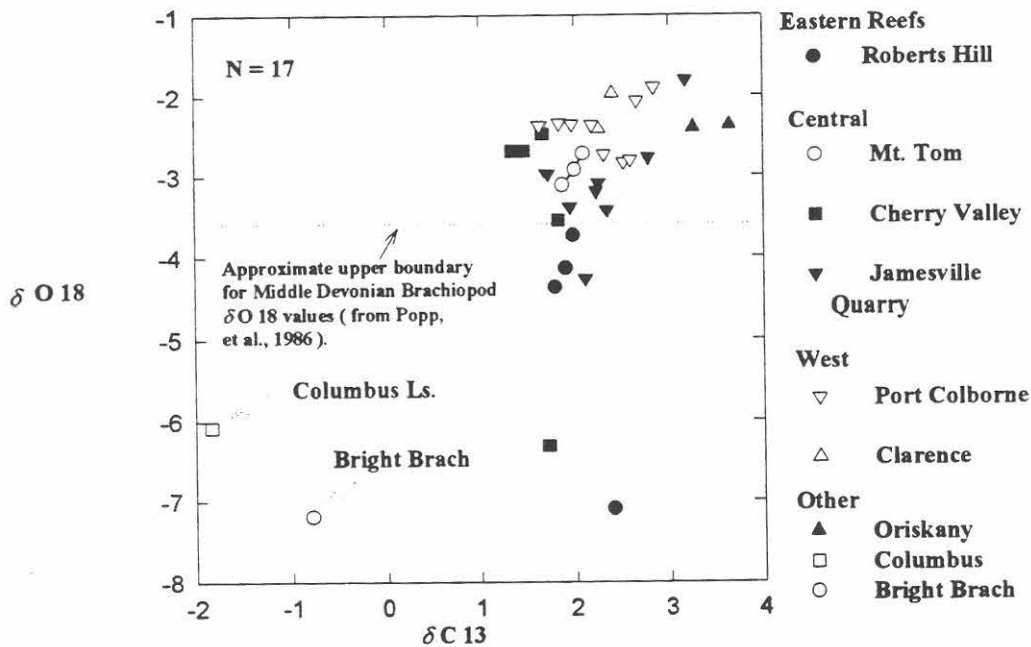


Figure 1. Isotope analyses for non-luminescent brachiopods. Locations as mentioned in text. "Bright brachiopod" (luminescent) sampled from thin dolomite bed at base of Edgecliff in Port Colborne, Ontario, Canada; and Columbus Limestone sample are for comparison.

**CARBON AND OXYGEN ISOTOPE DATA
FROM DULL LUMINESCENT CEMENTS**

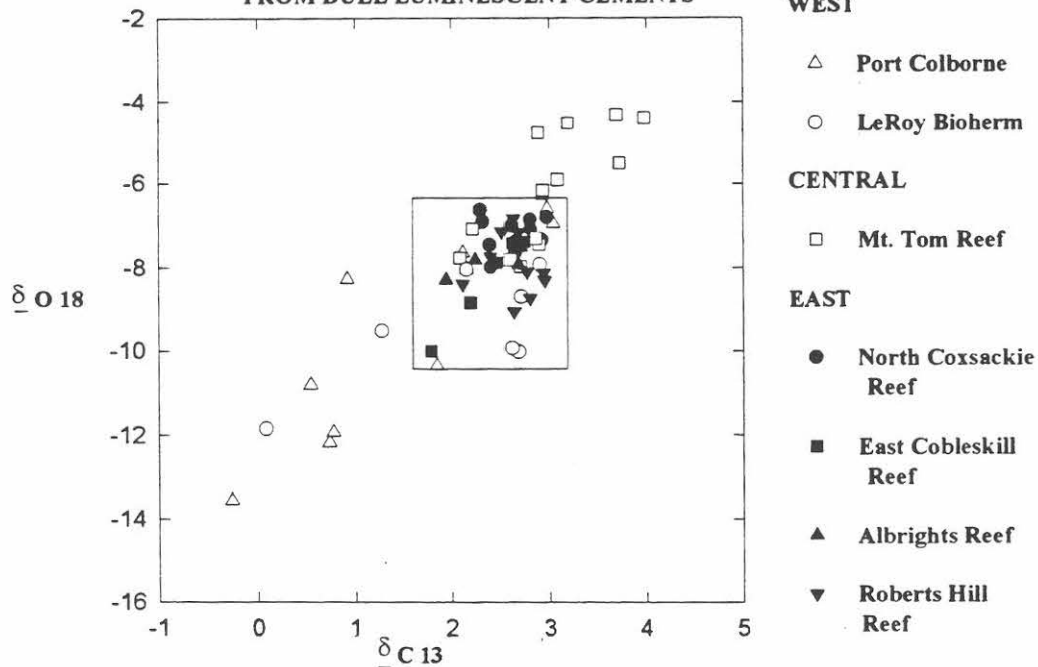


Figure 2. Isotopic analyses of dull luminescent pore filling cements. Locations as noted in text. Area enclosed in box represents restricted range of data from eastern bioherms.

Comparison of these data with the Devonian data of Popp, et al., (1986) and Brand (1989) clearly illustrates that the Onondaga and Oriskany samples are isotopically heavier for δO^{18} , suggesting a cooler Edgecliff depositional environment (Figure 1). The Oriskany data is in good agreement with that of Rush and Chafetz (1990), who interpreted the isotopically lighter signatures of the underlying Helderberg limestones as diagenetically altered. The similarity of the Onondaga data to the Oriskany data may, however, indicate that both the Onondaga and Oriskany were deposited under temperate water conditions as compared to the tropical Helderberg (which includes stromatoporoid-rich facies). Recently, Bates and Brand (1991) and Lavoie (1994) have presented δO^{18} isotope data for the Lower and Middle Devonian which are similar to those from the Onondaga; but in both cases the brachiopods analyzed were from deep water communities, and the interpreted lower temperatures are attributed to greater water depth.

Stromatoporoid Distribution

Stromatoporoids are assumed warm water organisms (Stock, 1990; Nestor, 1984, 1990) and important Devonian reef builders. Because of their rarity in the Onondaga all occurrences of stromatoporoids have been noted and when possible specimens collected as part of an ongoing study of Edgecliff bioherms and depositional environments (Wolosz, 1992a). These data reveal an apparent trend of increasing size and abundance of stromatoporoid colonies from east to west (Figure 3 and Table 2). Eastern New York stromatoporoids are rare and have most often been noted in thin-sections as either small, juvenile colonies or small encrusting colonies. In western New York and Ontario, Canada stromatoporoids become locally much more common and also more diverse. St. Jean (1986) reported seven species representing three genera collected from large blocks of limestone adjacent to an abandoned limestone quarry at Empire Beach along the shore of Lake Ontario, roughly ten kilometers east of Port Colborne, Canada. Large lamellar colonies (three species and two genera) have been collected by the author from an abandoned quarry to the west of Port Colborne, Ontario. Domal colonies are also common within the Onondaga near Hagarville, Ontario; and Lindemann (1988) noted large lamellar stromatoporoids in the capping facies at the LeRoy Bioherm.

Based upon current paleogeographic reconstructions (Scotese and McKerrow, 1990; Witzke and Heckel, 1990), the increase in stromatoporoid abundance across New York and into Ontario would mark a south to north trend of increasing abundance towards the paleo-equator. The paleo-geographic interpretation of Heckel and Witzke (1979) hypothesizes a current flow direction along the same Devonian south to north trend toward the paleo-equator. Gradual solar heating of the originally cool temperate waters flowing towards the equator could explain the apparent trends in stromatoporoid abundance observed across New York and into Ontario.

SHALLOW WATER FACIES

Recent evidence suggests that the assumed lack of peritidal or shallow subtidal facies in the Edgecliff is actually a lack of shallow tropical carbonate facies. Wolosz (1990, 1991) interpreted Edgecliff exposures near Port Colborne, Ontario, Canada as shallow subtidal to near peritidal environments. Figure 4 illustrates a model for Edgecliff shallow water deposition based on data from the Ridgemount bioherm and the Quarry Road exposures west of Port Colborne, Ontario.

In the Ridgemount bioherm quarry, a linear coral ridge is made up almost entirely of displaced and/or fragmentary rugosan and favositid colonies interpreted by Wolosz (1990) as evidence of shallow water storm damage. The coral ridge directly overlies and interfingers with a bioturbated dolomite bed interpreted as a former lime mud. The sparse presence of large crinoid columnals which are considered characteristic of the Edgecliff (Oliver, 1954) confirms that this dolomite is basal Edgecliff.

Table 1. Isotope analysis data by sample.

Location	Genus	N	Sample	δO^{18} (PDB)	δC^{13} (PDB)
Roberts Hill Reef	Meristina	2	A	-4.15	1.90
			A	-3.74	1.98
			A	-4.38	1.79
			B	-7.10	2.40
Mt. Tom Reef	Amphigenia	1	A	-2.73	2.09
			A	-2.93	2.00
			A	-3.12	1.87
Jamesville Quarry	Amphigenia	3	A	-3.43	2.34
			A	-4.28	2.11
			B	-1.81	3.18
			B	-3.20	2.23
			B	-2.78	2.78
			C	-3.39	1.95
			C	-2.98	1.72
Cherry Valley	Megakoz-lowskiella	2	D	-2.48	1.67
			B	-2.69	1.48
			B	-2.69	1.35
	Pentam-erella	1	C	-6.31	1.72
			C	-3.55	1.83
Clarence member Williams-ville NY	Megakoz-lowskiella	1	A	-1.98	2.40
			A	-2.42	2.25
Ridgemount Bioherm	Acro-spirifer	1	A	-2.07	2.66
			A	-1.90	2.84
Port Colborne	Meristina	4	E	-2.38	2.18
			E	-2.37	2.18
			F	-2.74	2.31
			C	-2.35	1.84
			C	-2.36	2.52
			D	-2.84	2.52
			D	-2.81	2.59
Oriskany SS.	Unident. frag.	1	A	-2.38	3.65
			A	-2.41	3.26

Table 2. Distribution of stromatoproids based on field observations and collections by author.

LOCATION	LARGE COLONIES	SMALL COLONIES	MICROSCOPIC
Roberts Hill	none	1	5
North Coxsackie	none	none	1
Mt. Tom	1	1	
LeRoy Bioherm	present (Lindemann, 1988)	5	
Port Colbourne Quarry	common	common	
Hagarsville	common	common	
Formosa Reefs	abundant	abundant	

STROMATOPOROID DISTRIBUTION

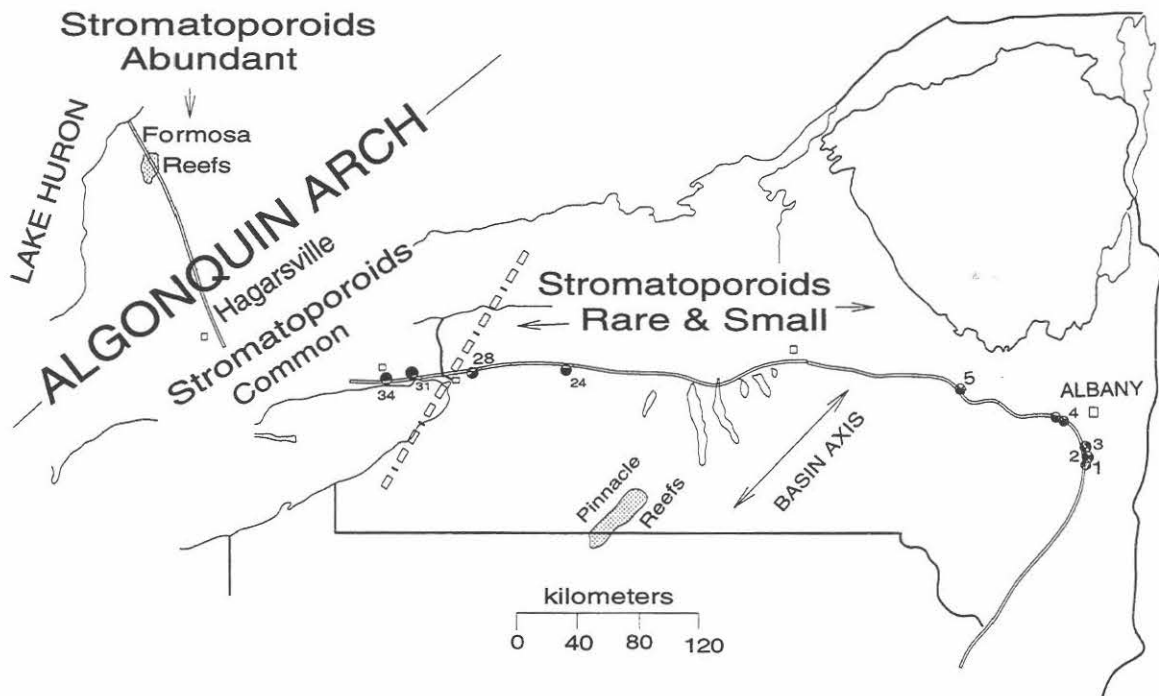


Figure 3. Stromatoporoid abundance across New York State and into Ontario, Canada. Bioherm exposures numbered as follows: Roberts Hill Reef (1), Albrights Reef (2), North Cocksackie Reef (3), Thompson's Lake Bioherm (4), Mt. Tom (5), LeRoy Bioherm (24), Buffalo Country Club Reef (28), Ridgemount Bioherm (31), Quarry Road Mounds (34). Formosa Reefs are approximately equivalent in age to the Edgecliff bioherms.

To the west of Port Colborne (Quarry Road bioherm Stop of Wolosz, 1990), extensive biostromal deposits which vary from dark gray to light gray to buff limestone with varying densities of green clay seams, are characterized by abundant *Cystiphyllodes*, a solitary rugose coral, along with tabulate and colonial rugose coral. This biostrome is well exposed along the east wall of the west quarry but grades westwards across the quarry pit and northwards to the active quarry on the north side of Route 3 (former Law Quarry) into greenish, shaley limestone with clay seams and sparse to common coral which are interpreted as near-shore muds (Figure 4). Oliver (1976, p.10 and 143) noted the presence of these shaley limestone beds and identified the upper 16 feet of this 20 foot thick unit as Edgecliff based upon the sparse coral fauna.

The *Cystiphyllodes*-rich biostromal units are overlain by a grainstone bed with abundant coral and stromatoporoids, followed stratigraphically by "reefy" beds containing numerous Small Rugosan Mounds (see reef classification below).

Edgecliff Shallow Water Facies Port Colborne, Ont.

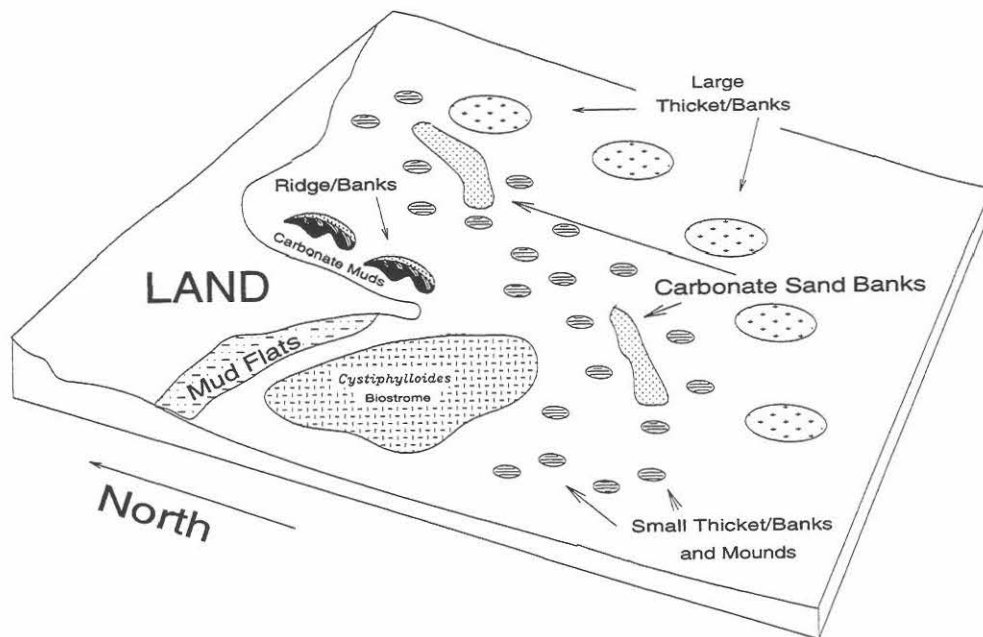


Figure 4. Model of shallow water Edgecliff facies as interpreted from exposures near Port Colborne, Ontario, Canada (see text for details).

Additional evidence for extreme shallow water conditions in the Edgecliff were noted by Wolosz and Paquette (1994) in a study of the LeRoy bioherm in western New York. They interpreted extensive erosion of the bioherm core prior to deposition of the flank beds as due to shallow water (subaerial?) erosion.

REEF PATTERNS AND DISTRIBUTION

Reef Communities

Most Edgecliff bioherms include two distinct communities - the Phaceloid Colonial Rugosan Community and the Favositid/ Crinoidal Sand Community.

The phaceloid colonial rugosan community is made up almost exclusively of colonial rugosans. Common genera include *Acinophyllum*, *Cylindrophyllum*, and *Cyathocylindrium*; with *Eridophyllum*, *Synaptophyllum*, and possibly phaceloid colonies of *Heliophyllum* as accessories. The dense growth of these rugosan colonies appears to have restricted most other organisms to only minor roles, with favositids (both domal and branching) being small and rare, brachiopods uncommon, and bryozoans mainly fragmentary encrusters.

The favositid/crinoidal sand community displays a much higher diversity than the rugosan community. This community is more biostromal than biohermal. Large sheet-like to domal favositids are abundant, but never form a constructional mass. Solitary rugose corals are also extremely abundant as are fenestrate bryozoan colonies. Single colonies of the mound building phaceloid rugosans are occasionally found.

Brachiopods and other reef dwellers are also common although never extremely abundant. Stromatoporoids and massive colonial rugosans, while extremely rare in the Edgecliff reefs, when found are part of this community. The crinoids were the greatest contributor to this community - ossicles making up the bulk of the rock and indicating abundant growth of these organisms - but complete calyces are never found.

Reef Classification

Wolosz (1992a) noted that Edgecliff bioherms represent a continuum of growth pattern in which the two above described paleocommunities are the pure end members (the only exception being the LeRoy bioherm "calcsiltite mounds" (see Wolosz and Paquette, 1994)). As a result, the following simple classification of these bioherms was suggested:

Mounds - distinct high relief mounds of the Phaceloid Colonial Rugosan Paleocommunity. Subdivided into: 1) Successional Mounds, and 2) Small Mounds.

Composite Structures - structures formed through interbedding or intergrowth of the two communities. Subdivided into: 1) Mound/Bank, 2) Ridge/Bank, and 3) Thicket/Bank. The term "bank" follows the definition of Nelson, et al. (1962, p.242): "a skeletal limestone deposit formed by organisms which do not have the ecologic potential to erect a rigid, wave resistant structure."

Biostrome - bedded Favositid/Crinoidal Biostrome, typical bedded Edgecliff, with no evidence of relief above the sea-floor. Banks of pure Favositid/Crinoidal Paleocommunity have not been found, although some Thicket/Bank structures are, volumetrically, very close to this state.

Rugosan Mounds

The Phaceloid Colonial Rugosan Paleocommunity is the dominant biota. This paleocommunity produced a high relief mound which is overlapped by the bedded grainstone/packstone of the Favositid/Crinoidal Sand Paleocommunity with dips ranging from 8 to 15 degrees.

Successional Mounds. - These structures reach thicknesses of up to approximately 15 m., and are dense accumulations of phaceloid colonial rugosans in a matrix of fine, bioclastic calcsiltite. The mound building colonial rugosan genera exhibit a distinct succession (Wolosz, 1985, 1992b), starting with Acinophyllum at the mound base and progressing upwards and outwards through a Cylindrophyllum and then a Cyathocylindrium dominance stage. As the reef progresses through these dominance stages coral diversity increases as the previous dominants become accessory. Wolosz (1985, 1992b) has argued that these successions are controlled by the degree of water turbulence. Roberts Hill and Albrights reefs (Field Trip Stops 1 and 2) are examples of this reef type, as are the main mounds in Mound/Bank structures (see below). Small Mounds. - Small mounds are commonly monogeneric; but when more than one genus of colonial rugosan is present the placement of colonies is random, with no evidence of rugosan succession. Some small rugosan mounds are bedded, suggesting only small relief above the sea floor and a shallow water environment (Wolosz, 1990), while field relationships in others suggest that these are small satellite mounds which formed in protected backreef environments (Wolosz, et al., 1991; and Field Trip Stop 4 and 7).

Composite Structures

Composite reef structures formed through repetitive (possibly cyclic) intergrowth or interbedding of the two most common paleocommunities.

Mound/Bank. These are the largest of all the Edgecliff reef structures reaching thicknesses of up to 60 m. and areal extents of up to 3 km. by 2.5 km. The repetitive development of Rugosan Mounds interbedded with the Favositid/Crinoidal Sandstone facies led to the development of a large reef structure with flanking biostromal

beds having dips of up to 25 degrees. This category includes the Mt. Tom Reef (Wolosz, et al., 1991; and Fieldtrip Stop 5) and the subsurface pinnacle reefs (Wolosz and Paquette, 1988; Coughlin, 1980, pp.139-163).

Ridge/Bank. The Rugosan Paleocommunity and occasional large favositids form a series of small mounds which coalesce laterally to form a ridge-like structure. Biostromal onlap produces a topographically large linear structure, the development of which may have included Rugosan Paleocommunity/Biostrome Cycles. Interpreted as a very shallow water structure by Wolosz (1990), the only known example of this type of Edgecliff Reef structure is the Ridgemount Bioherm located in a quarry approximately 2.4 km. due west of Fort Erie, Ontario, Canada (Cassa and Kissling, 1982; Wolosz, 1990).

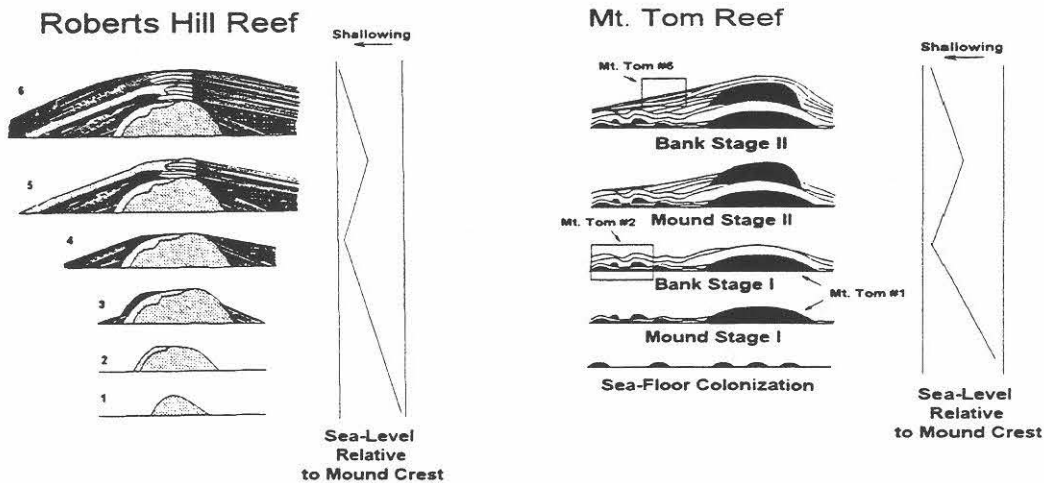


Figure 5. Comparison of development of Roberts Hill and Mt. Tom Reefs. At Roberts Hill, stages 1 - 3 represent the development of the mound through a succession of colonial rugosan genera, stage 4 is a bank stage dominated by the favositid/crinoidal sand community, stage 5 a recolonization of the bank by rugosans, and stage 6 the final bank stage with termination of reef growth. Development of Mt. Tom is similar, but greater subsidence allows development of a second mound stage.

Thicket/Bank. The Favositid/Crinoidal Sand Paleocommunity makes up the main mass of these buildups in the form of gently dipping (5 to 12 degrees), bedded packstone and grainstone with abundant large sheet to domal favositids. The rugosan paleocommunity is reduced to thickets, now less than roughly .3 m. thick, which covered the entire bank, and are now interbedded with the biostromal deposits (Wolosz, 1992b, in press). The resultant structure is a large, low relief, shield shaped mound. The North Cocksackie Reef (Fieldtrip Stop 3) is an excellent example of this reef type, as is the flanking stage of the LeRoy Bioherm (Wolosz and Paquette, 1994; Wolosz, in press).

Tabulate "Calcsiltite Mounds"

This facies is known only from the LeRoy Bioherm complex (Lindemann, 1988; Wolosz and Paquette, 1994). Unlike the other Edgecliff reefs, phaceloid rugosans are absent from the central mound which is dominated by small branching tabulate corals. No bioherms of this type are known in eastern New York.

GEOGRAPHIC DISTRIBUTION AND ECOLOGY

Wolosz (1992a) noted that the geographic distribution (Figure 5) of these bioherms support the hypothesis that the above described patterns of reef growth were controlled by

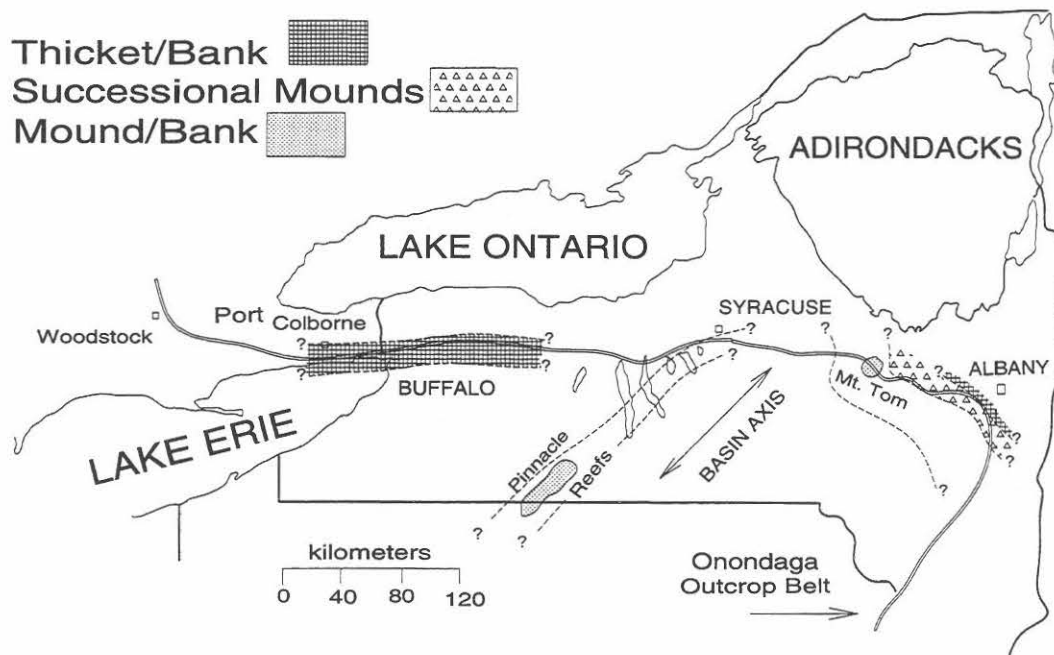


Figure 6. Interpreted distribution of Edgecliff bioherms by type. Mound/bank structures rim the area of major subsidence. Dashed lines represent interpreted boundaries between areas of distinct bioherm types, while patterning indicates confirmed presence of bioherms in either outcrop or subsurface. Question marks indicate limits of interpretable area due to lack of exposure or subsurface information or due to erosional removal of Onondaga (from Wolosz, 1992a).

water depth and rate of basinal subsidence. Wolosz (1985, 1992a, 1992b, in press) and Wolosz and Paquette (1988) have argued that the dominant reef community in Edgecliff Bioherms was controlled by the level of water turbulence at the crest of the reef. Following this model, shifts between the Favosiid-Crinoidal Sand Paleocommunity and the Colonial Rugosan Paleocommunity mark "catchup/fall back" growth cycles as the reef community tried to maintain itself at a constant water depth during basinal subsidence. These cycles are illustrated by a comparison of the development of Roberts Hill and Mt. Tom Reefs (Figure 5). The similarity in the interpreted sea-level curves for each reef would support control of reef growth by basinal sea-level fluctuation. This balance of growth versus subsidence resulted in the great thickness of the pinnacle reefs.

The pinnacle reefs are large mound/bank structures which rim the axis of major basinal subsidence as located by Lindholm (1967) and Mesolella (1978). Shorewards of the pinnacle reefs lie the successional mound bioherms followed by the thicket/bank structures, then the small mounds and finally the ridge bank structures (Figure 5). In the field trip area, the four southeasternmost reefs (Roberts Hill, Albrights, North Coxsackie, Thompson's Lake Bioherm, Figure 1 #'s 1-4) illustrate the off-shore to on-shore trend from successional mounds (Robert's Hill and Albrights Reefs) to thicket/banks (North Coxsackie Reef) to shallow water small rugosan mounds and *Cystiphyllloides* biostrome (Illustrated in part in Figure 6). Roberts Hill and Albrights reefs are rooted in the deeper water C1 Edgecliff facies, while the North Coxsackie reef and Thompson's Lake bioherm are rooted in the typical grainstone/ packstone of the Edgecliff. Further, the transition from the North Coxsackie thicket/bank structure to the small rugosan mounds associated with *Cystiphyllloides* biostromes at the Thompson's Lake exposure represent the same shallow water facies as found near Port Colborne, Ontario (Figure 4).

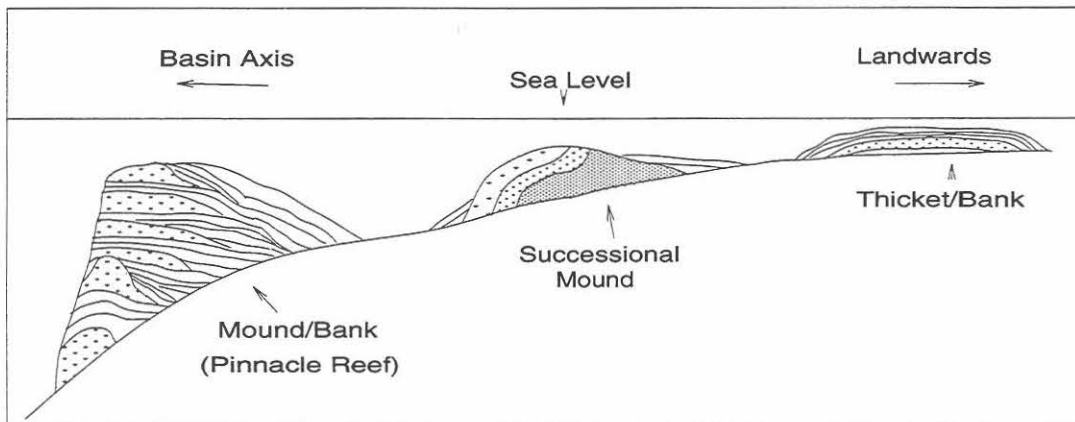


Figure 7. Depth controlled off-shore to on-shore transect showing relative positions of various bioherm types. Shallow subtidal facies (Figure 4) would lie in on-shore (to right) of thicket/bank bioherm (From Wolosz, 1990).

SUMMARY

- 1) Isotope data derived from nonluminescent brachiopods and analysis of stromatoporoid distribution in the Edgecliff, when added to previously published paleontologic and paleo-geo-graphic studies, support the hypothesis that the Edgecliff bioherms grew in cool temperate waters.
- 2) Shallow subtidal facies are present in the Edgecliff, but they are characterized by *Cystiphylloides* dominated biostromes, small rugosan mounds, and shaley, sparsely fossiliferous limestone deposits - not by classic tropical peritidal facies.
- 3) The Edgecliff bioherms display patterns of development which follow distinct onshore to off shore trends.
- 4) The cyclic development of Edgecliff bioherms was controlled by water depth (turbulence) over the crest of the bioherm.

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**MIDDLE DEVONIAN TEMPERATE WATER BIOHERMS OF WESTERN NEW YORK STATE (EDGECLIFF
MEMBER, ONONDAGA FORMATION)**

ROAD LOG

CUMULATIVE MILEAGE	MILES FROM LAST POINT	ROUTE DESCRIPTION
0.0	0.0	Exit 21B from NYS Thruway. Left turn onto Route 9W
0.6	0.6	Right turn onto Schiller Park Road
2.0	1.4	Right turn onto Limnkiln Road
3.0	1.0	STOP 1. Roberts Hill Reef Park cars and proceed east onto hill.

NOTE: Private Property. **DO NOT** enter without first obtaining permission at house on north side of hill.

Roberts Hill Reef

Roberts Hill Reef is the best known example of a Successional Mound in the Edgecliff, and has been described in detail by Wolosz (1985). This exposure allows for examination of the typical Edgecliff bioherm growth pattern of interbedded Colonial Rugosan and Favositid/crinoidal Sand Paleocommunities (Figure 9).

The outer edge of a central rugosan mound is exposed along the east side of a small valley which runs southwards across the hill (refer to Figure 8), while the west side of the valley exposes bedded packstone flank beds with large favositid colonies. Further to the east, south and southeast a second stage of rugosan growth (Recolonization Stage of Wolosz (1985)) can be observed to cap the flank beds. This cyclic intergrowth of paleocommunities was controlled by sea-level fluctuation (see Figure 5; and Wolosz, 1985, 1992b). Evidence for changing turbulence levels during reef growth can be directly observed at the outcrop and includes the following: 1) development of rubble apron along the northeast side of the hill which wedges out into the flank beds between the two stages of rugosan growth. This facies is dominated by the small solitary rugosan *Cystiphyllodes*, and is interpreted as the result of an extreme shallow water stage during which the top of the mound was covered by a meadow of these small solitary rugosans. Interpretation of water depth is based on the similarity of this facies with the shallow water *Cystiphyllodes* biostromes at Thompson's Lake Bioherm and in the Port Colborne area. 2) A rubble ring surrounding the central rugosan core and large overturned favositids in the adjoining flank beds. 3) Pods of colonial rugosan rubble at the toe of the Rugosan Recolonization Zone at the southernmost edge of the reef along the east cliff face of the hill. The large colony fragments in these rubble pods indicate elevated turbulence levels during the demise of the reef.

This reef is also tightly cemented with little observable porosity as is common for the eastern reefs of the Edgecliff.

3.4	0.4	Return to cars. Continue north. Right turn onto Reservoir Road.
4.1	0.7	Left turn onto Roberts Hill Road.
5.5	1.4	STOP 2. Albrights Reef. Park cars and proceed east onto reef.

NOTE: Private Property. **DO NOT** enter without first obtaining permission at house on north side of hill.

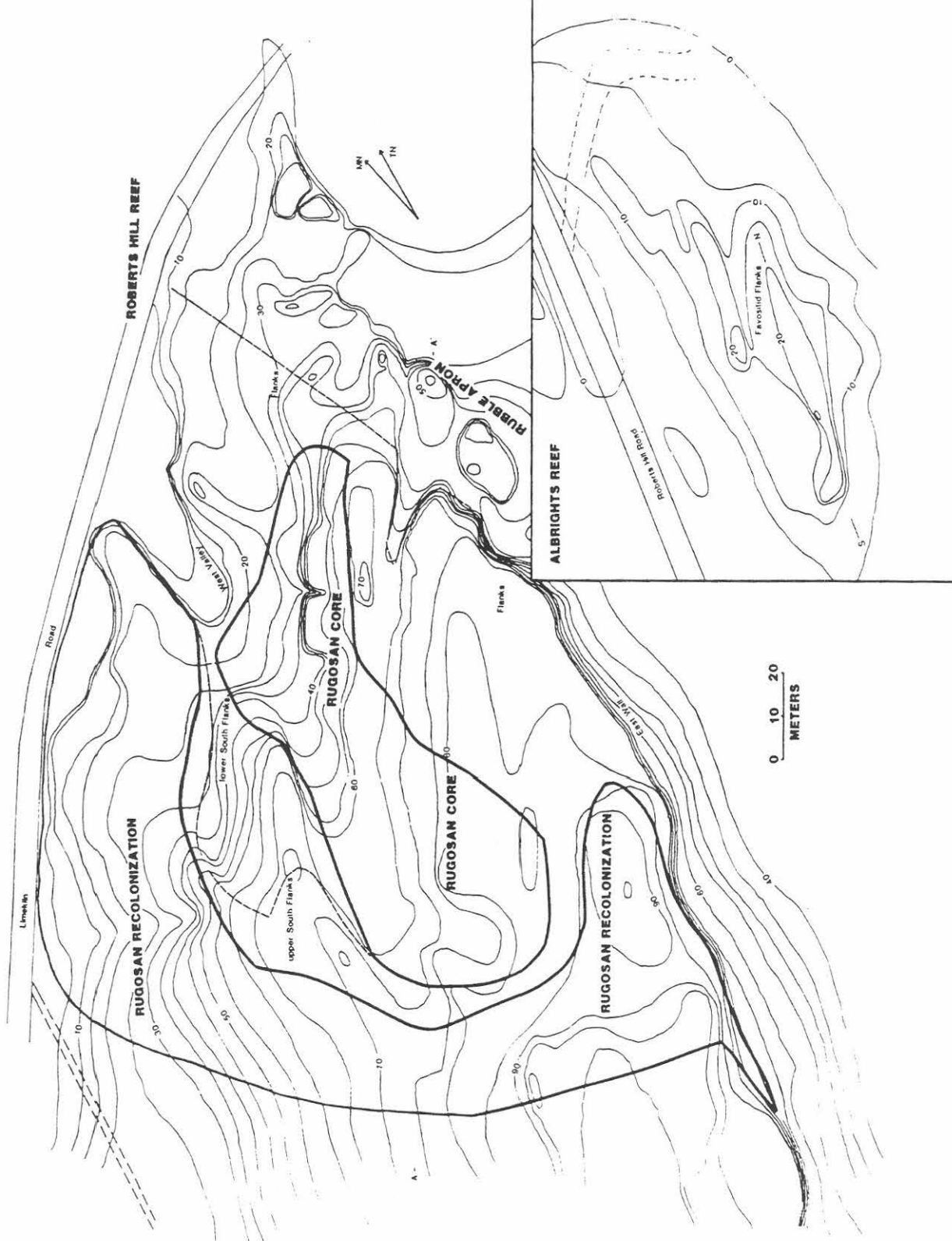


Figure 8. Map of Roberts Hill and Albrights reef illustrating location of core facies and flanking beds. Rugosan Recolonization facies marks re-establishment of colonial rugosans on mound. (From Wolosz, 1985).

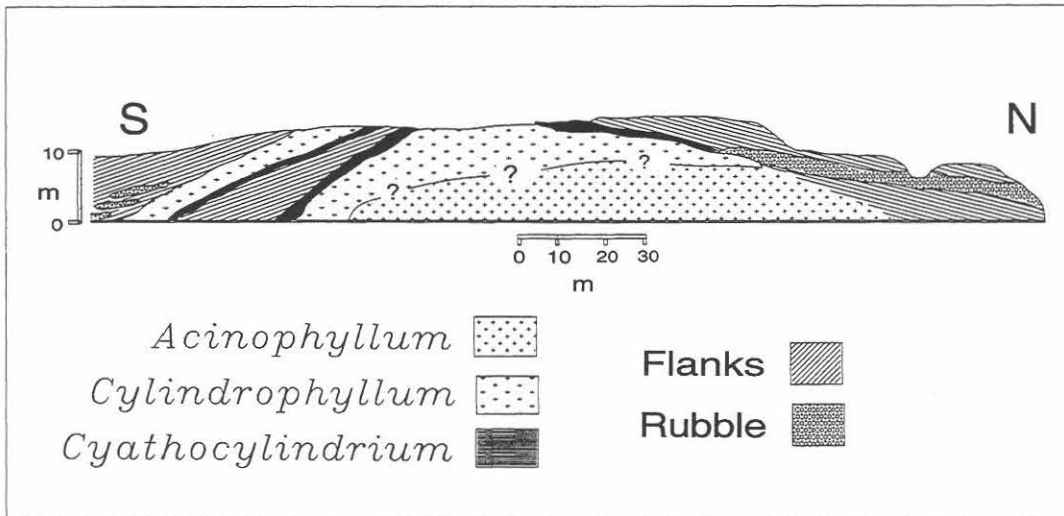


Figure 9. Fore- to back-reef cross-section through Roberts Hill Reef. Core facies is a rugosan successional mound. Note sequential development of colonial rugosan biofacies. (From Wolosz, 1992).

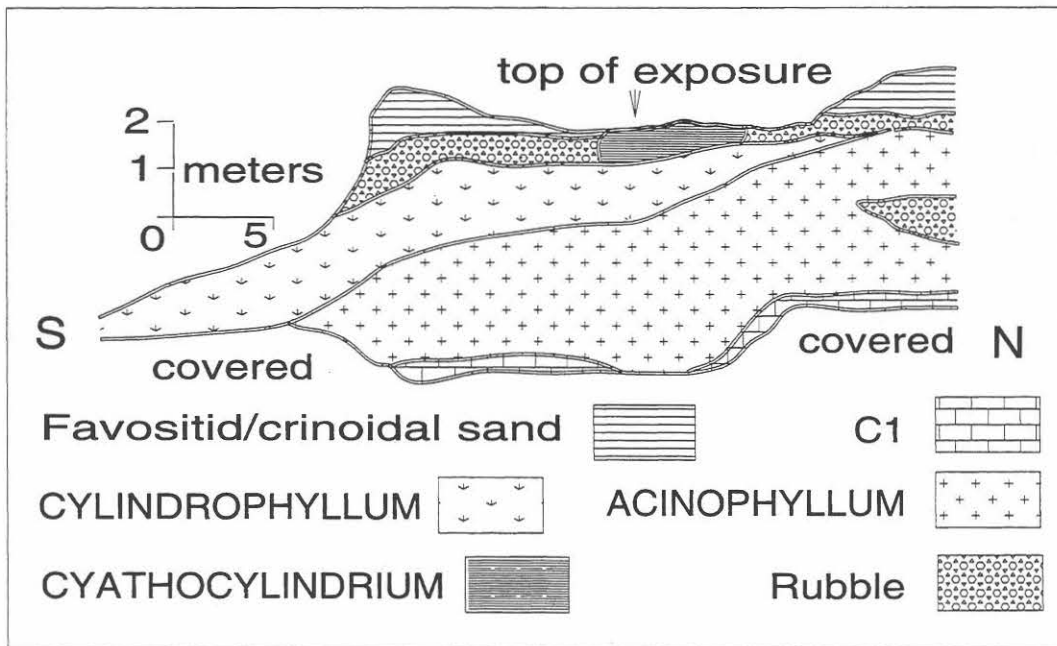


Figure 10. Interpretative cross-section of eastern cliff face core exposure at Albright's Reef. Note successional facies patterns of rugosa community development. (From Wolosz, 1985).

Albrights Reef

This exposure is a small erosional remnant of a large successional mound similar to Roberts Hill (Wolosz, 1985). It is an important exposure because it offers an excellent cross-section through the rugosan mound itself, exposes the basal contact of the mound with the underlying Edgecliff C1 facies, and allows direct observation of the successional patterns among the rugosan fauna during development of the mound (see Figure 10).

		Return to cars. Continue north.
5.9	0.4	bear left on Route 51
6.0	0.1	bear right onto West Dean's Mill Road
		bear right
7.65	1.65	
8.05	0.4	left turn onto Dean's Mill Road
8.5	0.45	left turn onto Aquetuck Road
9.15	0.65	left at stop onto Route 143
10.15	1.0	bear right onto Route 102
12.15	2.0	bear left onto Route 102
12.3	0.15	Stop 3. North Coxsackie Reef. Enter using dirt road on south side of road.

NOTE: Private Property. **DO NOT** enter without first obtaining permission at house on north-west side of intersection of Route 102 and 106 (approx. 0.7 mile south).

North Coxsackie Reef

The North Coxsackie reef (Figures 11 and 12) is rooted in Edgecliff crinoidal grainstone/packstone - the C1 unit is absent. The main mass of the reef is a Composite Thicket/Bank reef roughly 280 meters long (north-south) by 220 meters wide (east-west) with an estimated thickness of 15 meters. Volumetrically, the Favositid/Crinoidal Sand Paleocommunity dominates this reef, having formed a bank, with beds dipping gently (5 to 12 degrees) away from the center of the structure. Interbedded with the favositid/crinoidal sand facies are two roughly 0.3 meter thick horizons of the Phaceloid Colonial Rugosan Paleocommunity - each representing a single thickening event which covered the entire mound. These thickets are dominated by Cyathocylindrium, although Cylindrophyllum and Acinophyllum are common.

A cliff along the northwestern edge of the reef exposes an approximately 6 meters thick by approximately 60 meters long Acinophyllum mound, underlain by crinoidal grainstone/-packstone. This small structure is located to the west of, and stratigraphically just above, a well defined rubble horizon exposed along the northeastern edge of the main reef, indicating that it grew on the down-current side of the mound. This mound is similar to the early Acinophyllum stage of reef growth exposed at Albright's reef, with the exception that the calcisilt-packstone matrix surrounding the corallites is more finely bioclastic, containing debris transported from the larger mound to the south.

		Return to cars. Continue North on Route 102.
15.2	2.9	left turn onto Route 396
15.9	0.7	straight onto Route 301 (396 ends)
19.7	4.5	right onto Route 301
21.2	1.5	left onto route 443
23.1	1.9	right turn onto Route 85
25.25	2.15	left turn onto Route 157
32.75	7.5	right turn onto Ketchum Road
32.9	0.15	Stop 4. Thompson's Lake Bioherm

NOTE: Private Property. **DO NOT** enter without first obtaining permission at house to the southwest of hill on Route 157.

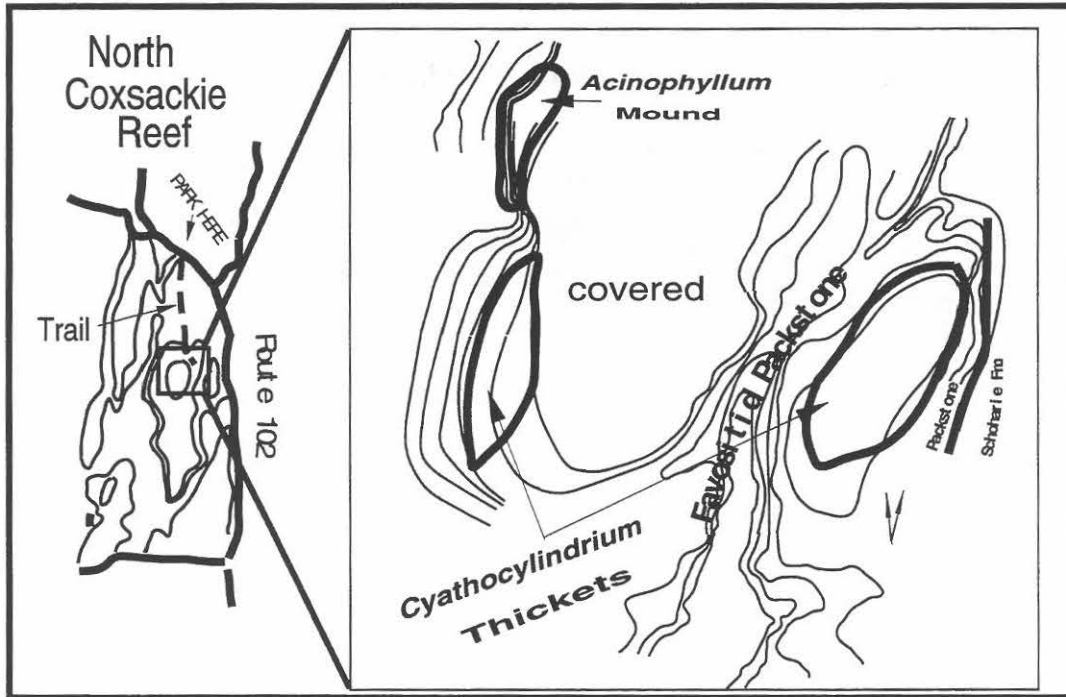


Figure 11. Map of North Cocksackie Reef.

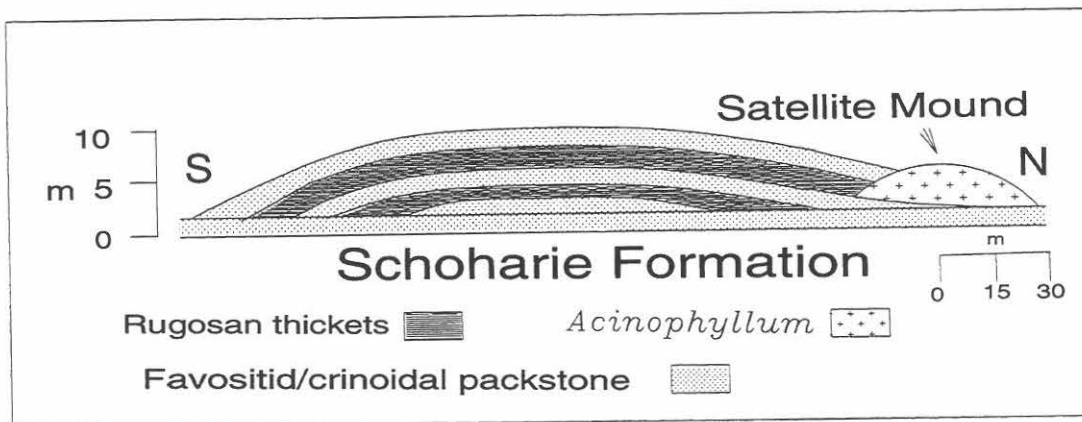


Figure 12. Cross-section of North Cocksackie Reef. Dark horizons (with thickness greatly exaggerated) represent thickets of *Cyathocylindrium*. Other beds are Favositid/crinoidal sand facies. Note presence of *Acinophyllum* mound in back-reef (From Wolosz, 1992a).

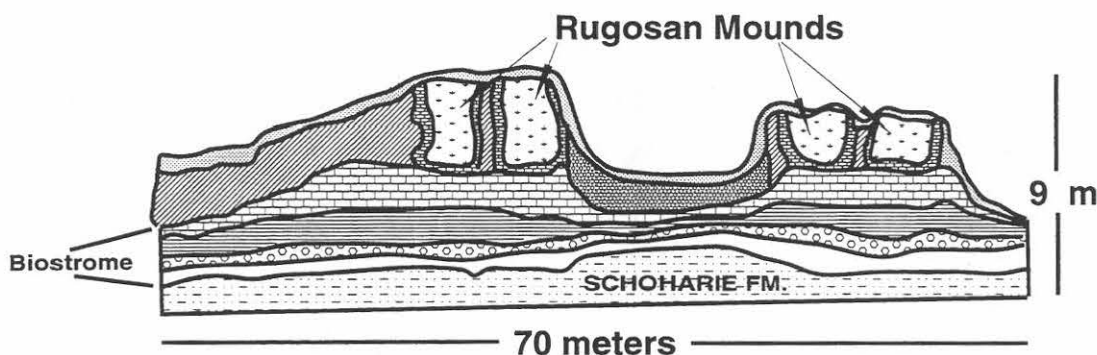


Figure 13. Thompson's Lake Bioherm. View looking south from Ketchum Road. Redrawn from, and with patterning following, Williams (1980).

Thompson's Lake Bioherm

This exposure has been described by Williams (1980), who divided the exposure into eleven different micro-facies, discussion of which is beyond the scope of this report. Simply put, most of the small elongate hill is made up of packstone with abundant favositid colonies. Along the roadcut two small rugosan mounds can be observed (Figure 13), underlain by roughly 2.6 meters of Edgecliff biostomal deposits, including a well developed *Cystiphyloides* biostrome which can be followed around the entire hill as a recessive bed. These facies are analogous to the shallow water deposits found near Port Colborne, Ontario, Canada as described in the text.

The underlying contact with the Schoharie Formation is easily observable, and represents initial shallow water deposition with grainstone as the basal Edgecliff facies.

Return to cars. Turn around and proceed west on Ketchum Road.

33.05	0.15	right turn onto Route 157
34.75	1.7	right turn onto Route 156
38.15	3.4	left turn onto Route 146 West
38.45	0.3	right turn onto Route 397
41.55	3.1	left turn onto Route 20
86.75	45.2	Right turn on Route 80.(see Figure 14)
88.25	1.5	Left turn onto Koenig Road.
88.85	0.6	Bare left onto Mt. Tom Road
88.95	0.1	STOP 5. Mt. Tom Reef. The reef makes up the large hill to the south of the road.

NOTE: Private Property. **DO NOT** enter without first obtaining permission at house at intersection of Koenig and Mt. Tom roads. Examination of Mt. Tom #6 requires permission from farm on Koenig Road to the north of hill. Examination of Mt. Tom #2 requires permission from Mercy Hill Farm.

MT. TOM REEF

NOTE: Private Property. **DO NOT** enter without first obtaining permission at house at intersection of Koenig and Mt. Tom roads. Examination of Mt. Tom #6 requires permission from farm on Koenig Road to the north of hill. Examination of Mt. Tom #2 requires permission from Mercy Hill Farm.

Paquette and Wolosz (1987) argued that three reef exposures - the Mt. Tom, Mt. Tom #2, and Mt. Tom #6 (Figure 14) - represent the extensively eroded remains of a single large mound/bank reef. The .6 by .8 km. areal extent of this structure would make Mt. Tom both the largest exposed Edgecliff reef and the only known surface exposure of this reef type (these reefs are commonly referred to as pinnacle reefs in the subsurface).

The mound/bank nature of Mt. Tom #1 is displayed in the cliff face along the southeast side of the hill (Figure 15). The reef is underlain by the basal Edgecliff calcisiltite, with the base of the reef marked by thickets of *Acinophyllum*. Small phaceloid colonial rugosan mounds (again, mainly *Acinophyllum*) can be observed along the cliff near the base of the reef. These small mounds and thickets coalesced to begin the formation of the larger structure. Dominance of the initial large mound shifted between *Acinophyllum* and *Cylindrophyllum* prior to onlapping by the crinoidal sands of the favositid/crinoidal sand paleocommunity. A second mound stage made up of *Cylindrophyllum* thickets overlies these grainstones and packstones. In turn, the second mound stage is itself onlapped and eventually swamped by the favositid/crinoidal sand paleocommunity (exposed further back on the top of the hill, not shown in Figure 2). Overall, Mt. Tom #1 is roughly 18m thick as preserved.

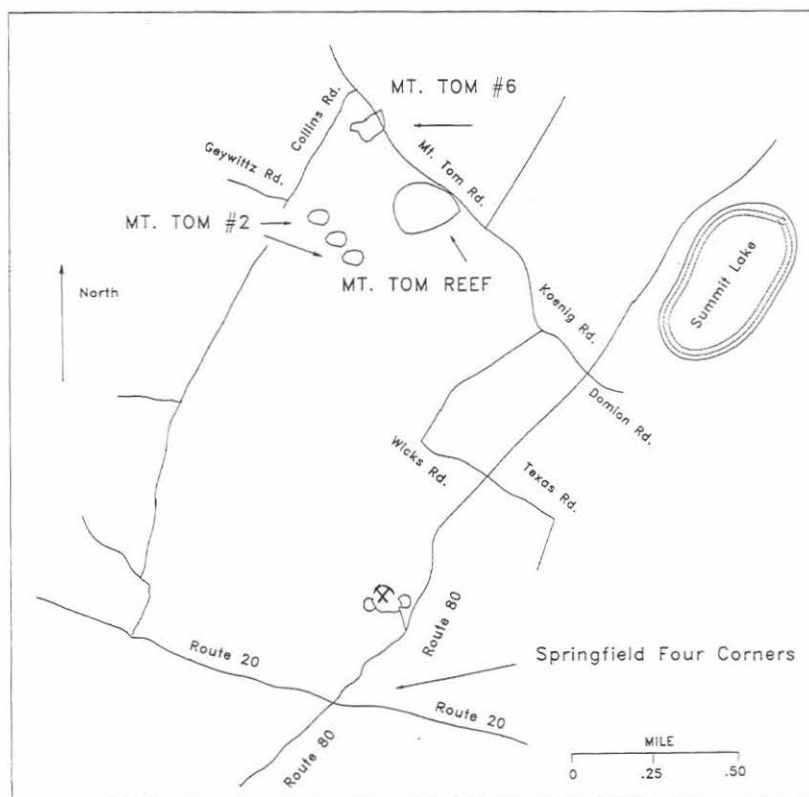


Figure 14. Location map of Mt. Tom reef #'s 1, 2 and 6. Note relative positions of reefs.

Wolosz and Paquette (1988) have interpreted this mound/bank/mound/bank pattern as catch-up/fall back cycles controlled by fluctuations in water depth above the top of the reef. It is important to note that the second mound building stage at Mt. Tom #1 (Figure 2) does not drape the entire pre-existing structure, but is instead restricted to the top of that structure. In effect, during bank stage, the reef was a high relief platform on the sea-floor with its top within the ecologic mound building zone of the colonial rugosans. Upward growth of the reef is mainly due to the repetitive establishment of new mounds on the top of the platform. As sea-level was approached, the mound building colonial rugosans were overwhelmed by increased turbulence conditions and the mounds overlapped by encroaching crinoidal sands producing a bank stage; but with sea-level rise the mounds became re-established. This shifting between rugosan mound/thicket construction and the favositid/crinoidal sand paleocommunity has been attributed to a water turbulence controlled community succession (Wolosz, 1992b).

Following the initial mound building stage, lateral growth of Mt. Tom appears to have been due mainly to deposition of crinoidal debris flanks with occasional small mound structures (satellite mounds) growing in those flanks (see discussion of Mt. Tom #2).

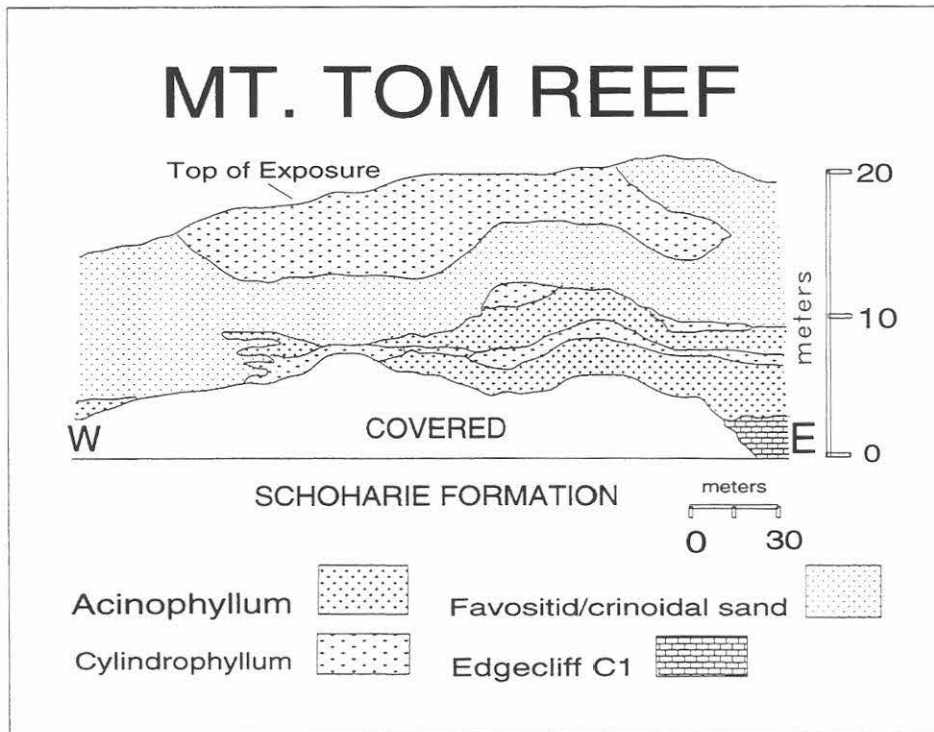


Figure 15. Cross-section of cliff face at Mt. Tom reef illustrating a mound/bank structure. Lower mound develops through two colonial rugosan successional cycles before being swamped by crinoidal sands. Second appearance of rugosan paleocommunity occurs as a mound within area of crinoidal sandstone, and does not drape entire reef structure (from Wolosz, 1992).

Diagenetic patterns in Mt. Tom are similar to those in subsurface pinnacle reefs. The initial mound facies (base of reef) is tightly cemented, as is Mt. Tom #2; while the upper parts of the reef and Mt. Tom #6 exhibit a small primary porosity.

89.45

0.5

Return to cars and continue northwest on Mt. Tom Road.

STOP 6. Mt. Tom #6 forms the low, wooded ridge to the southwest of Mt. Tom Road.

Mt. Tom #6

Mt. Tom #6 is a small ridge which consists mainly of bedded crinoidal grainstone/packstone. Randomly distributed small overturned favositids are common as are both solitary and phaceloid rugosans, but no evidence of mound formation is present. Topographically, Mt. Tom #6 is at the same elevation as the present top of Mt. Tom. Since the regional southwest dip of about 18 meters/kilometer (Rickard and Zenger, 1964, p.5) would not greatly alter this topographic relationship, the elevations of the Mt. Tom #6 exposure and the top of Mt. Tom were probably also equivalent at the time of deposition. However, when one observes Mt. Tom #6 from Collins Road (see map, Figure 14), the quassa-like nature of this small ridge is evident, with the dip slope pointing to the north-northwest, directly away from the main mass of Mt. Tom.

Paquette and Wolosz (1987) cited this as evidence that the two exposures are parts of one reef, with Mt. Tom #6 consisting of distal flank beds. Mt. Tom reef would then be at least 0.8km. long on an northwest axis from Mt. Tom #1 to Mt. Tom #6.

89.6

0.15

Return to cars and continue northwest on Mt. Tom Road.

Left turn onto Collins Road.

90.1

0.5

STOP 7. Mt. Tom #2.

Intersection of Collins and

Geywitz Roads. Mt. Tom #2 Reef.

Mt. Tom #2 Reef.

Leave cars and proceed east from the intersection. Mt. Tom #2 forms the low hill to the south of the small creek, and numerous small outcrops may be examined along the south side of the creek valley or on the hill itself. A small quarry on the northwest edge of the hillside exposes bedded Edgecliff facies, while a small rugosan mound is located just to the southeast of the quarry among the trees. In contrast to Mt. Tom #2 lies to the west of Mt. Tom #1 and is topo-graphically roughly 18 meters below #6. Stratigraphically older beds can be examined here, with the Edgecliff/Carlisle Center contact marked by the appearance of a spring just east of the intersection of Collins and Geywitz Roads. A small quarry visible from the road exposes bedded Edgecliff with overturned colonial coral. To the southeast of this quarry is an exposure of a small colonial rugosan mound roughly 17 meters across and of indeterminate thickness. East from the quarry, along the south side of the creek, there are numerous outcrops of bedded crinoidal grainstone/packstone with abundant favositids. Small patches or lenses of colonial rugosans within the bedded packstones are common, and represent small satellite thickets or mounds which appear to range stratigraphically from near the C1/C2 contact (roughly the point at which growth of Mt. Tom #1 began), upwards to about 6 meters above that contact. The packstones surrounding these upper mounds dip away from Mt. Tom #1 at roughly 15 degrees.

TYING THE EXPOSURES TOGETHER -DEVELOPMENT OF THE MT. TOM PINNACLE REEF (FROM WOLOSZ, ET AL., 1991)

Figure 5 illustrates an interpreted developmental history for the Mt. Tom (small) pinnacle reef. As sea-level dropped from possible deep water conditions of Carlisle Center deposition through the early Edgecliff (C1), abundant small rugosan thickets and mounds began to form in the late C1 calcisilts. By the beginning of C2 deposition these thickets and small mounds had begun to coalesce to form the initial large mound at Mt. Tom #1 (Mound Stage I), while an abundance of other small mounds dotted the crinoidal sand sea-floor as satellites to the growing reef. Crinoidal debris of the favositid/crinoidal sand paleocommunity lapped up onto

the large mound, eventually forming flank beds which spread outward from the main mass of the reef. Small satellite mounds continued to develop along distal flank beds (Mt. Tom #2), contributing to the overall volume of the reef structure, but never coalescing into a large central structure similar to Mt. Tom #1. Continued sea-level drop resulted in the cessation of rugosan mound growth and the eventual swamping of the mound by the crinoidal sand beds, resulting in Bank Stage I. A second cycle of sea-level rise resulted in the establishment of new rugosan thickets and mounds on the top of the bank (Mound Stage II), but later shallowing over the crest of the reef again caused the demise of the colonial rugosans and the re-establishment of the favositid/crinoidal sand paleocommunity in Bank Stage II.

End of trip. Return to cars follow Collins Road back to Route 20.

THE GEOLOGY OF CLARKSVILLE CAVE, ALBANY COUNTY, NEW YORK

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SECTION A: Modification of Pre-Woodfordian Caves by Glacial Meltwater Invasion in East-Central New York

ABSTRACT

Periods of high glacial meltwater have altered some preglacial cave-passage configurations. Floodwater and fossil karst features, whose formation cannot be explained based on available water from the surrounding watershed, are found superposed on actively forming cave passages. These features may be recognized through correlation of watershed boundaries, peak-runoff observations through a cave system, the presence of anomalous in-cave and surface features, and the geomorphic interpretation of the area in question. Knowledge of minimum rates of karstification may be used to infer climatic conditions, making possible the reconstruction of the hydrology associated with deglaciation.

Clarksville Cave, situated in the hamlet of Clarksville, New York, provides an excellent example of invasion by Wisconsinan meltwater on a pre-Woodfordian cave system. Vadose development of a major part of the explored cave has occurred preferentially aslant a thrust-fault ramp, often along a calcite bed/limestone contact created by pressure solution. Other fault-related features include slickensides, extension veins, fault-bend folds, stylolites and the repeated basal Onondaga Limestone and impermeable Schoharie Formation thrust below the Onondaga Limestone stratigraphic column. An imbricate thrust east of the cave has upthrown the Esopus Shale against the Onondaga Limestone, forcing the development of an inefficient resurgence at the baselevel Mill Pond.

During the Wisconsinan glacial stage, subglacial meltwater formed a series of now abandoned bedrock channels and paleogorges that, due in part to topographic controls, found outlets along and over the flank of the Helderberg Escarpment. Some of this meltwater was pirated into Clarksville Cave where inefficient outlets resulted in the formation of higher in-cave "intermittent phreatic" levels not controlled by the thrust fault. These levels abruptly truncate and grade to lower vadose passages. The character of these upper levels, the paleogorge and related caves, and elevated paleo-insurgence points correlate with described alpine karst settings.

PHYSICAL SETTING

Clarksville Cave is nestled under the flank of a low wooded ridge virtually in the center of the hamlet of Clarksville, New York (Fig. 1). It is formed in the lower subunits of the Devonian Onondaga Limestone that were deposited approximately 380 million years ago. Its large passage size, up to 15 feet high and 40 feet wide, complete with multiple levels, makes it unique among other, usually smaller, Onondaga caves.

The Clarksville area lies at an elevation of 600 to 800 feet msl. It is situated within the foothills of the Helderberg Plateau, a part of the Appalachian Plateau physiographic province. Meyerhoff (1972) attributed the present day drainage pattern of this region to the normal erosive processes of stream adjustment to structure. The Helderberg Plateau has been modified by stream incision, physical weathering, glacial and postglacial erosion, and deposition during the Cenozoic era (Dineen, 1987).

In Garver, J.I., and Smith, J.A. (editors), Field Trips for the 67th annual meeting of the New York State Geological Association, Union College, Schenectady NY, 1995, p. 251-273.

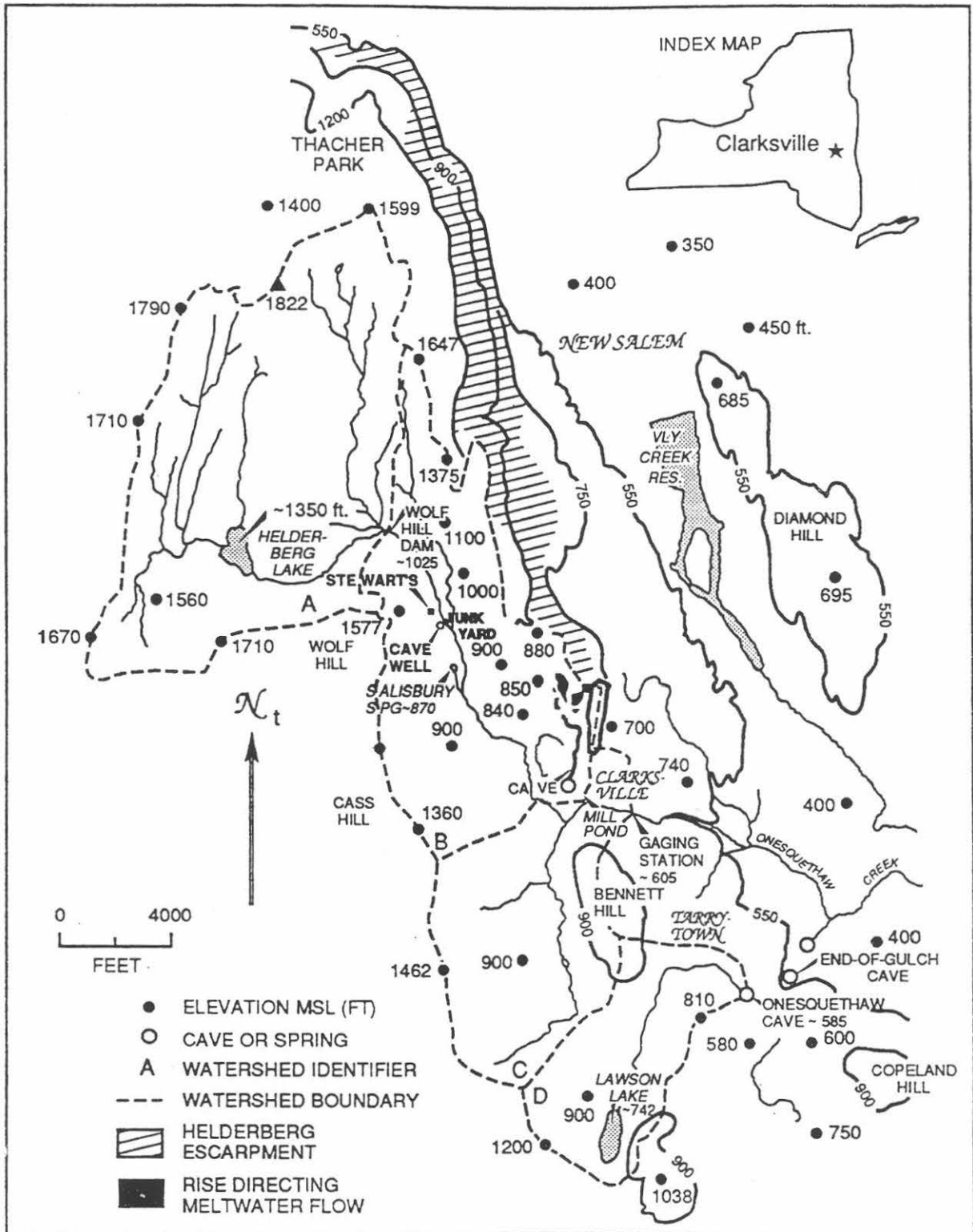


FIGURE 1: TOPOGRAPHY, DRAINAGE BASINS, AND SELECTED FEATURES ALONG THE HELDERBERG ESCARPMENT, ALBANY COUNTY, NEW YORK. WATERSHED BOUNDARIES BASED ON TOPOGRAPHY, GEOLOGY, AND DYE TRACES. CONTOUR INTERVALS VARIABLE FOR CLARITY.

Dineen (1987) has determined that present-day drainage trends in the Hudson Valley were established before the Wisconsin glacialiation, sometime prior to 70,000 years ago. Glacial striations in two locations near Clarksville further indicate that today's drainage was in place prior to inundation by Woodfordian ice. The direction of glacial movement was almost exactly north-south (S13°W), with a maximum ice thickness on the order of one mile about 22,000 years before present (Dineen, pers. comm.). Late Pleistocene drainage along the Onesquethaw Creek was probably little different from what it is today. Evidence presented in this paper argues for pre-Woodfordian cave development.

STRUCTURAL SETTING

The hamlets of Clarksville, Tarrytown and Feura Bush have all been subjected to extensive faulting. Marshak (1986), Marshak and Engelder (1987), and Cassie (1990) discuss structural deformation within parts of the Hudson Valley Fold-Thrust Belt (HVB). The HVB extends roughly from Kingston to Albany, New York, extending to a maximum of 20 kilometers east and west of the Hudson River (Marshak and others, 1986). The deformation may have occurred during the Acadian (Cassie, 1990) or Alleghanian orogenies (Geiser and Engelder, 1983), or during Mount Marion deposition (Murphy and others, 1980).

Faulting and deformation of the Esopus Shale, Schoharie Formation, and Onondaga Limestone, throughout the Clarksville area, may represent the farthest northwestern exposure of the Hudson Valley Fold-Thrust Belt. The extensive structural deformation present throughout Clarksville and the previously documented southern parts of the HVB are characteristic of deformation of sedimentary rock under relatively low pressure and temperature conditions (Marshak and Engelder, 1985). Mapping of the structural or bedrock geology in the area, both on the surface and in the cave, reveals that faulting in the Clarksville area is characteristic of either an imbricate thrust zone or a duplex.

At least six elongate ridges, trending north-south approximately along strike of the faults, are unevenly spaced throughout the Clarksville area. They often exhibit extensive fault-bend folding, slickensides, and in places an anticlinal structure. These limestone ridges, which are underlain by one or more basal thrust faults, can be mapped for distances of up to one mile. One such deformed ridge, situated at the eastern end of Clarksville, has been breached by Onesquethaw Creek. Perhaps the most prominent example is found in the upper Onesquethaw Creek gorge.

The upper Onesquethaw Creek gorge exhibits the best out-of-cave exposure of the repeated basal Onondaga Limestone and the impermeable Schoharie Formation (a quartz rich limestone) thrust below the Onondaga Limestone stratigraphic column. Here much of the bed of Onesquethaw Creek is guided by fault-zone features. The thrust-fault ramp, associated thick calcite bed, and fault-bend folds are the same as those along which Clarksville Cave has developed, except that they are farther south along strike.

STRUCTURAL FEATURES INFLUENCING GROUNDWATER FLOW IN THE KARST AQUIFER

Faulting in and immediately east of Clarksville Cave has resulted in thrusting, deformation, and upward movement of impermeable bedrock units underlying the Onondaga Limestone (Schoharie Formation and Esopus Shale) into a position that makes the eastern escape of groundwater impossible. The gentle southwesterly dip of the bedrock of the Mill Pond aquifer (Fig. 1) fails to direct all subsurface flow in this direction. Instead, significantly higher surface topography to the southwest (e.g., Wolf Hill and Cass Hill) retards dissolution in this direction, in favor of the 1.3° apparent dip between Wolf Hill Dam and the base-level discharge point at Mill Pond. Tracer studies generally verify this predicted flow path, at least during periods of low discharge. However, tracer studies also document an unexpected

easterly diversion of moderate- to high-discharge waters through Pauley Avenue in Clarksville Cave. This is significantly farther north than the Mill Pond. This easterly deflection of floodwaters may occur in response to an inefficient outlet and conduit leading to the Mill Pond.

Pauley Avenue floodwaters flow easterly until they become perched on a thin bed of impermeable Schoharie Formation that has been thrust below the basal, or lower, non-cherty subunit of the Onondaga Limestone. Cave diver John Schweyen (pers. comm.) reports the presence of the Schoharie Formation overlying the lower non-cherty subunit of the Onondaga Limestone approximately 700 feet west of the north-south trending Clarksville Cave. This number reflects a minimum westerly displacement of beds above the fault ramp. Floodwaters remain perched, flowing down the apparent dip of the Schoharie Formation, until they encounter a fractured zone along a more steeply inclined part of the fault ramp. Here, subsurface water is deflected sharply to the south and aslant the strike and dip of the inclined fault plane, with the possible localized exception of following a horse for 200 feet north of the Lake Room.

Pirated surface water must rise at the Mill Pond, because the impermeable Esopus Shale is thrust vertically upward against the cavernous Onondaga Limestone. The leading edge of this upthrown shale formation, an imbricate thrust sheet separate from the fault zone that Clarksville Cave formed along, trends roughly north-south (Fig. 2). The Esopus Shale and thrust-fault-induced fault-bend folds in the Onondaga Limestone, present slightly west of the upthrown Esopus, ultimately form a wall or barrier to easterly karstic groundwater flow for a distance that is well in excess of one-half mile.

However, this geologic barrier has only retarded eastern groundwater movement in two locations: (1) east of the known parts of the Waterfall Passage and (2) at the Mill Pond spring resurgence. Formation of most of the north-south oriented cave occurred preferentially aslant an inclined ramp of a thrust fault. Deformation along this thrust plane has produced a fault zone with at least three easily discerned, slickensided surfaces. Although separate, they occur within a few feet of one another. In much of the cave, this fault ramp is accented by one or more prominent calcite beds, often accompanied by a zone of stylolites and calcite-filled extension veins. This calcite bed is podiform in shape, with a central thickness ranging up to eight inches. Whereas the calcite forms a continuous bed of variable thickness aslant the strike of the thrust fault, it is most pronounced in the Gregory Section of the cave, where the vadose part of the cave is steeply inclined along the fault ramp. The ledge forming the outlet of the Waterfall Passage is the same buff brown to black, weathered Schoharie Formation, with underlying calcite bed, as seen in the upper Onesquethaw Creek gorge. The thicker zones of the calcite bed, where dissolution and crystallization are greatest, coincide with the more inclined segments of the thrust ramp, where the stress was highest. Occasional remnant calcite blocks, up to eight inches in thickness, in the Ward's Section of the cave provide the only evidence of the former presence of the thick calcite bed.

Ramsay (1980) provides evidence that similar "extension veins are formed by an accretionary process involving the formation of a narrow fracture followed by the filling of the open space by crystalline material, a mechanism termed crack-seal." Such stress-induced chemical transfer, or pressure solution of materials, seems to be relatively common (Ramsay, 1980). The characteristic crack-seal mechanism of repeated tectonic stress (Ramsay, 1980) is best illustrated in Onesquethaw Cave, situated 2 miles southeast of Clarksville, where calcite infilling aslant the ramp of a thrust fault reaches a maximum thickness of 27 inches. Here, insight into the fault style and repeated activation in the area is suggested by the presence of multiple calcite-vein infilling events along the fault ramp.

Successive cracks often occur along vein-matrix contacts of a previously sealed crack system, because this is mechanically the weakest surface in the rock (Ramsay, 1980). Fractures have been found to increase towards major faults. The higher the fracture frequency, the higher the percentage of calcite-filled fractures (Carrio-Schaffhauser and Gaviglio, 1990). It is a combination of this mechanically weaker calcite bed/Onondaga limestone boundary and related fault partings, all present along this inclined thrust ramp, that have served to orient the north-south segment of Clarksville Cave. Further structural

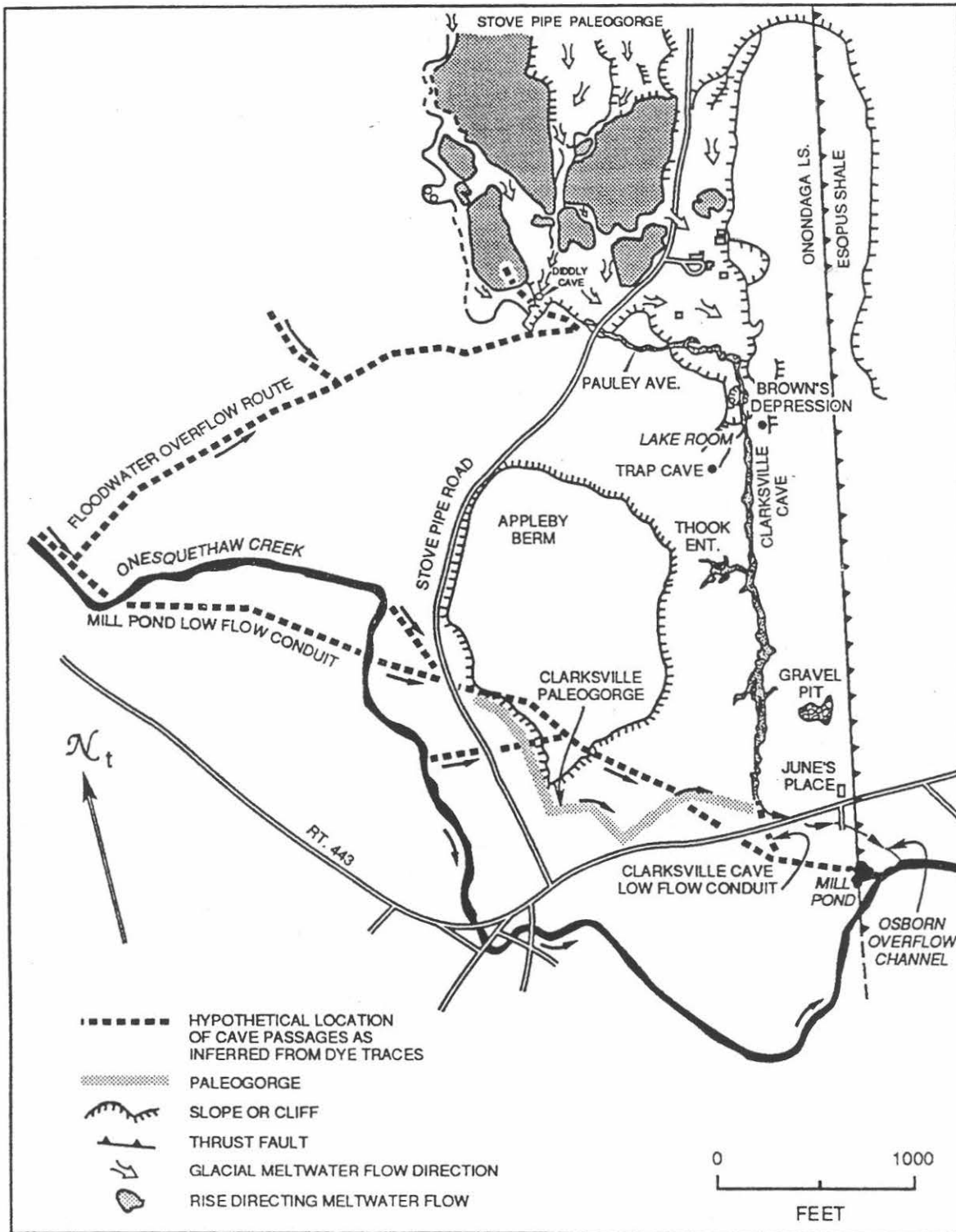


FIGURE 2 CONFIGURATION OF DRAINAGE IN THE VICINITY OF CLARKSVILLE, NEW YORK. SHOWN ARE CLARKSVILLE CAVE AND PRESENT AND PAST ROUTES OF FLOW. DYE TRACES ARE OFTEN IMPORTANT FOR DETERMINING SUBSURFACE FLOW ROUTES ESTABLISHED IN FORMER GEOLOGIC PERIODS.

and hydraulic control of cave-forming waters may also be locally attributed to perching on a fault-thinned Schoharie Formation. Similarly, much of Onesquethaw Cave has developed down and along the mechanically weaker vein-matrix contact. Both caves exhibit characteristic fault-bend folds, stylolites, and extension veins adjacent to the prominent thrust plane.

Of major importance to the development of both caves was Cenozoic structural deformation which provided a preferential solutional pathway along the inclined surface of a fault ramp. A steep hydraulic gradient was thus set up between infiltrating waters and their resurgence points along fault ramps. These faults may then be considered as both negative and positive influences on groundwater flow and cavern development: negative in the sense that downward dissolution did not readily penetrate far below the fault zone (Kastning, 1977 and 1984), and positive in the sense that almost the complete trend of the caves follows a structurally weakened zone of increased permeability.

INTERMITTENT PHREATIC PASSAGES

Meltwater invasion of pre-Woodfordian passages in Clarksville Cave occurred during glaciation, significantly enlarging the cave and its tributary conduits within the aquifer. Observation of the degree of flooding within the cave during major storm and runoff events reveals that only the lowest levels of the cave carry water along the fault zone. Two abandoned upper-level passages, both with relatively consistent ceiling elevations, were identified via a leveling survey. The level of these passages is determined by the relative uniformity of their ceiling heights. The highest of these two upper-level passages extends from the Lake Room, through the Big Room, until its truncation in the Pixie Passages immediately above the Corkscrew (*see* detailed map of Clarksville Cave in Fig. 3: Section B). This 714-foot level can roughly be characterized as the meandering upper level of Perry Avenue. In places, the lower ceiling elevation is controlled by chert beds. The 698-foot level extends from the Bathtub Passage through Upper Cook Avenue, where the passage is truncated by breakdown, and where flow had been diverted down a steeply dipping, fault-plane-controlled tube leading to Lower Cook Avenue.

These large upper-level passages are generally high and dry, and are sometimes tubular in cross-section, suggesting a water-table formation; they undulate upward and downward as is typical of phreatic passages, and have formed with only limited influence from the fault plane. However, they lack the low hydraulic gradient typical of phreatic or water-table origin, are discontinuous in size and extent, and truncate suddenly or become much reduced in cross-sectional area, grading systematically to active pre-Woodfordian vadose passages. The relatively smaller drain size at the down-gradient end of these passages, compared with the cross-sectional area of these floodwater conduits, resulted in temporary phreatic conditions within the cave, quite dissimilar from the conditions of normal phreatic water. Palmer (1991) documents similar floodwater formation of conduits behind local passage impediments such as collapse debris, insoluble beds, or sediment fill, where aggressive water results in rapid passage enlargement. Formation of these "intermittent phreatic" floodwater conduits, coincident with the direct influx of large quantities of subglacial meltwater, occurred under alpine karst conditions.

A third solutionally developed upper level is also identifiable in the cave at 739 feet msl. This discontinuous level is represented in only a short segment of the cave proximal to the Root Room (northeast of the Lost Rock Hammer Room) and some nearby domes. It clearly represents the maximum flood level attained in the cave. Solution domes at the 739-foot level are characteristic flood water-injection features. Like the two lower abandoned levels, it is systematically graded to the actively forming lowest level. These three abandoned levels are interpreted as reflective of different glacial discharges (perhaps seasonal fluctuations) coursing through the cave, rather than different time-based developmental stages. Variable discharges, perhaps influenced by variable outlet efficiencies and climatic conditions during glaciation, are inferred for formation of the levels.

There is no obvious mechanism present today that would explain both the elevation of these upper-level passages and their configuration, that exhibits little fault control. Other than the passage alignment, fault control appears to be a significant developmental factor only in the low-discharge lower-level passages. A ^{14}C date ($27,350 \pm 750$ yrs BP) obtained from a wood rat bone sample excavated from upgradient Diddy Cave provides evidence for 1) a den site used in pre-Woodfordian time; and 2) pre-Woodfordian cave development (Steadman, pers. comm.). Further argument for pre-Woodfordian origin of the vadose-level passages in the cave stems from the recognition that the mean annual precipitation and the size of the Mill Pond watershed has probably not changed significantly in the last 10,000 years. Funk (1989), through the interpretation of archaeological sites, established that climatic conditions within the last 10,000 years were at times either dryer or similar to that of today. Thus, the availability of the significant discharges necessary to form the upper-level passages was not there postglacially.

These upper-level passages and related higher solution features on ceilings (upward to 739 feet) indicate that they are younger than many lower passages, having formed in response to aggressive glacial floodwaters behind an inefficient outlet. It is hypothesized that the lower-level Clarksville Cave passages served as a natural *in situ* drainage system for glacial meltwaters underneath warm-based Wisconsinan ice. Higher-level cave passages (*e.g.*, the 698- and 714-foot levels) formed in response to the massive influx of subglacial meltwaters behind inefficient, perhaps partially ice-blocked outlets. Similarly, the formation of high-ceiling solution domes, anastomoses, pendants, spongework-like dissolution, and diversion passages may be attributed to floodwater invasion. The gradation of upper-level conduits tributary to the lower levels of the cave, from relict meltwater infiltration points, also lends supportive evidence for a pre-Woodfordian origin of the linear (N12°E) Clarksville Cave passage.

RELICT KARST

A number of relict karst features are present both proximal to Clarksville Cave and to the northwest within the same watershed. These include a number of small shafts and caves (*e.g.*, Trap Cave, North and Thook entrances) that receive only minor amounts of direct meteoric or snowmelt infiltration today. The most important relict karst feature is the Stove Pipe Paleogorge (Fig. 2). This abandoned rockcut gorge grades directly into the pre-Woodfordian Clarksville Cave via the North Entrance, Brown's Depression area, Trap Cave, and the Thook Entrance. Its channel is well defined for most of its course. The morphology of upper reaches of the gorge is characteristic of an ice-marginal meltwater channel with small-scale hanging valleys, rather than a well-graded streambed which would be expected of a former channel of the Onesquethaw Creek. Sugden and John (1976) describe the ice-flow dynamics which cause favorable formation of drainage routes in bedrock versus ice. In some places, the channel configuration is such that only large quantities of water would have been capable of filling the channel sufficiently high enough to overflow into sub-parallel channels. This paleogorge is sharply truncated to the north by a downwardly sloping limestone cliff. A negligible catchment area is present (Fig. 1), certainly too small to carry any significant quantities of water or sediment into the cave as suggested by thick sediment banks and upper-level phreatic passages.

A second paleogorge, the Clarksville Paleogorge, proximal to Osborn Cave, may represent either a pre-Woodfordian drainage route of the Onesquethaw Creek or a channel carved around the Appleby Berm by glacial meltwaters. Gorges of this nature can form in a relatively short time if sufficient abrasive material is carried through it. Von Engeln (1911) documents the rapid formation of a rockcut marginal gorge at the outlet of the Hidden glacier in the Yakutat Bay Region of Alaska.

Brown's Depression (739 ft msl) is an important location in that it received significant paleo-streamflow from the northwest (Fig. 2). Stove Pipe Paleogorge streamflow, originating from a vast subglacial watershed to the north and northwest, incised a channel into the Onondaga Limestone from

the northeast until it reached the Hunter's Fissure Cave and Diddly Cave area. Here, this paleo-streamflow was responsible for the formation of these caves. From Hunter's Fissure Cave the paleo-streamflow spread out to the southeast over the gently undulating topography. However, its course was partially constrained by the elevationally higher surface topography to the west, north and east. Thus, much of the Stove Pipe Paleogorge streamflow was funneled southeast into the Brown's Depression/North Entrance (above Lake Room) area, where it entered Clarksville Cave.

During periods of low- to moderate-glacial discharge, meltwaters converged proximal to Brown's Depression where much of the southeastern discharge was retarded from flowing east by a low north-south trending limestone ridge (747 ft. msl). These meltwaters were pirated into Clarksville Cave through the Diddly Cave and Brown's Depression/North Entrance areas. Diddly Cave was recently dug open, increasing the length from 5 to 550 plus feet. A dive push, a short distance into the cave, led to a master conduit which is hydrologically linked to Clarksville Cave. The exposed limestone pavement, coupled with the steep hydraulic gradient present between insurgence and resurgence points, provided a favorable avenue for subsurface piracy.

High-discharge meltwater, with a glacial hydrostatic head, encountered the Clarksville Cave ridge and sought the most efficient outlet, flowing southward and over the ridge barrier. However, the even higher surveyed elevations of a number of abandoned surface stream infiltration points and their conduits leading to Clarksville Cave provide information on the large magnitude of subglacial discharge necessary for their formation. Some of these features include Trap Cave (753 ft. msl), the Thook Entrance (766 ft. msl), and a deep solutionally-enlarged joint near the Ward's Entrance (761 ft. msl) [excavation recently exposed cave passage at its base]. During periods of high glacial discharge, meltwaters probably flowed both within and outside the Stove Pipe Paleogorge channel. This meltwater splayed outward around, and possibly over, the Appleby Berm (Fig. 2). The alignment of the Thook Entrance passages and the Pixie Passages suggests that meltwater coursing around both sides of the Appleby Berm sought to enter the pre-Woodfordian Clarksville Cave through the most direct pathway. Meltwater thus entered the Clarksville and Stove Pipe paleogorges, encountering the Gregory Entrance to the cave, the Ward's Entrance, the Thook Entrance, Trap Cave, the North Entrance, Brown's Depression, and the jointed pavement above the cave.

SEDIMENT FILL

Thick deposits of sediment in the cave provide direct evidence of the quantity of material carried down the Stove Pipe Paleogorge by glacial meltwaters. The point of entry of this material was largely through the Brown's Depression/North Entrance area. The finding of sediments in the newly discovered northwestern segment of the cave (Pauley Avenue) also argues for input via Hunter's Fissure and Diddly Cave. Thick remnant sediments reveal that at least the Ward's Section of the cave was once sediment filled. Because the physical opening of these sediment input points is believed to have been formed by subglacial meltwaters, a sedimentation, passage infilling, and re-excavation history may be constructed.

The thickness of the sedimentary column in contact with bedrock suggests that significant cave enlargement had occurred prior to sediment infilling, possibly during Illinoian and/or Kansan times. The basal deposits on bedrock include imbricate shale-clast-rich sediments with small cobbles, indicative of rapid infilling. Once much of the cave became filled, floodwaters stagnated, leaving their signature in finely laminated sand, silt, and clay layers. These layers are seen near the ceiling in the Ward's Section and are interlayered with courser sediments in the Gregory Section (to at least the 714-foot level). Sediment deposition may have occurred during pre-Woodfordian time, with re-excavation during Woodfordian or post-glacial times. Lack of significant sediment cover and fill in paleo-channels and abandoned insurgence points supports this hypothesis.

Partial plugging by sediment of Clarksville Cave's overflow outlets may have contributed to the degree of backflooding and upward passage development in the cave. Evidence is found for this in large glacial cobbles cemented in a clay matrix now terminating The Hidden Room. It is possible that sediments washing down the Clarksville Paleogorge partially or totally blocked the Gregory and Osborn Entrance overflow outlets for a period of time prior to being washed free again. Alternately, these outlets may have formed as a floodwater modification behind the inefficient Clarksville Cave low-flow outlet.

GLACIAL GEOLOGY

Two and possibly three glaciations are documented as far south as Corinth, New York (LaFleur, 1991, unpublished report). Approximately 14,700 yrs BP, the Wisconsinan ice sheet receded from the Helderberg Plateau (DeSimone and LaFleur, 1985). Dineen (1986) gives an extrapolated bog-bottom date of $15,060 \pm 1,000$ yrs BP for the Great Bear Swamp situated somewhat west of Clarksville. This date further confirms the timing of the deglaciation of the Clarksville area. DeSimone and LaFleur (1985) provide a date of approximately 14,700 yrs BP for the recession of the Pine Swamp ice front from the Clarksville area. They depict the ice front as a lobe or tongue projecting southward to Stuyvesant, New York, with Clarksville situated along the southwestern flank of the ice margin.

Dineen (1986) documents ice thinning during glacial stagnation over the Helderberg Escarpment. Large quantities of meltwater flowed southward proximal to the southwestern flank of the Hudson Champlain Lobe of the Schoharie ice margin. Dineen describes deposition of sediments in multiple meltwater tunnels under stagnant ice. It thus appears that deglaciation from the Stove Pipe Road area was characterized by a southward thinning ice cover, with a southward meltwater flow direction. Free-surface flow was probably present in the paleogorge prior to the final retreat of Woodfordian ice from the Clarksville area.

IMPLICATIONS OF WISCONSINAN CLIMATES

Of even greater importance than the physical presence of the Stove Pipe Paleogorge is its relationship to Hunter's Fissure and Diddy Cave and the implication this has on interpretation of Wisconsinan, and possibly Illinoian and Kansan climate. Hunter's Fissure and Diddy Cave formed along the abandoned Stove Pipe Paleogorge. The presence of small scallop wavelengths in joint-controlled Diddy Cave indicates rapid streamflow along the base of one or more Wisconsinan ice sheets. Rounded stream cobbles in walking-sized passages in Diddy Cave provide clear evidence that a large stream once flowed through the paleogorge. The recent finding of bones of a varying hare and the extinct passenger pigeon within clay deposits in the cave provide important scientific information on these species' recolonization following deglaciation (e.g., Layer 1: $4,350 \pm 60$ yrs BP; Layer 2: $9,040 \pm 70$ to $10,470 \pm 60$ yrs BP; Steadman, pers. comm.).

The Clarksville Paleogorge has probably not had a stream in it for the last 14,700 years, coincident with retreat of Woodfordian ice in the mid-Hudson Valley (DeSimone and LaFleur, 1985). Furthermore, Wisconsinan ice had retreated from the lower Hudson Valley 15,000 or 16,000 years ago (Connally and Sirkin, 1986; Dineen, 1986), with the ice front retreating to the St. Lawrence Valley by 13,000 years ago. Therefore, the maximum time frame for possible ice front stagnation in the Clarksville area during active deglaciation is on the order of 1,000 years.

Solutional cave formation will occur only where a pre-existing network of integrated openings connects the recharge and discharge areas (Palmer, 1991). This is a process that requires a minimum of 10,000 years (Palmer, 1984; Dreybrodt, 1987, 1990; Palmer, 1991) before passageways obtain sufficient size for human entry. Additional passage cross-sectional size requires additional time. A shorter time

period, on the order of 5,000 years, may be possible depending on joint widths present in the bedrock prior to infiltration by glacial meltwaters. Thus, subglacial meltwaters apparently were not only responsible for the formation of Diddly Cave, but must have flowed for a minimum of 5,000 to 10,000 years in order for the cave to form. Because the maximum amount of time the retreating ice front could possibly have stayed in the Clarksville area was on the order of 1,000 years, it follows that Diddly Cave, Hunter's Fissure Cave, the Brown's Depression area and other southern infiltration points were receiving meltwater from below warm-based glacial ice for at least 5,000 to 10,000 years. It is likely that temperate climatic conditions were present during the early and later Wisconsinan.

MELTWERter FEATURES IN OTHER NEW YORK STATE CAVES

Many New York State caves need to be re-examined for evidence of glacial meltwater modification. Several caves in east-central New York exhibit features characteristic of meltwater invasion. Examples include Skull, Knox, Ella Armstrong, McFail's, Howe Caverns, Single X, Schoharie, Gage, Onesquethaw, and Surprise (Mystery) caves, all of which have one or more passage segments superposed above passages receiving Holocene peak floodwaters. For example, the upper levels of Skull Cave have aragonite speleothems that are incapable of surviving floodwater invasion. Similarly, caves such as Skull, Knox, and Ella Armstrong have watershed sizes too small to account for the volume of water necessary to form their observed vertical and areal extent. Other caves, such as McFail's and Surprise, carry underfit streams, yet exhibit anomalously large passage sizes. An artificially enlarged subglacial watershed would have been capable of providing the necessary recharge. Related fossil karst includes abandoned and sometimes glacial-debris-covered sinkholes and shafts which once served as significant infiltration points, but now serve only to focus localized drainage into the epikarst, funnelling runoff and infiltrating meteoric waters to deeper conduit flow routes. Anomalous in-cave features such as abandoned pits and multiple level, ungraded passages may also reflect meltwater invasion (e.g., Surprise Cave). Similarly, significant cave development proximal to the headwaters of a drainage basin (e.g., Gage Caverns and Phoebe Pit) may also reflect meltwater invasion from an expanded ice-sheet watershed. Other relict caves, swallow holes and solution conduits such as Knox, Salamander, several Saugerties-area caves, and Joralemon's (this volume) are now abandoned and largely waterfree. Their derangement from active drainage patterns may portray development during a previous interglacial period or, more likely, may be a result of modification by glacial meltwaters.

The characterization of cave modification by glacial meltwater invasion poses many exciting geomorphic questions for researchers in New York State. Speleothem- and sediment-dating techniques may shed light on karstic evolution and modification through multiple glacial periods. A complete geomorphic interpretation must include an assessment of flow conditions and geologic features in a defined watershed, both on the surface and in the subsurface.

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SECTION B: **Flow Characteristics and Scallop-Forming Hydraulics within the Mill Pond Karst Basin, East-Central New York**

ABSTRACT

This is a study of the hydrology of Clarksville Cave and the headwaters of the Onesquethaw Creek, situated in the hamlet of Clarksville, New York, specifically the Mill Pond karst basin. During most of the hydrologic year, water entering that part of the watershed that is downstream of the Wolf Hill Dam is pirated into the Onondaga Limestone. Tracer tests and in-cave stream gaging indicate that extreme conduit conditions are present in the aquifer, with a maximum water velocity on the order of 5.3 km/hr.

It has been hypothesized that a submerged conduit must lie covered by breakdown blocks at the cave's northern terminus. Having established a known peak flow, a modified version of the Darcy-Weisbach equation was used to accurately calculate the minimum diameter of this conduit. Knowledge of the structural geology throughout the watershed, coupled with a detailed leveling survey in the cave, permitted reasonable estimates to be made for the two unknowns in the equation. A submerged conduit was subsequently opened and explored.

Scalloped cave walls are present in Perry Avenue (Fig. 3) at a key stream gaging location. Backflooding occurs behind inefficient passage constrictions a short distance downstream of, but not up to, this station. Evidence exists that documents that only long return-interval flood stages cause backflooding to this station. This situation permits a reasonable estimate of the maximum discharge and flow velocity responsible for scallop formation. Scallop wavelengths were measured below the elevation of peak floodwaters. By inputting measured values for discharge and flow velocity into published equations, it was possible to back calculate scallop Reynold's numbers that favorably correlate with measured flow velocities and discharges. A possible revision of the scallop Reynold's number is suggested when it is utilized in the determination of paleoflow velocities. It also appears that scallop wavelength is partially determined by the properties of the rock comprising the walls of the conduit.

LOCATION AND WATERSHED BOUNDARIES

A broad carbonate aquifer is present in the Clarksville area. Its boundaries extend north and northwest of the Mill Pond, situated less than 120 meters south of the restaurant, June's Place (*see* Fig. 1: Section A). The farthest boundary of the Mill Pond karst basin lies about 3.9 kilometers to the northwest, proximal to the Wolf Hill Dam on the Onesquethaw Creek. The elevation of the basin ranges from 1822 feet msl atop the Helderberg escarpment to approximately 645 feet msl at the Mill Pond. The boundaries of the catchment basin are depicted in bold dashed lines. These boundaries were defined through the use of low-altitude stereo aerial-photography, U.S.G.S. topographic maps, tracer studies and, in places, detailed structural geologic mapping.

The Mill Pond watershed may be subdivided into two parts: A) that part of the watershed located upstream of the Wolf Hill Dam (1,245 hectares), and B) that part of the watershed located downstream of the Wolf Hill Dam (829 hectares). The downstream part of the Mill Pond watershed exhibits features characteristic of karst terranes. These include sinking streams, limited surface drainage, solutionally enlarged joints, sinkholes, and the Clarksville/Diddly cave system. Structural deformation throughout the region has resulted in extensive jointing and faulting, providing solutional pathways for infiltrating waters.

PIRACY OF ONESQUETHAW CREEK WATERS

Most of the Onesquethaw Creek downstream from the Wolf Hill Dam and upstream of the Mill Pond is a losing stream, with a substantial amount of surface flow lost to solutionally enlarged joints in the streambed. During most of the hydrologic year little or no surface flow occurs in the area downstream of the Wolf Hill Dam, located nearly on the Marcellus Shale/Onondaga Limestone contact. Subsurface piracy of water into the Onondaga Limestone below the Wolf Hill Dam occurs via numerous joints in the streambed. The water briefly surfaces at the Salisbury Spring, only to again sink into joints in the stream bed. The volume of water flowing in the streambed and the relative efficiency of the often partially sediment-choked joints governs the distance water may be found flowing on the surface downstream from the Wolf Hill Dam and the Salisbury Spring. The greater the discharge of the stream, the farther its flow is capable of traveling prior to complete subsurface piracy. During periods of low or moderate discharge, all Onesquethaw Creek surface flow is pirated into the karst network prior to where the bed of the Onesquethaw Creek passes beneath Rt. 443 (*see* Fig. 2: Section A). Only large storm and

CLARKSVILLE CAVE
(Ward-Gregory Cave)
Osborn to the Lake Room
Albany Co., NY

J. Brown, T. Engel, W.
Froto, A. Fleischmann, E.
Kastning, G. Saunders, J.
Schwonen, R. Smith, & many
others.

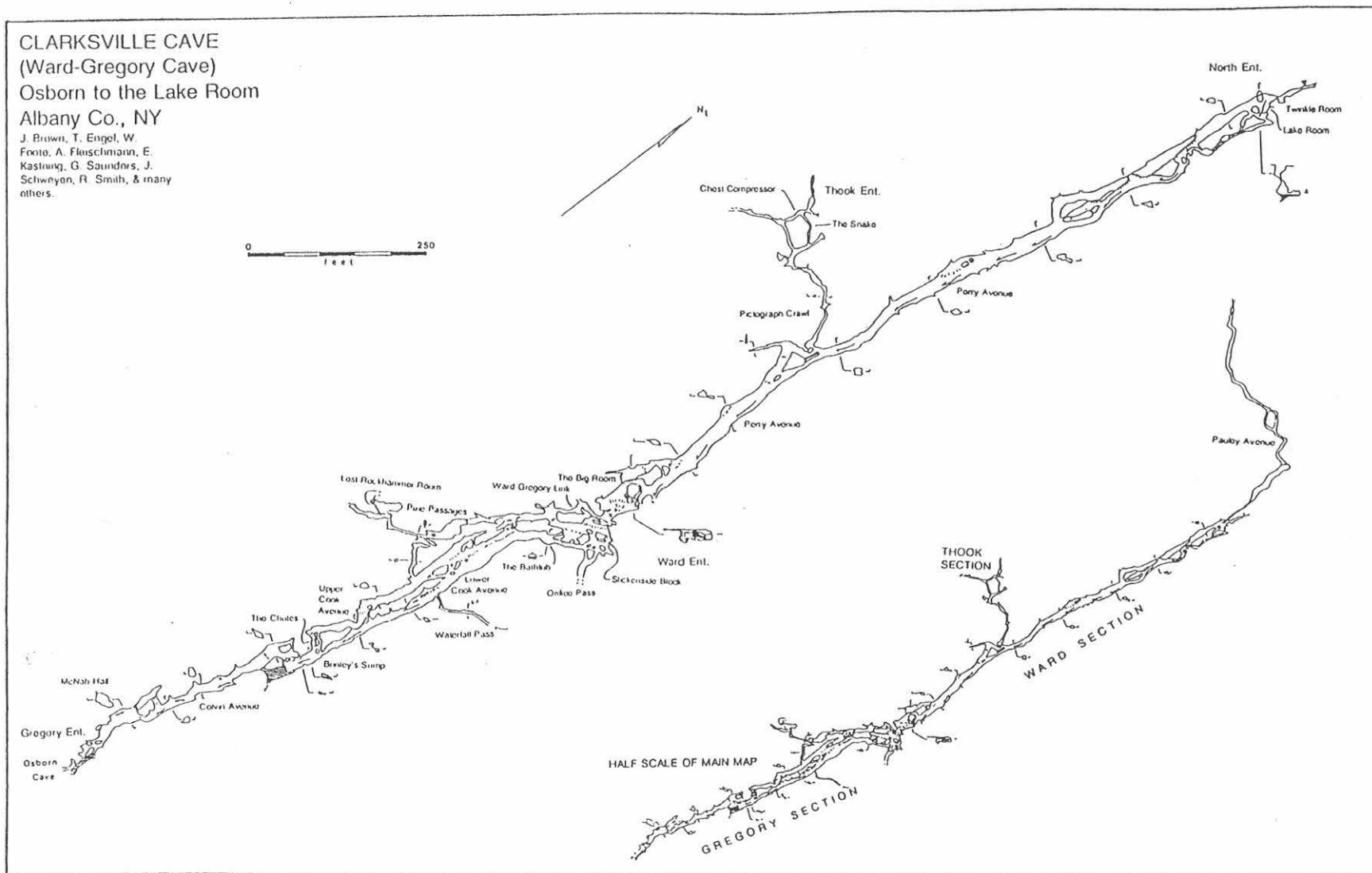


Figure 3 Map of Clarksville Cave, Albany County, New York. Notes: 1) This map is based on map of Kastning (1975). New passages added are Pauley Avenue, the Pinch Passage, the Thook Section, the Orifice Passage, and part of the Bathtub Feeder. 2) Stream flow is perennial, but varies from a low of 2 gpm to in excess of 60,000 gpm. 3) The Gregory Entrance, Brinley's Sump, and the stream access immediately east of McNab

Hall are subject to flooding. During periods of extreme flood, the Bathtub and the downstream end of Perry Avenue also sump shut. 4) From the Lake Room south, the entire cave is Grade 5 except the Pinch Passage (grade unknown) and the Orifice Passage (Grade 2). Upstream of the Lake Room the underwater section is Grade 3; the rest as far as the Loop is Grade 5. Beyond the loop it is Grade 1.

snowmelt events generate enough surface flow in the watershed to cause the Onesquethaw Creek to flow throughout its course. This represents a very small part of the hydrologic year. Surface-stream flow is short-lived even after major storm events.

TRACER TESTS

A series of uranine-tracer tests have permitted partial delineation under varying conditions of discharge of the subsurface flow paths throughout the Mill Pond drainage basin. Uranine is a non-toxic tracer frequently used in karst investigations (Smart, 1984). It was injected into various joints in the Onesquethaw Creek streambed that were pirating water. Activated-carbon detection bugs were placed at all likely resurgence points, collected later, and chemically elutriated with Smart solution (Quinlan, 1986).

Tracer testing has revealed that dye injections from 3.2 km upstream of the Mill Pond resurgence remain perched above the Upper Cherty Subunit for at least 1.6 km before breaching chert beds that overlie the lower, more massive, non-cherty subunits. Water pirated into the Onesquethaw Creek streambed immediately downstream of the Wolf Hill Dam and upstream of Rt. 85 remains in one or more subsurface conduits, until surfacing briefly at the Salisbury Spring, only to again sink into joints in the stream bed downstream. The Salisbury Spring is located on the western side of Rt. 443, approximately 0.6 km southeast of Rt. 85. It is set back some distance from the road. It is likely that piracy of the Salisbury Spring discharge into the bed of the Onesquethaw Creek is roughly coincident with the point at which this water breaches the Upper Cherty Subunit of the Onondaga Limestone. Thus, one major tributary conduit to the system is likely to become physically impassable within 1.4 km northwest of the Lake Room in Clarksville Cave (Fig. 2). However, stream gaging and tracer studies indicate the presence of a second low-flow conduit entering the known parts of Clarksville Cave from the large, heavily jointed watershed to the north-northwest.

All subsurface flow resurges at the Mill Pond. The relative inefficiency of the outlet of the Mill Pond conduit may be due to structural problems resulting from the upward thrusting of the impermeable Esopus Shale against the cave-bearing Onondaga Limestone (*see* Section A). The presence of impermeable Esopus Shale in the bed of the Onesquethaw Creek at and immediately downstream of the Mill Pond forces all subsurface flow from the karst aquifer to surface at the Mill Pond. This author established a gaging station downstream of this point (*see* Fig. 1: Section A).

Tracer tests and discharge measurements throughout the watershed indicate that during periods of low discharge, pirated Onesquethaw Creek waters do not travel through Clarksville Cave. Surface and subsurface stream gaging and tracer tests establish the intersection of the pirated Onesquethaw Creek low-flow conduit with the Clarksville Cave low-flow drainage conduit to be located between the southern end of the cave and the Mill Pond (*see* Fig. 2: Section A). Hereafter, the conduit that resurges at the Mill Pond and is physically separate from the Clarksville Cave conduit north of Osborn Cave, is referred to as the Mill Pond low-flow conduit. Although the exact elevation of the lowest drain point in Clarksville Cave remains to be surveyed, it lies slightly below an elevation of 660 ft msl. The hypothesized flow routes of unentered parts of the network are portrayed in Figure 2 of Section A.

Tracer tests verify that after a certain critical subsurface discharge is reached, coincident with piracy of increasingly greater amounts of surface flow into the subsurface conduit system, the efficiency of the Mill Pond low-flow conduit is exceeded and surplus water is shunted to the Lake Room in Clarksville Cave. The Mill Pond low-flow conduit utilized today, which bypasses Clarksville Cave, may be the original flow route, with the flow route leading to the Lake Room (via Pauley Avenue) forming as a floodwater-overflow route. Alternatively, the flowpath to the Lake Room may represent the original subsurface flow route that was later abandoned due to further stream piracy, possibly coincident with

lowering of the regional base level. Under this genetic interpretation, diversion of waters from the Mill Pond low-flow conduit to Pauley Avenue would occur behind an immature drain. During periods of base flow, it appears that only water from the north-northwestern part of the Mill Pond watershed rises in the Lake Room. Moderate and high discharge in the subsurface causes a significant backup of water behind the Mill Pond low-flow conduit, resulting in large overflows to the Lake Room. The greater the flow in the Onesquethaw Creek, the more water is lost through joints in the streambed, and the greater is the discharge that appears in the Lake Room.

From the Lake Room the water flows south through the cave where some of it joins, in a tributary manner, the Mill Pond low-flow conduit somewhere between the downstream end of the cave and the Mill Pond (*see* Fig. 2: Section A). As the discharge of floodwaters within Clarksville Cave increases, the hydraulic efficiency of the branched conduit leading to the Mill Pond is exceeded. The remaining water that cannot be handled by Clarksville Cave's low-flow subsurface conduit and the Mill Pond low-flow conduit backs up within the cave as temporary storage. After a critical flow on the order of 2.7 cfs is reached, excess floodwaters are discharged along the Osborn Cave overflow route (677 ft msl) to the surface. Osborn Cave is situated directly south of the Gregory entrance and is physically connected to Clarksville Cave by a water filled conduit. Figure 2 of Section A shows the Osborn overflow channel, that sometimes carries large quantities of water.

KARST BASIN CHARACTERIZATION

Tracer tests conducted in parts of the Mill Pond aquifer reveal that all subsurface waters reappear or resurge at the Mill Pond. During periods of low flow, all surface and groundwater downstream of the Salisbury Spring and upstream of the bridge crossing Rt. 443 discharge through conduits in the carbonate aquifer at the Mill Pond. During periods of moderate to high-subsurface discharge, part of the subsurface flow is shunted through Clarksville Cave. All flow throughout the Mill Pond watershed thus surfaces either in the Mill Pond or Osborn Cave, where, for much of the year, it comprises the headwaters of continuous surface flow of the Onesquethaw Creek. At times this flow is supplemented by water from the Clarksville South Road and western Bennett Hill Road sub-watersheds.

Water in the Onesquethaw Creek, from that part of the watershed upstream of Wolf Hill Dam that is not artificially diverted to the Vly Creek Reservoir, also sinks into the subsurface downstream of the Wolf Hill Dam. Much of the flow in the karst network originates as diffuse infiltration outside the Onesquethaw Creek Corridor. Virtually all meteoric water and snowmelt contacting the heavily jointed, generally thin-soil-mantled limestone pavement within the Mill Pond watershed is pirated into subterranean limestone conduits. Geologically, water entering the soluble Onondaga Limestone must stay within it because it is underlain by approximately 1 m of the Schoharie Formation (a quartzitic limestone) and approximately 30 m of impermeable Esopus Shale.

Physically unentered segments of the conduit network may be envisioned as being similar to a tree, where all branches coalesce downstream toward the trunk. Palmer (1991) describes such branchwork caves as the most common type. Water infiltrating from different segments of the aquifer's recharge area converges as higher-order passages that decrease in number and generally increase in size in the downstream direction. It is likely that the large northwestern part of the Mill Pond aquifer is branchwork in nature, with many tributaries coalescing downstream toward larger, master passages. It is also likely that segments of the conduit system directly underlie the bed of Onesquethaw Creek, whereas others extend far to the northwest. Still other segments must enter from the northwest where runoff from the Marcellus and Hamilton beds of Wolf and Cass hills sinks near the Onondaga Limestone contact and is rapidly pirated into the system. In the early 1990s, exploration via the newly opened Diddly Cave entrance yielded approximately 0.5 km of large stream passages extending north into the Mill Pond karst basin. These passages are branchwork in character, and if connected to Clarksville Cave would bring the

cave's length to greater than 2 km. The dashed lines on Figure 2 of Section A portray a simplified version of the hypothesized configuration of conduits in the eastern end of the system.

SUBSURFACE TRAVEL TIMES

The combined flow from stream losses and diffuse fracture infiltration is documented as moving very rapidly through the karst system. Although effort has not been made to absolutely quantify the rate of subsurface flow in the aquifer, the timing of two tracer tests provides some insight on the situation. Under moderate flow conditions present on February 23, 1990, uranine tracer was injected into a joint in the bed of the Onesquethaw Creek, 3.2 km northwest of the Mill Pond. At this time, all surface flow in the upper reaches of the Onesquethaw Creek was being pirated into this joint. Tracer-detection bugs were collected from Clarksville Cave at 4:00 p.m. on February 24, 1990, about 22 hours after the tracer injection. All were positive for uranine. Thus, a subsurface groundwater transit time in excess of 150 meters per hour was documented.

A similar trace was conducted in October 1988 under low-flow conditions. In this instance, the tracer injection and sinking of the stream occurred farther northwest than during the above trace. In this second test the tracer-detection bug was removed from a location proximal to the Mill Pond 27 hours after tracer injection. After elutriation, the detection bug was positive for uranine. A subsurface groundwater transit time in excess of 120 meters per hour was documented for low-flow conditions.

In contrast, during a time of peak flow within the aquifer (March 15, 1986 at 1:45 a.m.), the discharge and velocity of flow within Clarksville Cave were measured. The velocity was recorded as 1.48 meters per second. This equates to 5,328 meters per hour (5.3 kms/hr) and may be considered as indicative of the peak velocity of potential groundwater movement within the aquifer and of extreme conduit conditions. At times of peak flow, groundwater may move from end to end through the karst aquifer, a distance of approximately 3.9 km, in less than one hour.

The rapid hydraulic response to significant precipitation or snowmelt within the watershed has been repeatedly documented with stream hydrographs both in Clarksville Cave and in the Onesquethaw Creek. Subsurface conduit flow in the Mill Pond aquifer is roughly analogous to open channel flow in a surface stream. A thin soil-moisture bank over much of the watershed's limestone pavement further permits rapid infiltration of meteoric waters and snowmelt, thus bolstering subsurface transit times. Flood pulses throughout the karstified system are flashy, providing evidence of mature conduit development. Rapid flow characteristics present within the 2,074-hectare Mill Pond watershed, especially that part downstream of the Wolf Hill Dam, make it and the Onesquethaw Creek extremely sensitive to infiltration of contaminants.

During much of the hydrologic year, discharge from the Mill Pond acts as the sole source of water to the upper reaches of the Onesquethaw Creek. During periods of base flow this discharge has been gaged at less than 0.1 cfs. The recent zoning of land central to the carbonate aquifer as rural commercial may have severe effects on both the aquifer and Onesquethaw Creek if untreated waste streams or septic infiltration are permitted (Rubin, 1990b, 1992).

IN-CAVE AND ONESQUETHAW CREEK FLOW CALCULATIONS

Measurements of discharge and streamflow velocity have been made periodically in Clarksville Cave since 1983. Over 99% of the water flowing through Clarksville Cave rises in the Lake Room. This water has been gaged during both low and high flow at discharges ranging between 0.002 and 111 cfs.

A maximum water depth of 63.5 cm was measured during the storm of March 15, 1986. Discussions with Ed Gregory revealed that the flood discharge component in the cave, associated with the 1938 failure of the Helderberg Lake Dam, was significantly greater than the above maximum-gaged amount. Gregory reported that floodwaters were ponded to an elevation of approximately 719 feet msl, a short distance down the entrance slope inside the Ward's Entrance. The elevation of the cave passage in upstream Perry Avenue, approximately 11 meters south of the Lake Room, lies between 715 and 708 feet msl, thus indicating that all of Perry Avenue was flooded during this event. Confirmation of this flood level, and possibly another in 1903, is manifested in a thick mud film covering historic names and dates chiseled near the passage ceiling.

A gaging station was established in the Onesquethaw Creek (*see* Fig. 1: Section A) in order to examine the relationship between in-cave discharge and surface-watershed discharge. This was monitored twice daily for 15 months, more frequently during flood events, and periodically for 4 years thereafter during major runoff events. Stream discharge was gaged at 13 different stages. Curvilinear regression was then utilized to establish a series of multi-order equations that could be used to correlate stage height with discharge. The greatest discharge recorded for the Onesquethaw Creek during the course of this study was approximately 1337 cfs. This occurred on March 15, 1986 at 3:00 am following heavy rains (≈ 7.0 cm) on a 38-centimeter snow pack. Temperatures up to 40° F accompanied the coastal storm of March 13-14, 1986. Daily monitoring of stream stage in Clarksville Cave for the same 15-month period revealed that a direct correlation exists between this discharge and that in the Onesquethaw Creek. Approximately 8 percent of flood-peak discharge in the Onesquethaw Creek flows through Clarksville Cave.

Knowledge of expected flood-return intervals and their magnitude in the Mill Pond karst basin was found to be essential to both the interpretation of how abandoned upper-level passages in Clarksville Cave formed and an understanding of the dynamics controlling scallop formation. The limited data for statistical comparison among hydrologic years in the Mill Pond karst basin necessitated examination of another roughly comparable basin in order to assess flood-return intervals. The farthest headwater gaging station on Schoharie Creek at Prattsville was selected. Many inherent differences occur between the basins, notably elevation, geology, regolith thickness, size, and location. The Prattsville and Onesquethaw Creek gaging stations are approximately 48 kilometers apart. However, the Prattsville and Mill Pond watersheds are comparable under conditions of a saturated soil-moisture bank, high runoff, and similar storm systems. Eighty-two years of data were examined at the Prattsville, New York station.

A Log-Pearson-Type-III and Gumbel-distribution statistical comparison of historic peak flow of Schoharie Creek gaging data with this study's hydrograph information for the Onesquethaw Creek indicates that the largest Onesquethaw Creek peak of record (March 15, 1986) has a return interval on the order of 30 to 47 years (Fig. 4). This corresponds to a Prattsville hydrologic-year peak discharge of 54,900 cfs. Thus, if 40 years was the expected flood return interval, 25 floods of this magnitude could be expected every 1000 years. Reconstruction of the 1903 peak discharge at Prattsville (approximately 63,000 cfs), the highest on record, further reveals that a cave discharge well in excess of 111 cfs may also have occurred in 1903. The 1903 discharge, three standard deviations greater than the mean-annual Prattsville peak flow, has a predicted flood return interval on the order of 47 to 90 years. Although the magnitude of this flood was larger than the 1986 flood, it probably was not as great as during the dam-failure flood in 1938. These infrequent storm or runoff events reasonably represent a near-maximum quantity of water available in the watershed under ideal, thin-soil-mantled, rapid-infiltration conditions. Therefore, it is difficult to explain the "intermittent phreatic" upper-level passages in Clarksville Cave without a substantially greater quantity of water. A subglacially enlarged watershed, as discussed in Section A, appears to be the only viable explanation.

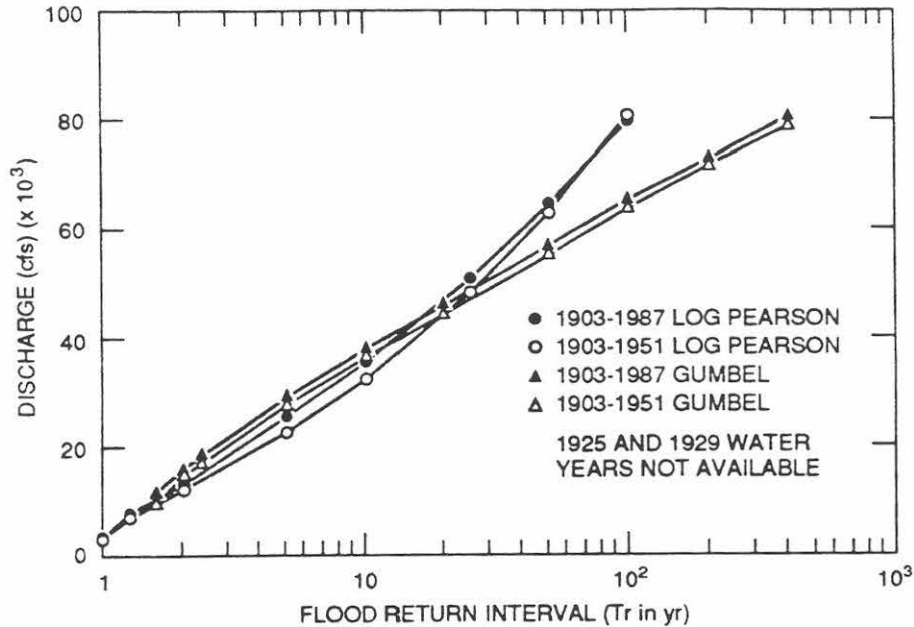


Figure 4: Schoharie Creek at Prattsville, New York. Log-Pearson type III and Gumbel distributions utilizing historic water year peak flow data. The flood return interval for the largest Onesquethaw Creek peak discharge of record correlated to the Prattsville March 15, 1986 flood of 54,000 cfs. The range of two methods shows a Tr of 30 to 47 years. Similar comparisons may be made for low flows in geologically comparable basins. Analyses of this type can be useful tools in predicting flow velocities, contaminant arrival times, and contaminant dilutions. Knowledge of peak or base flow return periods can sometimes be correlated with water chemistry results to help assess chemical loading both in the karst system and to stream receptors.

LAKE ROOM SUBMERGED CONDUIT

Measured, statistically predicted, and inferred (e.g. mud-covered historic names and dates) high discharges rising from the Lake Room indicated that an obscured conduit was present that must be capable of transmitting large discharges. It was thus hypothesized (Rubin, 1989) that a submerged conduit must lie covered by breakdown blocks, below the water surface in the Lake Room. By making a number of reasonable assumptions, it was possible to calculate the minimum diameter of an assumed circular conduit capable of discharging a given flow. A modified version of the Darcy-Weisbach equation

$$r = \left(\frac{Q}{\pi \sqrt{\frac{4g}{f}} \sqrt{\frac{\Delta h}{L}}} \right)^{2/5}$$

was successfully utilized to examine the size of the, until recently, unentered upstream segments of cave conduit, north of known parts of the cave. Calculations were confined to a circular conduit capable of discharging between 111 and 222 cfs (Q). The latter value was considered a reasonable approximation for the 1938 dam-failure discharge. A friction factor (f) of 0.1 was used. The two unknowns in the equation were the change in head (elevation of water upstream of the lake versus the elevation of the lake, Δh) and the length of flooded passage upstream of the lake (L) during flood events. A wide range of values of 1.5 to 30 meters, and 6 to 1219 meters were tested, respectively, for these unknowns. Although some of the values tested were likely to be extreme in nature, they were selected based on knowledge of the structural geology within the watershed, coupled with a detailed leveling survey throughout the cave. It was believed possible that significant backflooding might be occurring behind the Lake Room breakdown.

Insertion in the modified Darcy-Weisbach equation of a reasonable range of values for the change in head and the length of flooded passage suggested that the minimum diameter of a circular-conduit tributary to the Lake Room is between 0.6 and 2.4 meters. Recent excavation and penetration of a formerly blocked and water-filled conduit extending north and west of the Lake Room verified the calculations (Rubin, 1990a). The length of the water-filled passage was found to be 61 meters. The actual Δh value is probably no more than 3 meters. The smallest diameter found in these newly discovered passages was 1.4 meters. Maximum cross-sectional area found in the approximately 366 meters of conduit beyond the Lake Room that have been entered thus far is on the order of 8 square meters. These passage dimensions attest to mature conduit development in the carbonate aquifer within the catchment basin.

The assumed friction factor of 0.1 was found to accurately reflect the flow conditions through the Lake Room breakdown. Approximately two vertical meters of clean washed angular breakdown, interspersed with minor quantities of rounded glacial cobbles, were excavated. The heterogeneous mixture of breakdown blocks ranged in size from several centimeters in length, width and height to approximately one meter. The water's approach angle, toward the lake surface, rises at approximately 30 degrees for the last 3 meters before reaching an irregular constriction (1.2 meters by 0.5 meters) in the breakdown. Prior to the last 3 meters, the submerged conduit is generally horizontal. The maximum conduit depth below the surface of the lake was found to be approximately 4.3 meters.

SCALLOP-FORMING HYDRAULICS

Phases of this study focused on defining the Mill Pond karst basin, the relationship between flow in the carbonate aquifer versus that in Clarksville Cave, the expected return interval of peak discharge inside and outside Clarksville Cave, and flow conditions peculiar to Clarksville Cave. Specifically, a range of stream discharges and velocities were measured in an air-filled segment of linear passage and rectangular cross section. Water depth was recorded, as well as scallop wavelengths within the zone of the 30- to 47-year flood-return interval.

Paleoflow information that researchers hope to reconstruct, based on scallop wavelengths and dimensions of an abandoned passage, is either empirically measured or reasonably constrained. Measurements in Clarksville Cave permitted a cave-specific evaluation of Blumberg and Curl's (1974) scallop Reynold's number.

One potential problem with characterization of the physical conditions under which scallops form is defining the discharge, or range thereof, responsible for scallop development. It was possible to define minimum and maximum discharge limits leading to scallop formation at a key stream-gaging location in Perry Avenue. Here, cave walls in a fossiliferous sparite are scalloped. Streamflow across the width of the cobble floor does not become deep enough, or of sufficient discharge to form scallops until the water is approximately 18 centimeters deep. The 30 to 47-year flood (111 cfs) of March 15, 1986 resulted in

a stream depth of 63 centimeters, but decayed in 34 hours to 10 cfs with a stream depth of less than 13 centimeters.

Backflooding occurs behind inefficient passage constrictions at the Big Room, approximately 120 meters downstream of this same key stream-gaging location. The level of backflooded waters, as measured on March 15, 1986 at the stream's surface, was only 48 centimeters lower in elevation than ponded water at the Perry Avenue gaging station. A small additional discharge amount, such as that probable in 1903, or certainly in the 1938 flood, would have substantially reduced the stream velocity here and its ability to form scallops. Thus, the 111 cfs measured on March 15, 1986 represents a value that is close to the maximum possible for discharge capable of forming scallops at the gaging station. This situation allows for a reasonable estimate of the maximum discharge and flow velocity responsible for scallop formation.

Statistical analysis of Clarksville Cave flood-return intervals indicates that cave discharges in excess of two standard deviations about the mean-annual peak discharge may be sufficiently short-lived and of infrequent recurrence to form the observed scallops. The frequency of shorter term flood intervals is greater, and perhaps it is these events which are recorded as scallops rather than very short duration, high-discharge long-return-interval floods. Although the shorter-return-interval floods are also relatively short-lived, it may be the combined contact time of water with bedrock of many similar magnitude floods that is of importance. The relationship among stream depth, discharge, and flood-return interval in the watershed, as partially indicated in Table 1 below, suggests that scallop formation in Clarksville Cave may occur during flood intervals that range between one and two standard deviations of the mean-annual peak discharge.

<u>Channel Width (m)</u>	<u>Stream Depth (cm)</u>	<u>Discharge (cfs)</u>	<u>Velocity (cm/sec)</u>
a) 3.3	18.3	14	68.6
b) 3.3	27.2	30	96.5
c) 3.3	63.4	111	147.8

Table 1: Three flow conditions in Perry Avenue, Clarksville Cave.

Scallop wavelengths were measured below the elevation of peak floodwaters. By inputting measured discharge and flow velocity numbers into published equations, it was possible to back-calculate scallop Reynold's numbers that favorably correlated with measured flow velocities and discharges. This procedure involved measuring scallops, streamflow, and stream velocity and examining the likely range of scallop-forming conditions utilizing published equations. For the rectangular Perry Avenue conduit:

The Sauter mean was used to calculate mean scallop wavelengths of scallop groups within 63.5 cm of the cave floor:

$$L_{32} = \frac{\sum li^3}{\sum li^2}$$

(Curl, 1974)

A range of in-cave flow conditions was examined. The three flow conditions presented in Table 1 bracket the minimum and maximum stream discharge and velocity believed to be responsible for scallop formation at the key Perry Avenue gaging station.

"The scallop Reynold's number, N_R^* based on friction velocity, is a universal constant for scallop formation and was determined from model experiments (Blumberg and Curl, 1974) to have the numerical value $N_R^* = 2200$ " (White, 1988).

Scallop formation is controlled, in part, by a dimensionless Reynolds number:

$$N_R = \frac{vL_{32}\rho}{\eta}$$

where: v = mean velocity of fluid flowing past scallop in cm/sec
 L = mean scallop length in cm
 ρ = density of fluid ≈ 1.0 gm/cm for 5°C and 10°C
 η = fluid viscosity ≈ 0.015 gm/cm/sec for 5°C
and ≈ 0.013 gm/cm/sec for 10°C

Thus, examining the specific flow conditions in a), b), and c) above (see Table 1) using $L_{32} = 7.49$ cm and $\eta = 0.015$ gm/cm/sec., a range of site specific Reynold's numbers was obtained:

- a) $N_R = 34,254$
- b) $N_R = 48,186$
- c) $N_R = 73,801$

Curl (1974) provides the limiting geometry for a rectangular cave passage:

$$N_R = N_R^* \left[\frac{2.5 \left(\ln \frac{D}{2L_{32}} - 1 \right) + B_L}{2L_{32}} \right]$$

By inserting the range of N_R 's above into Curl's equation, we can examine N_R^* , the scallop Reynold's number based on a range of actual flow velocities:

- a) $N_R^* = 2,341$
- b) $N_R^* = 3,293$
- c) $N_R^* = 4,337$

Blumberg and Curl (1974) derived a universal constant for the scallop Reynold's number, based on plaster model studies, of 2200. The wall material subject to scallop formation may influence the value of the scallop Reynold's number. Different types of surfaces, like limestone, ice, and plaster, may respond differently to water scour. Based on this study, it appears that N_R^* may actually not be a constant, but instead may best be characterized by a range of values. These values, based on this cave-specific study of a rectangular conduit, appear to be from 1 to 2 times the accepted constant. Empirical observation of the flow dynamics in Clarksville Cave, coupled with a characterization of flood-return intervals within the catchment basin, suggest that a scallop Reynold's number on the order of 3300 better approximates New York State cave-specific conditions. A scallop Reynold's number of 3300 was used for the determination of paleoflows at the Hollyhock Hollow Sanctuary and Joralemon Park (this volume).

It is possible that constants in accepted equations may lead to an underestimate of paleo or recent flow velocities and discharges. It should be noted that B_L , another constant in the Reynold's number equation (which deals with wall roughness) was accepted on face value. Further studies of the actual hydrologic conditions in which scallops form are warranted.

ACKNOWLEDGMENTS

Heartfelt thanks are extended to the many northeastern cavers who have contributed to the various activities associated with the study of the Clarksville/Diddly Cave system. Different aspects of the project have included stream gaging, tracer tests, photography, leveling, digging, diving, surveying, drafting, and wall scrubbing. Kevin Downey and Kevin Harris deserve special thanks for their many hours of appreciated cave photography. Clayton Pauley, now deceased, was a true friend with whom I spent many fine evenings leveling through Clarksville Cave's labyrinth. Special thanks go to Thom Engel, the unsung hero who is always there to help survey, stream gage, draft maps and formulate ideas. The study of the Mill Pond karst basin is northeastern caving at its finest.

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ROAD LOG

CLARKSVILLE CAVE

<u>Total Miles</u>	<u>Miles From Last Point</u>	<u>Route Description</u>
0.0	0.0	Take the exit from the New York State Thruway at Interchange 22 (Selkirk). Proceed through the tollbooth, turn right, and proceed south on Route 144.
0.4	0.4	Junction of Route 144 with Route 396 to Selkirk. Turn right and proceed west through Selkirk on Route 396.
6.4	6.0	Junction of Route 396 with Route 102.
13.2	6.8	Continue west on Route 396. Route 396 becomes Route 301. Continue, crossing the Onesquethaw Creek in the hamlet of Clarksville, until the end of Route 301. Junction of Route 301 with Route 443. Turn left and proceed west on Route 443.
13.3	0.1	<u>STOP #1</u> - Turn right into parking area for June's Place restaurant. Proceed to upper parking area behind June's. Clarksville Cave is a popular caving location in the Albany area. Always cave with experienced cavers, dress warmly, wear a hardhat with a chin strap, and carry three sources of light (one preferably mounted on your helmet to free your hands for climbing). Do not attempt to pass through Brinley's Sump when there is little or no air. Under these flow conditions, the Gregory Entrance is sumped shut and a through trip is not possible. Failure to follow standard safety procedures has resulted in numerous rescues, mobilization of multiple rescue teams, ambulances, and the press.

BUILDING STONES OF SCHENECTADY, NEW YORK

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Introduction

The building stones of downtown Schenectady contain some of the best examples of rock types and medium- and small-scale geologic features that the general public can readily view. Within easy walking or driving distance are high-quality rocks of many types, some of them carefully polished and beautifully displayed. There are buildings made with igneous, sedimentary, and metamorphic rocks that include oxide-rich gabbro, porphyritic granite, crossbedded sandstone, fossiliferous limestone, marble, gneiss, and partially melted migmatite. During this half-day trip, we will tour the downtown area to view the most interesting building stones, to see what minerals, fossils and other features they contain, to infer the geologic processes that formed them.

The aim of this trip is to help people to learn more about geology, just by spending a few hours in downtown Schenectady (Figure 1). Anyone who is interested in geology can observe rock types and features on his own, without a guide, and without traveling miles on rural roads, bushwacking through woods and swamps, or stopping next to busy highways. Those who examine urban building stones can sharpen their powers of observation and learn to recognize a wide variety of rock types, minerals, features, and fossils. **BRING HAND LENSES!**

Each block of building stone, or each panel of facing stone, is one small window into the past. However, there are some important things to remember when looking at building stones. First, remember that all of the rocks you are looking at came from somewhere else. The geologic history that they record within them happened in another place. Second, many buildings contain rocks of several types, so as to juxtapose geologic events from different places and different times. No matter how closely adjacent stones may seem to match one another, we can not, in a strict sense, prove that they were ever connected in their original positions. However, it is a reasonable assumption that similar stones from the same building came from the same formation or body of rock, and possibly came from the same quarry. While the overall arrangement of the stones in a building has no geological meaning, we can gain knowledge about the original rock as the sum of observations made on individual stones.

Building Stones as Geological Materials

Of the numerous varieties of rocks that exist, only a small number are prized as building material. Depending on use, building stone is selected for some combination of strength, ease of working, proximity to the building site, beauty, weather resistance, and availability. These properties depend on many factors, including mineralogy (Table 1), grain size, textures, and overall structure.

Clay-rich sedimentary rocks such as shale and siltstone tend to break up quickly when exposed to the weather, and so tend to make poor building materials. The reason is that clays swell when wet, and shrink when they dry. The constant flexing of the rock causes cracks to form, which speeds up disintegration. Although sandstones (Figure 2) are made of grains of quartz and other weather-resistant minerals, the grains are held together by some kind of intergranular "cement". Some sandstones have clay cement, which causes the sandstones to decompose relatively rapidly like shale and siltstone. Calcite and dolomite are also common cements and are resistant to the effects of wetting and drying, but they are relatively water soluble. As a result, calcite-cemented sandstones tend to weather away grain by grain as the cement dissolves over the years. Quartz is the most weather-resistant of common cements. Quartz-cemented sandstones commonly form prominent ridges and cliffs due to their weather resistance, and can last centuries on a

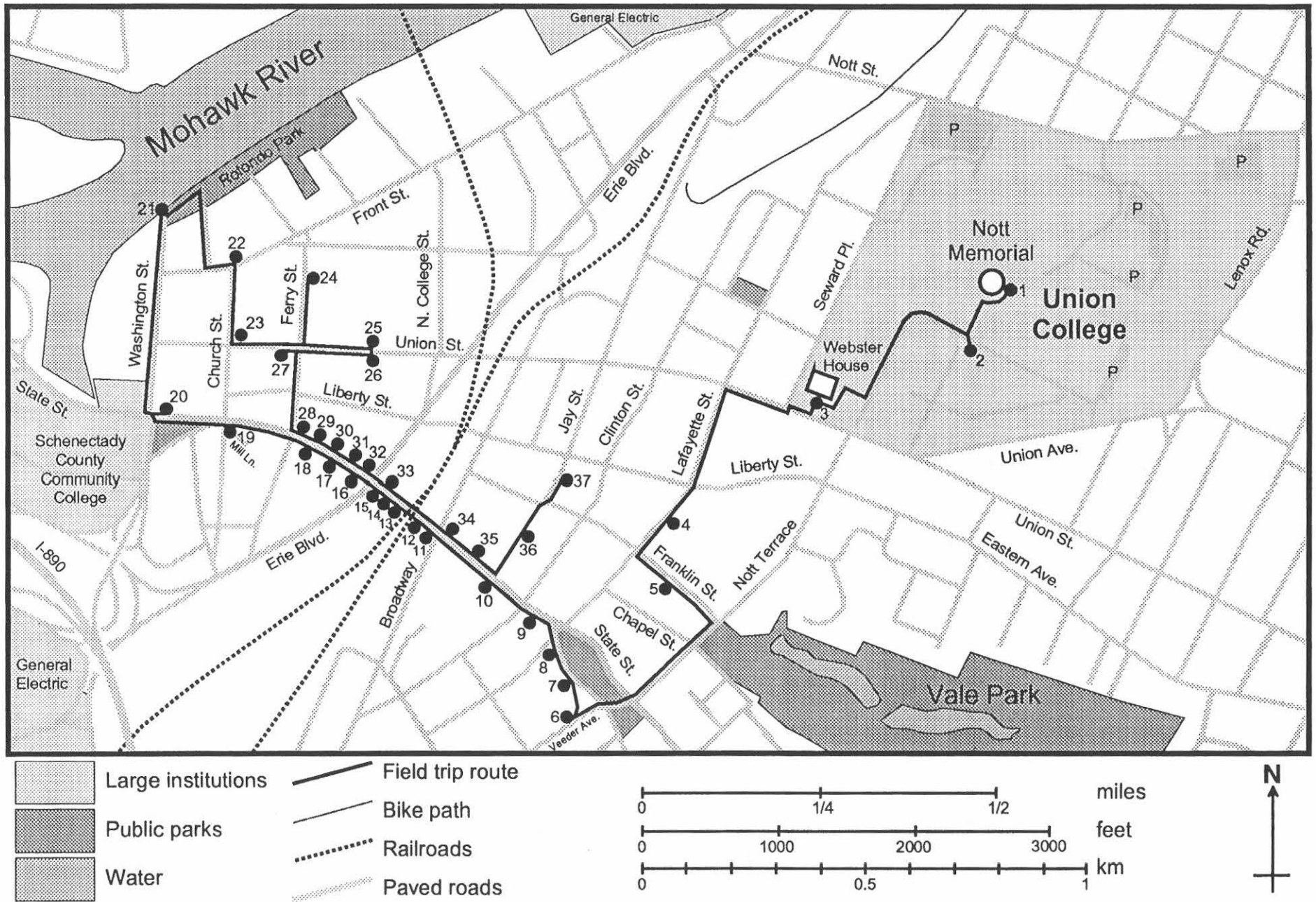


Figure 1. Street map of downtown Schenectady, showing trip route and numbered stops.

building face with little change. Limestones, although they are mostly made of calcite and other relatively soluble minerals, are commonly used for building stone since they are easily worked, aesthetically attractive, and tend to weather evenly. Limestone also commonly contains numerous fossils (Figure 3) and interesting sedimentary features (Figure 4) that can be interpreted in terms of the environment and processes of deposition. Sedimentary rocks are composed largely of minerals that are intrinsically colorless, such as quartz, feldspars, and calcite. The variety of colors one finds in sedimentary rocks usually depends on small quantities of coloring agents that coat or contaminate the mineral grains that make up most of the rock (Figure 5). Like paint on a house, it commonly takes very little pigment coating the grains to color them.

Table 1. Common minerals seen on this trip.

Mineral	Formula
Silicates	
Quartz	SiO_2
K-feldspar	KAlSi_3O_8
Albite	$\text{NaAlSi}_3\text{O}_8$
Plagioclase	$(\text{Na,Ca})\text{Al}(\text{Si,Al})_3\text{O}_8$
Biotite	$\text{K}(\text{Fe,Mg})_3\text{AlSi}_3\text{O}_{10}(\text{OH})_2$
Phlogopite	$\text{KMg}_3\text{AlSi}_3\text{O}_{10}(\text{OH})_2$
Muscovite	$\text{KAl}_2\text{AlSi}_3\text{O}_{10}(\text{OH})_2$
Talc	$(\text{Mg,Fe})_3\text{Si}_4\text{O}_{10}(\text{OH})_2$
Serpentine	$(\text{Mg,Fe})_2\text{Si}_2\text{O}_5(\text{OH})_4$
Garnet	$\text{Fe}_3\text{Al}_2\text{Si}_3\text{O}_{12}$
Tremolite	$\text{Ca}_2\text{Mg}_5\text{Si}_8\text{O}_{22}(\text{OH})_2$
Actinolite	$\text{Ca}_2(\text{Mg,Fe})_5\text{Si}_8\text{O}_{22}(\text{OH})_2$
Diopside	$\text{CaMgSi}_2\text{O}_6$
Clinopyroxene	$\text{Ca}(\text{Mg,Fe})\text{Si}_2\text{O}_6$
Orthopyroxene	$(\text{Mg,Fe})_2\text{Si}_2\text{O}_6$
Olivine	$(\text{Mg,Fe})_2\text{SiO}_4$
Sphene (titanite)	CaTiSiO_5
Tourmaline	$(\text{Na,Ca})(\text{Mg,Fe,Al})_6\text{Al}_3\text{Si}_6\text{O}_{18}(\text{BO}_3)_3(\text{OH})_4$
Carbonates	
Calcite	CaCO_3
Dolomite	$\text{CaMg}(\text{CO}_3)_2$
Oxides	
Magnetite	Fe_3O_4
Hematite	Fe_2O_3
Ilmenite	FeTiO_3
Rutile	TiO_2
Sulfides	
Pyrite	FeS_2
Native elements	
Graphite	C

shale, splits easily along its foliation and so is used for roofs and paving for floors and walkways. Marble is metamorphosed limestone. Fossils rarely survive metamorphism, but marble can have large crystals of calcite and a variety of metamorphic minerals including tremolite, diopside, tourmaline, phlogopite, sphene, and graphite. Gneiss is most commonly derived from a felsic igneous rock such as granite or granodiorite, or their volcanic equivalents rhyolite or dacite. Gneisses are composed of feldspar, quartz, and dark iron- and magnesium-rich minerals such as biotite. Gneisses, however, tend to be layered and folded as a result of deformation during metamorphism. Some gneisses have undergone partial melting, so some parts look like gneiss, and other parts have an igneous appearance. These "mixed rocks" are called migmatites.

Igneous rocks are generally thought of as being hard and strong. However, most are not suitable for use as building stone. Volcanic rocks are usually too porous, fragmented, or brittle. Plutonic igneous rocks (Figure 6) are hard and strong, but not all of these are suitable for building stone either. Plutonic igneous rock is very hard and difficult and expensive to work. Igneous rock that is used for building stone must therefore have special aesthetic qualities to justify the extra difficulty and expense. Many igneous rocks also contain substantial quantities of easily weathered minerals such as pyrite, olivine, and orthopyroxene. Although, the weathering of these minerals usually does not reduce the strength of the rock, the iron and manganese weathered from these minerals can discolor the stone and other parts of the building.

Metamorphic rocks (Figure 7) are derived from igneous or sedimentary protoliths by the action of heat, pressure, and deformation. The wide variety of possible protoliths, and the wide range of conditions under which metamorphism can occur, result in a similarly wide range of metamorphic rock types, colors, and textures. The common metamorphic building stones are slate, marble, gneiss, and *verd antique*. Slate, a fine-grained low-grade metamorphic rock derived from

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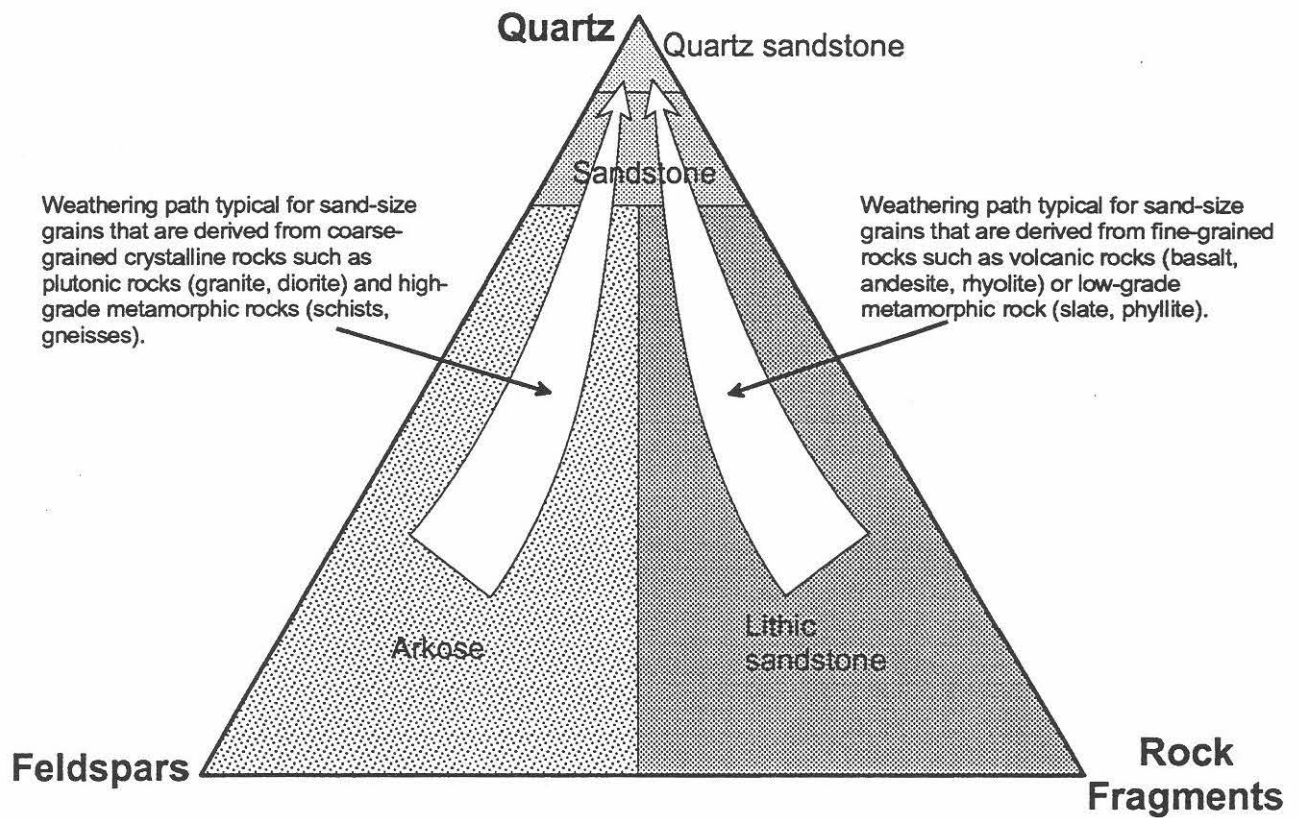


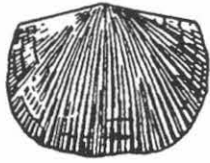
Figure 2. Classification scheme for sandstones. Sandstones have grains that are mostly in the range of 1/16 to 2 mm. Sandstones that were deposited close to their source tend to have unstable grains such as feldspars (usually white or pink) and rock fragments (usually black). With time and transport distance, quartz (glassy clear) eventually dominates.

BUILDING STONES OF SCHENECTADY, NEW YORK

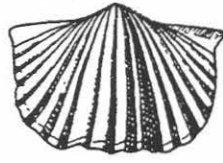
TRIP LOG

- Start Assemble at the east side of the Nott Memorial on the Union College Campus
- Stop 1 Nott Memorial. Note the various building materials that comprise this structure, from foundation to roof. The foundation appears to be made of gray granodiorite. The abundant wall stones are gray lithic sandstone (Figure 2) that may be Devonian in age, from the Catskill Delta, or Ordovician, from the Schenectady Formation. The columns are of pink granite and gray granodiorite. The light-colored cross bedded quartz sandstone on the steps is probably the Beriea sandstone from Ohio.
- Stop 2 Memorial Chapel. This building, constructed after World War I, has been here long enough to demonstrate the effects of weathering on rock. The columns in front of the building are made of calcite marble. Notice that the surfaces of the marble facing the building are smooth with sharp edges. The surfaces facing away from the building are rougher and the edges more rounded, a result of chemical and mechanical weathering on the side exposed to the elements. The steps are gray granitic rock, made of quartz, biotite, and feldspar.
- Stop 3 Webster House, southwest corner of Union College campus. This building was once the library for the City of Schenectady. The steps and foundation are made of muscovite granite with at least one inclusion of foliated muscovite granite. The sides of the steps are made of a fossil hash limestone, composed of broken pieces of shells, crinoids, and bryozoans. The kinds of fossils and the relatively uniform grain size suggest that they were deposited in a shallow marine environment by relatively uniform currents. The columns on the front of the building are pink granite similar to that quarried on Deer Isle, Maine. The granite has a "rapakivi" texture, which consists of pink microcline phenocrysts with rims of white albite. Rapakivi is a Finnish name meaning approximately "mud rock", after the crystal-bearing soil that commonly develops on rapakivi granite. Other minerals include black biotite and quartz.
- Stop 4 600 Liberty Street (on the Lafayette Street side between Liberty and Franklin Streets). This building has decorative facing stones of muscovite quartzite. This is a metamorphic rock composed mostly of quartz, and with some muscovite, tourmaline, and pyrite. Rocks similar to these are found in the Adirondacks and in New England. The original rock, or protolith, was most likely a quartz-rich sandstone or quartz conglomerate that contained some clay, iron sulfide, and probably organic matter. During metamorphism, under the influence of heat and pressure, the quartz grains in the sand recrystallized into quartzite, while the clays and associated sulfides were transformed into layers composed of muscovite, tourmaline, and pyrite.
- Thanks to the varying placement of the stones, one can observe the layers from the side and also face-on. Most of the stones have been placed with the layering horizontal, so that you see contrasting layers that are relatively rich in quartz or in muscovite. Some stones are oriented with the layering vertical and parallel to the building face, so that the stones look sparkly and the long, black tourmaline crystals are easily visible on the surface.
- Some of the quartzite is colored pinkish red by weathering pyrite, which leaves cube-shaped cavities behind. One can still see pyrite cubes in some of the stones. Into other cavities, quartz crystals have grown and their prismatic shapes are visible. On the southernmost section of quartzite facing, quartz veins can be seen, some containing quartz boudins – strong quartz layers around which other, weaker rock has deformed.
- Stop 5 600 Franklin Street. This one-story modern building has a facade of sandstone that is stunningly fossiliferous. Probably from a Paleozoic shoreline environment, where rich shell beds lay near a sandy beach, this rock is full of fossils. Look for large and whole brachiopods, pelecypods, and gastropods (up to 6 cm across). Note that many of the blocks are darker (brownier or redder) and more porous near their periphery. This "weathering rind" probably formed along joint surfaces through which ground water flowed. The ground water gradually dissolved away some calcite, making the rock porous with abundant fossil molds. The joints also broke the rock into slabs, which were later subdivided into blocks for building.

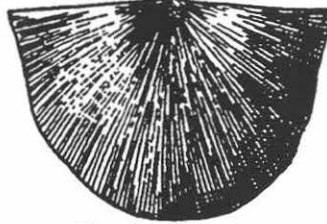
Brachiopods



Glyptorthis
Ordovician



Platystrophia
Ordovician



Strophonella
Silurian and Devonian



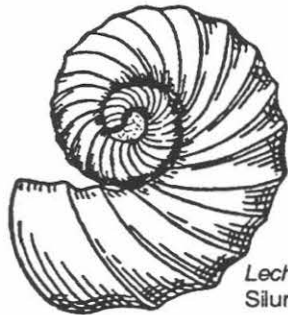
Mucrospirifer
Devonian

Brachiopods are bivalves that usually have calcareous shells. They are characterized by bilateral symmetry of each valve, but the two valves are typically dissimilar in size and shape.

Cephalopods



Kionoceras
Silurian



Lechtricoceras
Silurian

The cephalopods include squids, octopi, and the Nautilus. Most fossil cephalopods have calcareous, chambered shells that were either straight or coiled. The coiled varieties are symmetrical about a plane that is perpendicular to the coil axis.

Figure 3. Examples of the kinds of fossils that can be found in building stones in downtown Schenectady. Most of these fossils occur in limestone, but some can be found in some unusual calcareous sandstone (adapted from Moore et al., 1952, reproduced with permission from McGraw-Hill book Company, Inc.). All examples are approximately life-size unless otherwise indicated.

Gastropods



Lecanospira
Ordovician

Snails and related organisms having shells that usually coil along a particular axis, rather than symmetrically about a plane.

Pelecypods



Orthodesma
Ordovician

Pelecypods include the common modern bivalves, mussels, and clams. The two valves are usually very similar or identical (mirror images), but each shell is not bilaterally symmetrical.

Bryozoans



Stictoporella
Ordovician

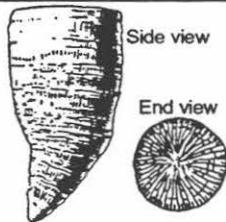


Rhinidictya
Ordovician

1 mm

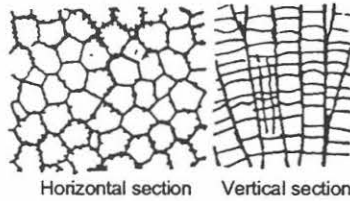
Small leaf-like or branching forms with many small pores, in which these colonial bryozoan polyp organisms lived.

Corals



Streptelasma
Ordovician

Rugose corals are cone-shaped and have a solitary coral polyp. Radial ribs decorate the inside of the cone.

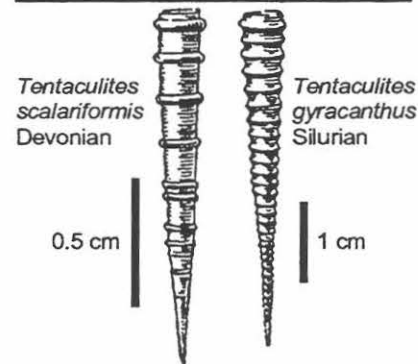


Nyctopora
Ordovician

Tabulate corals are layered or plate-shaped and are made up of numerous small cells. The cells were where the many little colonial polyps lived.

0.5 cm

Tentaculites



Tentaculites scalariformis
Devonian

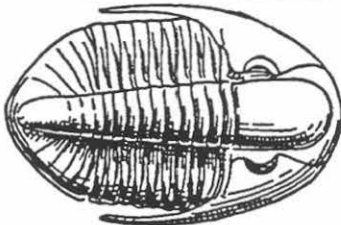
Tentaculites gyracanthus
Silurian

0.5 cm

1 cm

Tentaculites is a variety of invertebrate of uncertain zoological affinity. Shells are typically chitin mineralized with calcium phosphate.

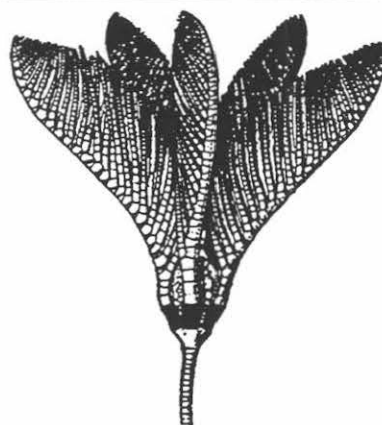
Trilobites



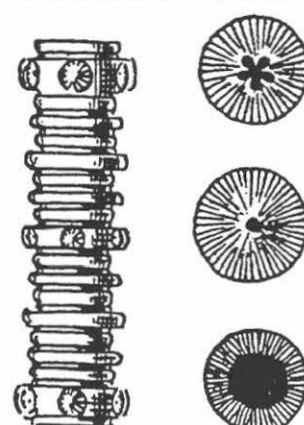
Bathyrus
Ordovician

Trilobites were arthropods whose closest modern relative may be the horseshoe crab. Though extinct since the Permian, trilobites dominated the community of mobile bottom-dwelling organisms in the Paleozoic seas. The mineralized chitinous exoskeletons commonly appear as thin, black plates in rock.

Crinoids



Ptilocrinus
Modern



Close-up of
crinoid stalk

Close-up of
crinoid segments

Crinoids are delicate, flower-like organisms that consist of a segmented stalk, a plate-covered body, and numerous segmented tentacles. The segments are made of calcite and usually become disarticulated during decomposition of the soft parts of the organism. Most crinoid fossils are disarticulated disk segments.

- Stop 6 Schenectady County Office Building, Veeder Avenue and State Street. The facade of this modern building contains three types of building stone. At the Veeder Avenue entrance, the steps and low enclosing wall are made of coarse, gray granite which contains white plagioclase, pink K-feldspar, hornblende, and gray quartz.

The red facing stone around the Veeder Avenue entrance is an oxidized granite. The oxidized iron, as hematite, gives it a red color. Look closely at the quartz to see its bluish cast, an effect that is commonly produced by the presence of sub-microscopic rutile crystals that scatter blue light. The presence of blue quartz suggests a special temperature history for the rock. Titanium, in the original magma, was included as a minor dissolved impurity in quartz as it crystallized. If the pluton had cooled relatively quickly, over tens of thousands of years, titanium would have remained dissolved in the quartz. However, this pluton either cooled much more slowly, or the pluton was reheated to moderate temperatures and slowly cooled long after the pluton was solid. Over this long time at moderate temperatures (~400°C), titanium dissolved in the quartz nucleated and grew as tiny rutile crystals.

The retaining wall and the low walls parallel to Veeder Avenue and enclosing the Albany Street entrance are made of gray diorite (Figure 6), a plutonic igneous rock that contains much less quartz and K-feldspar than granite, and more sodium- and calcium-rich plagioclase. This diorite contains fine-grained xenoliths, which look like dark blobs. They may be "auto-inclusions", inclusions of finer-grained (more rapidly cooled) portions of the same magma, which was later fractured and mixed with adjacent magma.

- Stop 7 County Courthouse, Albany Street. The walls next to the sidewalk are biotite granite. The dark, sparkly flecks are the mica biotite, which are the next most abundant mineral after feldspar and quartz. The steps of the Courthouse are muscovite granite, which is similar to the biotite granite but it also has shiny flakes of colorless muscovite.

The columns and pediments of the Courthouse are made of limestone. First, look at the middle pediment on the left as you face the Courthouse. Notice the layers in this rock. In the lower half of this pediment, the layers are slanted; in the upper half, the layers are horizontal. The slanted layers are called cross-beds. Inspection of the rock with a hand lens shows that it is made largely of small fossil fragments. Many of the fossil fragments are rounded, suggesting abrasion during transport in water. The fact that most of the particles are of similar size suggests deposition by relatively even currents water. These observations might lead us to interpret that this rock originated as deposits in tidal channels or deltas on a carbonate bank, such as might be found in the present day Bahamas.

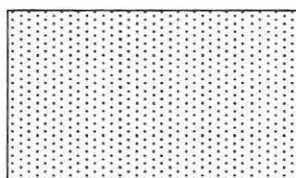
- Stop 8 St. Joseph's Church, Albany Street. This brick church has four types of building stone. The lowest foundation stone is a gray-green sandstone. Notice the pattern of closely spaced horizontal lines made by cutting tools on many of the block surfaces. Notice, too, that some of the blocks have another, less regular pattern of horizontal lines or grooves, thicker than the tool-cut pattern. These are "plane beds", horizontal layers made when sand is deposited in fast-moving streams or currents. These may be Devonian sandstones from the Catskill Delta.

Capping the lower foundation stone is a lighter-colored fossiliferous limestone that contains bryozoans. The steps of the church are made of coarse-grained, gray, biotite granite. Atop the stones that flank the steps are blocks of fossil hash limestone.

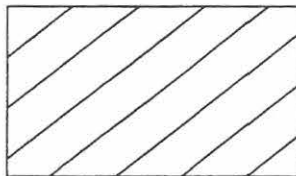
- Stop 9 504 State Street. The large panels of facing stone that cover the entryway are made of marble. Look for panels that contain large and dramatic folds that are highlighted by gray patches and layers in the white rock. The black specks and gray areas are graphite, formed by the metamorphism of organic matter that was present in the original limestone. The sparkly mineral is the brown mica, phlogopite, a magnesium-rich biotite. The elongate, colorless to light brown crystals in the marble are the mineral actinolite, an amphibole similar to hornblende. Its presence indicates that high temperatures were reached during metamorphism.

Outside the entryway, note the foundation stones composed of fossil hash limestone. Notice that this limestone is coarser – the fossil fragments are larger – than in fossil hash limestones at previous stops. Notice, too, the shapes of the fragments. Many of them are rounded, indicating that the fragments have been abraded during transport.

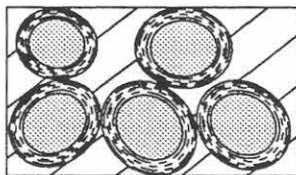
Around the corner on Clinton Street, look at the blocks of limestone and the orientations of the layers within them. Notice that some blocks contain slanted cross-beds. In one block, a flat-



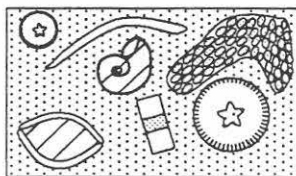
Carbonate mud. Fine-grained calcite with crystals too small to see, even with a hand lens, starts as muddy material. The rock made from carbonate mud is fine-grained and usually colored in shades of gray. The carbonate mud is produced by certain kinds of algae, from abrasion of larger calcite fossils, and from inorganic precipitation.



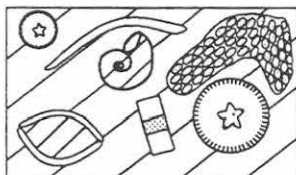
Crystalline calcite cement. Void space, such as the space between fossil grains, can be filled in by calcite crystals that are precipitated from groundwater circulating through the rock. Nice calcite crystals can sometimes be seen in some cavities that were not completely filled in.



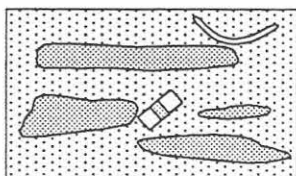
Oolites. An oolite starts out life as a seed grain (a bit of fossil, quartz sand, fish bone). The seed grain is rolled by tidal currents in an area with carbonate mud. The fine mud grains stick to the seed grain to form a layer. When the tide turns, the oolite rolls the other way and another layer is built up. Eventually the oolite becomes too large to roll. Oolites are therefore ovoid grains composed of a central seed surrounded by thin layers of fine-grained carbonate mud. Oolites are typically 1 mm in diameter.



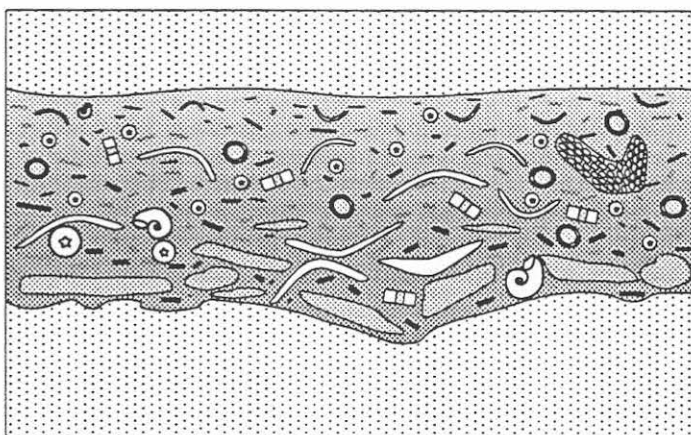
Here is an example of fossils embedded in carbonate mud. The fossils shown here include crinoid fragments (doughnuts), a bryozoan (celled structure), a gastropod (snail), and some brachiopods (arc-shaped bivalve shell fragments). The void spaces inside the paired bivalve shell and the gastropod have been filled by crystalline calcite cement.



In this example, the fossils are the same as above, but have been cemented together by coarse-grained calcite cement that was precipitated from groundwater.



Rip-up clasts. During storms or other heavy wave action in near-shore environments, pieces of local carbonate rock can be torn up and deposited elsewhere. In this example, carbonate rock clasts have been mixed with and deposited in a matrix of fossils and carbonate mud.



Graded bed of coarse-grained fossil-bearing limestone. Graded beds are those that vary systematically in grain size from bottom to top. Graded beds in limestones typically represent storm deposits that were formed in shallow offshore water or in a nearshore or lagoon environment. The beds are made of coarse-grained carbonate materials that were transported and deposited in an environment different from where they formed. The bottoms of graded beds are usually irregular, as a result of erosion of the sediments by the storm waves and currents. As the currents slowed, large rip-up clasts and big fossils were deposited first, followed by particles of smaller and smaller size. The example here shows a graded bed between two layers of carbonate mud. This example is typical of storm surge deposits in a lagoon environment in which carbonate mud is usually deposited.

Figure 4. Examples of typical limestone compositions and textures.

lying fossiliferous layer truncates, or cuts off, the cross-beds in a lower layer. This sedimentary feature is typical of channel fills – channels that have cut into underlying layers that are then filled in with sediment.

- Stop 10 Key Bank, 436 State St. The foundation stones on the outside of this building are made of rapakivi granite. Look closely and you will see large crystals of feldspar, much larger than the other crystals in the rock. These large crystals are called phenocrysts, and they crystallized from the slowly-cooling magma. Later, more rapid cooling caused crystallization of the smaller crystals that make up the matrix to the phenocrysts.
- Stop 11 340 State Street. The facing stone on this building is made of a granitic migmatite that contains the minerals K-feldspar, quartz, plagioclase, and biotite. This rock is composed of two parts: a metamorphic component and an igneous component that are closely associated. The metamorphic component is a layered gneiss that is separated into blocks by dikes and other small bodies of unlayered granite and granitic pegmatite. The gneiss has the same mineralogy as the granite, and it contains fine-grained dark inclusions that look like igneous xenoliths. This evidence suggests that the gneiss was once a granite, was then metamorphosed to form the gneiss, and metamorphism reached high enough temperatures so that the rock partially melted (~800°C). Deformation while the rock was partly liquid caused segregation of the liquid into fractures, which became the granite and pegmatite dikes when they crystallized during cooling.
- Stop 12 332 State St. The base of this building is made of a fine-grained biotite granite, over which is a fossil hash limestone.
- Stop 13 TrustCo Bank, 320 State St. This building has columns of pink biotite granite and slabs of polished red granite. The latter is particularly interesting. It contains large, gray to white plagioclase crystals that contain concentric layers within them having slightly different appearance. The layers formed as the crystals grew while suspended in the granite liquid. Some of the pink K-feldspar crystals have rims of white albite, which is the rapakivi texture described above. The zoning and rapakivi textures indicate physical changes in the magma during growth of these crystals, changes that may have included temperature, pressure, water content, or the chemical composition of the liquid during mixing with other magma or during crystallization of the liquid.
- Stop 14 Rudnick's, 308 State St. This storefront is faced with a granitic gneiss: a metamorphosed light-pink or tan granite. If you look carefully, you will see that the rock has a weak foliation that is defined by the parallel orientation of elongated crystals, including feldspar augen (Figure 7). Augen formed during deformation from pre-existing large crystals in the granite protolith.
- Stop 15 Fleet Bank, 306 State Street. Although the source of this rock is unknown, it looks just like the Whiteface Mountain facies of anorthosite from the high peaks region of the Adirondacks. Anorthosite is an igneous rock composed mostly of dark- to light-gray plagioclase with small quantities of black pyroxene and perhaps a few other minerals. Undeformed Adirondack anorthosite contains large dark-gray plagioclase crystals up to 30 cm long, and commonly 10 cm long. The dark color is caused by dust-sized ilmenite that formed in the plagioclase during cooling. When the original anorthosite was deformed, the edges of the gray plagioclase, and sometimes whole crystals, recrystallized into smaller light-gray or white plagioclase crystals that are free of ilmenite dust. In some of the less deformed parts of this rock you can see nice, big dark-gray igneous plagioclase crystals, and interlocking plagioclase and pyroxene crystals that have the original igneous texture. If you look closely at the dark-gray plagioclase crystals, you may see the blue iridescence that this variety of plagioclase is famous for.
- Stop 16 Olender Furniture, 260 State Street. The front of this building is made of polished muscovite-biotite granite. The interesting thing about this rock is that, if you look closely with a hand lens, you will see occasional dull pinkish-tan crystals up to 2 mm long. These are crystals of allanite, a member of the epidote family that is extremely rich in rare earth elements and the radioactive elements uranium and thorium. Over geologic time, the uranium and thorium cause sufficient radiation damage to the allanite crystal lattice to destroy it. The result is a glass that weathers easily to the pink material you see. Fresh allanite is dark brown and shiny. This granite is

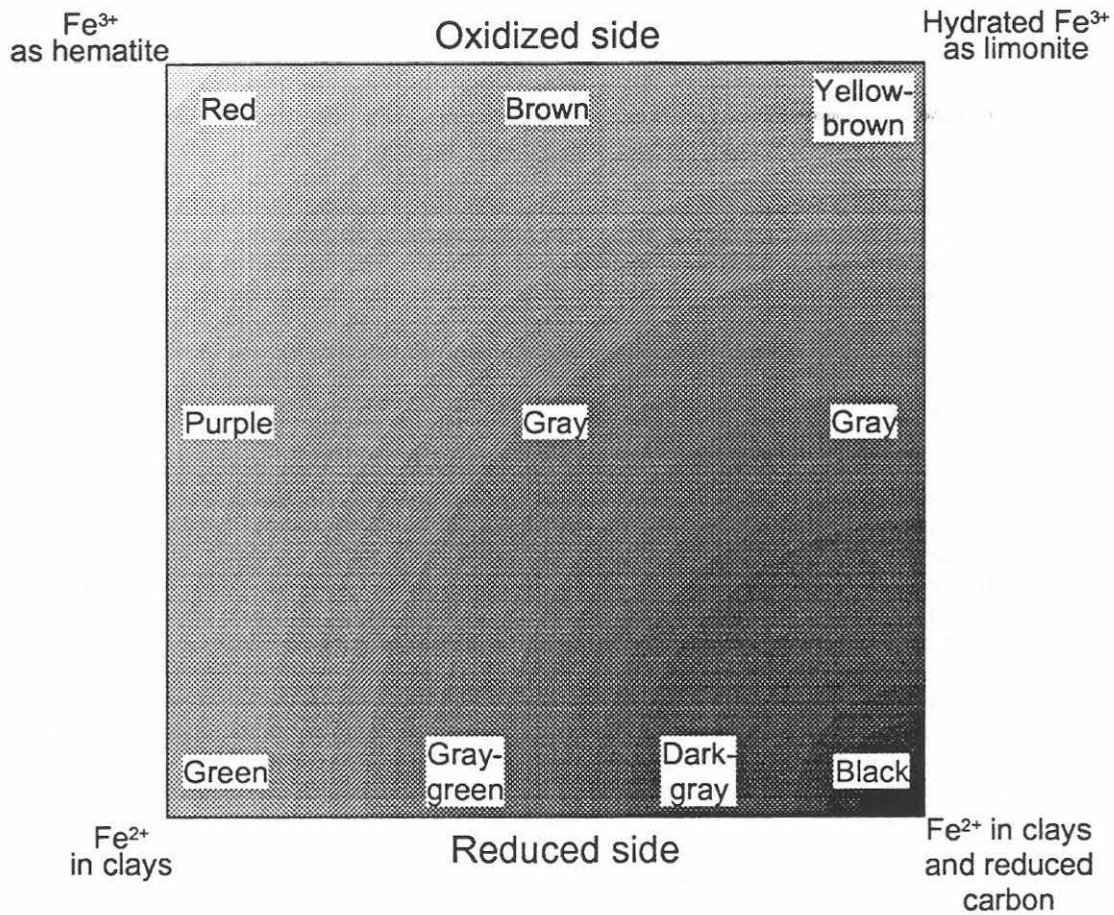


Figure 5. Diagram illustrating the origin of color in many sedimentary rocks. Most minerals in sedimentary rocks are essentially colorless, yet sedimentary rocks have a wide variety of colors. The color of many sedimentary rocks is caused by small quantities of coloring agents that include the forms of iron in the rock and the abundance of reduced carbon. Depending on its oxidation state and the minerals it is in, iron can impart red, yellow-brown, or green colors. Reduced carbon usually imparts a gray to black color. Rocks with little or no iron or reduced carbon are usually light-gray or white.

probably no more radioactive than the other rocks you have seen today, it just has larger (but fewer) crystals that contain these ubiquitous radioactive elements. This rock also contains small xenoliths, or perhaps auto-inclusions, and big, white K-feldspar phenocrysts.

Stop 17 224 State Street. The face of this building is composed of pink biotite granite that contains several xenoliths. Besides these and the zoned feldspars, the most interesting aspect of this rock is that it contains numerous, brownish-red crystals of sphene up to 3 mm long that are shaped like elongated diamonds. The name sphene is derived from the Latin word for wedge, which these pointy crystals resemble. Sphene is useful for radiometric dating, and is one of the second author's favorite minerals.

Stop 18 Fleet Bank, 216 State St. The facing stone of this entryway is composed of an almost black gabbro, an iron- and magnesium-rich rock that is rare as a building stone. The bulk of the rock is composed of dark-gray plagioclase, up to 5 cm long, and black clinopyroxene. These form an interlocking, igneous texture. If you look at a low angle at the highly polished surface, you will see that some patches of crystals are very shiny, like metal. If you look closely at these metallic patches, you will see that there are two different minerals present. One is more shiny and better polished. This is ilmenite which, in large concentrations, is the principle ore mineral of titanium. The other is somewhat duller and more poorly polished. This mineral is magnetite, and its poor polish is caused by its four good cleavage planes. The intersecting cleavages make numerous tiny triangular pits on the surface that are difficult to polish away. Ilmenite and magnetite both crystallized from the magma, are both oxide minerals, and are both natural semiconductors.

This building also has fossil hash limestone and biotite granodiorite. The granodiorite contains phenocrysts of white feldspar.

Stop 19 Curb stone at the intersection of Mill Lane and State St. This rock is a granitic gneiss that has a peculiar texture. Most of the mineral grains are rather small and form closely-spaced layers. The most obvious grains, however, are white feldspars that have angular or broken shapes. This is a particular kind of gneiss known as a mylonite. Mylonites are fault rocks in which fault deformation was by slow ductile flow instead of by brittle fracture. This rock was deformed enough to recrystallize the quartz and micas, but the large, strong feldspars remain intact enough to give an indication of the original grain size of the granite.

Stop 20 YMCA, on the south side facing State St. The lower part of the building is made of highly fossiliferous limestone, in which can be seen bryozoans, crinoids, brachiopods, and gastropods. Some of the stones have layering, both horizontal plane beds and cross beds. These indicate that this rock was deposited in a near shore carbonate bank environment in which currents, possibly driven by the tides, transported and deposited the fossils.

Stop 21 Washington St. at Rotondo Park. The stone wall at this Mohawk River overlook is composed mostly of fine-grained dolomitic limestone. The limestone is made of fine-grained, gray calcite mud (Figure 4), with irregular layers and patches of coarser-grained, brownish dolomite. The dolomite fills ancient worm burrows and bioturbated horizons in the rock.

In addition, the lower part of the wall on the west side contains some blocks of fossil-rich limestone that contain large brachiopods and tabulate coral. There are also some blocks of gray Schenectady Formation lithic sandstone (Figure 2), one of which has exposed flute casts that were formed by the swift turbidity currents that deposited the sandstones.

Stop 22 21 Front St. This house contains an excellent example of a red sandstone. With a hand lens you will be able to see that the rock contains abundant translucent white and pink feldspars in addition to gray glassy quartz and mica. The abundant feldspars mean that this sandstone is classified as an arkose (Figure 2), the grains for which were derived from coarse-grained igneous or metamorphic rocks, and were transported a relatively short distance prior to final sedimentation and lithification into rock. The red color strongly indicates that this sediment was originally deposited on land, which allowed oxidation of the iron to form the red hematite pigment. This rock was probably quarried from Mesozoic red beds in the Connecticut River basin or in the Newark-Delaware basin farther south.

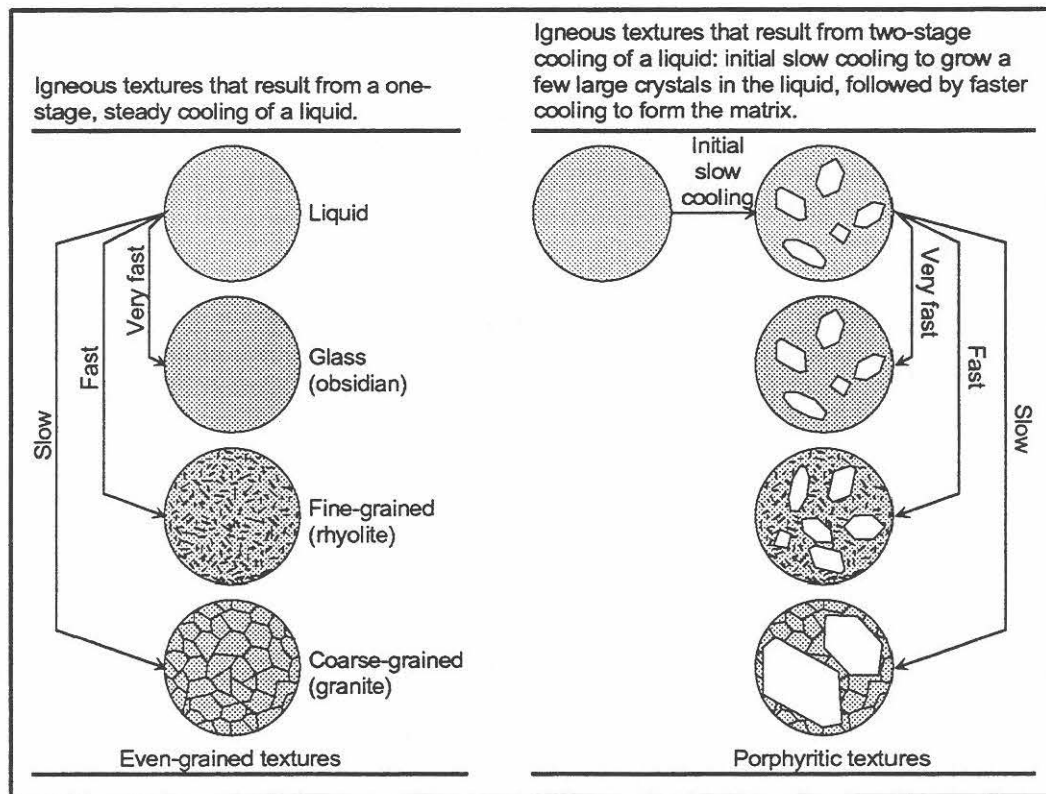
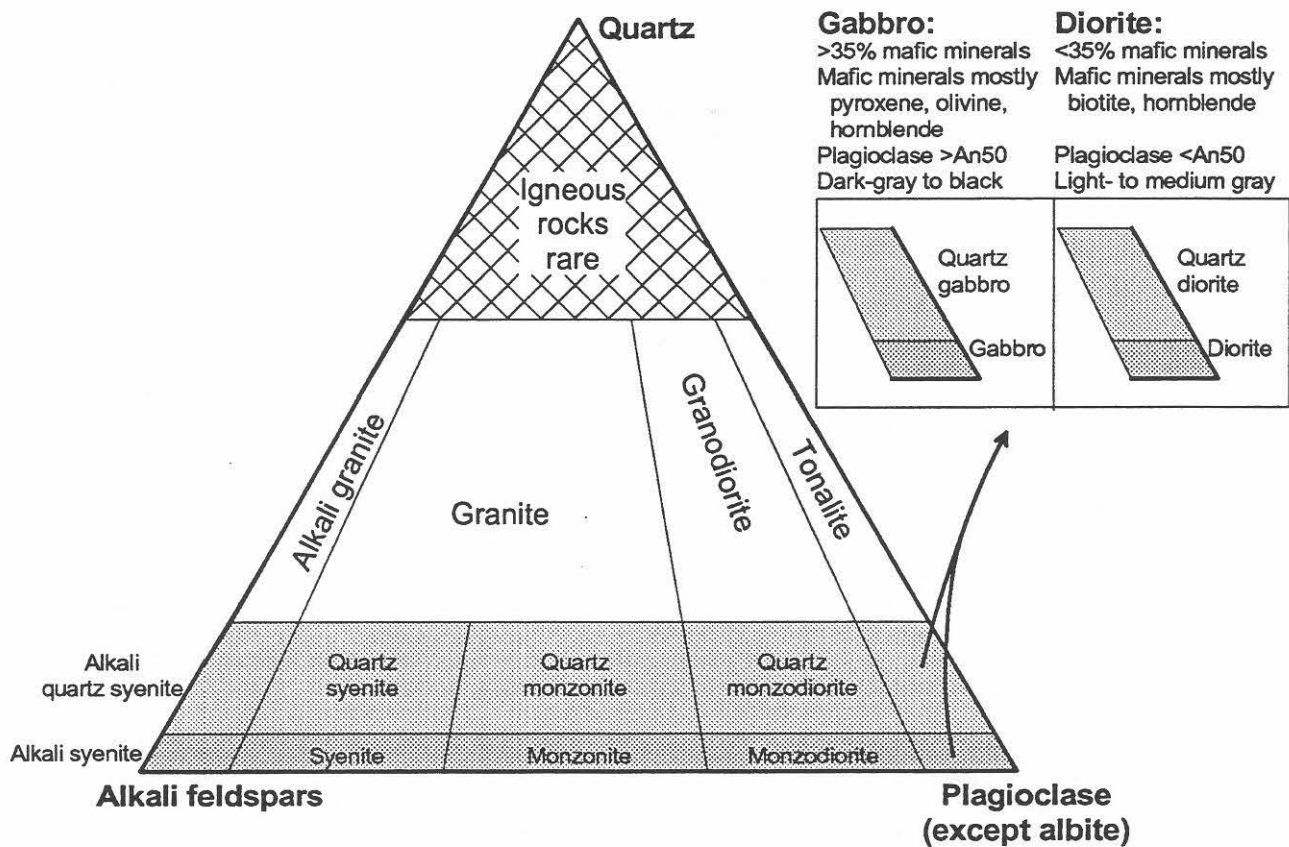
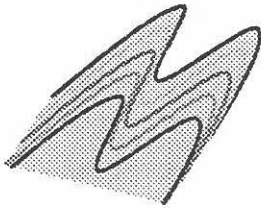


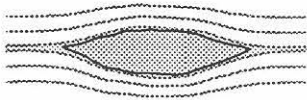
Figure 6. Classification scheme for the common plutonic igneous rocks. The alkali feldspars include the Na-feldspar albite, the K-feldspars sanidine, orthoclase, and microcline, and the Na-K-feldspar perthite. The K-feldspars can be pink or white, whereas albite and the plagioclase feldspars are rarely pink and usually white or gray. Anorthosite is a type of gabbro or diorite having >90% plagioclase.

- Stop 23 First Reformed Church, at the intersection of Union and Church Streets. This church was originally built in 1862, but was damaged by fire and rebuilt in 1948. The foundation to this church is made of highly fossiliferous limestone that contains abundant bryozoans, brachiopods, tabulate and rugose corals, crinoids, and cephalopods.
- The front of the church has steps made of slate, probably from the Taconics in New York or Vermont. The pillars are made of pink granite, containing pink K-feldspar, white albite, clear gray quartz, and black mafic minerals. Several xenoliths can be seen. The coarse-grained red sandstone surrounding the doors is feldspar-rich arkose similar to that at Stop 22. The brown sandstone, which is carved into the forms of corn, grapes, wheat, pine, and oak, is a quartz sandstone in which most of the grains are glassy quartz. In both sandstones the coloring agent is hematite.
- The step below the doorway on the east side near the front of the church is a different fossiliferous limestone that contains abundant gastropods up to 4 cm across. This rock also contains thin, black fragments of trilobite shells and sub-spherical ostracods (?).
- Stop 24 St. Georges Episcopal Church. In the cemetery next to this church one can see the results of differential weathering. Compare in particular the weathering of marble and red sandstone of stones of similar age.
- The stone around the front door is made of fossiliferous limestone that includes beautiful examples of brachiopods, rugose coral, bryozoans, trilobite fragments, crinoids, and fine-grained rip-up clasts that were probably broken up and transported during storms. Some of the rip-up clasts contain thin layers that were probably deposited by algal mats on the floors of quiet lagoons. Most of the church is made of gray sandstone that weathers brown.
- Stop 25 First Presbyterian Church, 209 Union St. This building contains red quartz sandstone and blocks of gray sandstone, probably from the Schenectady Formation. The building next door is the Turnbull House. It is made of a red quartz sandstone that has numerous peculiar pockmarks up to 4 mm across. Many of the pockmarks have a rhombic (diamond) shape that is reminiscent of the crystal form of calcite. Calcite was probably the first cement in this rock, which was then followed by a more durable cement such as quartz. When the sandstone blocks were exposed to the weather, the soluble calcite gradually dissolved, and the sand grains which the calcite held in place fell out, leaving the pits.
- The steps of this building are made of red arkose, in which can be seen numerous black grains which are sand-size grains of the fine-grained metamorphic rocks slate and phyllite. The presence of both feldspars and low-grade metamorphic rock fragments indicates a nearby source area for the sand that was made up of igneous and low-grade metamorphic rock.
- Stop 26 220 Union St., Van Antwerp house This old house has a variety of different kinds of slate on its roof. The colors of the slates include black, red, gray, green, and purple (Figure 5). The different shapes, thicknesses, and edge treatments clearly show that the slates had different sources, and were probably periodically replaced to make repairs. The rusty-weathering slates are probably from Pennsylvania, whereas the others are probably from the Taconics in Vermont or New York.
- Stop 27 108 Union, the old Schenectady Court House. The steps are made of gray sandstone that has interesting parallel grooves on their sides. These grooves are formed as the plane beds in the sandstone weather out at different rates because of different grain sizes or different cements in alternating layers. The blocks to the sides of the steps are made of fossiliferous limestone that contains tabulate and rugose corals, brachiopods, and other fossils. An excellent tabulate coral can be seen to the left of the steps, and a large cephalopod fragment can be seen just to the right of the steps.
- Stop 28 215 State Street. The facing stone beneath the windows is marble. This marble is particularly interesting because it contains stylolites – very wiggly, sub-parallel sets of black lines in the rock. Stylolites form as fractures or other surfaces in the rock, along which water flowed during burial or low-grade metamorphism. The water dissolved some of the calcite, and left the dark insoluble material behind. Dissolution does not form a flat surface, but rather develops interdigitating calcite fingers on either side of the solution surface. This is what makes the stylolites so irregular. The fingers point in the direction of the maximum compressive stress that was acting on the rock during stylolite formation. This building also has muscovite-biotite

Metamorphic grade ↓	Resulting metamorphic rock types			
Very high	Aluminous Gneiss	Quartzite	Marble	Pyroxene gneiss
High	Biotite schist	Quartzite	Marble	Amphibolite
Medium	Muscovite schist	Quartzite	Marble	Epidote amphibolite
Low	Muscovite phyllite	Quartzite	Marble	Chlorite-actinolite schist
Very low	Slate	Quartzite	Marble	Greenstone
Unmetamorphosed Protolith →	Shale	Sandstone	Limestone	Basalt



Folds are very common in metamorphic rocks, since most metamorphic rocks have been deformed. Although many protoliths are layered to begin with and remain layered after metamorphism, layering can also form during metamorphism from unlayered protoliths. After severe deformation, inhomogeneities, such as veins, dikes, or pebbles can be smeared out so that they appear to be layers.



Augen (German for "eyes"), are formed from preexisting crystals in a metamorphic rock. Deformation smears out the crystal into a lens (eye) shape, with a fine-grained recrystallized mineral rim, and tails that extend both directions from the crystal in the plane of the rock layering.

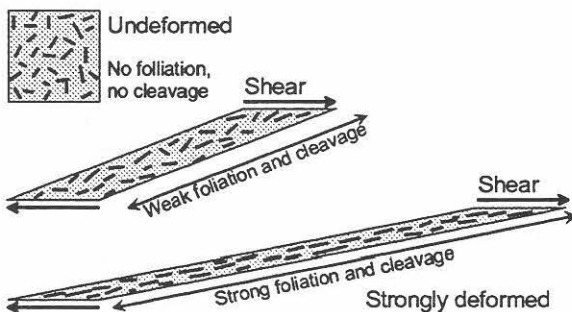


Illustration of how foliation and cleavage develop in a rock. The first block represents an undeformed rock having randomly oriented minerals such as the sheet silicates biotite, muscovite, and chlorite. These minerals are usually plate-shaped and have a good mineral cleavage. Although the minerals have cleavage, the rock initially has no cleavage or foliation. As the rock is deformed by shear (shown here) or flattening, the crystals gradually align themselves parallel to the direction that the rock is being elongated. The parallel arrangement of plate-shaped minerals is the rock foliation. The parallel alignment of the cleavage in these minerals results in a rock having good cleavage also.

Figure 7. The common metamorphic rocks that form at different metamorphic grades from four common protoliths. Metamorphic rock mineralogy and textures can be very complex, and depend on the protolith, which gives the rock its chemical composition, in addition to metamorphic pressure and temperature (grade), degree and type of deformation, and details of the metamorphic history of the rock.

granite facing some portions, and blocks of very fine-grained fossil hash limestone with visible bedding.

- Stop 29 225 State St. The facing below the windows of this building is black slate. The surface of the slate is the natural texture that formed when the slate was split along its excellent cleavage. The cleavage of slate is caused by the parallel alignment of muscovite and chlorite, which themselves have an excellent mineral cleavage. The sections of the building between the windows are faced with biotite granite, which contains rare, large feldspar phenocrysts.
- Stop 30 Mr. James Salon, 249 State St. The beautiful polished green stone on the outside of this building is a chlorite schist, rare in building stone because it is quite soft and hard to polish. If you look closely, you will see dark red garnets up to 4 mm across in this lovely rock. The most striking thing about this rock is its wavy "grain". The principal grain is a strong foliation of parallel-aligned chlorite crystals. This foliation was developed by deformation during metamorphism (Figure 7). At the time it was formed, this foliation was probably quite planar. The rock was later deformed twice to form two sets of small "crenulation" folds that are similar in size, shape, and appearance to the corrugated paper of a cardboard box. This rock therefore gives clear evidence of three episodes of deformation during metamorphism. Different episodes of deformation probably indicate different episodes of the mountain building event that caused the metamorphism.
- Stop 31 Schenectady Federal Savings, 251 State St. This entryway is faced with slabs of spectacular red granitic migmatite. The visible minerals include pink K-feldspar, white plagioclase, glassy-clear gray quartz, and black biotite. As described above, migmatite means "mixed rock" in which some parts appear to be metamorphic rock and others appear to be igneous. The layered portions of this rock are gneisses that have been highly metamorphosed and deformed. The igneous rocks include the entire range from fine grained granite to granitic pegmatite (very coarse-grained). In some places gneissic layers or granitic dikes have been offset by small shear zones (very local and short faults), that formed when the rock was hot, ductile, and probably partly liquid. Some of the granite dikes are obviously intrusive and crosscut layering or other dikes, but some appear to have formed in place by the partial melting of the metamorphic portions. On the polished surface one can see metallic gray and metallic yellow minerals, which are iron oxides and pyrite, respectively.
- Stop 32 267 State Street. This building has polished facing stone that looks identical to the dark gabbro at Stop 18, only more fine-grained. The Photo Lab, at 273 State St. has a few small slabs of gray quartz monzonite porphyry (Figure 6) on the lower part of the foundation.
- Stop 33 Wall Street, at the intersection of State Street and Erie Boulevard. The lower part of the wall facing Erie Boulevard is composed of fossiliferous limestone. Many of the features that can be seen here are illustrated and explained in Figure 4. Some layers in the rock are composed of very fine-grained carbonate mud that contains few fossils. This may have been deposited in a quiet, possibly hypersaline lagoon environment. The coarse-grained layers that are interbedded with the carbonate mud are made mostly of whole and broken fossils and chunks of rock that look like the carbonate mud. These were probably formed by masses of water that were forced over the beach area into the lagoon during severe storms. The fast-moving, turbulent water carried fossils from the near offshore and beach environments into the lagoon where they were deposited as the currents slowed. In some cases the storm surge currents broke up lithified sediments into smaller stones, which were transported and deposited with the fossils as "rip-up clasts". The storm surge deposits are typically coarse-grained on the bottom and grade to finer-grained on top. Using this rule, you can see that some of the blocks are upside-down from their original orientation. Fossils visible here include gastropods, brachiopods, and corals. Red sandstone and fine-grained fossil hash are also present in this wall.
- Stop 34 401 State St. The facing stone on the outside of this building is an almost black diabase, a rock of basaltic composition having a grain size and texture intermediate between that of gabbro and basalt, but having a similar chemical and mineral composition. It is formed in small plutons that cool relatively quickly (in ~1 to 100 years), and in the centers of thick lava flows and lava lakes. This rock contains long, lath-shaped plagioclase crystals that enclose black pyroxenes and

metallic iron-titanium oxides. The textures are best seen by looking at a low angle at light reflected off the polished surface.

Stop 35 Center City and CVS, 433 State St. The facing stone here is composed of a "hypersolvus" granite. The granites we have seen previously have had discrete crystals of K-feldspar (pink or white) and plagioclase (usually white in granite, with twinning). This granite has only one feldspar with a composition intermediate between plagioclase and K-feldspar. During cooling, this intermediate feldspar composition became unstable and unmixed into very thin, alternating layers of K-feldspar and plagioclase. The unmixed feldspar is known as perthite. If you look closely with a hand lens you will see the alternating, irregular layers of unmixed plagioclase and K-feldspar in the feldspar crystals. Perthite only forms in slowly cooled rocks. If the rock had cooled quickly, it would still have only a single feldspar.

Stop 36 Jay Street pedestrian mall. Some stores have *verd antique*, green serpentinite, and green to maroon to brown altered serpentinite. The original material for both was probably ultramafic rock, such as a dunite or peridotite, which may have been part of the deep oceanic crust or upper mantle. During metamorphism, the rock was infiltrated by aqueous fluids that turned anhydrous olivine and pyroxene into water-bearing serpentine, talc, and amphiboles, and deposited silica and white calcite. Deformation during metamorphism shattered the rock mass, forming irregular blocks and veins. The altered serpentinite formed when later circulation of oxygen-rich water through the rock caused some of the iron in the silicates to turn to hematite, staining parts of the rock red.

Stop 37 City Hall. This building is among the most splendid in downtown Schenectady, and is a fitting location for the end of the trip. The steps are made out of gray muscovite granite, similar to that found in many other buildings we have seen.

The columns are made of white Vermont marble. These columns show especially well the effects of weathering, similar to that seen on the columns of Memorial Hall on the Union College campus (Stop 2). The side of the columns facing the elements are rather rough and worn-looking. The side of the columns facing the building is contrastingly smooth with sharp edges to the carved flutes. The softness, excellent cleavage, and solubility of calcite in the marble is what makes it so susceptible to weathering.

The historical placards on the front of the building are made of altered serpentinite, similar to that seen along Jay St. at Stop 36.

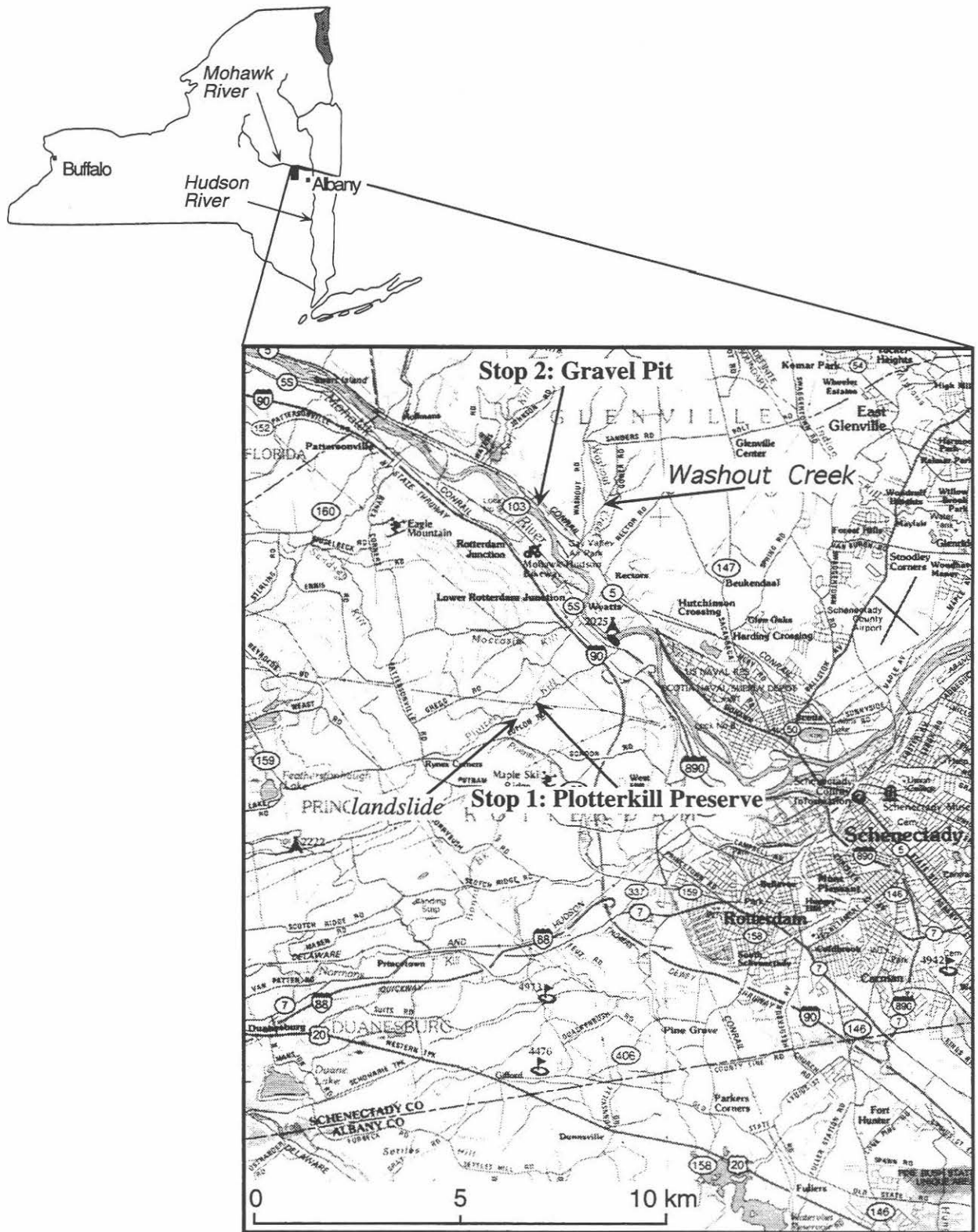


Figure 1. Location of Stops 1 and 2 in Schenectady County, New York.

FLUVIAL GEOMORPHOLOGY OF THE PLOTTERKILL PRESERVE

The valley floor along the Plotterkill reveals up to three prominent fluvial terraces, which are underlain by very poorly sorted deposits of sand, gravel, and angular boulders up to 70 cm long. Clasts in these deposits are angular and were derived locally from the sandstone units of the Schenectady Formation. These clasts show a pronounced imbrication in which their a-b planes are inclined upstream, suggesting deposition by fluvial processes.

The presence of terraces along the Plotterkill indicates that the rate of incision of the Plotterkill has not been constant since deglaciation of the region ca. 13 ka. Fluvial terraces represent remnants of paleofloodplains and the development of floodplains requires a period of equilibrium between sediment yield to a stream and stream power (Ritter et al., 1995). During such a period of equilibrium, channels neither aggrade nor degrade and fluvial erosion is dominantly lateral, which produces broad flat valley floors beveled on bedrock or alluvium. A subsequent increase in stream power or a decrease in sediment yield, or both will cause a river to incise into its floodplain. This paleofloodplain may thus be preserved as the tread of a terrace along the margins of the modern stream; the older the terrace, the higher it is above the modern stream (Fig. 2).

In several localities along the Plotterkill the boulder lag of angular sandstone that underlies terrace treads is less than 2 m thick and rests directly on a horizontally beveled planar surface of Schenectady Formation shale. This indicates that at least some of the Plotterkill terraces are erosional (strath) terraces rather than depositional terraces. An erosional origin for the terraces along the Plotterkill is supported by the observation that the Plotterkill terraces are unpaired; that is, terrace elevations differ from one side of the Plotterkill to the other (Fig. 2). Unpaired terraces are commonly associated with lateral and vertical erosion rather than deposition (Ritter et al., 1995). The distinction between erosional and depositional terraces is significant in that depositional terraces require a period of aggradation followed by fluvial incision whereas strath terraces reflect progressive fluvial incision punctuated by periods of equilibrium during which the stream neither aggraded nor degraded its channel.

Progressive incision by the Plotterkill may have been caused by a progressive decrease in local baselevel, a decrease in sediment yield, and/or an increase in stream power. Progressive drainage of Glacial Lake Albany ca. 13-12 ka (see Wall and LaFleur, this volume) and subsequent incision of the Mohawk and Hudson Rivers into deposits of Glacial Lake Albany would have lowered local baselevel progressively. Tributary streams affected by this decreasing baselevel would be forced to incise their channels. The observation of 3 prominent fluvial terraces along Washout Creek, which joins the Mohawk River directly north of the Plotterkill-Mohawk River juncture (Fig. 1), suggests that fluvial incision in the eastern Mohawk Valley was a regional phenomenon and may have been caused by a lowering of local baselevel in response to a step-wise drainage of Glacial Lake Albany. This hypothesis is consistent with the observation of multiple levels of Glacial Lake Albany (e.g., Clark and Krakow, 1984; Pair and Rodrigues, 1993; Wall and LaFleur, this volume). If correct, this hypothesis would imply that drainages that did *not* drain directly into Glacial Lake Albany did not incise their channels to the same degree as did drainages that drained directly into the Lake.

GEOMORPHOLOGY OF AN ACTIVE LANDSLIDE IN THE PLOTTERKILL PRESERVE

Simultaneous downcutting and lateral erosion by the Plotterkill has eroded the toes from many hillslope reaches within the Preserve and this, in turn, probably caused one large complex landslide (Fig. 1). The lower part of the slide is devoid of vegetation and is actively eroding into the Plotterkill. In contrast the upper half of the slide is characterized by immature trees, numerous transverse cracks, and minor scarps (Fig. 3). The top of the slide comprises the main scarp, a ca. 2 m nearly vertical slope. The presence of a prominent scarp, several back-tilted surfaces, numerous transverse cracks and scarplets suggests a rotational slide (slump) origin for the upper part of the slide. Active and continuous erosion of the lower part of the slide suggests that earthflow is the dominant process. Springs at the approximate boundary between the slump-dominated and flow-dominated parts of the slide have been noted on several occasions and probably reflect groundwater flow along a shear surface beneath the slump-dominated part of the slide (Fig. 3). It is likely that the flow dominated part of the slide is caused by high pore water pressures in the foot and toe slopes.

The obvious instability of this landslide contrasts markedly with the more stable hillslopes in other parts of the Plotterkill Preserve and reflects the influence of lithology on hillslope stability. The location of this slide within the Preserve corresponds precisely with a change in the lithology of the

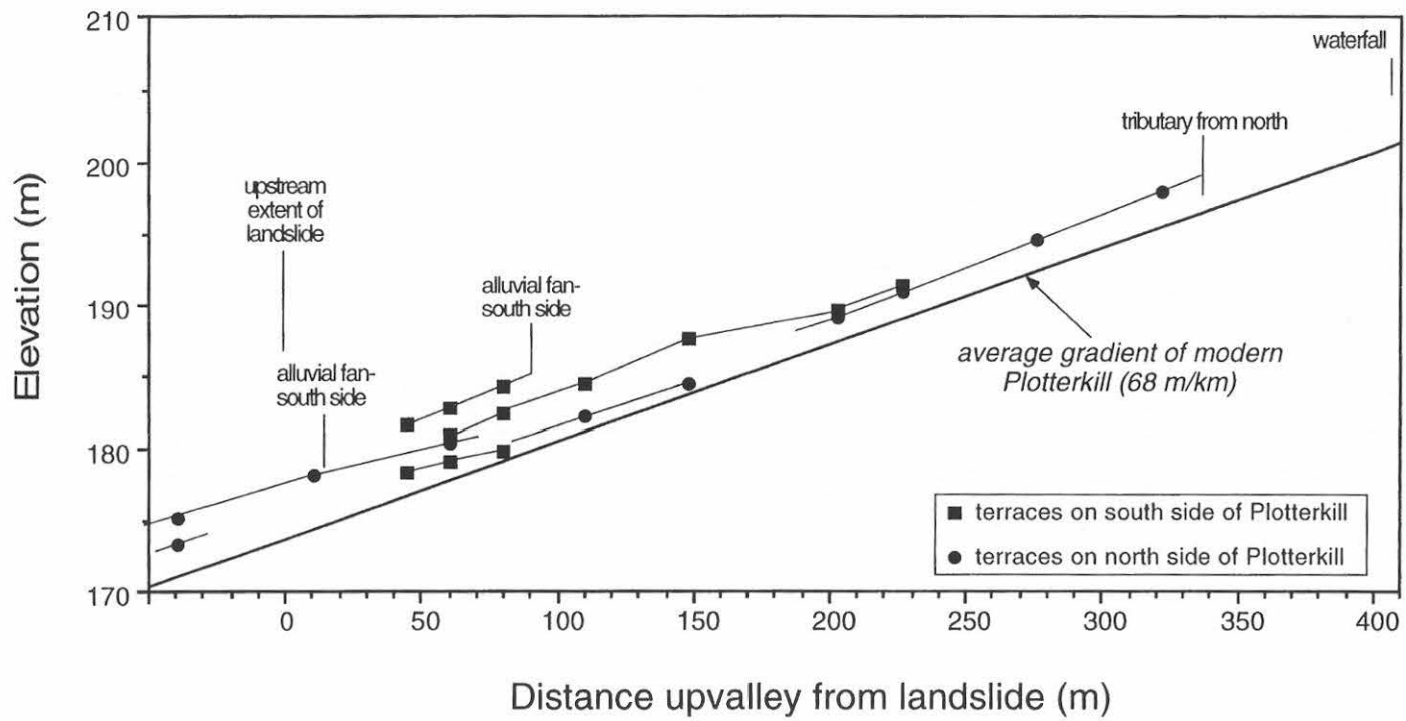


Figure 2. Longitudinal profile of fluvial terraces along the Plotterkill upstream from the landslide (see Fig. 4).

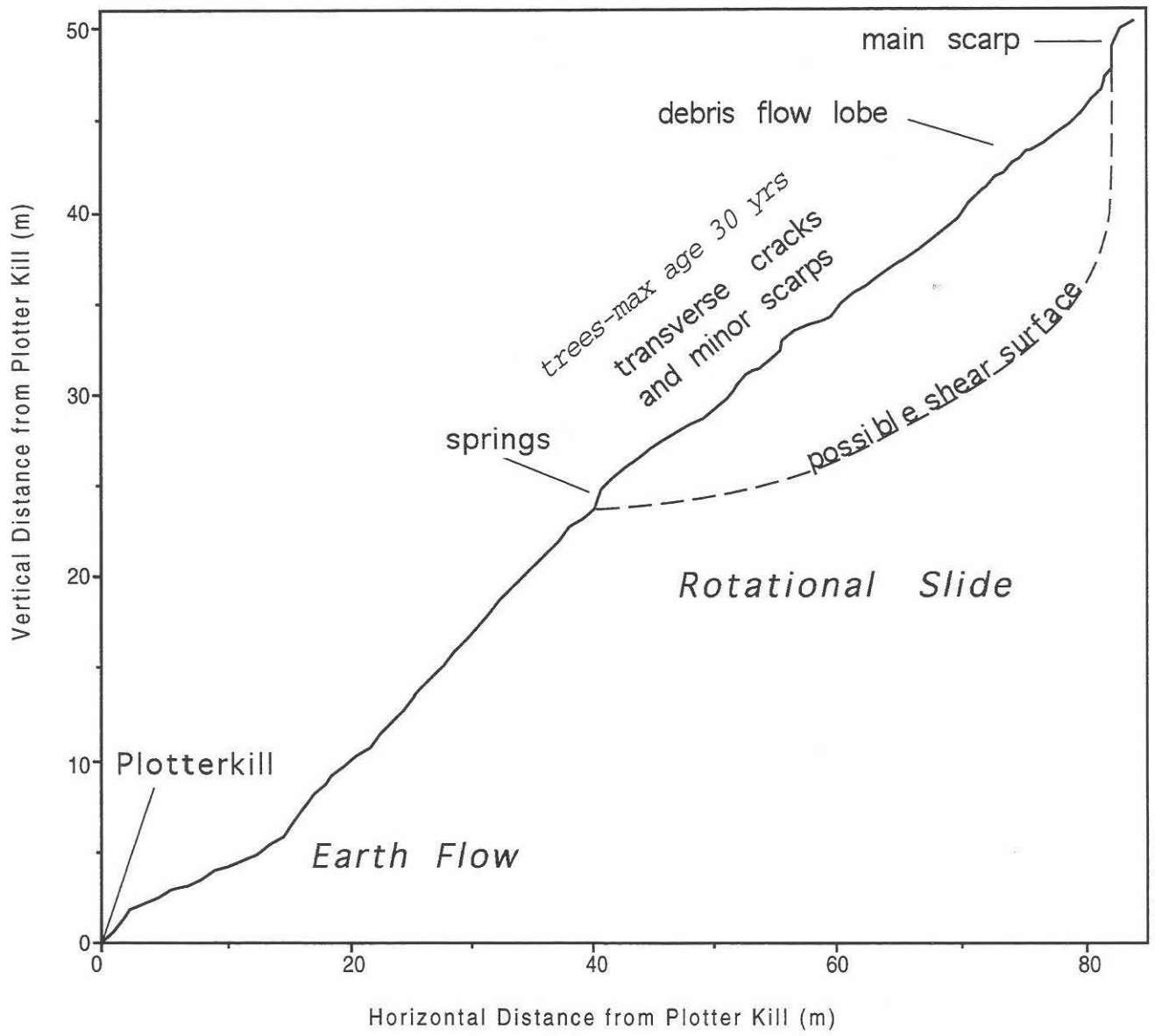


Figure 3. Profile of landslide on the south side of the Plotterkill Preserve as of May, 1995

material composing the hillslopes. Whereas most of the hillslopes in the Preserve are underlain by the Schenectady Formation or colluvium derived from the Schenectady Formation, the slide is composed of numerous exotic lithologies such as granite, gneiss, quartz sandstone, and volcanic rocks. Pebble counts of exotic lithologies (non-Schenectady formation) in the stream bed upstream and downstream from the slide and from the slide surface itself reveal clearly that the slide is composed of material that is anomalous to much of the rest of the Preserve (Table 2). The material composing the landslide is dominated by exotic lithologies and was probably derived from till, which either thickly mantles the south side of this part of the valley or infills a paleochannel that runs obliquely to the modern course of the Plotterkill.

TABLE 2. Number of Exotic Clasts >2 cm found during 3 Minute Pebble Counts

Upstream from the Slide	On the Slide Surface	Downstream from the Slide
28	163	75

The distribution and age of trees on the slide surface suggest several periods of landslide activity. Isolated tree islands on the lower part of the slide which are actively being eroded or buried by earthflow activity may be relicts of a vegetated surface which once extended across the entire surface of the slide/flow. The numerous trees on the upper part of the slide may be remnants of this same surface. Tree cores indicate that stabilization of this surface began more than ca. 30 years ago. Initial movement of the slide, which may have occurred anytime from 10^2 - 10^4 years ago, altered the course of the Plotterkill and produced an abandoned channel immediately downstream from the slide (Fig. 4). This was followed by at least one period of hillslope stability in which the slide surface became revegetated. In order to monitor the activity of the landslide, an array of stakes have been placed on the slide surface and these have been surveyed precisely with an electronic distance meter from the base of the slide (Hays, 1995).

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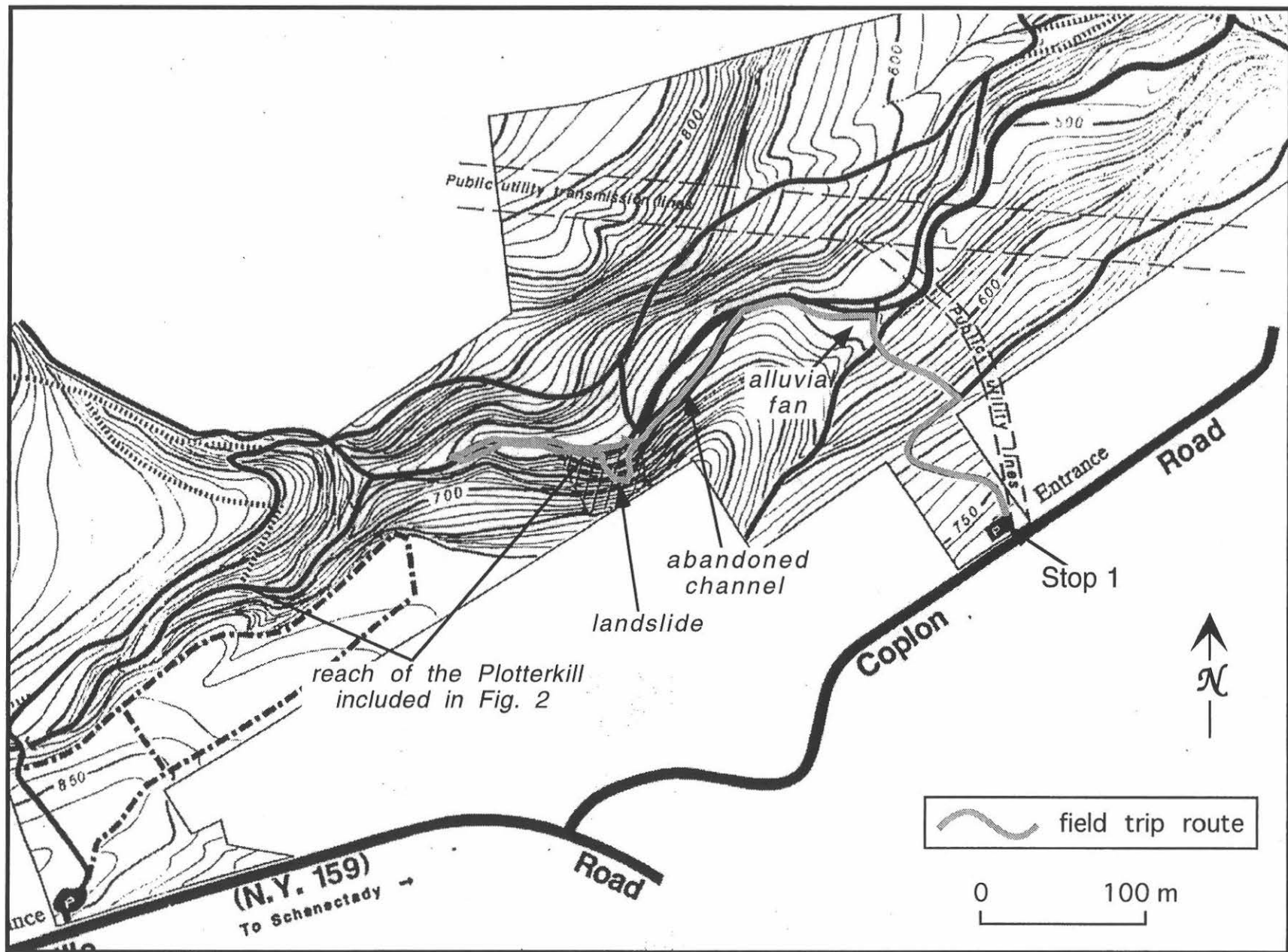


Figure 4. Topographic Map of a part of the Plotterkill Preserve, Schenectady County, New York (see Fig. 1).

FLUVIAL AND HILLSLOPE GEOMORPHOLOGY OF THE PLOTTERKILL PRESERVE, ROTTERDAM, NEW YORK

ROAD LOG

Miles from start	Miles from last point	Description
0.0		Start at Union College Nott Street Gate; turn left onto Nott Street, proceed west
0.5	0.5	Turn left onto Erie Blvd.; proceed southwest, cross State Street
1.4	0.9	enter 890
1.6	0.2	enter 890W
3.2	1.6	exit 890W at Campbell Road; stay on Campbell Road and pass first entrance to the Rotterdam Square Mall
4.1	0.9	take right onto Putnam Road, drive west up steep hill
6.0	1.9	take right onto Rt. 159; pass Maple Ski Ridge on left
7.9	1.9	take right onto Coplon road
8.3	0.4	take left into Plotterkill Preserve parking lot

From here we will spend several hours walking along the Plotterkill; the trails are well marked but some are quite steep.

We will begin by walking down the trail to the Plotterkill (Fig. 4). As you approach the valley floor (ca. 8 minutes) and before you reach the alluvial fan (Fig. 4) you will notice the curved trunks of some trees on the hillside to your left. Tilted trunks are a good indication that soil creep is active on the hillside

When you reach the relatively flat valley floor, you will be standing on an alluvial fan (Fig. 4) which is graded to a level that is much higher than the modern Plotterkill. As you continue down the axis of the fan you will cross two scarps, which demarcate distinct terrace levels cut onto the toe of the alluvial fan by the Plotterkill.

The trail turns to follow the south side of the Plotterkill and enters the downvalley end of an abandoned channel of the Plotterkill. Well preserved terraces can be seen on either side of this paleochannel. At the downvalley end of the abandoned channel, there is a good exposure into one of the prominent terraces in this part of the valley. This exposure reveals ca. 1-2 m of a very poorly sorted deposit of sand, gravel, and angular boulders up to 70 cm long. Clasts in these deposits are angular and were derived locally from the sandstone units of the Schenectady Formation. These clasts show a pronounced imbrication in which their a-b planes are inclined upstream, reflecting fluvial transport. In addition, one can observe that the modern Plotterkill flows on a beveled surface of Schenectady Formation shale. In places this surface is mantled by ca. 1 m or more of coarse and poorly sorted alluvium, which is very similar to the material that makes up the deposits that underlie the terrace surface.

Follow the paleochannel until the trail crosses the Plotterkill. You will see the large landslide to your left (south). The upvalley end of the paleochannel corresponds precisely with the downvalley end of the landslide. It is likely that initial movement of the slide forced the course of the Plotterkill to the north until it avulsed its channel, thus forming the paleochannel. Subsequent fluvial incision into the slide material and southward migration of the cut bank of the Plotterkill has isolated the inlet to the paleochannel.

Continue up the bed of the Plotterkill. Watch your step! On the south side of the Plotterkill, approximately 75 m upstream from the landslide is an excellent example of a flight of fluvial terraces (Fig. 2). Here, three distinct terraces can be differentiated on the basis of elevation. In addition, exposures into the alluvium that underlies the terrace surface can be seen. In several locations one can see clearly that the alluvium rests unconformably on a beveled surface which has been cut onto the shales of the Schenectady Formation. This surface is a rock cut terrace or a strath (Ritter et al., 1995) and is overlain by alluvium. The observation of a planar erosional surface that mirrors the terrace tread and which is overlain by alluvium the thickness of which does not exceed the normal scouring depth of the river meets the criteria for designating a terrace an erosional or strath terrace (Mackin, 1937). The presence of strath terraces suggests that the Plotterkill has been incising its channel progressively and that this incision was punctuated by periods of stream equilibrium during which time the Plotterkill laterally eroded the underlying Schenectady Formation. Step-wise incision may have been caused by episodic increases in stream power, decreases in sediment yield to the stream, decreases in local base level, or a combination of these three factors.

Return to the base of the landslide. Standing here, one can clearly see the abundance of exotic lithologies that make up the material composing the landslide (Table 2). This material is till and its prevalence in this part of the valley suggests that hillslope instability at this particular point in the valley is due, in part, to the influence of lithology on hillslope stability.

Walk up the slide. The slide surface can be divided into an upper and lower half (Fig. 3). The lower part of the slide is devoid of vegetation except for several small tree islands; these may be remnants of a continuous vegetated surface. Tree ring studies suggest that this surface is more than 30 years old. This part of the slide is continuously eroding into the Plotterkill and during times of high water tables, springs develop in the uppermost part of the lower half of the slide. These springs probably reflect ground water flow along shear planes that underlie the upper half of the slide. Saturation of the lower half of the slide is responsible for this nearly continuous earth flow. The upper half of the slide is probably dominated by rotational slide (slump) activity as is evidenced by transverse tensional cracks, minor scarps and rotated segments. The uppermost part of this part of the slide is characterized by a prominent near-vertical scarp.

An array of stakes (steel rebars) have been placed on the surface of the landslide and surveyed precisely using an electronic distance meter in April, 1995. We plan to monitor the movement of the landslide annually.

Return to vehicle

8.8

0.5 turn right on Rt.159

- 10.2 1.4 turn right on Rynex Corners road
- 15.5 5.3 turn right on Rt. 5S
- 17.4 1.9 turn left on Rt. 103
- 17.7 0.3 cross the Mohawk River on Erie Canal Lock 9
- 17.9 0.2 right on Rt. 5N
- 18.4 0.5 gravel pit on north side of road (pit is owned by Scotia Sand and Stone Company- 518-346-5749)

This gravel pit provides excellent exposures of the Scotia Delta (see Wall and LaFleur, this volume) that prograded eastward into Glacial Lake Albany. We are stopping here to briefly see the unequivocal evidence that these exposures reveal for higher local base levels ca. 12-13 ka. The elevation of the foreset-topset transition here (88 m) can be used to estimate of the surface of Glacial Lake Albany ca. 12.5 ka (see Wall and LaFleur, this volume) and hence of local baselevel at that time. Progressive and step-wise lowering of local base level to the modern elevation of the Mohawk River at this point of the Mohawk Valley (67 m) may have been responsible for episodic incision by the Plotterkill and other creeks in the region (e.g., Washout Creek, Fig. 1).

- 22.3 3.9 here we can see a lower level of the Scotia delta to the right (south) of Rt. 5N. There are at least 3 levels of the Scotia delta in this part of the Mohawk Valley. The elevations of these levels are 90 m, 85 m, and 70 m.

To return to Union College, continue southeast on Rt. 5N

- 24.5 2.2 left onto S. Church Street
- 24.6 0.1 right onto Union Street
- 24.9 0.3 left onto Erie Blvd.
- 25.4 0.5 right onto Nott St.
- 25.8 0.4 right into Union College Nott Street entrance

SEQUENCE STRATIGRAPHY OF PLATFORM CARBONATES: DEVONIAN LIMESTONES OF JOHN BOYD THACHER STATE PARK, SOUTHWEST OF ALBANY, NY

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ABSTRACT

The Lower Devonian strata that crop out on the Helderberg Escarpment illustrate the characteristics of marine platform parasequences in carbonate rocks. In the hierarchy of stratigraphic units, parasequences are a relatively conformable succession of genetically related strata bounded by surfaces of erosion. The parasequences exposed in this escarpment, which include stromatoporoid reefs, stromatolites, and lithified lime mud (micrite) show regressive facies separated by unconformities representing transgressive episodes. The lowermost part of the section exhibits classical karst-generated solution collapse breccia of the kind that hosts oils and gas in the subsurface. The Middle Devonian Onondaga facies at this site are full of coral-reef debris. In the subsurface, Onondaga reefs form gas reservoirs and now serve as reservoirs for gas storage. Sir Charles Lyell visited this classic site, part of the Helderberg Mountains, in 1841 and a monument commemorates his visit and that of Sir William Logan, James Hall, Amos Eaton, and others of the heroic age of Geology.

INTRODUCTION

In the hierarchy of stratigraphic units, parasequences are a relatively conformable succession of genetically related strata bounded by surfaces of erosion. The Lower Devonian parasequences exposed in this escarpment, which include stromatoporoid reefs, stromatolites, and lithified lime mud (micrite) show regressive facies separated by unconformities representing transgressive episodes. The Middle Devonian Onondaga facies Stratigraphically above the rocks at this site is full of coral-reef debris.

HISTORY

This classic site is on hallowed ground. A plaque erected in 1933 in memoriam of those pioneer geologists whose researches in the Helderbergs in the nineteenth century made this region classic ground includes not only almost all American pioneer geologists, but in addition, lists pioneers of British and Canadian geology. Among those listed are Amos Eaton (1776-1842), the John Gebhards Sr. and Jr. (life-cycle dates not available), James Hall (1811-1898), William W. Mather (1804-1859), Lardner Vanuxem (1792-1848), James Eights (1797- 1882), Sir Charles Lyell (1797-1875), Benjamin Silliman (1779-1864), Edouard de Verneuil (1805-1873), James D. Dana (1813- 1895), Henry D. Rogers (1808-1866), William B. Rogers (1804- 1882), Ferdinand Roemer (1818-1891), Louis Agassiz (1807-1873), and Sir William E. Logan (1798-1875).

Sir Charles Lyell visited the "Helderberg Mountains", as he called them, in September 1841 and although he rejoiced, noting that "the precipitous cliffs of limestone, render this region more picturesque than is usual where the strata are undisturbed" (Lyell, 1845, p. 67), he was more concerned in his account with the "Helderberg war" between Van Rensselaer and his tenants. On his return to the "Helderberg Mountains" in May 1846 the "Helderberg war" absorbed him again because he states that "the anti-renters have not only set the whole militia of the state at defiance, but have actually killed a sheriff's officer, who was distraining for rent." (Lyell, 1849, p. 260).

The definitive studies of the Lower Devonian carbonates of the Helderberg Escarpment exposed at John Boyd Thacher Park date to the early New York State Geological Survey and were written by Vanuxem (1842), Mather (1843), and Hall (1843). Their reports were supplemented and complemented later in the nineteenth century

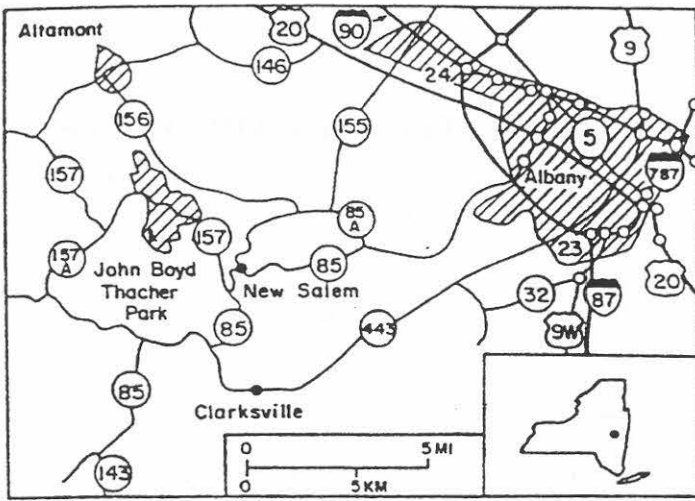


Figure 1A. Location of John Boyd Thacher-State Park (Fisher, 1987). The Indian Ladder Trail is located on Route 157.

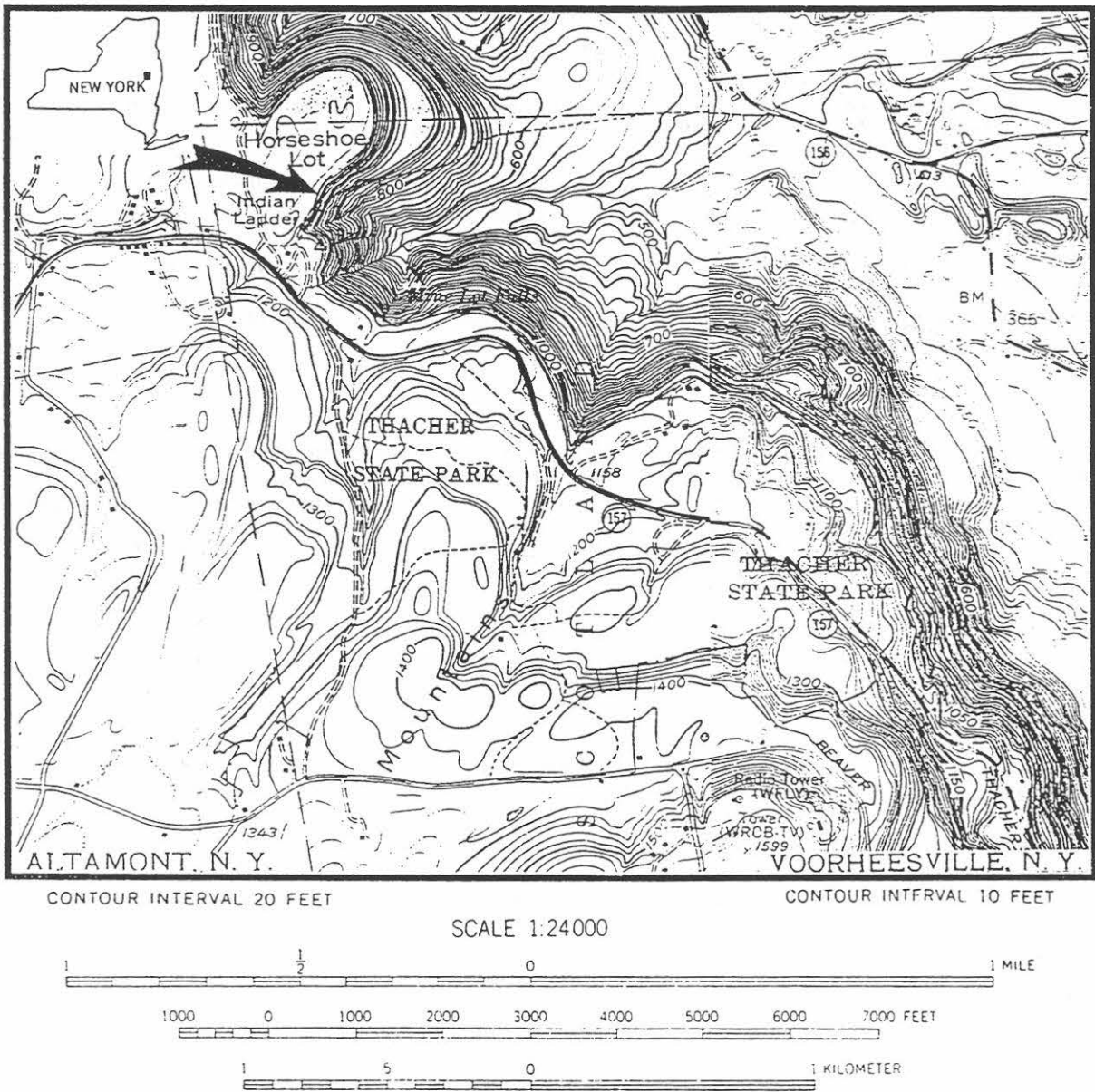


Figure 1B: Topographic map of John Boyd Thacher-State Parking showing location of Indian Ladder Trail.

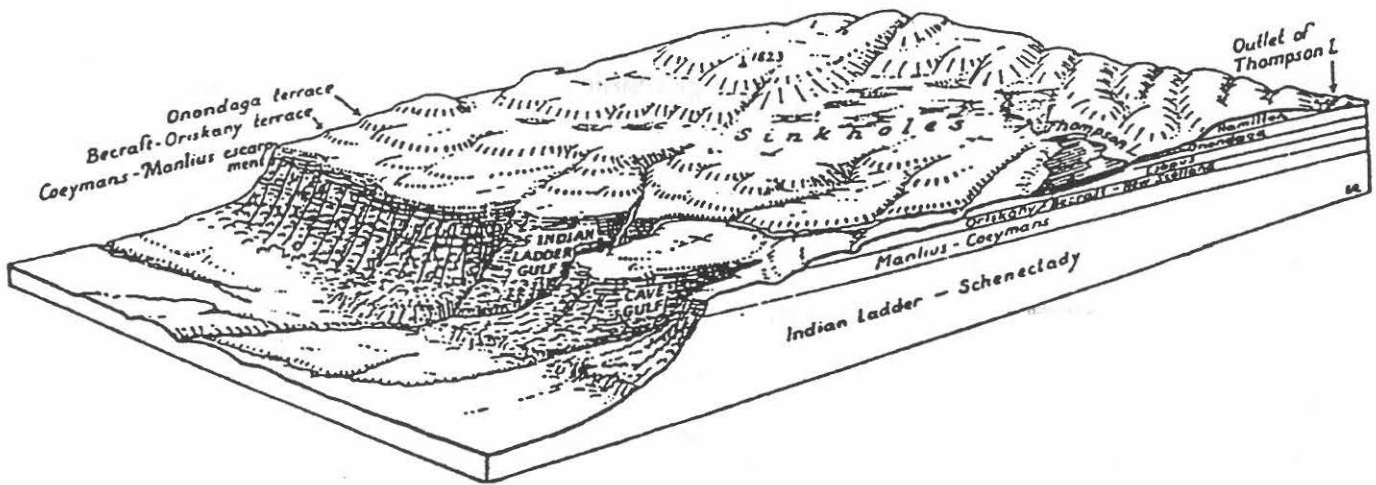


Figure 2. Block diagram of Helderberg Escarpment showing Indian Ladder location (labeled Indian Ladder Gulf), dip slope of Paleozoic strata, prominent terraces, and sinkhole topography (H.F. Cleland, 1930: modified by Winifred Goldring, 1935).

by the many volumes and papers by James Hall detailing stratigraphy and paleontology, followed in this century by many others, including Fisher (1987), Goldring (1935, 1943), Goodwin and Anderson (1982, 1985), Laporte (1967, 1969), Rickard (1962, 1975) Friedman, (1990;1991), and Friedman et al. (1992, p.180). Gurney and Friedman (1986) described vertical parasequences of Lower Devonian limestones as transgressive-regressive cycles in Cherry Valley, approximately 35 miles west of Thacher Park.

The Indian Ladder Trail site provides an unusual opportunity to study a vertical cliff of limestone strata: a vertical exposure of approximately 80 ft or 24 m exposed in the cliff is accessible by stairway and footpath; hand railings assure safety. One can view the entire sequence of the rocks at close quarter, including by hand lens; comparable physical settings in quarries never allow such close inspection.

Why the name Indian Ladder? Verplanck Colvin, one of the earliest men to write about the Helderbergs, in 1869 wrote:

“What is this Indian ladder so often mentioned? In 1710 this Helderberg region was a wilderness; nay all westward of the Hudson River settlement was unknown. Albany was a frontier town, a trading post, a place where annuities were paid, and blankets exchanged with Indians for beaver pelts. From Albany over the sand plains ... “Schenectada” (pine barrens) of the Indians ... led an Indian trail westward. Straight as the wild bee or the crow the wild Indian made his course from the white man’s settlement to his own home in the beautiful Schoharie valley. The stern cliffs of these hills opposed his progress; his hatchet fell a tree against them, the stumps of the branches which he trimmed away formed the round of the Indian ladder.”

The trail ended where the cliff did not exceed twenty feet in height. Here stood “the old ladder.” In 1820 this ladder was still in daily use (Goldring, 1935). The modern stairway crosses the old Indian ladder road which ran to the top of the escarpment where the trail begins.

LOCATION

Figure 1 shows the location of the John Boyd Thacher State Park, where the Indian Ladder Trail reveals the vertical sequence of Lower Devonian limestones that rest unconformably on the Ordovician Indian Ladder beds and Schenectady Formation which can be seen, locally, in gullies below the escarpment (Fig. 2). Entering Thacher State Park from Albany on Route 157 stop at the "Cliff Edge Overlook" for a view of the Taconic and Berkshire Mountains, Adirondacks, Hudson River, and City of Albany; then drive to the next parking lot which has a sign La Grange Bush Picnic Area - Indian Ladder Trail. The trail is open from May 1 to November 1, weather conditions permitting. Descend here for study of the Lower Devonian carbonate facies. Examine also the memorial plaque near the cliff edge at the Mine Lot Creek parking lot which has been attached to a vertical rockwall. It says "in memory of those pioneer geologists whose researches in the Helderbergs from 1819 to 1850 made this region classic ground." The names of these pioneers have been cited under History.

THE SEQUENCE STRATIGRAPHIC COLUMN OF THE HELDERBERG GROUP

The cliff face exposes an excellent case history of sequence stratigraphy. Lower Devonian limestone of the Helderberg Group reveal sets of parasequences which may be recognized among the exposed formations (Rondout, Manlius, and Coeymans formations) (Fig. 2). Parasequences are the building blocks of vertical sequences. A parasequence is defined as a relatively conformable succession of genetically related beds bounded by surfaces (called parasequence surfaces) of erosion, nondeposition, or their correlative conformities (Van Wagoner, 1985). Each sequence is initiated by a eustatic fall in sea level rapid enough to overcome subsidence (Van Wagoner, 1985) or by epeirogenic upward motion. A parasequence surface commonly is an unconformity surface.

Below the Indian Ladder Trail, where a waterfall known as Minelot Falls spouts across the path, sandstones and shales of the Middle Ordovician Schenectady Formation are mostly concealed beneath a cover of blocks of Devonian limestone forming a talus slope. At the waterfall, a major unconformity just below the trail separates the Ordovician strata from the Rondout Formation, exposed at the base of the cliff. The nonfossiliferous Rondout Formation is latest Silurian or earliest Devonian (Fisher, 1987). The rocks of this formation consist of brecciated dolostone cemented by gypsum. They display spectacular karst-generated solution-collapse features of the kind that hosts oil and gas in the subsurface. Evaporite minerals, now all dissolved, were present in this deposit. Note the concentration of springs which relates to the pore space created by ground-water dissolution of the evaporites.

Figure 3 shows the columnar stratigraphic section, the parasequence surfaces, and facies distribution of the Lower Devonian carbonates. The Rondout Formation at the base is overlain by the Manlius Formation, and the top of the section extending to the break in slope at the top of the cliff is occupied by the Coeymans Formation. The stratigraphic section exposed on this trail is the type locality for the Thacher Member of the Manlius Formation, proposed by Rickard (1962).

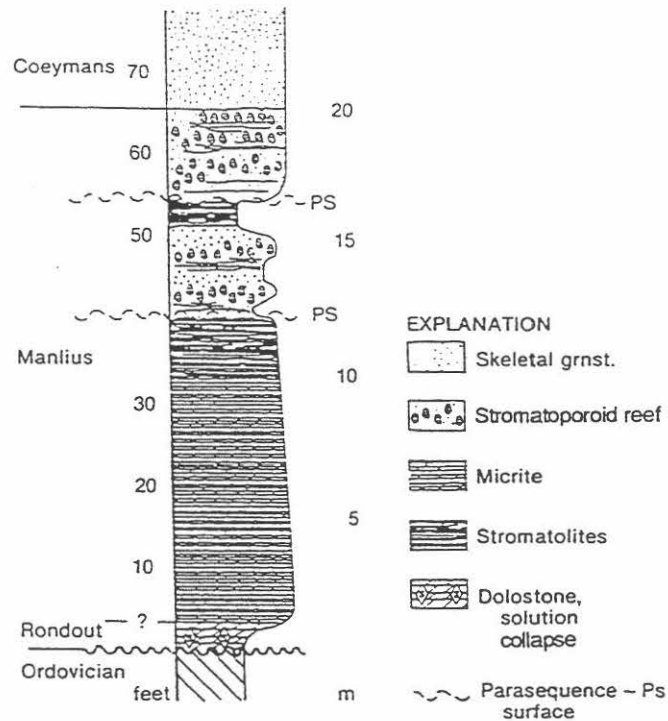
The Lower Devonian strata that crop out on the escarpment at John Boyd Thacher State Park illustrate the characteristics of marine parasequences in carbonate rocks (Figure 3). Two of the several parasequences are constituted as follows: a skeletal grainstone is overlain by stromatoporoid reefs and these, in turn, are overlain by interreef grainstone [interval between the upper PS (parasequence surface) and the scale mark for 20 m on Figure 3]. An underlying parasequence consists of skeletal grainstone that grades up into stromatolites (algal-laminated mudstone). Each of these two parasequences consists of strata that were formed when a depositional slope prograded seaward. The surfaces bounding the parasequences (labeled PS in Figure 3) are inferred to have resulted from rapid submergence. A set of several repeating parasequences, as shown in Figure 3, is known as a parasequence set, defined as "a relatively conformable succession of genetically related parasequences bounded by surfaces (called parasequence set boundaries) of erosion, non-deposition, or their correlative conformities."

Concepts identical to those just set forth, and developed independently of the definition of parasequence in seismic stratigraphy were formulated under the name of Punctuated Aggradational Cycles (PACs). What have been named PACs are thin (1-5 m) upward- shoaling cycles whose boundaries are defined by the depositional products of episodes of rapid submergence (Goodwin and Anderson, 1982).

Outcrop Guide

Studies of vertical sequences should normally be worked from the base of the section upward. However, at this exposure it is best to work the section downward following the stairway from the edge of the cliff.

Figure 3. Columnar stratigraphic section, parasequence surfaces, and facies distribution of the Lower Devonian carbonates. The Rondout Formation at the base is overlain by the Manlius Formation, and the top of the section extending to the break in slope at the top of the cliff is occupied by the Coeymans Formation. The stratigraphic section exposed on this trail is the type locality for the Thacher Member of the Manlius Formation, proposed by Rickard (1962). The Ordovician deposits at the base of the section are Indian Ladder beds-Schenectady Formation (see Fig. 2).



The top of the section is composed of skeletal grainstone (locally skeletal packstone) in which fossils, especially brachiopods, and crinoids, are evident (Fig. 3); the pentamerid *Gypidula coeymanensis* is prevalent. This facies is part of the Coeymans Formation. Its lower contact is sharp and obvious in the field. Below this contact is the Manlius Formation which underlies most of this escarpment. A stromatoporoid reef with locally intercalated skeletal grainstone represents the top of this formation (Fig. 3). The stromatoporoids show their distinctive globular concentric structures resembling cabbage heads. Previous authors (Fisher, 1987; Rickard, 1962) have termed this reef facies a biostrome, presumably because its geometry in outcrop is sheetlike rather than mound-shaped. In my experience with reefs of all ages, I have observed that most large reefs are flat on top and bottom, especially on the scale of this exposure. Other geologists share this experience, thus Shaver and Sunderman (1989) note "virtually all large reefs seen on outcrop have eroded, flattened tops, whereas smaller reefs that were not naturally aborted and that were unaffected by erosion as seen on outcrop have convex-upward rounded tops."

Close examination of the reef facies reveals a fine-grained matrix between the framework-building stromatoporoids. This matrix resembles micrite, a lithified lime mud; hence this facies may be misinterpreted as representing a low-energy setting. However, in modern reefs, cement forms millimeters to centimeters beneath the living part which, in thin section, is finely crystalline (cryptocrystalline) and semi-opaque. Hence the matrix in such reefrock looks just like low-energy micrite (Friedman et al., 1974). Case histories abound where unwary geologists have confused high-energy reefrock with a supposed "low-energy" lime-mud facies (Friedman, 1975). Therefore the observation of a fine-grained matrix between the framework builders does not deter, in fact confirms, the interpretation that this part of the section formed as a high-energy reef facies, and not in a low-energy setting.

The stromatoporoids are massive which in the ecologic zonation of Devonian reefs represents the shallowest-water zone of a subtidal setting.

Below the reef facies occurs a stromatolitic (finely laminated) facies which is recessed back creating a near cavelike morphologic feature (Fig. 4). This recessed feature can be traced throughout Thacher State Park and is known as "Upper Bear Path". By analogy with modern environments the stromatolitic facies represents a low-energy intertidal or supratidal setting. The sharp contact between the intertidal or supratidal low-energy stromatolitic facies and overlying subtidal high-energy reef facies represents a parasequence surface (Fig. 4). Downward from the stromatolites, a stromatoporoid reef facies is present, separated by bedded skeletal grainstone from the stromatolites; in fact, the reef facies is present twice (Fig. 5). Hence downward the setting changes from intertidal or supratidal to subtidal shallow water. Below this double-reef section, the change is again interpreted to be intertidal or supratidal stromatolites. Hence, once again, a parasequence surface separates the subtidal high-energy reef facies from the underlying intertidal to supratidal stromatolites (Fig. 3). Interestingly, this stromatolitic facies is resistant to erosion (Fig. 5), and it projects outwards in the cliff face, whereas the upper stromatolite facies is recessed almost cavelike. Below this lower stromatolite facies the lithology and facies are that of a low-energy, thin-bedded micrite with local

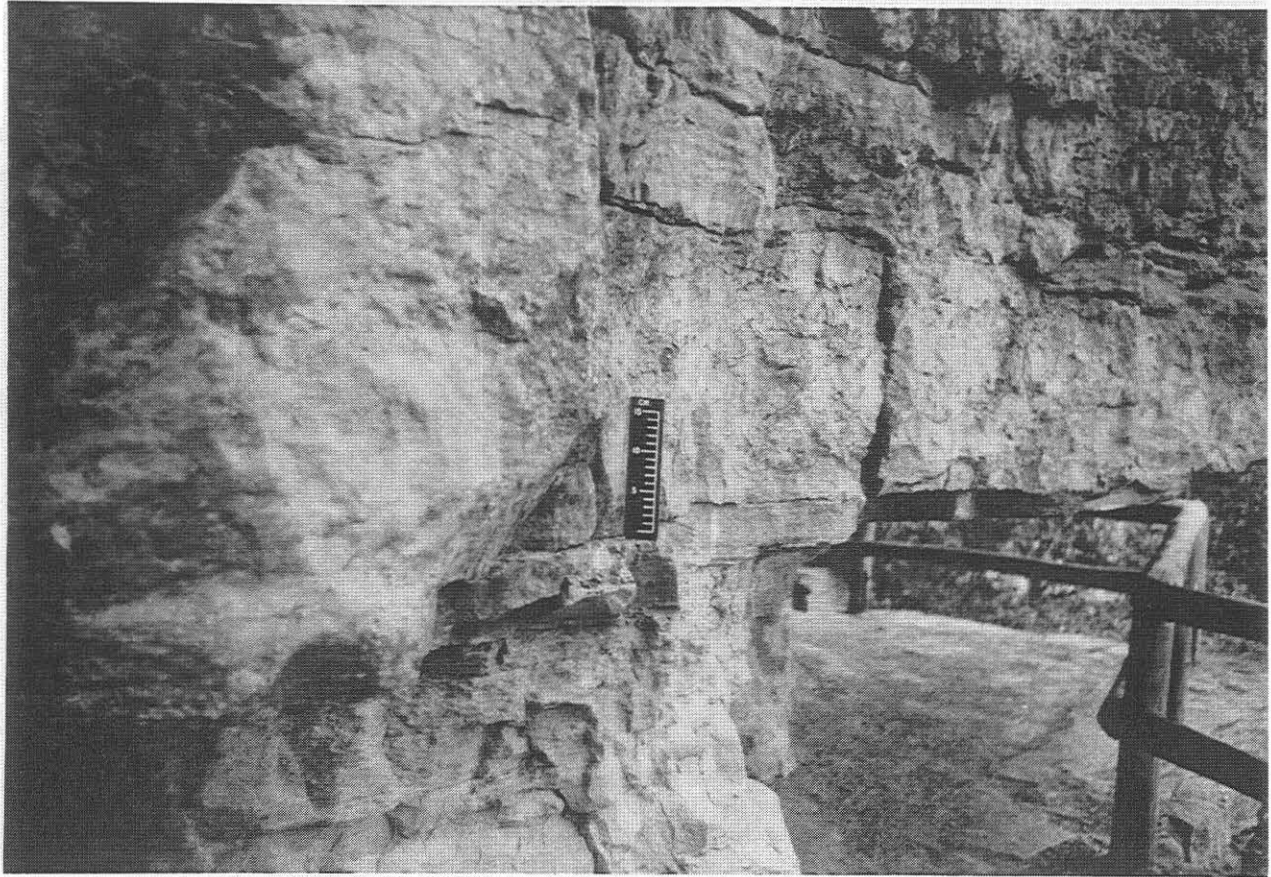


Figure 4. Photograph showing recessed underlying stromatolitic (finely-laminated) facies and overlying stromatoporoid reef facies of the Manlius Formation. Sharp contact between the two facies on which scale rests is parasequence surface (see Fig. 3).

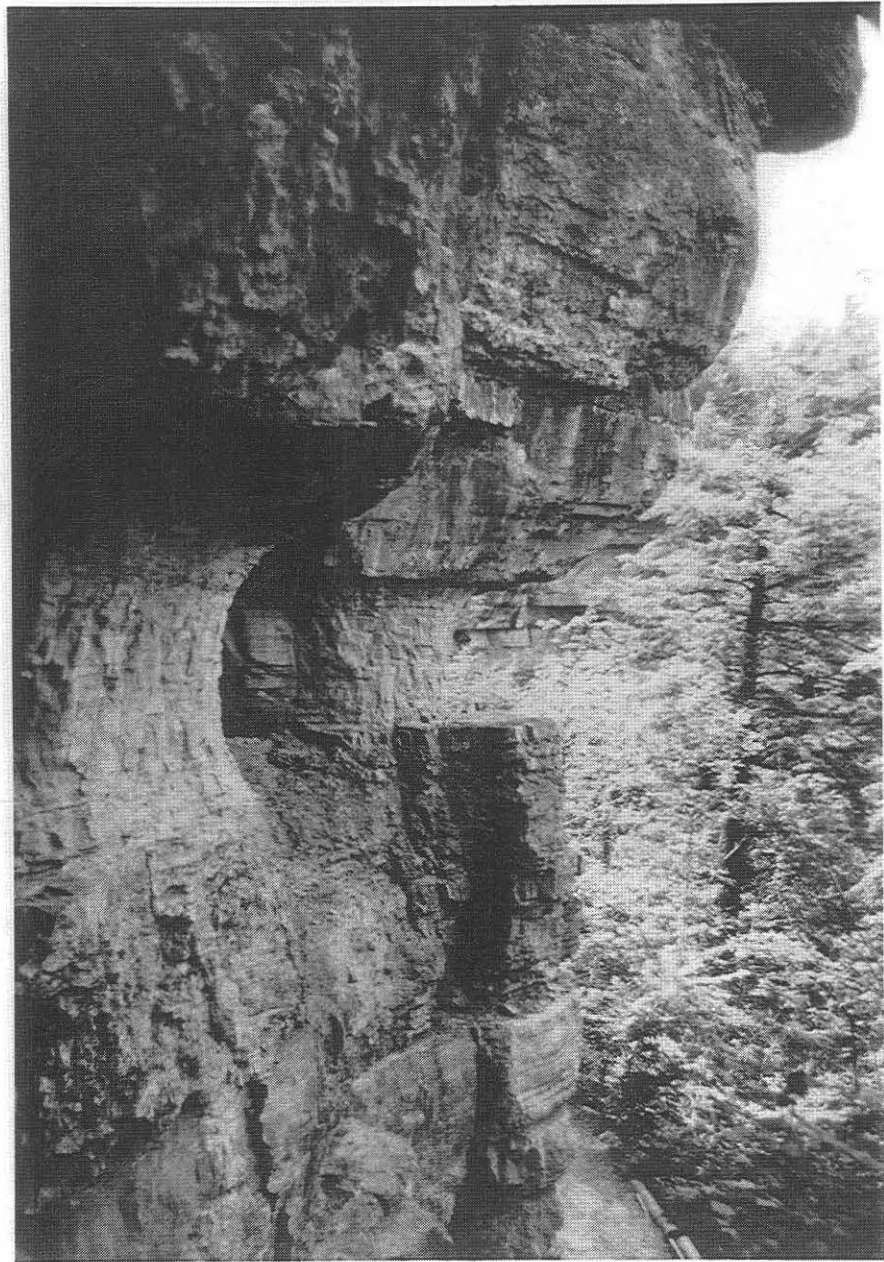
skeletal grainstone occurring as finely interbedded couplets, scour-and-fill structures, local cross-bedding, and some beds containing abundant spiriferid brachiopods, tentaculitids, ostracodes, and bryozoans. Near the base of the Manlius Formation occur several thicker beds, up to about 20 cm in thickness.

Near the base of the section is the Rondout Formations, a recessed zone at the base of the cliff characterized by brecciated carbonate rock cemented by gypsum. Its exact contact with the overlying Manlius Formation is subject to debate. In the columnar section (Fig. 3) the Rondout Formation is identified where solution-collapse features are prominent and the lithology changes to dolomitic, especially dolomitic stromatolites, with sporadic intercalated calcitic laminae and shale laminae, an interpreted supratidal facies. Clasts of solution-collapse breccia are prominent together with gypsum-filled veins. The angular clasts of the collapse breccia may have resulted from collapse and brecciation of overlying carbonate strata when evaporites underlying them were dissolved. It represents a karst setting. Springs and caves, which are present here, are a function of dissolution of evaporites by groundwater. Karst-type openings were created during subaerial emergence. The Rondout Formation is commonly known as Rondout Waterlime. Its base is at or below the trail.

THE ONONDAGA FORMATION

In John Boyd Thacher State Park the Onondaga Terrace (Fig. 2) exposes carbonate rocks of the Onondaga Formation which in the subsurface produces gas and serves as gas-storage reservoirs. Take Rt. 157 southwest to exit of park and continue to Indian Ledge Road (right turn-off before NY 85) (Fig.1). Turn right on to Indian Ledge Road and drive 0.7 mile, park on right of road just above uppermost limestone outcrop. The outcrop exposes the Edgecliff Member of the Onondaga Formation.

Figure 5. Photograph showing from top downward: massive projection reef separated by parasequence surface from underlying recessed stromatolites ("Upper Bear Path"). Below recessed stromatolite facies note in descending order bedded skeletal grainstone, nonbedded reef, bedded skeletal grainstone, reef, parasequence surface, and non-recessed stromatolite facies recognizable as a well-bedded facies. For detail compare with Figure 3.



This limestone is not reefal. It is a skeletal limestone and ranges in this exposure from micrite to skeletal grainstone. Crinoids and coral fragments are abundant. Of interest in this outcrop is the coarse-grained, moldic facies in mid-section. Dissolution of fossils has led to high-porosity facies of a kind that would make excellent reservoirs for oil or gas. Decide for yourself whether this high-porosity zone continues into the subsurface or is only a surface feature. This porous facies vies in its porosity with the best of reservoir facies in the subsurface. The accumulation of this coarse debris resulted from an episodic event, perhaps a storm or even a tsunami. Note the prominent erosional surface which underlies this deposit (Fig. 6). Lindemann (1979) described the biofacies of this exposure.



Figure 6. Photograph of exposure of Onondaga Formation, Indian Ledge Road, Onondaga Terrace (Fig. 2). Note prominent erosional truncation on mid-section. High-porosity moldic storm coquinite overlies surface of truncation.

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**LOWER AND MIDDLE DEVONIAN FORELAND BASIN FILL IN THE CATSKILL
FRONT: STRATIGRAPHIC SYNTHESIS, SEQUENCE STRATIGRAPHY, AND THE
ACADIAN OROGENY**

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INTRODUCTION

Modern stratigraphic study has come to be increasingly based on high resolution subdivision of the sedimentary record through detailed event and cyclic analysis. The recognition and correlation of unique single event horizons (e.g., altered volcanic ashes, shell marker beds, faunal epiboles and immigration events) and genetically-related depositional sequences permits microstratigraphic subdivision of the rock record at a finer scale than is possible through classical methods of bio- and lithostratigraphy. "Layer cake" stratigraphy, once the bane of stratigraphic thought, has again emerged in modified form through the application of "Event" (Kauffman, 1988) and "Sequence Stratigraphic" (Van Wagoner et al., 1988) methods.

The construction of a refined microstratigraphic framework of strata permits detailed analysis of the timing and nature of events in the rock record. Short term dynamics in the history of a basin are recorded by: sharp transitions between intrabasinal and extrabasinal sedimentation; clustered volcanogenic strata; condensed beds and minor unconformities; abrupt changes in faunas and short-term incursions of exotic faunas; cyclic sedimentation; and eustatic- and tectonic-controlled changes in relative sea level. A detailed, correlatable framework of large and small scale events helps unravel the history of a sedimentary basin.

The late Early and Middle Devonian was a very dynamic time on the margin of the Eastern Interior of North America. A continent-continent collision, subsequent subsidence and infilling of an adjacent sedimentary foreland basin, and the evolution and migration of major faunas are recorded in the rocks of the Catskill Front in eastern New York State. The interaction of tectonically-induced flexure of the foreland basin and changes in eustatic sea level had a profound influence on the Northern Appalachian Basin at that time. Alternating tectonically active and quiescent stages of the Acadian Orogen (Acadian "Tectophases" of Effensohn, 1985) resulted in major periodic alterations of subsidence and uplift, sedimentary style, faunas, and paleoecology. Superimposed over the tectonic influences are apparent changes in eustatic sea level (e.g., Johnson et al., 1985). As a result, the upper Lower and Middle Devonian foreland basin fill of New York is a complex succession of tabular to distinctly wedge-shaped bodies of carbonates and siliciclastics, shifting depocenters and basin centers, and marked unconformities.

This paper and field trip will examine the sedimentary record of the upper Lower and Middle Devonian rocks in the Catskill Front (see Figure 1). In addition to a review of member- and formation-level stratigraphy, the authors will present: 1) important new details and interpretations of strata of the Tristates Group, the Onondaga Formation, and the lower part of the Hamilton Group; 2) a discussion of the major and minor unconformities within the study interval; 3) the sequence stratigraphic framework for the upper Lower and Middle Devonian in New York State; and 4) stratigraphic trends and their implications for the evolution of the Northern Appalachian Basin and the Acadian Orogeny.

The Appalachian Basin and the Acadian Orogeny

The Appalachian Basin was a retroarc foreland basin that formed on the periphery of the eastern interior of North America associated with late Early to Late Paleozoic convergent margin tectonism (Quinlan and Beaumont, 1984). Collision of the Cambrian-Ordovician North American passive margin with an island arc in the Late Ordovician (Taconic Orogeny) resulted in formation of the Appalachian foreland basin. Relatively low-level tectonism throughout most of the middle to late Paleozoic Era was punctuated by large scale events, which include continent-continent collisions during the Devonian (Acadian Orogeny)

In Garver, J.I., and Smith, J.A. (editors), *Field Trips for the 67th annual meeting of the New York State Geological Association*, Union College, Schenectady NY, 1995, p. 313-356.

and Pennsylvanian (Alleghenian Orogeny). In each of these latter cases the Appalachian Basin was reorganized into a rapidly subsiding retroarc foreland basin into which synorogenic sediments were shed.

The Acadian Orogeny was the most significant event in the Devonian of eastern North America (Osberg et al., 1989; Roy and Skehan, 1993; Fail, 1985; Bradley, 1983). Collision of the North American landmass with a microcontinent or a series of terranes (Rast and Skehan, 1993; i.e., Avalon terrane) resulted in the formation of a large-scale mountain belt (referred to here as the Acadian Orogen) and renewed subsidence of the cratonward Appalachian foreland basin. Formation of a magmatic arc along the eastern margin of North America, best preserved in New England and Maritime Canada, was associated with major volcanism and plutonism, metamorphism, and deformation. Apparent southwestward migration of the orogeny through time (Rogers, 1967) is thought to have been associated with oblique convergence of Avalon with separate promontories on the eastern margin of North America with possible transpressional fault movement along a major system of dextral strike-slip faults (Williams and Hatcher, 1982; Ettensohn, 1985, 1987; Ferrill and Thomas, 1988). Alternating pulses of siliciclastic- and carbonate-dominated sedimentation in the adjacent Appalachian foreland basin indicate three to four active phases of Acadian tectonism occurred during the late Early Devonian to Early Mississippian (Tectophases I-IV of Ettensohn, 1985). Pulses of active tectonism, marked by overdeepening of the foreland basin and deposition of transgressive dark gray to black shales and volcanoclastic strata, alternate with periods of relative quiescence that are associated with regression, clastic progradation, and/or a return to more carbonate-dominated deposition.

Ettensohn (1985, 1987) proposed a model for the evolution of the Acadian Orogeny in eastern North America. Based on the stratigraphic record of the Appalachian foreland basin-fill, he projected three to four phases of active tectonism in the late Early Devonian to Early Mississippian associated with oblique convergence of the eastern margin of North America and a landmass termed Avalon. Successive collisions of Avalon with separate promontories in Quebec, New York, Virginia, and Alabama resulted in tectonic rejuvenation and the onset of separate "tectophases." Each of Ettensohn's tectophases is comprised of a progression from stages of active tectonism to quiescence, recorded in the rock strata by a succession of clastic- to carbonate-dominated deposition.

STRATIGRAPHY

Carbonate-dominated strata of the uppermost Silurian and lower part of the Lower Devonian (Rondout Formation and Helderberg Group respectively) are succeeded in eastern New York by a thick and complex interval of upper Lower and Middle Devonian siliciclastic and carbonate rocks. Strata of the Tristates Group (upper Lower Devonian) and the Onondaga Formation and Hamilton Group (Middle Devonian) are greater than a kilometer thick in the Catskill Front (see Figure 2) but thin to less than 25 m in parts of central New York. These rocks were subdivided by Rickard (1975) into five stages. These stages in ascending stratigraphic order are the Deerpark, Sawkill, Southwood, Cazenovian, and Tioghnogian Stages; the Sawkill and Southwood stages are referred to as the "Onesquethaw Stage" by some authors (e.g., Dennison, 1961, 1962; Inners, 1976). The overall vertical succession of rock consists of: (1) quartz arenites and carbonates (Oriskany, Glenerie, and Connelly Fms.), overlain by (2) siliciclastics (Esopus Formation), and mixed (3) siliciclastics and carbonates (Schoharie Formation). Overlying limestone of the Onondaga Formation is succeeded by thick, fine- to medium-grained, marine to non-marine clastic rocks of the Union Springs, Mount Marion, Ashokan, Plattekill, and overlying formations. The stratigraphy, lithology, biostratigraphy and fauna, and distribution of the individual units are discussed below.

Figure 1. Physiographic map and cross-section of the Hudson Valley and Catskill Front, showing bedrock and relation to topography. Cross section at the southern boundaries of the Catskill and Kaaterskill quadrangles (Ulster Co.). Stratigraphic terminology of Figure 1 after Chadwick (1944); changes (from Rickard, 1975) include: Catskill Shaly=New Scotland; Kiskatom=Plattekill and Manorkill; Onteora=Oneonta; and Katsberg=Walton Formations. Onandaga=Onondaga. Note that the Mount Marion Formation includes most of the strata shown as Ashokan Formation. Original drawing by Alan McKnight.

Tristates Group: Deerpark Stage (Lower Devonian)

Oriskany Formation. The Oriskany Formation (Vanuxem, 1839) is the basal transgressive sandstone of Sloss' (1963) Kaskaskia Supersequence York. It overlies the Wallbridge Unconformity (Sloss, 1963), one of six major unconformities in the Phanerozoic of North America. The unit consists dominantly of mature quartz arenite that was deposited under relatively widespread, shallow-marine conditions across much of eastern North America (Boucot and Johnson, 1967). These shoreface to nearshore sandstones are characterized by abundant large, robust brachiopods ("Big-Shell Community" of Boucot and Johnson, 1967). The Oriskany Formation in New York ranges from 0 to < 3 m along the New York outcrop belt (Hodgson, 1970; up to ~30 m-thick in subsurface of Tioga Co., N.Y.; Rickard, 1989), and is up to 100 m-thick to the south in parts of Pennsylvania, Maryland, and West Virginia.

The Oriskany Sandstone sharply and unconformably overlies older Lower Devonian or Upper Silurian rocks along its outcrop belt in New York. The formation is thin in eastern New York and thin to absent in central New York, and is generally absent in the western part of the state. Locally, quartz sands of the Oriskany Formation fills fissures and karstic cavities in underlying carbonates (Brett and Ver Straeten, 1994). In eastern New York, the unit laterally grades into the Glenerie Limestone and Connelly Conglomerate (see below). The Oriskany Formation and equivalent strata are assigned to the *Icriodus huddlei* conodont zone and the *Costispirifer arenosus* subzone of the *Rensselaeria* brachiopod range zone (Lower Devonian; Klapper, 1981; Dutro, 1981). Previous work on the Oriskany Formation includes Hodgson (1970), Eaton (1921), and Oliver and Hecht (1994).

Glenerie Formation. The Glenerie Formation (Chadwick, 1908) represents limestone-dominated facies in eastern New York laterally equivalent to the Oriskany Sandstone. Siliceous to cherty limestones characterize the formation along its outcrop from southern Greene County (Catskill area; ~3 m-thick) southward through the Hudson Valley and to the Tristates area of New York, New Jersey, and Pennsylvania (~45 m-thick; Rickard, 1989). The formation extends in the subsurface across much of southeastern New York and northeastern Pennsylvania (Rickard 1989). In southern Greene County the base of the formation is marked by a phosphate pebble-rich lag immediately above the Wallbridge Unconformity. Southward (e.g., Kingston area), the basal-lag bed becomes quartz pebble-rich as the lower part of the Glenerie Limestone interfingers with the Connelly Conglomerate. In the Tristates area, the Glenerie Formation conformably overlies limestones of the Port Jervis Formation (Rickard, 1975).

The Glenerie Formation is generally richly fossiliferous; for example, ~100 species have been recorded from the Kingston area (Van Ingen and Clark, 1903), and Clarke (1900) reported 113 forms from a Devonian outlier across the Hudson River from Catskill. The fossils, which are commonly silicified, range from large, robust, Oriskany Sandstone-type brachiopods, to small, delicate forms that lived in more offshore facies represented by the Glenerie Limestone. Important references on the Glenerie Formation include Clarke (1900), Chadwick (1944), and Hodgson (1970).

Connelly Formation. The Connelly Formation (Chadwick, 1908) comprises quartz pebble-rich conglomerates in southeastern New York that are laterally equivalent to the Oriskany Sandstone. The unit crops out in the Kingston area (Ulster Co.), where it interfingers with lower strata of the Glenerie Limestone. The Connelly Conglomerate also occurs in the Skunnemunk Outlier, ~35 km southeast of the main outcrop belt in southeastern New York (Orange County; see below). In the outlier, the formation composes the entirety of the Oriskany interval, and has a maximum known thickness of 14 m (Boucot et al., 1970). Along the main outcrop belt in Ulster County, the unit is reported to range from 6.3 m (near Bloomington; Hodgson 1970) to a feather edge north of Kingston; in that area it unconformably overlies limestones of the Port Ewen Formation (Lower Devonian Helderberg Group). The fauna of the Connelly Formation is characterized by typical robust Oriskany brachiopods (Boucot, 1959). Previous studies of the Connelly Formation include Boucot (1959), Boucot et al. (1970), and most notably, Hodgson (1970).

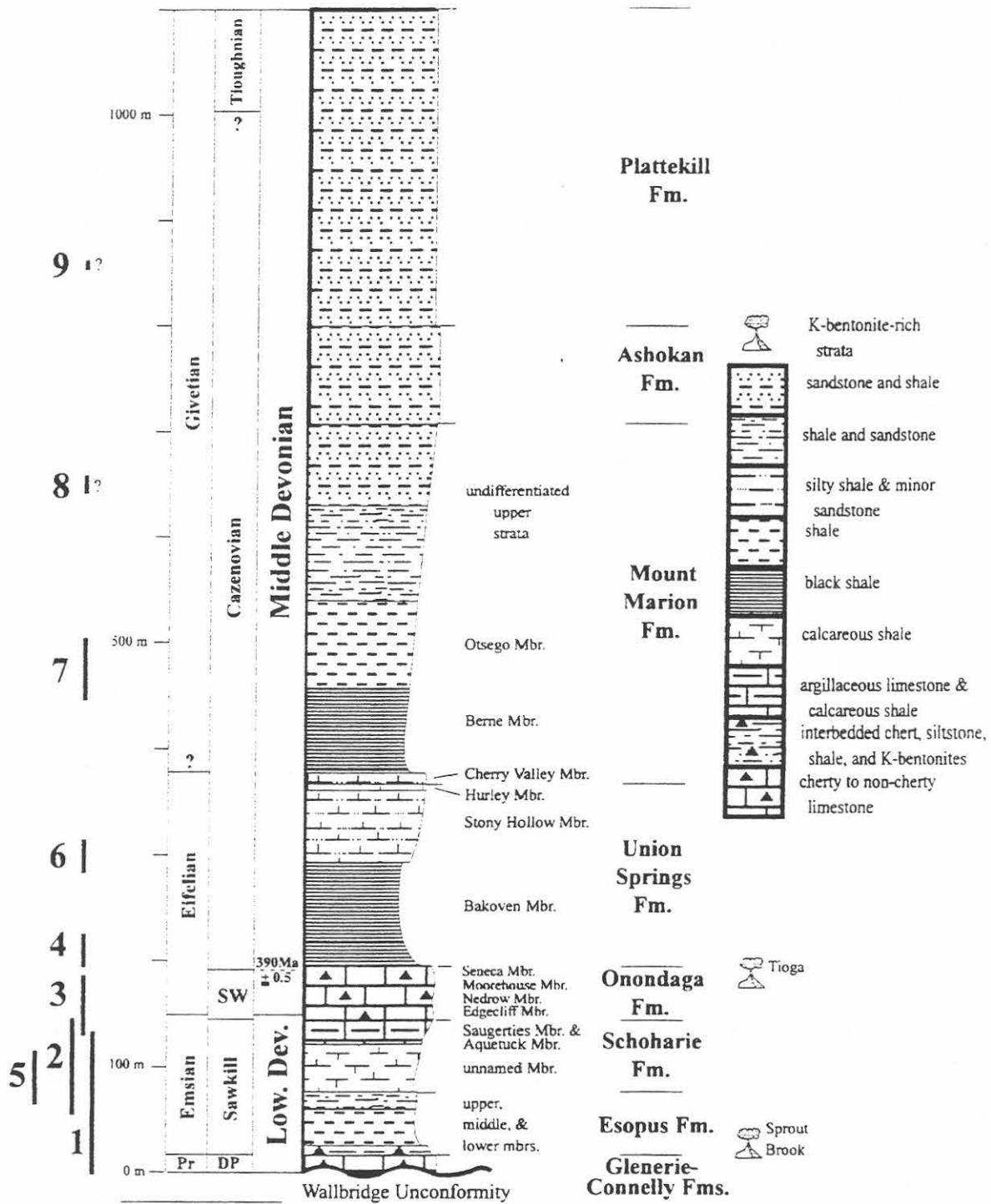


Figure 2. Composite section of upper Lower and Middle Devonian rocks in the vicinity of Kingston, N.Y. Bars on left side of diagram indicated intervals exposed at field trip stops 1-9. Absolute age of the "Tioga B" K-bentonite from Roden et al. (1990). Compiled from Ver Straeten (field notes), Rickard (1989), Hodgson (1970), Rehmer (1976), Feldman (1985), and Fletcher (1967).

Tristates Group: Sawkill Stage (Lower Devonian)

Esopus Formation. The Esopus Formation (Darton 1894; previously termed the "Cauda-galli Grit;" Vanuxem, 1842) of eastern New York, northern New Jersey, and eastern Pennsylvania is dominantly composed of dark gray shales and siltstones with lesser amounts of fine sandstone, chert, and K-bentonites. The formation crops out from east-central NY (Herkimer County) to eastern New York and southwestward into eastern Pennsylvania (Carbon Co.). The Esopus Shale is 0-100 m-thick along the outcrop belt (Rehmer, 1976), and is over 120 m in the subsurface (New York-Pennsylvania border area; Rickard 1989). Boucot et al. (1970) and Marintsch and Finks (1982) report as much as 185 m of Esopus in the Skunnemunk Outlier in southeastern New York (however, see Skunnemunk Outlier section below).

The Esopus Formation is generally unfossiliferous. Except for the Skunnemunk Outlier (see Boucot 1959; Boucot et al. 1970), the unit is typically barren, although it locally has a low-diversity macrofauna characterized by the brachiopod *Atlanticocoelia* (new genus, Koch in press; =*Pacificocoelia* of Boucot and Rehmer, 1977). The Esopus Formation is commonly extensively bioturbated, with common *Zoophycos* and/or *Chondrites* traces; Marintsch and Finks (1982) report over 16 different trace fossils from the Skunnemunk Outlier.

The biostratigraphy of the Esopus Formation is generally poorly documented; it is assigned to the *Etymothyris* brachiopod range zone (Lower Devonian; Dutro, 1981). Key references for the Esopus Formation include Rehmer (1976), Fenner (1971), Fenner and Hagner (1967), Boucot (1959), Boucot et al. (1970), Marintsch and Finks (1978, 1982).

Informal member-level stratigraphic schemes for the Esopus Formation have been proposed by several workers; Fenner and Hagner (1967), Fenner (1971), and Rehmer (1976) reported three subdivisions of the formation in the main outcrop belt. Boucot et al. (1970) proposed four members for the Esopus Formation in the Skunnemunk Outlier in Orange County. In each case the members were poorly delineated, most notably in the main outcrop belt. The authors herein propose a more rigorously defined member-level stratigraphy of three members for the Esopus Formation (main outcrop belt), informally termed the "lower, middle, and upper members."

The lower member of the Esopus Formation overlies the coeval Oriskany, Glenerie, and Connelly Formation along the New York State outcrop belt except in westernmost outcrops. Three subdivisions are recognizable in the lower member along the outcrop from Kingston to Cherry Valley: 1) a lower interval of interbedded siltstones, impure cherts, thin dark shales, and altered volcanic ash beds of the Sprout Brook Bentonites (see below); 2) a middle subunit of dark gray-black shales that commonly features large calcareo-phosphatic concretions; and 3) an upper siltstone to sandstone, generally 1-2 m-thick (to as thin as 0.1 m in western outcrops), with some fossils (including *Atlanticocoelia* and *Zoophycos*). Scattered glauconite has been noted in the upper subunit of the lower member.

These three subdivisions of the lower member of the Esopus are correlatable all along the main outcrop belt in New York (e.g., Kingston to Cherry Valley). Outcrops in the Hudson Valley range from 6-12.5 m in thickness. Along Rt. 23A near Catskill (Stop 2 of this fieldtrip) the lower member is 12.5 m-thick. The member thins in the northern Helderberg area, but thickens to the west. At Cherry Valley, the lower member (~4.3 m-thick) is represented by 3.6 m of interbedded siltstones, cherts, and k-bentonites that are overlain by rusty-weathering silty shales with scattered phosphatic grains (0.6 m). The upper siltstone-sandstone submember is represented by a thin (~10 cm-thick) siltstone bed. Overlying dark gray to purple and olive green shales and mudstones (1.7 m) at Cherry Valley represent lower strata of the middle member of the Esopus Formation.

The base of the middle member of the Esopus Shale is generally characterized by fine-grained, dark-gray to black shales and mudstones. The member grades upward through silty shales to a coarser cap of *Zoophycos*-churned argillaceous siltstones that immediately underlie the base of the upper member of the formation. Subtle-weathering dark and light bands can be noted on some outcrops; in general, however, the middle member appears massive and uniform.

Several workers including the authors (e.g., Inners 1975) have noted a distinctive interval of thinly laminated strata (>5m-thick) toward the upper part of the Esopus Formation between Kingston and Catskill. Near Catskill, the lower part of the interval consists of dark gray to black, rusty shales that pass upward into finely laminated ("pin-striped") shale-siltstone couplets. The upper part of the pinstriped unit in the Catskill area tends to be slightly coarser, and the thin siltstone to fine sandstone beds thicken and show planar to

low-angle cross laminae. The upper part of the laminated unit in the Catskill area is also siliceous and resistant to weathering.

The stratigraphically highest rocks of the Esopus Formation consists of massive, gray to dark brownish gray siltstones to sandstones with pervasive, well-preserved *Zoophycos* spreiten. This unit abruptly overlies the laminated unit. The authors propose to place the base of the upper member at the bottom of the rusty shales in the lower part of the laminated unit, immediately above the previously noted *Zoophycos*-churned siltstone of the middle member. This places the base of each of the three members of the Esopus at a relatively abrupt transition to finer-grained strata (i.e., base of a sedimentary cycle) and, importantly, defines contacts that are distinct and recognizable in the field.

The authors' recent studies show that the laminated unit changes position with respect to the overlying contact with the Schoharie Formation. At Kingston (O'Reilly Street railroad cut, Stop #3 of Oliver et al., 1962) the top of the laminated unit lies 22.5 m below this contact. Thirty five km to the north, however, at Catskill (Rt. 23, Stop 1B of this fieldtrip) the laminated unit is directly overlain by the Schoharie Formation and the coarse, *Zoophycos*-churned sandstones that comprise the upper part of the upper member are absent. Twelve km farther north, near Climax, the laminated unit is also absent; basal strata of the Schoharie Formation overlie the *Zoophycos*-churned siltstone cap of the middle member of the Esopus Formation.

To the north and west across northern Greene County and Albany and Schoharie Counties the upper siltstone unit of the middle member forms the caprock of the Esopus Formation. At Cherry Valley (Otsego Co.; ~110 km along the outcrop belt) the upper siltstone unit is absent, and only the lower 1.7 m of the middle member remains. The middle and upper members are both absent west of Cherry Valley, and the Esopus Formation is represented by the lower member only. Twenty five km west of Cherry Valley (southwestern Herkimer Co.) the Esopus Formation pinches out (Rehmer, 1976).

Schoharie Formation. The Schoharie Formation (Vanuxem, 1840) comprises strata in eastern New York between the Esopus Shale and the overlying Onondaga Limestone. The Schoharie Formation is distributed across east-central to eastern New York, northwestern New Jersey, and eastern Pennsylvania. In eastern New York the strata generally consist of gray- to buff- weathering calcareous mudstones to argillaceous limestones; carbonate content increases upward through the formation. Three members are recognized in the Hudson Valley: 1) a lower unit of gray-weathering, bioturbated, calcareous silty mudstone to argillaceous limestone ("unnamed member" of this report = Carlisle Center Member of Johnsen, 1957, and Carlisle Center Formation of Rickard, 1975; see discussion below); 2) a middle unit of calcareous shale and nodular limestone, locally cherty/siliceous (Aquetuck Member); and 3) interbedded, buff- and cream-weathering, calcareous shales and limestones (Saugerties Member) that in the Hudson Valley grade upward into the Onondaga Limestone. A fourth member, recognized to the north in the Albany to Schoharie region, is composed of calcareous, quartz-rich sandstones (Rickard Hill Member). West of the Albany region undifferentiated Schoharie strata are herein assigned to the Carlisle Center Member (restricted; see below). Schoharie-equivalent strata in western New York are assigned to the Bois Blanc Formation; locally occurring, apparently equivalent strata in central New York have been termed the "Springvale Sandstone" (Baker, 1983; see Brett and Ver Straeten, 1994).

In outcrop, the Schoharie Formation and equivalent units range from zero to ~75 m between east-central New York and the Tristates area (Rickard, 1989); over 120 m of strata are reported from the subsurface of northeastern Pennsylvania (Rickard, 1989). The lower contact is generally abrupt, and is locally marked by glauconite, phosphate, and scattered quartz pebbles. The lower members exhibit low diversity/low density faunas that generally consist of ostracods (Fenner, 1971), small brachiopods, and dacroconariids; upper strata of the formation are increasingly fossiliferous, especially the sandstones of the Rickard Hill Member.

The Schoharie Formation is assigned to the *Eodevonaria arcuata* subzone of the *Amphigenia* brachiopod assemblage zone, the *Aemulophyllum exigum* coral assemblage zone, and questionably to the *serotimus* conodont zone (Lower Devonian; Dutro, 1981; Oliver and Sorauf, 1981; Klapper, 1981). References on the Schoharie Formation in New York include Johnsen (1957), Oliver et al. (1962), Inners (1975), and Goldring and Flower (1942).

An important aspect to the authors' new interpretations of the Esopus and Schoharie Formations is the recognition of a relatively cryptic, but substantial unconformity between the two units along the New

York outcrop belt (see below). This unconformity, which we term the "Sub-Schoharie unconformity" (= "Pre-Upper Tristates unconformity" of Brett and Ver Straeten, 1994), was noted by previous workers as one of several glauconite- and phosphate-rich discontinuities in the lower part of the Schoharie Formation (Johnsen, 1957; Oliver et al., 1962). The authors herein informally propose a modification of the member-level terminology of Johnsen (1957) as outlined above.

Preliminary work on the Schoharie Formation in eastern to east-central New York suggests that strata previously termed the "Carlisle Center Member" in the Hudson Valley is not laterally equivalent to the Carlisle Center Member in its type area to the northwest (western Schoharie and Otsego Counties). Recent field study and recognition of the previously mentioned "Sub-Schoharie unconformity" suggests that the type Carlisle Center Member is equivalent to parts of the unnamed member, Aquetuck, and Saugerties Members of the Hudson Valley outcrop belt.

A part of Johnsen's (1957) reasoning for correlating strata of the type Carlisle Center Member with lower strata of the Schoharie Formation in the Hudson Valley was the apparent disappearance of the Aquetuck and Saugerties Members in the Albany area and west. Along NY Rt. 85 near Clarksville (Albany County) Johnsen (1957, p. 31-32) reported 6.1 m of "Carlisle Center" strata overlain by 0.9 m of richly fossiliferous, sand-dominated Rickard Hill Member. Restudy of the Rt. 85 outcrop, however, indicates that the lower unnamed member is present, but is much thinner than reported by Johnsen (1957). The Schoharie Formation there consists of: 1) 1.3 m of buff gray-weathering, bioturbated, calcareous, argillaceous siltstone (base of the member is covered at road level). A zone of chert nodules occur in the lower part of the unit, and scattered white quartz pebbles are found in the upper 0.5 m; 2) 4.7 m of increasingly calcareous silty to limestone-rich strata. The lower part of this unit (~2.6 m) consists of interbedded thin, silty, calcareous shales and bedded to nodular and knobby-bedded silty to argillaceous limestones. The strata are notably siliceous, and chert nodules occur in some beds. Strata above a 0.6 m-thick covered interval are characterized by interbedded calcareous shales and calcareous siltstone to limestone beds.; and 3) 0.86 m of richly fossiliferous calcareous quartz arenites of the Rickard Hill Member. The present authors interpret Unit 1 to be the lateral equivalent of the unnamed lower member of the Schoharie Formation of the Hudson Valley; the chert noted above may correlate with a distinctive black cherty bed reported from the unnamed member in the Hudson Valley (see below). Similarly, the white quartz pebbles appear to correlate with similar occurrences in the upper part of the unnamed member in the Catskill to Kingston region. The two subdivisions of Unit 2 appear to correspond to the Aquetuck and part of the Saugerties Members as they are developed in the Catskill vicinity and northward. Upper strata of the Saugerties Member is equivalent to the calcareous sandstone unit (Unit 3, Rickard Hill Member) that caps the Schoharie Formation in the Albany-Schoharie region.

The lower, unnamed member of the Schoharie Formation in the Hudson Valley is characterized by bioturbated, calcareous, silty shales and argillaceous siltstones. Basal strata of the unnamed member above the sub-Schoharie unconformity commonly feature abundant glauconite, phosphate pebbles, and scattered white quartz pebbles (see "unconformities" discussion below). A prominent dark siliceous band toward the center of the member ("black bed" of Oliver et al., 1962; ~0.6-1.0 m-thick) is correlatable throughout the Catskill to Kingston area, and may be recognized as far as Stroudsburg in eastern Pennsylvania (Inners, 1975). The unnamed member thickens southward through the Hudson Valley. At Catskill (Rt. 23, Stop 1 of this fieldtrip) it is 5.0 m-thick, but it is ~31.5 m-thick along N.Y. Rtes. 199/209 at Kingston (Stop 5 of this fieldtrip) and 43.75 m-thick in southwest Kingston, (O'Reilly Street railroad cut=Stop #3 of Oliver et al., 1962). The anomalous change in thickness toward the Kingston area is at present unknown, but may be associated with redeposition of upper Esopus sediments eroded from the north and northwest below the sub-Schoharie unconformity.

In the Catskill to Albany area, the overlying Aquetuck Member is characterized by fine-grained, dark gray- to buff-weathering cherty to siliceous shales with lesser carbonate. To the south, however, the siliceous aspect of the member disappears and the strata are more calcareous. One distinctive part of the member at Catskill is an interval of dark gray to buff siliceous shales 2.5-5.1 m above the base of the Aquetuck Member. These strata are represented by blocky-weathering, calcareous shales at Kingston (e.g., Rte 199-209) ~3-6 m above the base. This shaly interval in the lower part of the Aquetuck appears to be widespread, as it can be clearly distinguished in the middle of the Needmore Formation (submember 3 of the Hares Valley Member of Ver Straeten and Brett, in prep.) in central Pennsylvania. The same interval of dark

gray to black shales is locally reported from the Needmore Formation of Virginia and West Virginia (upper tongue of Beaverdam black shale facies of Dennison, 1960, 1961).

West of the Albany area, strata equivalent to the Schoharie Formation are increasingly silt- to sand-dominated, as seen in the type Carlisle Center Member in east-central New York. At Cherry Valley (Otsego Co.), 5.9 m of the Esopus Formation (lower and lowest part of the middle members) are overlain by 13 m of *Zoophycos*-churned, glauconitic, medium-bedded to massive, quartz-rich calcareous siltstones of the Carlisle Center Member. The basal contact is very sharp and is marked by well-defined trace fossils (Miller and Rehmer, 1980; see below). The upper contact is placed at a deeply weathered, glauconite- and phosphate-rich clay layer interpreted to mark a pre-Edgecliff unconformity (Wolosz et al., 1991, p. 382-383; numerous abraided- and pristine-appearing zircons found in the clay layer by the present authors indicate the bed may be an impure K-bentonite layer). As noted above, the type Carlisle Center Member appears to be equivalent to parts of the unnamed member, Aquetuck, and Saugerties Members of the Hudson Valley outcrop belt.

Southwood Stage (Middle Devonian)

Onondaga Formation. The Onondaga Formation (originally “Corniferous Limestone” of Eaton, 1828) in eastern New York ranges from coarse crinoidal grainstones to fine-grained micritic limestones, with abundant chert and locally developed reef facies. Equivalent limestone-dominated facies occur widely across eastern North America, from the James Bay region of northern Ontario to southeastern Quebec and Maine to the Virginias to the Illinois Basin (Koch, 1981). Across New York, the Onondaga Limestone consists of four members (in ascending order): the Edgecliff, Nedrow, Moorehouse, and Seneca Members (Oliver, 1954, 1956). A fifth subdivision, the former Clarence Member (designation abandoned), is now recognized as a cherty facies of the Edgecliff Member (Brett and Ver Straeten, 1994). The lower subdivision, the Edgecliff Member is characterized by coarse crinoidal- and coral-rich to finer grained, chert-rich facies (Jamesville Quarry facies and Clarence facies, respectively; Brett and Ver Straeten, 1994). Biostromal to biohermal facies, including pinnacle reefs in the subsurface, are rooted in the lower part of the Edgecliff Member. In its type area, south of Syracuse, the overlying Nedrow Member is represented by calcareous shales and argillaceous limestones. In eastern New York, however, the Nedrow Member is characterized by argillaceous to fine-grained, generally non-cherty limestones that may be difficult to distinguish from underlying and overlying units. The third member (Moorehouse Member) is typified by chert-rich, fine to medium-grained limestones throughout New York State. The uppermost unit, the Seneca Member, features finer-grained, generally non-cherty limestones with thin shales and k-bentonite layers of the Tioga Bentonites cluster (see below). The Seneca Member is generally thin to absent in eastern New York (Rickard, 1975, 1989).

The Onondaga Limestone ranges in thickness from ~18-60 m in thickness across the outcrop belt in New York State; in the subsurface it may be as thin as 6 m (Rickard, 1989). The lower contact is disconformable across central to western New York, but is gradational with the underlying Schoharie Formation in the Hudson Valley outcrop belt in eastern New York.

The Middle Devonian Onondaga Formation is characterized by a diverse fauna dominated by brachiopods, bryozoans, corals, and echinoderm fragments, with additional trilobites, gastropods, and dacroconariids. The lower part of the formation (Edgecliff Member) has abundant corals and echinoderms with local coral bioherms.

Biostratigraphically, Nedrow and questionably the Edgecliff Members are assigned to the *Polygnathus costatus patulus* conodont zone and the Moorehouse and Seneca Members to the *Polygnathus costatus costatus* zone (Klapper, 1981). The Edgecliff, Nedrow, and lower to middle parts of the Moorehouse Members occur within the *Fimbrispirifer divaricatus* subzone of the *Amphigenia* brachiopod assemblage zone (Dutro, 1981). The Nedrow Member in New York features the goniatite *Foordites buttsi* (House 1978, 1981). Oliver and Sorauf (1981) assign the Edgecliff Member to the *Synaptophyllum arundinaceum* subzone and the Nedrow and Moorehouse Members to the *Eridophyllum seriale* subzone of the *Acinophyllum segregatum* coral zone. The Seneca Member is assigned to the *Paraspirifer acuminatus* brachiopod assemblage zone and an unnamed coral zone (Dutro, 1981; Oliver and Sorauf, 1981). The Lower-Middle Devonian (Emsian-Eifelian) boundary is presently placed at or near the base of the Edgecliff Member, although, due to poor conodont control, it could occur as high as the base of the overlying Nedrow Member (Kirchgasser and Oliver, 1993).

The Onondaga Formation of New York has been the focus of numerous studies. Key works on the formation in eastern New York include Oliver (1956), Oliver et al. (1962), Lindemann (1979, 1980), Lindemann and Feldman (1987), and Feldman (1985). Recent field study augments this body of work and examines the regional relationships of Southwood-age strata in New York and Pennsylvania.

In the Catskill Front the Onondaga Formation is best exposed at the Saugerties exit of the New York State Thruway. A nearly complete section of the formation (50 m; Feldman, 1985) includes strata of the Edgecliff, Nedrow, and Moorehouse Members. The contact with the underlying Schoharie Formation (Saugerties Member is gradational. The lower 1.8 m and a thin interval 2.5-2.7 m above the base of the Onondaga Formation are relatively chert-free; the remainder of the Edgecliff Member (14.6 m total) features abundant layers of gray-weathering chert in fossiliferous pack- to wackestones.

The contact with the overlying Nedrow Member is marked by the disappearance of chert, an abundance of pyrite, and a thin shaly seam. Several distinctive and widely correlatable marker units found in the Nedrow in central to western New York and equivalent strata in Pennsylvania (see Brett and Ver Straeten, 1994) have been recognized by the authors at Saugerties. They include: 1) a thin clay parting 1.6 m above the base of the member (which possibly represents a thin K-bentonite layer) that may correlate with a similar bed in central New York; and 2) a subtle but distinctive darker interval 3.4-4.2 m above the base, also present at Kingston, that correlates with a pair of black shales at the top of the Nedrow Member in central New York and equivalent strata of the Selinsgrove Limestone throughout central Pennsylvania ("lower and upper black beds" of Brett and Ver Straeten, 1994).

The overlying Moorehouse Member at Saugerties has not as yet been examined in detail. Several key marker beds found in lower Moorehouse and correlative strata across New York and Pennsylvania are, however, recognized in the Thruway exposures. More detailed work on the Onondaga Formation at Saugerties is in process.

To the southwest, in eastern Pennsylvania, Onondaga-equivalent strata are represented by limestones of the Buttermilk Falls Formation. The four members of this formation (Inners, 1975; Epstein, 1984) are laterally equivalent to the Edgecliff, Nedrow, Moorehouse, and Seneca Members of the Onondaga Formation in New York. In general, the strata of the Buttermilk Falls Limestone is finer-grained than the Onondaga Formation of eastern New York State. A lower, crinoid rich, chert-rich unit with corals and a chert-free interval at its base is overlain by calcareous shales, similar to the Edgecliff and Nedrow Members in the central New York Finger Lakes region. Overlying limestones are characterized by dark chert (=Moorehouse Member). A prominent K-bentonite bed in the upper part of the formation is equivalent to the Tioga B-OIN bentonite that occurs at the base of the Seneca Member in New York. The Buttermilk Falls Formation is ~83 m-thick in the vicinity of its type section (East Stroudsburg, Monroe Co., eastern PA.). Lithologic and faunal trends throughout the Buttermilk Falls Limestone indicate that the Stroudsburg area represented deeper-water environments than eastern New York throughout its deposition. Lithologic and faunal trends clearly show that the Stroudsburg region was the basinward trough of carbonate-dominated ramps that extended from the Albany area in eastern New York and the Harrisburg vicinity of central Pennsylvania.

Hamilton Group: Cazenovian Stage (Middle Devonian)

Marcellus "subgroup." Relatively thin, black Marcellus shales above the Onondaga Limestone in western New York are the lateral equivalents of a thick, complex set of fine- to coarse-grained siliciclastics in eastern New York. These eastern strata, ~600 m in thickness (Rickard, 1989), range from basinal black shales that shallow upward to nearshore sandstones and alluvial-dominated brackish to freshwater continental environments. Recent work on these rocks show that this thick succession is separated into two separate subdivisions with distinct faunas, lithologies, and sea level histories. Recognition of these significant differences between the two units has led to a revision of the Marcellus "Shale" (Hall, 1839) in New York (see Ver Straeten et al., 1994, in prep.) in which the Marcellus is raised to subgroup status, and the lower and upper subdivisions are assigned formational status. In this new stratigraphic scheme, the lower part of the Marcellus subgroup is termed the "Union Springs Formation". The upper subdivision is divided into two laterally-equivalent units, the "Mount Marion Formation" for the fine- to coarse-grained, progradational clastics in eastern New York, and the "Oatka Creek Formation" for black shale-dominated facies in western to central New York (see Ver Straeten et al., 1994 for more details). The contact of the Union Springs and

Mount Marion-Oatka Creek Formations lies at the base of the Cherry Valley Member across New York. Reasons for this stratigraphic revision include: 1) Recognition that lower and upper parts of the Marcellus subgroup each comprise a major sedimentary sequence, equivalent in stature to the other formations of the Hamilton Group (Skaneateles, Ludlowville, and Moscow Formations); 2) the lower part of the Marcellus subgroup features a unique fauna that is distinctly different from the overlying upper part of the subgroup and the remainder of the Hamilton Group.

Union Springs Formation. The Union Springs Formation (originally Union Springs “Member” of Cooper, 1930) marks the initiation of Middle Devonian clastic deposition in the Northern Appalachian Basin. Initial black shale-dominated strata (Bakoven Member) are generally in sharp contact with the underlying upper part of the Onondaga Formation; the contact is commonly marked by phosphate-rich fish bone beds. In eastern New York, the black shale facies of the Bakoven Member are overlain by buff- to dark gray-weathering, calcareous shales, siltstones, and sandstones of the Stony Hollow Member. The Union Springs Formation is capped across New York State by fossiliferous limestones and dark shales of the Hurley Member (Ver Straeten et al., 1994). The Union Springs Formation is best developed in southeastern New York (Hudson Valley, notably the Kingston area), where it is ~175 m-thick. The formation thins to the north and west and ranges from 10 m-thick to a feather edge across central to western New York. Union Springs strata are generally absent west of the Genesee River (Rochester area; for discussion see Rickard, 1984; Ver Straeten et al., 1994).

Barren to sparsely fossiliferous black shale-dominated facies of the Bakoven Member are characterized by pelagic to dysaerobic-type faunas dominated by dactyloconariids, nautiloid and goniatite cephalopods, and leiorhynchid brachiopods (Brower and Nye, 1991). More aerobic facies of the upper part of the formation (upper part of the Stony Hollow and the Hurley Members), however, feature a more diverse fauna, characterized by a unique assemblage of brachiopods, cephalopods, trilobites, corals, and crinoids. Elements of this fauna are associated with immigration of benthic forms dominantly from arctic Canada into the Appalachian, Michigan, Illinois, and Iowa Basins (Koch, 1978, 1988; Boucot, 1990; see Ver Straeten et al., 1994) and a global evolutionary event of pelagic cephalopods and dactyloconariids (Kacak-otomari Event; Truyllos-Massoni et al., 1990; Chlupac and Kukal, 1986; Boucot, 1990; see Ver Straeten et al., 1994). This Union Springs fauna is very distinctive from both that of the underlying Onondaga Formation and the overlying remainder of the Hamilton Group. Key elements include the brachiopods *Variatrypa arctica*, *Warrenella*, *Kaysarella*, and *Pentamerella*, the proetid trilobite *Dechenella haldemanni*, a rugose coral (*Guerichiphyllum*), and the microcrinoid *Haplocrinites*. Cephalopods of the Union Springs Formation associated with the global Kacak-otomari Event include *Cabrioceras* and *Agoniatites nodiferous*. A detailed study of this fauna by A.J. Boucot and others is presently in progress.

The Bakoven Member is assigned to the *Tortodus kockeliamus australis* conodont zone and the Hurley and questionably the Stony Hollow Members are placed within the *Tortodus kockeliamus kockeliamus* zone (Middle Devonian; Klapper, 1981). The Bakoven, Stony Hollow, and at least lower part of the Hurley Members feature *Cabrioceras plebeiforme* goniatite faunas; the upper part of the Hurley Member (Lincoln Park submember) features *Agoniatites vanuxemi nodiferous* (House, 1978, 1981). The Union Springs Formation is not assigned a brachiopod zone at present, but recent study shows it features a unique fauna that is characterized by the brachiopod *Variatrypa arctica*. Key references for the Union Springs Formation in eastern New York include Goldring (1935, 1943), Chadwick (1944), Rickard (1952, 1985), Storm (1985), Griffing and Ver Straeten (1991), Griffing (1994), and Ver Straeten et al. (1994).

Griffing and Ver Straeten (1991) presented an initial detailed report of the lower part of the Marcellus subgroup in eastern New York. The subsequent revisions presented by Ver Straeten et al. (1994, in prep.) distinguish upper strata that Griffing and Ver Straeten (1991) retained in the Stony Hollow Member and assigns them to the new Hurley Member. Griffing and Ver Straeten also presented a series of key beds within strata now assigned to the Union Springs Formation. Two richly fossiliferous limestone dominated beds (“Proetid Units” of Griffing and Ver Straeten, 1991) with proetid trilobites occur in the upper part of the revised Union Springs formation in eastern New York. The higher bed, the “Upper Proetid Unit” of Griffing and Ver Straeten (1991) is now assigned to the Chestnut Street submember of the Hurley Member. It is separated from the overlying Cherry Valley Member by dark shales and minor sandstones of the Lincoln Park submember of the Hurley Member. The “Lower Proetid Unit” occurs in the upper part of the Stony Hollow Member, where it underlies a massive, resistant sandstone (“Massive

Sandstone Unit" of Griffing and Ver Straeten, 1991) that occurs immediately below the top of the Stony Hollow Member.

Mount Marion Formation. The upper part of the Marcellus subgroup in eastern New York is represented by the Mount Marion Formation and part or all of the overlying Ashokan Formation (Rickard, 1975). The Mount Marion Formation (Grabau, 1917, 1919) comprises a general coarsening-upward, progradational marine succession from basinal black shales to nearshore sandstones. The base of the formation, as redefined by Ver Straeten et al. (1994, ms. in prep.), is placed at the bottom of coeval limestones and calcareous sandstones of the Cherry Valley Member. Succeeding black to dark gray shales of the Berne Member are overlain by a prominent coral-rich unit (Halihan Hill Bed of Ver Straeten, 1994) at the base of the Otsego Member. Dark gray mudstones and thin sandstones of the lower and middle part of the Otsego Member grade upward into increasingly coarser, sand-dominated strata in the upper part of the formation. No member level subdivision has been proposed for the upper part of the formation beyond application of central New York terms (Solsville and Pecksport Members; see Wolff, 1967, 1969; Pedersen et al., 1976; Grasso and Wolff, 1977; Wolff and Buttner, 1979) which may or may not be correlative. In the Catskill Front the upper part of the Marcellus subgroup (= Mount Marion Fm. and all or part of Ashokan Fm.; Rickard, 1975) is reported to be in excess of 425 m (Rickard, 1989); equivalent strata of the Oatka Creek Formation (revised by Ver Straeten et al., 1994, ms. in prep.) in central to western New York thin to as little as 5 m (Ontario Co.; Rickard, 1989). Based on well log data from southwest of Catskill (Rickard, 1989, pl. 4) strata of the Mount Marion are probably in excess of 330 m-thick in the Catskill Front.

West of Albany, the Cherry Valley Member is composed of relatively fine-grained, bedded to nodular limestone, ~0.5 to 1.0 m in thickness. A pelagic fauna of cephalopods and dactyloconariids, as well as small brachiopods and small tabulate corals (auloporids) are characteristic of the unit (Griffing and Ver Straeten, 1991). South of Albany, however, the Cherry Valley Member becomes increasingly rich in terrigenous sand, and ranges up to 10 m in thickness (Griffing and Ver Straeten, 1991). Above the Cherry Valley, black to dark gray shales of the Berne Member are dominated by sparse, low diversity faunas of cephalopods, nuculid bivalves, and leiorhynchid brachiopods. The coral-rich Halihan Hill Bed (=base of Otsego Member) features a varied and highly diverse coral and brachiopod-dominated fauna (>70 forms) with abundant echinoderm debris. The overlying lower part of the Otsego Member is typified by low-diversity brachiopod assemblages that diversify upward into more mollusc-dominated faunas in the upper middle part of the formation. Diversity drops in the upper part of the formation; coarser, sand-dominated facies appear heavily bioturbated (*Zoophycos*) to laminar bedded (flagstone facies). Finer-grained facies in the upper part of the formation are dominated by low diversity *Mucrospirifer-Camarotoechia-Schizophoria*- (brachiopods) and bivalve-dominated faunas.

The Cherry Valley Member of the Mount Marion (and laterally equivalent Oatka Creek Formation) is placed within the *Tortodus kockelianus kockelianus* conodont zone and the *Agoniatites vanuxemi vanuxemi* goniatite zone (Middle Devonian: Klapper, 1981; House, 1978, 1981). Overlying strata of the Mount Marion Formation are assigned by Dutro (1981) to the *Mucrospirifer mucronatus* brachiopod assemblage zone and by Rickard (1975) to the *Mediospirifer audacualus* brachiopod zone; the formation lies within the *Tornoceras uniangulare* goniatite zone (House, 1981). Conodonts are not known from the formation. Oliver and Sorauf (1981) place the post-Cherry Valley strata of the formation in the *Heterophrentis ampla* coral zone. References for the Mount Marion Formation include Goldring (1943), Chadwick (1944), Wolff (1967, 1969), Pedersen et al. (1976), Grasso and Wolff (1977), Wolff and Buttner (1979), and Ver Straeten (1994).

Ashokan Formation. The Ashokan Formation (Grabau, 1917) marks the transition in eastern New York from marine to subaerial, fluvial-dominated deposition that characterizes the remainder of the Devonian in the Catskill Front. The formation consists of cyclic alternations of medium-grained subgraywacke sandstones and olive- to brown-weathering mudstones with dark gray shales. The Ashokan Formation is distinguished by the presence of channel-form sandstone bodies, the absence of normal-marine fossils, and the lack of continental redbeds. The sandstones appear cross-bedded to laminated and in the past were an important source of "bluestone" flagstones, utilized in building. Wolff (in Pederson et al., 1976) reports fluvially- and tidally-influenced sedimentary structures from the sandstones. Thickness of the unit ranges

from ~150 m northwest of Kingston (type area) to a feather edge north of Catskill due to facies change with the overlying Plattekill Formation (Chadwick, 1944).

Fossils in the Ashokan Formation consist dominantly of plant material and ostracod and conchostracan crustaceans. The latter two forms are thought to be indicative of brackish water environments (Goldring, 1935) and occur through the overlying succession of fluvial-dominated environments (Chadwick, 1944, p. 120-121). Biostratigraphy of the formation is unknown. References on the Ashokan Formation include Chadwick (1944), Wolff (1969), and Pederson et al. (1976).

Hamilton Group: Upper Cazenovian-Tiogniogan Stages (Middle Devonian)

Plattekill Formation. The lowest Middle Devonian redbeds in the Catskill Front mark the base of the Plattekill Formation (Fletcher, 1962). The formation is composed dominantly of medium dark gray subgraywacke sandstones, siltstones, shales, and grayish-red shales, siltstones, and mudstones (Lucier, 1966). Fletcher (1964, 1967) and Lucier (1966) report a maximum thickness of 305 m in the Catskill Front, and subdivide the Plattekill Formation into two subdivisions, a lower unit (210 m-thick) dominated by medium dark gray shales, siltstones and sandstones with minor redbeds. The upper subdivision is almost entirely composed of grayish-red shales, siltstones, and claystones.

Willis and Bridge (1988) recognize two dominant facies for the Plattekill and overlying Manorkill Formations of the Catskill Front: 1) sandstone bodies, which represent channel bar deposits of aggrading, sinuous, migrating river channels; and 2) a sandstone-mudstone association, which was deposited over a low-relief alluvial plain during flood events. Bedding of the latter facies is generally obscured by bioturbation, desiccation cracks, and paleosol development.

Biostratigraphy of the non-marine Plattekill Formation is unknown. Key references for the formation include Chadwick (1933a, 1944; =lower part of his Kiskatom Fm.), Fletcher (1962, 1963, 1964, 1967), Lucier (1966), Willis and Bridge (1988), and Bridge and Willis (1994).

Overlying Middle and Late Devonian strata of the Catskill Front

Manorkill and Oneonta Formations. Subaerial, alluvial-dominated environments characterize the remainder of the Middle and Upper Devonian succession along the Catskill Front. Rocks of the overlying Manorkill and Oneonta Formations (including the Twilight Park Conglomerate) are, in general, similar to those of the Plattekill Formation but in general coarsen-upward, and have larger paleochannel sizes and thicker fluvial cycles (Willis and Bridge, 1988; Gordon and Bridge, 1987). These trends culminate in deposition of the Twilight Park Conglomerate, characterized by coarse pebble lithologies and thick sandstone bodies (Willis and Bridge, 1988; Bridge and Nickelsen, 1985). References for the post-Plattekill Middle and Upper Devonian strata of the Catskill Front include Chadwick (1933 a&b, 1944), Fletcher (1962, 1963, 1964, 1967), Lucier (1966), Bridge and Gordon (1985 a&b), Bridge and Nickelsen (1985), Bridge, Gordon, and Titus (1986), and Willis and Bridge (1988).

Strata of the Skunnemunk Outlier, Southeastern New York

Upper Lower and Middle Devonian strata also outcrop in an outlier ~35 km southeast of the main outcrop belt in Orange County, New York and northern New Jersey. The strata represent more shoreward equivalents to the strata in the Hudson Valley outcrop belt.

The Oriskany-equivalent Connelly Conglomerate lies at the base of the interval, where it sharply overlies sandstones of the Central Valley Formation (Lower Devonian Helderberg Group). Overlying dark mudstones, siltstones, and fine sandstones are assigned to the Esopus Formation. Four subdivisions of the Esopus Shale were recognized by Boucot et al. (1970) in the Skunnemunk Outlier, a lower siltstone (Mountainville Member), overlying dark-gray to black mudstones (Quarry Hill Member), fine sandstones and interbedded siltstones (Highland Mills Member), and an upper interval of black siltstones and mudstones (Eddyville Member). Overlying strata of the Pine Hill Formation (Boucot et al., 1970) consist of medium-bedded siltstones and fine sandstones (Woodbury Creek Member) and sandstones and conglomerates (Kanouse Member; Boucot et al., 1970). The Kanouse Member is shown by Kindle and Eidman (1955) and Boucot (1959, p. 734; pers. commun., 1995) to be laterally equivalent to the Middle Devonian Onondaga Formation of the main outcrop belt.

Boucot et al. (1970) reported four members of the Esopus in the Skunnemunk Outlier in southeastern New York. Lithological trends through their lower three members (Mountainville, Quarry Hill, and Highland Mills Members) mirror the relative coarser-finer-coarser trends of the lower, middle, and upper members proposed herein. The overlying fourth member (Eddyville Member) is finer-grained; this mirrors the trend in the main outcrop belt wherein strata of the lower part of the Schoharie Formation fine upward above the upper strata of the Esopus Formation. We herein project that Boucot et al.'s (1970) fourth member of the Esopus Formation in the Skunnemunk Outlier may represent lower strata of the Schoharie Formation (unnamed member) in the main outcrop belt. More work is needed to resolve this, however.

No limestone of Southwood (Onondaga)-age is reported in the Skunnemunk Outlier. The Onondaga Limestone is, at least in part, laterally replaced by quartz pebble conglomerates and quartz arenites of the Kanouse Member of the Pine Hill Formation (Boucot et al., 1970). The occurrence of large *Amphigenia elongata* brachiopods and the absence of smaller *Amphigenias* indicate the member is equivalent to the Onondaga Formation of the main outcrop belt (Kindle and Eaton, 1955; Boucot, 1959; pers. commun, 1995). Interbedded conglomerates and sandstones in the lower part of the member fine upward into quartz-rich sandstones. Only the lower 7.6 m of the Kanouse Member is exposed in New York State (NYS Thruway at Highland Mills, Orange Co.); black shales of the Marcellus subgroup (Cornwall Formation) are exposed above a ~45 m-thick covered interval. The covered interval is not known from anywhere in the Skunnemunk Outlier.

The black Cornwall shales are overlain by offshore dark gray shales to nearshore and non-marine sandstone-dominated strata (Bellvale Formation). The top of the Devonian succession in the outlier (Skunnemunk Formation) is characterized by subaerially-deposited, conglomeratic fluvial-dominated facies. Thicknesses for the Cornwall, Bellvale, and Skunnemunk Formations range widely (~60-300 m; ~300-900 m; and ~90-750 m, respectively; see Sulenski, 1969). No comprehensive stratigraphic study of the Hamilton Group of the Skunnemunk outlier has been reported (however, see Sulenski, 1969), and the member- and formation-level units of the Hudson Valley have not been recognized. Rickard (1975) however, correlates the Cornwall and Bellvale Formations with the Union Springs and Mount Marion Formations (terminology of this paper) and the Skunnemunk Conglomerate with the Plattekill and Manorkill Formations of the Catskill Front.

References for upper Lower and Middle Devonian strata of the Skunnemunk outlier include Kindle and Eidman (1955), Boucot (1959), Oliver et al. (1962), Finks (1968), Sulenski (1969), Boucot et al. (1970), and Marintsch and Finks (1982).

K-bentonites

Volcanic ash falls are a minor but very important part of the sedimentary rock record. Altered volcanic ash deposits of Paleozoic age, termed K-bentonites, can provide important information in a detailed basin analysis. K-bentonites represent isochronous deposits important to the stratigrapher and may yield absolute age dates for the timing of geological and biological events. Furthermore, they may represent the best record of ancient volcanism adjacent to deeply eroded mountain belts. They are a very important element in reconstructing the tectonics and evolution of an orogeny.

Two intervals of K-bentonite rich strata are known from the upper Lower and lower Middle Devonian rocks of the Northern Appalachian Basin. These strata represent multiple layers of altered volcanic ash that were erupted during Plinian-type volcanic eruptions in the adjacent magmatic arc of the Acadian Orogen. Dispersal of ash-laden clouds over the Appalachian foreland basin permitted deposition of water-lain ash beds that are generally thought to represent single eruptive events. Recognition and correlation of altered volcanic ash beds (K-bentonites of Paleozoic age) are key elements to detailed correlation of the foreland basin fill. Furthermore, they provide insights into paleovolcanism and orogenesis in the adjacent Acadian mountain belt.

Sprout Brook Bentonites (Esopus Formation). The lower member of the Esopus Formation is characterized by an interval of interbedded K-bentonites, cherts, siliceous siltstones, and dark shales. This recently discovered interval of altered volcanic ashes is termed the Sprout Brook Bentonites (Ver Straeten, 1992 a,b, in review; Ver Straeten et al., 1993). Up to 15 thin (mm-scale to ~15 cm-thick), tan-, green-, or gray-colored, soapy-feeling clay to claystone beds characterize the Sprout Brook Bentonites. The presence

of euhedral zircons and apatites, diagnostic of volcanogenic strata, permit positive identification of the beds as altered volcanic ashes. The Sprout Brook Bentonites are best developed in New York State, but the interval is recognized across the Appalachian Basin through Pennsylvania into Maryland and the Virginias (Ver Straeten, 1992b, in review). They mark the transition from the shallow marine sandstones and limestones of the Oriskany-Glenerie-Connelly Formations and deeper, more basinal dark silty shales of the Esopus Formation.

Ver Straeten (in review) discusses potential volcanic sources for the Sprout Brook Bentonites in the Appalachian Mountains. One distinct possibility occurs 400-500 km to the northeast of the Hudson Valley in the Piscataquis volcanic belt of north-central to western Maine. Five volcanic centers in that region form a massive body of upper Lower Devonian pyroclastic ash flow tuffs. The largest of these deposits, the Traveler Rhyolite, has an estimated preserved volume of ~800 km³ (Rankin and Hon, 1987). This is comparable in size to some of the larger Tertiary deposits in the western United States, which include the Timber Mountain and Paintbrush Tuffs of southwestern Nevada (900 km³ and 1000 km³ respectively) and the Lava Creek Tuff in the Yellowstone Caldera (1000 km³; Christiansen, 1979, p. 31). Subadjacent to the Traveler Rhyolite is its comagmatic pluton, the Katahdin Granite, which has a surface area of ~1350 km² (Griscom, 1976). The upper Lower Devonian pyroclastic rocks of the Piscataquis volcanic belt overlie shallow-marine sandstones of Oriskany age; marine rocks that overlie the pyroclastic rocks yield Schoharie-age marine faunas (Boucot, 1969). This large volume of Esopus-age pyroclastic source rocks and comagmatic plutons 400-500 km northeast of the Northern Appalachian Basin (Ver Straeten, in review) may have been the source area for the Sprout Brook Bentonites.

Tioga Bentonites (Onondaga Formation). The Tioga Bentonites (*sensu* Way et al., 1986) mark a second interval of Devonian pyroclastic-rich strata above the Wallbridge Unconformity. They have been the focus of numerous studies since their initial discovery in the 1940s (Fettke in Ebright et al., 1949; Fettke, 1952; Dennison, 1960, 1961; Dennison and Textoris, 1970, 1978, 1987; Conkin and Conkin, 1979, 1984; Way et al., 1986; Brett and Ver Straeten, 1994). The source area for the Tioga Bentonites is projected by Dennison and Textoris (1978) to be in the region of Fredericksburg in northeastern Virginia, ~500 km southwest of the Catskill Front.

Ten or more thin clay to claystone layers, which may feature abundant biotite grains (in addition to zircons and apatites), occur widely in the upper part of the Onondaga Formation and equivalent strata across eastern North America (Conkin and Conkin, 1979, 1984; Way et al., 1986; Brett and Ver Straeten, 1994). Brett and Ver Straeten (1994) discuss the Tioga Bentonites in New York State and the relationship to their occurrence in Pennsylvania, where they have recently been the focus of detailed work (Smith and Way, 1983; Way et al., 1986). The authors recognize ~eight Tioga beds in the Seneca Member and base of the Union Springs Formation in west-central to western New York. Additional thin K-bentonite beds occur in the underlying Moorehouse, Nedrow, and possibly the Edgecliff Members. Very prominent K-bentonites occur at the base of the Seneca Member (Onondaga Indian Nation Ash [OIN] of Conkin and Conkin 1979, 1984; Conkin, 1987; = Tioga B of Way et al., 1986) and at the base of the overlying Union Springs Formation (=Tioga F of Way et al. 1986) in west-central to western New York.

The Tioga Bentonites in eastern New York State are poorly developed and relatively undocumented. This is in part due to the progressive eastward bevelling of upper Onondaga strata, where most of the bentonites occur (Rickard, 1975, 1989 reports the Seneca Member as absent in the Albany area). Furthermore, outcrops of the upper part of the Onondaga Limestone are rare south of the Albany region, where the Seneca Member reappears and thickens into eastern Pennsylvania. The bentonites do not appear to pass laterally into the lower part of the overlying black shale of the Bakoven Member as projected by previous authors (Oliver, 1954; Rickard, 1975). Instead, they appear to be cutout at a submarine unconformity at the top of the Onondaga in eastern New York (Rickard, 1984; Ver Straeten et al., 1994). However, recent work has shown that thin Tioga Bentonites occur in the upper part of the Onondaga Formation in the Hudson Valley and the Helderberg Escarpment near Albany.

The authors have found two altered volcanic ash beds in the upper 2-3 m and at the top of the Onondaga on the Helderberg Escarpment southeast of Albany. Both K-bentonites are on the order of 8-10 cm-thick and have visible micas. Either bed could potentially be the Tioga B-OIN bed that occurs at the base of the Seneca Member from Cherry Valley to the west. However, the Tioga B-OIN bentonite is

consistently on the order of 20-25 cm-thick. On the other hand, the beds may also correlate to thinner bentonites in the upper part of the Moorehouse Member to the west.

At Catskill (Stop 4 of this trip) a thin (~2 cm-thick) K-bentonite is present 0.3 m below the top of the Onondaga, in strata that Rickard (1989) considers to be the Seneca Member. This bed, similar to the layers on the Helderberg Escarpment, cannot be correlated into central New York at this time. Conkin and Conkin (1984) reported the Tioga F bed (=their "Tioga restricted" bed, designation abandoned) at the Onondaga-Marcellus contact at Stop 4. The authors, however, have not found any K-bentonite layer at the contact.

Unconformities In Upper Lower and Lower Middle Devonian Rocks, Eastern New York State

Unconformities are surfaces of erosion and/or non-deposition that mark significant breaks in the stratigraphic record. They occur within a sedimentary basin at different temporal and geographical scales, and may be of subaerial or submarine origin (Shanmugan, 1988). Their recognition is a crucial part of integrated basinal studies and are important to the stratigrapher for interpreting changes in relative sea level, defining depositional sequences, determining the timing of tectonic activity and flexure of the basin, and predicting the occurrence of economic deposits.

At least nine unconformities, listed below, of varying scales of magnitude and geographic distribution occur in upper Lower to Middle Devonian rocks of eastern New York State. In addition, some relatively conformable successions in the eastern part of the state may laterally be represented by unconformities.

1. Wallbridge Unconformity. The Wallbridge Unconformity is one of six major unconformities that bound six Phanerozoic supersequences of North America (Sloss, 1963). It marks the boundary between Sloss' Tippecanoe and Kaskaskia Supersequences. In eastern New York, the unconformity represents a relatively short part of the Lower Devonian; in the Tristates area of New York, Pennsylvania, and New Jersey the rock record is continuous, and no unconformity is recorded. Westward, across New York and toward the craton, missing time represented by the unconformity increases; in central to west-central New York, the entire Lower Devonian is absent, and Middle Devonian rocks overlie Upper Silurian strata.

In the Hudson Valley, the unconformity is marked by a quartz- to phosphate-pebble rich bed at the base of the Glenerie Formation (Stops 1A & 2). Upper strata of the Helderberg Group and the basal unit of the Tristates Group are progressively chopped out below the unconformity northward along the outcrop belt (for discussion of the Wallbridge Unconformity in central to western New York, see Brett and Ver Straeten, 1994, p. 248-250).

2. Basal Esopus Unconformity. A relatively minor unconformity occurs at the base of the Esopus Formation along its outcrop in eastern New York. Rehmer (1976, p. 68) reports that the basal contact is "everywhere abrupt and disconformable." No significant lag deposit has been noted at this position; the break is generally marked by the change from underlying quartz arenites or limestones into shales, siltstones, and cherts. To the south and southwest, however, the base of the Esopus is indicated by a sharp lithologic change from quartz arenites or quartz pebble conglomerates to shales (e.g., Skunnemunk Outlier, New York; eastern Pennsylvania).

3. Sub-Schoharie Unconformity (=Pre-Upper Tristates Unconformity of Brett and Ver Straeten, 1994). A depositional break between the Esopus and Schoharie formations was recorded by Johnsen (1957) and Oliver et al. (1962), who reported glauconite and scattered quartz pebbles at the contact in the Hudson Valley (see Rickard, 1975). The magnitude of this subtle but important unconformity was not previously recognized. As noted above, upper strata of the Esopus Formation are progressively truncated below the unconformity northward of Kingston (Stop 5). A lag deposit of glauconite, scattered quartz pebbles, phosphate pebbles, and fish bone material at Catskill (Stop 1) sharply overlies the thinly laminated interval in the upper Esopus that occurs 22.5 m below the contact at Kingston.

Further evidence of the erosive nature of this unconformity is shown along U.S. Rt. 20 at Cherry Valley (Otsego Co.) in east-central New York. Very finely-detailed trace fossils (e.g., *Cruziana*, *Fustiglyphus*, and scratch markings) are present on the base of the Carlisle Center Member (restricted) at Cherry Valley (Miller and Rehmer, 1982). The preservation of detailed scratch marks and individual

trilobite appendage traces were interpreted to have been produced in semi-consolidated muds previous to deposition of the overlying glauconitic and pebbly sands of the Carlisle Center Member. Miller and Rehmer interpreted the break to represent a short depositional hiatus. A similar occurrence of "delicately sculpted" trilobite appendage traces and "remainie sediments" in the Middle Devonian of Ontario (Landing and Brett, 1987), however, clearly showed that preservation of the traces was directly related to excavation of disconformity-related, pre-compacted mud-firmgrounds at a major cycle boundary.

West of Cherry Valley the sub-Schoharie unconformity erosively cuts down through the middle member and then bevels away the entire Esopus Formation. It is possible that westward cutout of Esopus strata may in part be due to uplift of a peripheral bulge in central to western New York during Acadian Tectophase I/upper Tristates time. Its occurrence as far southeast as Kingston, however, appears to indicate some degree of eustatic influence is also involved. The unconformity, marked by a shell-rich fish bone and phosphate pebble lag bed, is also found at the correlative position in the lower part of the "calcareous shale member" (middle member) of the Needmore Formation in central Pennsylvania (e.g., Newton Hamilton, Mifflin Co., PA; Ver Straeten and Brett, in prep.).

4. Sub-Aquetuck Unconformity. The widespread occurrence of glauconite and scattered quartz pebbles at the contact of the unnamed and Aquetuck Members of the Schoharie Formation between Kingston (Ulster Co.) and Clarksville (Albany Co.) marks a relatively minor break in deposition (visible at Stop 1B).

5. Sub-Edgecliff Unconformity. The Schoharie-Onondaga contact across eastern New York is relatively conformable and represents a shallowing- to deepening-upward transition (Stops 1 and 3A & B). Across central to western New York, however, this unconformity represents a major break in deposition that locally becomes amalgamated with the pre-Schoharie and Wallbridge Unconformities (see Brett and Ver Straeten, 1994). In part, the unconformity in central to western New York is associated with an uplifted peripheral bulge from Acadian Tectophase 1 (Brett and Ver Straeten, 1994). The widespread transition from shallowing- to deepening upward in equivalent strata across the Appalachian Basin (Ver Straeten and Brett, 1994 a,b; Brett and Ver Straeten, 1994; Ver Straeten and Brett in prep.) appears to indicate that the unconformity is also related to a fall and rise in eustatic sea level.

6. Sub-Nedrow Unconformity. A subtle depositional break is recorded at the contact between the Edgecliff and Nedrow Members of the Onondaga Formation. This is generally indicated by a lithologic break from medium- to fine-grained limestones to more argillaceous strata, especially in central New York. Glauconite and pyritic nodules are found locally associated with the contact; pyritic crusts are also known to occur on the contact.

7. Onondaga-Union Springs Formational Contact. The contact of the Onondaga and Union Springs Formations represents a significant regional unconformity in New York State (e.g., Stop 4 of this trip). The contact is commonly marked by lag deposits of corrasion-resistant fish bone material and teeth and/or phosphate pebbles. The unconformity is shown to be older to the northeast by the cutout of upper Onondaga strata (Seneca Member) from the central Finger Lakes region and eastern Pennsylvania toward the Albany area (Rickard 1989). Along the New York outcrop Seneca Member strata are progressively cut out from the top as shown by the eastward absence of marker beds (bentonites, distinctive shell beds, etc.) toward the Albany area. Ver Straeten et al. (1994) recently discussed the Onondaga-Union Springs Unconformity in New York. It may be interesting to note that the contact across Pennsylvania is relatively conformable; sediment supply is presumed to have been more continuous in that area at that time.

8. Pre-Cherry Valley Unconformity. Across central to western New York the base of the Cherry Valley Member unconformably cuts downward through upper strata of the underlying Union Springs Formation, and generally lies directly on a resistant limestone ledge of the Hurley Member (Chestnut Street submember); south of Rochester, this erosional unconformity locally (at outcrop scale; see Ver Straeten et al., 1994, p. 293-294) cuts down through the Hurley and Bakoven Members to nearly rest directly upon the top of the Onondaga Formation. In eastern New York, however, the succession between upper strata of the Union Springs Formation (Hurley Member) and the base of the Cherry Valley Member is relatively conformable.

9. Top of Cherry Valley Unconformity. Similar to the previous horizon, this unconformity is poorly developed across eastern New York State. Across central to western New York it becomes more prominent, and southwest of Rochester it cuts out the entire Union Springs Formation and the Cherry Valley Member; lower strata of the Oatka Creek Formation in westernmost New York lie directly upon limestones of the Onondaga Formation.

DISCUSSION

Sequence Stratigraphy, upper Lower and Middle Devonian

The application of sequence stratigraphic methods to sedimentary basin studies provides a powerful tool for the analysis of time-rock relationships. It permits chronostratigraphic subdivision of the rock record into cyclic, unconformity-bound, genetically related successions of strata (Van Wagoner et al., 1988). A "depositional sequence" is the fundamental, meso-scale unit of sequence stratigraphy. A sequence is a coherent package of strata that is bound at bottom and top by unconformities or their correlative conformities (Mitchum et al., 1977). It is formed by a cyclic change in relative sea level through the interaction of tectonics, eustatic sea level change, and sedimentologic factors (Allen and Allen, 1990). A sequence can be subdivided into "systems tracts," composed of smaller scale cycles ("parasequences"), and are deposited during different stages of a transgressive-regressive cycle. Three systems tracts are recognized within a sequence: 1) A "Lowstand Systems Tract" at the base of a cycle, which consists of progradational to aggradational strata deposited during a fall to early rise in relative sea level. The lowstand systems tract of a sequence is not always well preserved; 2) a "Transgressive Systems Tract," which is deposited during a rapid rise in relative sea level. This results in onlap of sea level and sedimentation onto the basal unconformity of a sequence; depositional is retrogradational. Condensed sections are common within the transgressive systems tract and the lower part of the overlying highstand systems tract; 3) "Highstand Systems Tract" forms during the late stage of a rise to the early stage of a fall in relative sea level, which results in deposition of aggradational, sediment-starved to progradational strata. The base of highstand deposits occur associated with a "surface of maximum starvation" that may be marked by a smaller scale, submarine unconformity during a period of extreme sediment starvation.

Detailed field studies of upper Lower and Middle Devonian strata in New York and Pennsylvania indicate that the succession is composed of at least nine depositional sequences (see left side of Figures 3 and 4). Each sequence is bounded by basal unconformities or their correlative conformities at the base of the transgressive systems tract (e.g., unconformities 1,3,5 and 8 of the previous section); as noted by Brett and Baird (in press), the lowstand systems tract of the sequences are rarely preserved. Lesser unconformities may form at the base of the highstand systems tract (e.g., unconformities 2, 4, 6, 7, and 9 of the previous section). All strata discussed in this paper are part of Sloss's (1963) Kaskaskia Supersequence. The basal unconformity of DS1 represents the Wallbridge Unconformity-supersequence boundary.

Depositional Sequence 1. DS1 is composed of strata of the coeval Oriskany-Glenerie-Connelly Formations and the overlying Esopus Formation of New York State. The basal bounding surface of the sequence is the Wallbridge Unconformity (#1 of previous section). The lowstand systems tract of DS1 is not preserved, except potentially in the Tristates area near Port Jervis. The transgressive systems tract consists of the Oriskany Sandstone and its equivalents; it may also include in part lower strata of the Esopus Formation. The surface of maximum flooding may occur at the base of the Esopus Formation or within the lower member; the exact position is unclear at this time. In a general sense, the middle, relatively fine-grained member of the Esopus Formation comprises the early highstand of DS1; an overall coarsening-upward trend through the upper part of the middle member and the upper member is associated with late highstand conditions of siliciclastic progradation.

The Esopus Formation, however is also divisible into at least three major coarsening-upward successions (=lower, middle, and upper members). Each commences with dark gray to black shale and culminates in bioturbated argillaceous siltstone or fine-grained sandstone. The tops of the lower and middle members are capped by a sharply defined flooding surface. Smaller scale cyclicity may be manifest in parts of the formation by alternating 0.3-0.5 m-thick bands of lighter and darker gray-weathering mudstone.

Depositional Sequence 2. DS2 comprises strata of the Schoharie Formation (unnamed lower, Aquetuck, and Saugerties Members). The basal bounding unconformity of DS2 is the sub-Schoharie unconformity (#3 of previous section), that regionally erosionally truncates the underlying Esopus Formation. Again, a lowstand systems tract is not recognized at the base of the sequence. The unnamed lower member composes the transgressive systems tract. The discontinuity at the base of the Aquetuck Member (#4 of the above section) marks the surface of maximum flooding of DS2. Early highstand conditions characterize the Aquetuck Member; the maximum highstand of sea level appears to be represented by the widespread interval (New York and Pennsylvania) of dark shaly strata in the lower part of the member. A general shallowing upward trend through the upper part of the Aquetuck and Saugerties Members is indicative of late highstand conditions.

A particularly intriguing feature of DS2 is the strongly developed rhythmic bedding. This banding is similar to but better developed than that seen in parts of the Esopus Formation of DS1. Darker brownish buff calcareous mudstone layers alternate with cream-weathering, more carbonate-rich mudstone bands, calcareous nodules, or tabular limestone beds; these alternations show hints of bundling. This widely-known banded nature of the Schoharie Formation may represent small-scale, Milankovitch cycle rhythms comparable to those documented in Cretaceous and other age rocks (Kauffman, 1988, p. 639-644; Fischer, 1986, 1993).

Depositional Sequence 3. The third post-Wallbridge sequence consists of the Edgecliff, Nedrow and lower to middle parts of the Moorehouse Members of the Onondaga Formation of New York State. In eastern New York the base of the sequence is conformable; a laterally-equivalent erosive unconformity occurs across central to western New York. Lowstand conditions are not recognized, but may be found in the lower part of the Edgecliff Member, associated with initial growth of coral bioherms. The Edgecliff Member comprises the transgressive systems tract; the surface of maximum flooding is found at the base of the Nedrow Member (#6 of the previous section). Overlying fine-grained strata of the Nedrow member equal early highstand facies of DS3, which are succeeded by late highstand deposits of the lower to middle Moorehouse Member.

Smaller-scale cycles are not as readily recognizable in the Onondaga as in the Schoharie but may be discerned with careful study. Parasequence-scale cycles (~1 m-thick) reported in the Edgecliff Member in western to central New York (Brett and Ver Straeten, 1994) are difficult to distinguish in the coarser, less differentiated, chert-dominated facies of the member in eastern New York. The smaller scale banding seen in the underlying DS2 (Schoharie Fm.) is less evident in the Onondaga Limestone; rhythmic cherty and non-cherty couplets 0.1-0.4 m-thick could be a manifestation of the same type of cycles. The Nedrow Member in central New York displays alternations of more and less argillaceous carbonate beds that in some outcrops approach the weathered appearance of the Schoharie Formation.

Depositional Sequence 4. DS4 is marked at its base by a gradational change from shallowing- to deepening-up lithologic and faunal trends; the succession is conformable along the New York outcrop, associated with the widespread shallow ramp geometry of Onondaga strata across eastern North America. The laterally equivalent unconformity may occur in correlative shallower water deposits of the Detroit River Group in Michigan and Ontario. Upper Moorehouse strata may represent lowstand deposits of DS4; fining-upward trends through the overlying Seneca Member indicate a rise in relative sea level (transgressive systems tract). The prominent unconformity at the Onondaga-Union Springs contact in New York (#7 of previous section) represents a prominent marine flooding surface at the base of early highstand (note; this

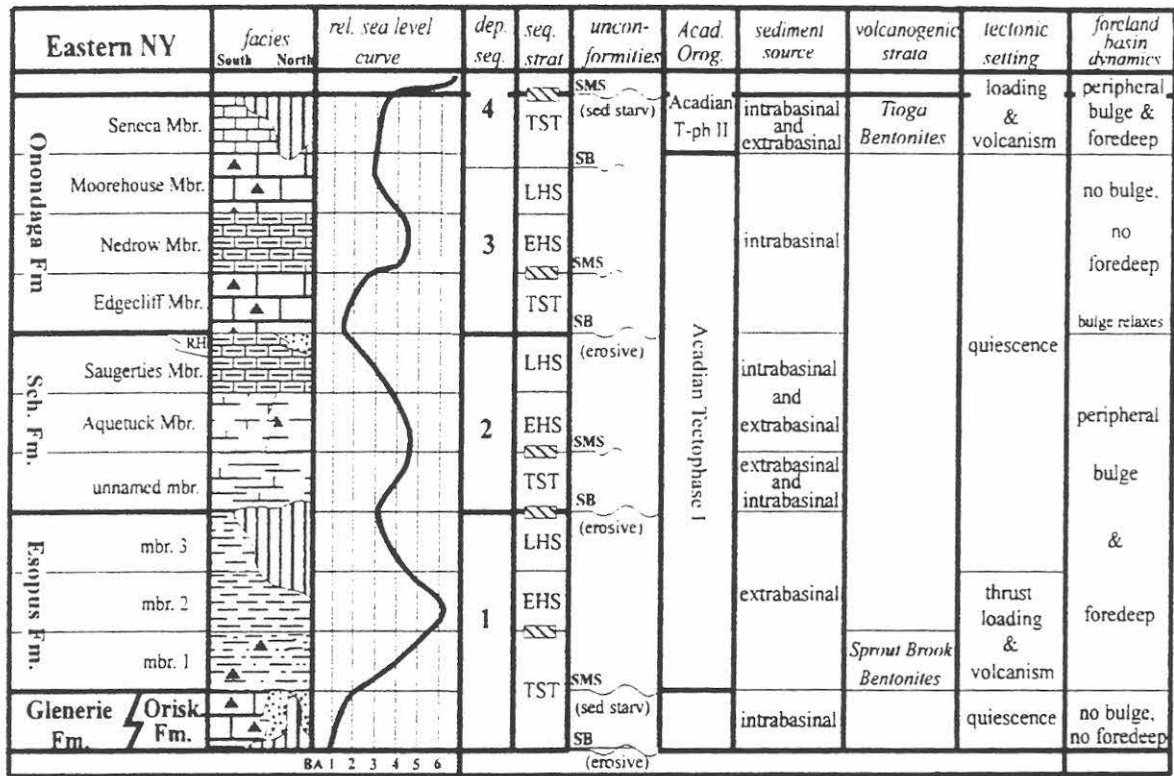


Figure 3. Stratigraphy, facies, sequence stratigraphy, and tectonic implications of upper Lower and lowest Middle Devonian rocks in New York State. Orisk.=Oriskany; Sch.=Schoharie; RH=Rickard Hill Member. WB=normal wave base; SWB=storm wave base; BA=Benthic Assemblages 1-6 of Boucot (1982). Dep.seq.=depositional sequence; TST=transgressive systems tract; EHS and LHS=early and late highstand systems tract; SB=sequence boundary; SMS=surface of maximum starvation of sequence stratigraphic terminology (see Van Wagoner et al., 1988). Erosive and sediment-starved unconformities as marked; cross-hatched boxes=condensed sections.

major flooding surface is associated in part with the onset of basin subsidence and siliciclastic deposition during Acadian Tectophase II).

Early highstand deposits of DS4 are composed of black shales of the overlying Bakoven Member (Union Springs Fm.). Late highstand conditions are represented by calcareous shales to sandstones of the Stony Hollow Member; several scales of shallowing-upward cycles are displayed in the member. The overlying Hurley Member has the character of an intermediate-scale sub-sequence (parasequence set?) in which the Chestnut Street submember (with several internal cycles) represents an analog of a transgressive systems tract and the overlying Lincoln Park submember shales represent highstand systems tract-like deposits. The upper part of the Hurley Member sub-sequence is truncated below the a sub-Cherry Valley unconformity across much of central to western New York State.

Depositional Sequence 5. The overlying fifth depositional sequence comprises the Mount Marion Formation in eastern New York and the coeval Oatka Creek Formation in central to western parts of the state. The authors questionably place the sequence boundary at the base of the Cherry Valley Member, but note the transgressive nature of the skeletal limestones of the Chestnut Street beds at the base of the Hurley Member. In eastern New York DS5 is conformable, but the basal contact (base of Cherry Valley Mbr.) erosionally truncates the upper part of the underlying Hurley Member across central to west-central New York. West of Rochester, the basal Cherry Valley unconformity becomes amalgamated with the overlying maximum flooding surface unconformity (#9 of the previous section), which in part erosionally truncates

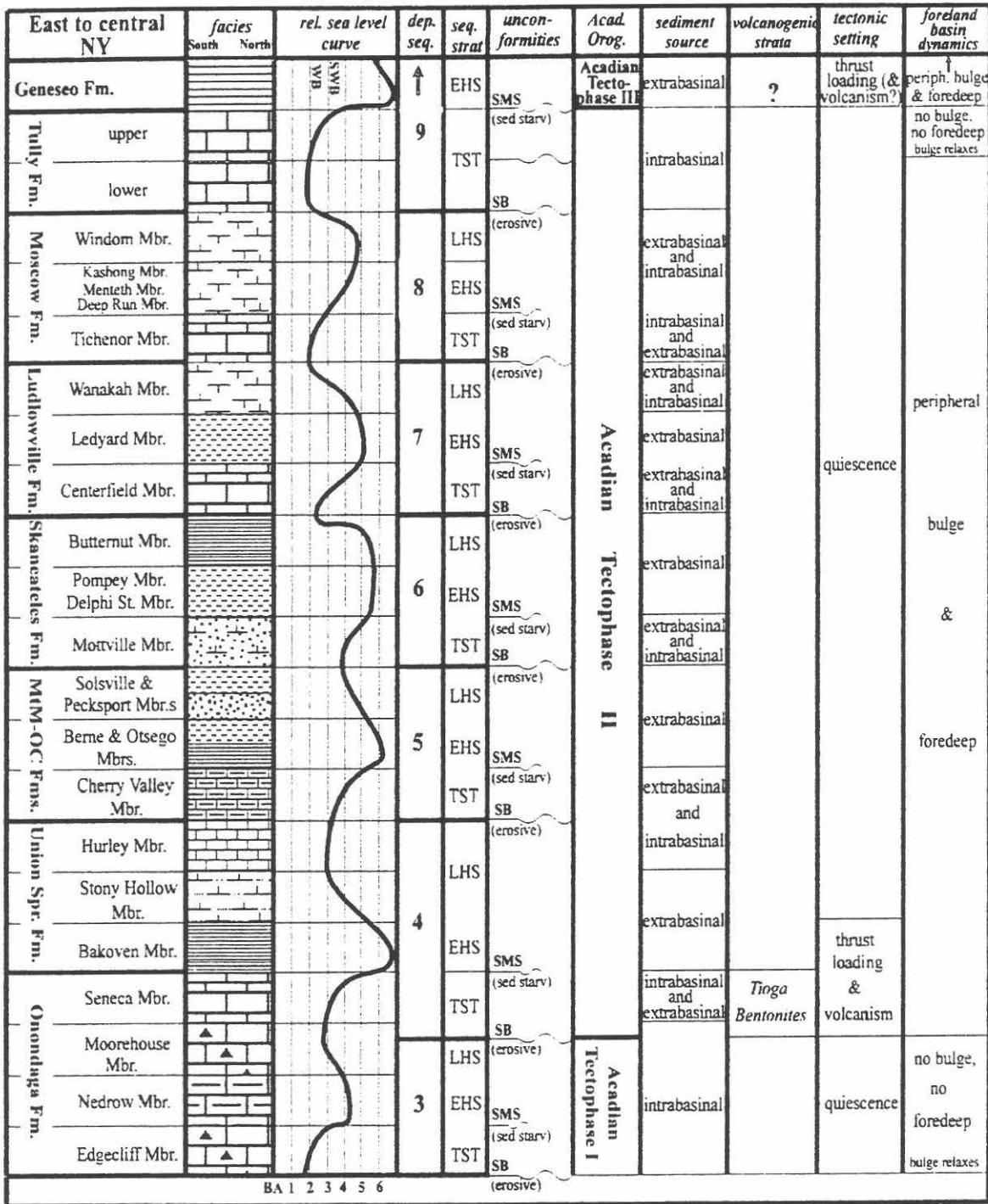


Figure 4. Stratigraphy, facies sequence stratigraphy and tectonic implication of Middle Devonian marine rocks in New York State. Abbreviations as in Figure 3. Skaneateles, Ludlowville, and Moscow Formations sea level-curve and sequence stratigraphy after Brett and Baird (in press).

both the Cherry Valley Member and the highly condensed western New York highstand deposits of DS4; in that area highstand deposits of DS5 (lower part of the Oatka Creek Fm.) directly overlie the transgressive systems tract of DS4 (Seneca Member of the Onondaga Formation).

Black to dark gray shales of the Berne Member (Mount Marion Fm.) in eastern New York represent early highstand deposits of DS5. Thick overlying deposits of the Otsego Member and undefined upper Mount Marion strata represent progradational infilling of the preserved eastern foredeep of the basin; uppermost strata of DS5 are represented by fluvial-dominated strata of the Ashokan Formation (Rickard, 1975).

Parasequence-scale cycles appear to be represented in the upper Berne and lower Otsego Members by 3-8 m-thick successions of dark gray mudstones capped by thin shell beds as reported by Ver Straeten (1994). The cycle-capping shell beds represent sediment-starved flooding surfaces at the base of the parasequences. The base of the coral-rich Halihan Hill bed (=base of Otsego Mbr.) is locally unconformable (Ver Straeten, 1994) and marks the bottom of a key, intermediate-scale sub-sequence within the Mount Marion Formation.

Depositional Sequences 6,7,8, and 9. Sequences 6-9 of the New York Devonian are not recognized to poorly defined (e.g., DS9; see Bridge and Willis, 1994) in the fluvial-dominated, subaerial redbed succession of the Middle Devonian Plattekill and Manorkill Formations in the Catskill Front. These sequences, which have been discussed in detail for equivalent marine strata (Brett and Baird, in press), comprise the Skaneateles, Ludlowville, Moscow, and combined Tully and Genesee Formations of central to western New York. These depositional sequences are characterized by a basal limestone-sandstone unit that overlies a sequence-bounding unconformity (e.g., Stafford-Mottville, Centerfield, and Tichenor-Menteth Members, DS6-8, respectively; Tully Formation, base of DS9). Flooding surfaces that cap the limestones are succeeded by dark, shale-dominated strata that in general coarsen upward to the base of the overlying sequence.

Basin Fill and the Acadian Orogeny

Foreland Basins Overview. Different mathematical and computer-generated models have been proposed in recent years to describe foreland basin dynamics and stratigraphy associated with orogenic episodes. Various approaches are based on thrust deformation and loading, sedimentary erosion and redistribution of the load, and an elastic (Jordan and Flemings, 1991; Sinclair, et al., 1991) or visco-elastic (Beaumont et al., 1988) flexural response of the lithosphere.

Quinlan and Beaumont (1984) and Beaumont et al. (1988) discuss a model of visco-elastic foreland basin flexure and its influence on stratigraphy of the basin fill over time. Loading of the lithosphere during episodes of tectonic thrusting leads to stress-induced subsidence of a proximal foreland basin and gentle uplift due to relaxation on a cratonward peripheral bulge. Subsequent periods of tectonic quiescence and unloading are marked by relaxation and uplift of the foreland. Beaumont et al. (1988) present a synthetic model of foreland basin fill and its implications for the magnitude of overthrust loads and the morphology of the inherited passive margin.

Sinclair et al. (1991) and Jordan and Flemings (1991) present computer-generated simulations of foreland basin stratigraphy that incorporate sedimentary erosion and distribution of tectonic loads. Sinclair et al. (1991) apply their model to the Cenozoic North Alpine Foreland Basin in Switzerland in an attempt to define possible controls on foreland basin stratigraphy during the evolution of an eroding thrust wedge. Assuming no eustatic sea level changes and elastic behavior of the lithosphere, they demonstrate that the foreland basin record of sedimentation and unconformities can be developed by changing the spatial distribution of a load.

Jordan and Flemings (1991) examine the foreland basin record through the interaction of subsidence, sediment flux, efficiency of sediment transport, and the period and amplitude of sea level changes. Variance of these parameters in the model generates distinct patterns of sedimentation and erosion associated with thrust-generated subsidence and changes of eustatic sea level.

Other recent work on foreland basin dynamics has focused on the sedimentary record of the basin. For example, Plint et al. (1993), in studies of upper Cretaceous strata in the Alberta foreland basin, note depositional patterns that include surfaces of erosive beveling at least 300 km cratonward of the present day Sevier deformation front. They interpret the regional truncation of strata to reflect forebulge uplift and erosion associated either with episodic loading/tectonic rejuvenation of an orogenic wedge or continuous loading of lithosphere of laterally varying flexural rigidity.

Patterns and implications of Acadian Sedimentary Fill, Northern Appalachian Basin. Figure 5 shows the geometry of the Oriskany, Esopus, Schoharie, Onondaga, and Union Springs Formations along ~450 km of the New York outcrop between Kingston and Buffalo. Distinctive wedge-shaped geometries characterize the Esopus and Union Springs Formations. This is in sharp contrast with the relatively tabular form of the Onondaga Limestone across New York. The Schoharie Formation, which lies between the Esopus and Onondaga Formations, exhibits an intermediate, modified wedge-shaped form. The geometry of the Oriskany and equivalent formations appears similar to that of the Schoharie Formation; the occurrence of Oriskany sands in karstic cavities and fissure fillings into underlying older Devonian and Silurian rocks across central to western New York appear to imply, however, a wider, and possibly more tabular distribution of the Oriskany Formation at the time of its deposition.

Strata equivalent to the Oriskany and Onondaga Formations are widespread across eastern North America. Rocks coeval with the Oriskany Sandstone occur continuously from Quebec and New Brunswick to Mexico and are found in the Appalachian, Michigan, and Illinois Basins (Boucot and Johnson, 1967, Figure 3a, p. 48). The Onondaga Formation is similarly widespread; it is found from the James Bay region of Ontario to southeast Quebec and Maine to Georgia and to Illinois (Koch, 1981; Oliver et al., 1967). The Schoharie Formation is also widely reported (Boucot and Johnson, 1967, Figure 5, p. 50), despite its absence across parts of central to western New York. In contrast, the Esopus Formation and equivalents are only reported from the Appalachian Basin (Esopus and Beaverdam shales; Ver Straeten, in prep.) and from the Lower Devonian of New England and Quebec (Boucot and Johnson, 1967, Fig. 4, p. 49; Rehmer, 1976, Fig. 12, p. 212).

The right side of Figures 3 and 4 summarize the succession of events during Acadian Tectophases I and II as recorded in the Appalachian foreland basin fill. The onset of tectonism during Tectophase I is indicated by: 1) a relatively abrupt change from widespread shallow marine, intrabasinal quartz arenites and carbonates (Oriskany-Glenier-Connelly Formations) to deeper, extrabasinal siliciclastics (shale and siltstones of the Esopus Formation) shed from rising Acadian highlands; 2) volcanogenic strata preserved at the transition (Sprout Brook Bentonites), apparently associated with an episodic increase in volcanism in the adjacent magmatic arc of the Acadian Orogen; and 3) rapid subsidence of the proximal foredeep of the basin, accompanied by uplift of a peripheral bulge in central to western New York (timing of the uplift of the peripheral bulge is poorly constrained at this time, but is definitely post-Oriskany and pre-Onondaga).

Progradation of coarser extrabasinal siliciclastics into the northern part of the Appalachian basin was limited during Tectophase I and is generally restricted to the upper part of the Esopus Formation. Deposition of the overlying Schoharie Formation was associated with decreased clastic input and a gradational upward increase in intrabasinal carbonate influence. This decrease in clastic input is interpreted to indicate reduced supply of extrabasinal sediments and lowered relief in adjacent Acadian source areas during a period of decreased tectonism.

A return to fully carbonate-dominated deposition (Middle Devonian Onondaga Limestone and equivalents) across much of eastern North America, including within the Appalachian Basin, is indicative of relative tectonic quiescence and low relief tectonic highlands in the Acadian mountain belt. Thin K-bentonites occur at several horizons through the lower and middle parts of the Onondaga Limestone, however, indicating that minor volcanism continued throughout this quiescent subphase.

Geometry of the basin floor underwent major changes during early Southwood (Onondaga) time; Brett and Ver Straeten (1994) discussed the subsidence and/or migration of the Tectophase I peripheral bulge in central and western New York. The area of apparent highest relief on the bulge during latest Sawkill-earliest Southwood time (e.g., central Finger Lakes region of New York) underwent subsidence and became the Onondaga basin axis by Nedrow Member time.

The onset of Acadian Tectophase II is first indicated in the upper part of the Onondaga Formation carbonates. Similar to Tectophase I, the foreland basin fill records the following: 1) an abrupt change from widespread, shallow marine, intrabasinal carbonates to deep water, extrabasinal siliciclastics (black shales of the Bakoven Member, Union Springs Formation); 2) volcanogenic strata preserved at the transition (Tioga Bentonites), associated with a second period of episodic increase in explosive volcanism in the adjacent

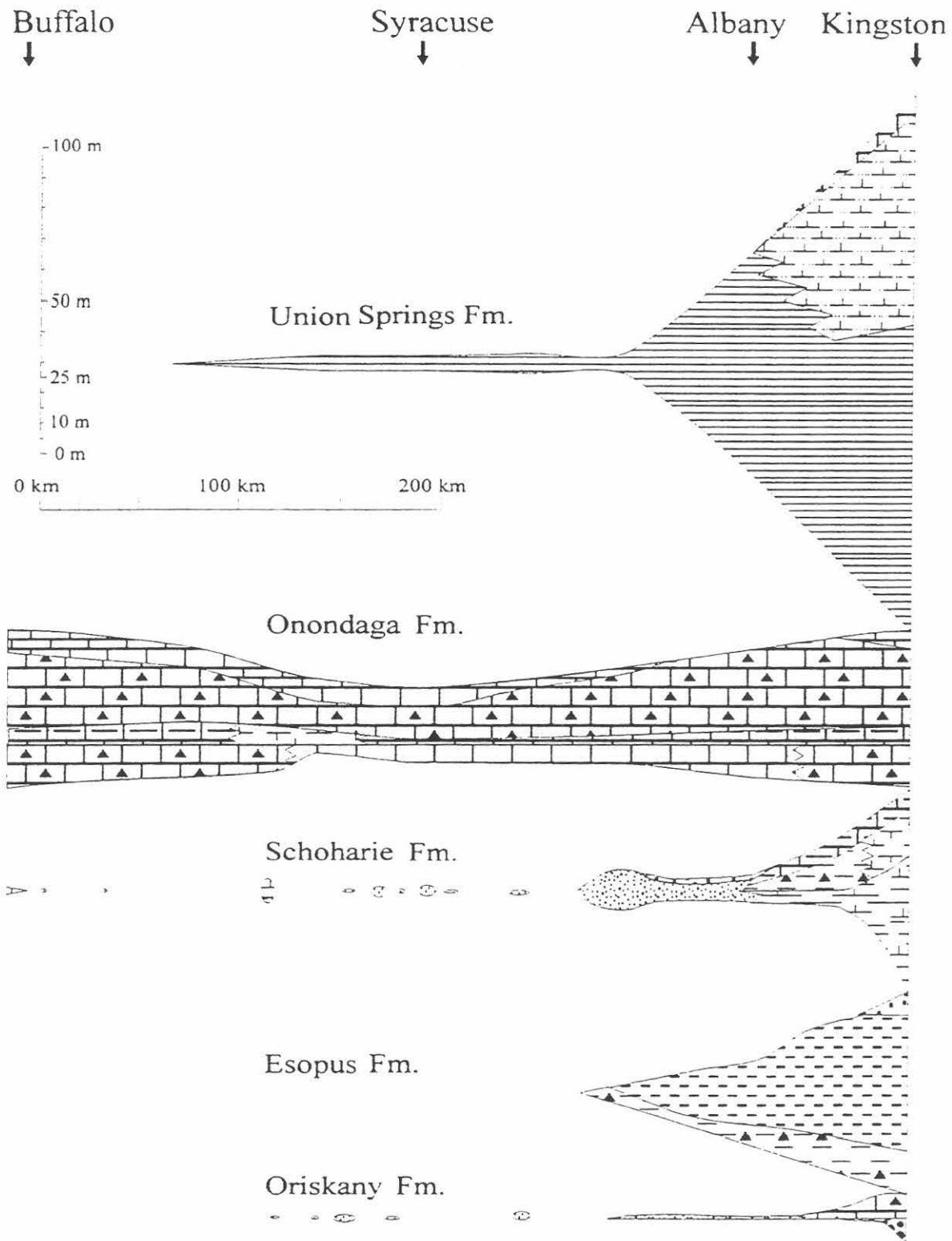


Figure 5. Geometry and generalized facies of upper Lower and lower Middle Devonian rocks along outcrop belt between Kingston (southeast) and Buffalo (west), New York. Note relatively tabular-shaped geometry of the Onondaga Formation in contrast to the distinct wedge-shaped geometry of the Esopus and Union Springs Formations. Thicknesses compiled from Ver Straeten (field notes), Hodgson (1970), Rehmer (1976), Johnsen (1957), Baker (1983), Oliver (1956), Feldman (1985), and Rickard (1989).

magmatic arc; and 3) rapid subsidence of the proximal foredeep (and associated sediment-starved conditions on the basin floor that led to the development of a submarine unconformity at the Onondaga-Union Springs formational contact) and uplift of a peripheral bulge in western New York and into southern Ontario. The uplift on the Tectophase II peripheral bulge appears to have begun in western New York at about the time of deposition of the Cherry Valley Member; Ver Straeten et al. (1994, p. 296) discuss multiple erosion surfaces in lower Marcellus strata and note coarse crinoidal grainstones in the Cherry Valley Member near Rochester. These occurrences are indicative of shoaling environments basinward of an area that was being uplifted as a general transgression occurred eastward along the remainder of the New York outcrop.

A decrease in tectonism during a medial stage of Acadian Tectophase II is marked by progradation and initial infilling of the foredeep of the basin by medium- to coarse-grained extrabasinal sediments in eastern New York (Mount Marion Fm.). Progradation of non-marine environments across eastern New York continued through deposition of the Hamilton Group (Ashokan, Plattekill, and Manorkill Fms.). An upward increase in intrabasinal carbonate content and dilution of extrabasinal clastics in equivalent marine rocks of the upper Hamilton Group (central to western New York) marks a progressive decrease in siliciclastic input from eroding highlands of Tectophase II. In this sense, the upper part of the Hamilton Group is similar to the Schoharie Formation during the early quiescent stage of Tectophase I.

A return to intrabasinal, limestone-dominated deposition of the Tully Formation is, similar to the older Onondaga Limestone, indicative of relative tectonic quiescence and lowered relief of tectonic highlands to the east in the Acadian orogen. Deposition of the marks the end of Tectophase II of the Acadian Orogeny.

Heckel (1973) reported a north-south striking anticlinal structure in central New York on the eastern margin of Tully Limestone deposition. This feature was a broad, low-angle area of topographic relief (~40-50 km across) that was active during deposition of the lower part of the Tully. In upper Tully time this feature subsided to form a topographic basin over the core of the breached anticline Heckel, 1973, p. 156).

This arch-like uplift to topographic basin progression during Tully time is strikingly similar to the dramatic subsidence of the Tectophase I peripheral bulge during early Onondaga time (central Finger Lakes region; Brett and Ver Straeten, 1994, see above). The authors postulate that Heckel's (1973) "anticline" of lower Tully time represents an uplifted Acadian Tectophase II peripheral bulge that subsided and/or migrated away by deposition of the upper part of the Tully Limestone.

The onset of Acadian Tectophase III is signified by a third abrupt transition from relatively widespread, shallow marine carbonates (Tully Formation) into overlying thick black shales (Geneseo Formation). This succession mimics those of the Oriskany to Esopus and Onondaga to Union Springs Formations. One exception to the previous depositional patterns is the absence of volcanogenic strata (K-bentonites) associated with this major transition. Closer inspection of this interval may yield another key cluster of Devonian K-bentonites associated with the initiation of the thick, Upper Devonian, Tectophase III clastic wedge of the Catskill Delta.

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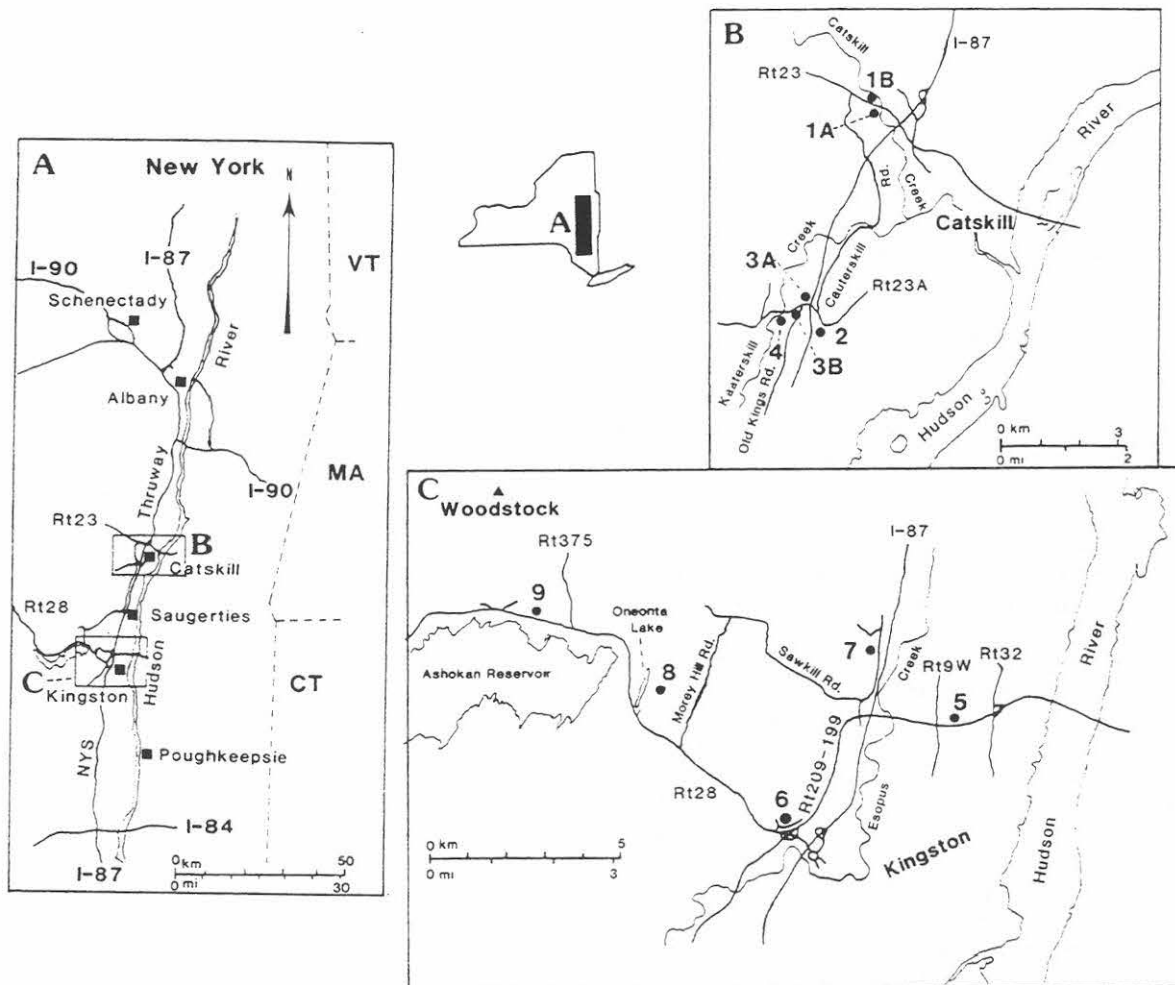


Figure 6. Maps of field trip area. A=eastern New York region with major highways. B=detailed map of Catskill (Greene Co.) vicinity; Stops 1A-4. C=detailed map of Kingston (Ulster Co.) area; Stops 5-9.

FIELD TRIP LOG

0.0	0.0	Leave parking lot opposite gymnasium; proceed to intersection with Union Avenue; <u>turn right</u> onto Union Avenue.
0.0	0.2	<u>Bear right</u> at stoplight onto Union Street.
0.3	0.1	<u>Turn left</u> onto Nott Terrace at stoplight.
0.5	0.2	Observe west wall of Mohawk Valley in distance on right.
0.7	0.2	Proceed through intersection with State Street; Nott Terrace changes to Veeder Avenue.
0.9	0.2	Veeder Street angles to right and becomes Miller Street; proceed downhill.
1.1	0.4	<u>Bear left</u> onto Broadway Avenue.
1.2	0.3	<u>Proceed into left lane</u> to enter Interstate 890 East.
1.3	0.1	<u>Turn left</u> onto entrance ramp for I-890 East.
1.5	0.2	<u>Merge</u> with I-890 East.
4.5	3.0	View of Helderberg Escarpment, capped by Lower and Middle Devonian strata, straight ahead.
5.15	0.65	Tollbooth for New York State Thruway; get toll card and <u>proceed to left lane</u> for Interstate 90/NYS Thruway eastbound. <u>Enter Thruway eastbound</u> and proceed through the Albany Pine Plain, on late-glacial windblown sand dunes near the northwest shore of glacial Lake Albany.
6.3	1.1	Guilderland Service area on right.

10.3	4.0	Proceed straight on NYS Thruway past Exit 24; note that the Thruway changes from I-90 to Interstate 87.
11.6	1.3	Highway continues through the Albany Pine Plain; note vegetated sand dunes along Thruway.
15.4	3.8	Taconic Mountains visible in distance ahead; remnants of Ordovician overthrust highlands of the Taconic Orogeny.
16.5	1.1	Exit 23; proceed straight on I-87 South.
17.2	0.7	Cross over Normanskill Gorge, near type section of Ordovician Normanskill Group.
17.6	0.4	Views of Helderberg Escarpment to right for next 17.5 mi.
19.8	2.2	Roadcuts in flysch-type sediments of Ordovician Austin Glen Formation (Normanskill Group) along NYS Thruway for next 15 mi.
23.9	4.1	Exit 21A (connector to I-90 and Massachusetts Turnpike; proceed straight on I-87.
31.9	8.0	Catskill Mountains, middle to Upper Devonian strata, visible ahead. Helderberg Escarpment on right, with limestone outcropping near crest of ridge.
32.9	1.0	Exit 21B (Coxsackie). Classic outcrop of "Dinosaur Leather", sole marks on base of Austin Glen Fm. turbidite bed visible 1.0 mi. north of Coxsackie exit along west side of N.Y. Rt. 9W.
35.1	2.2	Outcrops of Lower Devonian Kalkberg Limestone (Helderberg Group) visible on both sides of Thruway as you rise up onto the Helderberg-Kalkberg Escarpment. Exposures of Helderberg Group (Manlius, Coeymans, Kalkberg, New Scotland, and Becraft. Fms., in ascending order) for next 2.5 mi.
38.2	3.1	Long outcrop of middle part of Lower Devonian Esopus Formation (dark shale; Tristates Group) on right.
38.2	0.2	Strongly-dipping outcrop of Lower Devonian Glenerie Formation? (Tristates Group) on left.
38.9	0.7	New Scotland and Becraft Formations on left.
39.1	0.2	Esopus Shale on left.
39.6	0.5	Schoharie Formation on right; note dark band.
39.8	0.2	Low anticline on right.
40.2	0.4	Cherty Onondaga Formation (limestone) on both sides of I-87 for next 0.4 mi.
40.8	0.6	Cuts in Helderberg Group (Kalkberg, New Scotland, and Becraft Fms.) for next 3.0 mi.
42.8	2.0	<u>Bear right and exit I-87/NYS Thruway at Exit 21-Catskill</u> ; note additional outcrops of New Scotland and Becraft Limestones along exit.
43.3	0.5	View of Catskill Mountains/ Catskill Escarpment in the distance to right.
43.5	0.2	Tollbooth at Catskill Exit; New Scotland Limestone to left.
43.7	0.2	Junction with NY Rt. 23B; <u>turn left</u> onto highway.
43.8	0.1	New Scotland, then Kalkberg Formations on right.
44.0	0.2	<u>Bear right</u> onto entrance ramp for NY Rt. 23 West. Exit ramp to left (east) shows angular unconformity of Ordovician Austin Glen Formation with uppermost Silurian Rondout Group; outcrop continues through Lower Devonian Manlius, Coeymans, Kalkberg, and lower part of the New Scotland formations. Faulted section along right side of entrance ramp repeats Manlius, Coeymans, and Kalkberg Formations.
44.2	0.2	<u>Merge</u> with Rte, 23; observe outcrops of Helderberg Group Limestones to right and left. For detailed discussion of structure of rocks along Rt. 23, see Marshak, 1990.
44.35	0.15	Bridge over I-87-NYS Thruway.
44.6	0.25	Crest of anticlinal structure in roadcuts.
44.7	0.1	Bridge over Catskill Creek; note high outcrop of Esopus Shale on right.
45.0	0.3	<u>Pull over to right and park</u> near green mile marker sign past guard rail.

Stop 1A (Optional). Wallbridge Unconformity in Catskill Creek at Austin Glen. To get to Optional Stop 1, walk back along Rt. 23 to bridge and walk down to creek along side of bridge. Turn right at bank of Catskill Creek and follow around to the natural dam in creek south of the bridge.

The natural dam at the mouth of Austin Glen is formed by the resistant base of the Lower Devonian Glenerie Limestone, underlain successively by limestones of the Port Ewen and Alsen Formations at the top

of the Lower Devonian Helderberg Group. The basal bed of the Glenerie Formation is sand-rich, and features a prominent zone of reworked phosphate pebbles at its base. This phosphatic lag immediately overlies a prominent disconformity, the Wallbridge Unconformity (Sloss, 1963), which is one of six major unconformities in the Phanerozoic of North America. To the north and west, into central to western New York, this unconformity overlies progressively older rocks, as old as Late Silurian in western New York. South of Catskill the unconformity overlies progressively younger rocks in the upper part of the Port Ewen and Port Jervis Formations. In the Tristates area of southeastern New York, near Port Jervis, the succession from the Port Jervis Formation into the overlying Glenerie Formation is continuous and conformable (Rickard, 1975).

Downstream of the natural dam Catskill Creek runs through a gorge (Austin Glen) of the New Scotland and lower limestones of the Helderberg Group. The lower part of the gorge is the type section for the Ordovician Austin Glen Formation, seen previously on this trip in outcrops along the New York State Thruway. Upstream of the natural dam and the Rt. 23 bridge a high cliff of Esopus Shale is visible.

Return to the anticlinal outcrop adjacent to the vehicles.

Stop 1B. Esopus, Schoharie, and Onondaga Formations. Cuts on north side of N.Y. Rt. 23, west of bridge over Catskill Creek.

Highway cuts along Rt. 23 expose the upper part of the Esopus Formation, a complete section of the Schoharie Formation (unnamed lower, Aquetuck, and Saugerties Members), and the base of the Edgecliff Member of the Onondaga Formation in an antiformal fold west of Catskill Creek. The lower part of the exposure displays dark, rust-spotted, blocky-weathering silty shales of the Esopus Formation (Upper part of Depositional Sequence 2 of this paper). Along Rt. 23 the Esopus Shale is capped by a distinctive, 6.8 m-thick laminated unit that forms a distinctive marker at the base of the upper member of the formation. At Catskill the laminated unit is abruptly overlain by the Schoharie Formation; at Kingston, however, the top of the laminated unit underlies the Esopus-Schoharie contact by 22.5 m. The Esopus and Schoharie Formations thin laterally between Kingston and Catskill; however, the absence of upper Esopus strata above the laminated unit at Catskill appears dominantly to be associated with erosional truncation at a sequence-bounding unconformity below the Schoharie Formation (=base of Depositional Sequence 2 of this paper).

The sub-Schoharie unconformity at the base of Depositional Sequence 2 is marked by a sharp lithologic change and abundant glauconite, phosphate pebbles, fish bone material, and scattered white vein quartz pebbles. Five meters of calcareous, bioturbated argillaceous siltstone to silty mudstone of the lower unnamed member (formerly Carlisle Center Formation) overlie the contact. This contrasts sharply with up to 44 m of the unnamed member at Kingston. A 60 cm-thick, dark to black, rusty weathering, calcareous to cherty bed ("black bed" of Oliver et al., 1962) occurs near the middle of the member at both Catskill and Kingston. Glauconite and scattered quartz pebbles, apparently mixed downward from above by burrowing organisms, occur again in the upper part of the unnamed member at and below a lesser unconformity (=surface of maximum starvation in sequence stratigraphy terminology) at the base of the Aquetuck Member.

The Aquetuck Member along Rt. 23 (~12.7 m-thick) is characterized by fine-grained, buff- to olive-banded, cherty and siliceous to calcareous shales. An interval of interbedded olive and dark gray strata 2.6-5.2 m above the member base represent the most basinal conditions in the Schoharie Formation. This interval at Kingston is represented by ~3.25 m of dark, blocky-weathering, calcareous shale and is correlative with a widespread interval of dark gray to black shale in the middle of the Needmore Formation in central Pennsylvania (see body of paper). Carbonate content increases and the siliceous aspect of the strata decreases through the upper part of the Aquetuck Member at Catskill. The lower part of the member is unfossiliferous; small brachiopods and other forms appear in the upper part of the member, and diversify and increase in abundance upward through the overlying Saugerties Member.

The contact with the Saugerties Member is gradational and is placed at the lowest continuous, light-weathering limestone band. Nine meters of fossiliferous, interbedded buff calcareous shales and light-colored limestones comprise the Saugerties Member at Catskill. The lowest strata of the Onondaga Formation (Edgecliff Member) are exposed near the west end of the outcrop.

Return to cars and proceed straight (west) on NY Rt. 23.

45.2	0.2	<u>Turn left</u> onto Cauterskill Rd.-Greene Co. Rt. 47 and proceed ahead.
45.6	0.4	Junction with Vedder Road; continue ahead on Cauterskill Rd.
45.7	0.1	Intersection with Vedder Mountain Rd.; continue straight ahead. Outcrops of Union Springs and Mount Marion Formations on the Hoobergeberg Ridge to right.
46.5	0.8	Bridge over NYS Thruway; outcrops of Esopus to left along Thruway.
46.55	0.05	Outcrops of Schoharie Formation on left and right; note prominent cleavage.
47.5	1.0	Fine brown glacial sands exposed on right (possible shoreline sands of glacial Lake Albany?).
47.7	0.2	Bridge over Kaaterskill Creek; falls over Ordovician Austin Glen Formation on left.
47.75	0.05	<u>Turn right</u> and follow Cauterskill Road- Rt. 47 along Kaaterskill Creek.
48.4	0.65	Intersection with Cauterskill Creek Rd.; continue ahead.
48.7	0.3	Steeply-dipping Onondaga Formation on right.
49.0	0.3	Outcrop of Moorehouse and underlying Nedrow and Edgecliff Members of the Onondaga Limestone on left. <u>Follow road to left</u> past outcrop.
49.35	0.35	<u>Fork to left</u> on Cauterskill Rd.
49.4	0.05	<u>Turn left</u> onto N.Y. Rt. 23A
49.6	0.2	<u>Pull over to right</u> and <u>park</u> at abandoned NYS Thruway exit.

Stop 2. Port Ewen, Glenerie, and Esopus Formations, including the Wallbridge Unconformity.

Abandoned exit of New York State Thruway at N.Y. Rt. 23A. The lower and middle parts of the outcrop are accessible; the upper part of the outcrop is along the New York State Thruway and can only be observed with written permission of the Thruway Authority.

This Lower Devonian outcrop exposes uppermost strata of the Helderberg Group, the Wallbridge Unconformity, the Oriskany-equivalent Glenerie Limestone, and the lower part of the Esopus Shale. The exposure extends southward ~300 m along the New York State Thruway, where middle and upper Esopus strata are overlain by more calcareous strata of the Schoharie Formation.

The Wallbridge Unconformity, one of six major North American Phanerozoic unconformities (Sloss, 1963) caps argillaceous limestones of the Port Ewen Formation as seen at Optional Stop 1A. The unconformity is marked by a thin (~10 cm-thick) phosphate pebble-rich conglomeratic lag at the base of the Glenerie Formation. This bed lies at the base of Sloss' (1963) Kaskaskia Supersequence, which extends from the upper Lower Devonian through the Mississippian Period, and marks the base of Depositional Sequence 1 of this paper. The Glenerie Formation at Rt. 23A (3.5 m-thick) consists of relatively fine-grained, fossiliferous, drab brown limestones and dark gray cherts. Oriskany-age brachiopods are visible in cross-section in some limestone beds; the highly diverse and abundant fauna known from the Glenerie Limestone locally, however, appears absent at this outcrop.

The limestones and cherts of the Glenerie Formation are gradationally overlain by strata of the Esopus Formation at Stop 2. The lower member of the Esopus, as defined herein, consist of a tripartite subdivision of strata: 1) a lower interval of interbedded, thin siltstones, cherts, shales, and K-bentonites. This interval includes the Sprout Brook Bentonites of Ver Straeten (1992b, in review; also Ver Straeten et al., 1993), which at this outcrop consist of up to 15 thin, altered volcanic ash layers; 2) a middle subunit of dark gray to black, banded shale with scattered medium-size (~10-20 cm), dark blue-gray calcareous-phosphatic concretions; and 3) an overlying coarsening-up interval of medium dark gray silty shales to buff-gray-weathering, slightly calcareous siltstones. Small brachiopods (*Atlanticocoelia* and/or *Leptocoelia*) and other uncommon fossils occur in the upper subunit, with scattered glauconite.

The lower submember (~6.6m-thick) of the member is highly deformed at this outcrop; disharmonic folds, shear zones, and prominent fractures and cleavage are associated by Marshak (1990) with movement of a detachment fault between the Glenerie and Esopus Formations. Bentonite beds along the outcrop pinch and swell between more resistant beds and form horizons along which faults slide. The overlying middle dark gray-black shale subunit (2.0 m-thick) in the upper part of the cliff shows prominent slaty cleavage. Subunit 3 of the lower member of the Esopus Formation along Rt. 23A totals 4.0 m in thickness; its capping buff-gray siltstone bed forms a prominent ledge around the corner from the anticlinal exposure.

The overlying middle and upper members of the Esopus Formation extend southward along the outcrop and consist of rusty, dark gray silty mudstones to argillaceous siltstones that total over 40 m in

thickness. The upper member of the Esopus, unlike the previous stop, consists of the laminated unit and ~2.9 m of strongly *Zoophycos*-churned siltstones to fine sandstones. The base of the Schoharie Formation is marked by a prominent glauconite-rich bed with quartz pebbles. The overlying unnamed member of the Schoharie Formation is 9.2 m-thick along the Thruway cut and shows distinctive banding, as does the overlying Aquetuck Member (10.7+ m exposed; thicknesses for the Schoharie Formation reinterpreted from Johnsen, 1957). This banding may be related to Milankovitch cycles similar to those reported from Cretaceous and other age rocks (see discussion of Sequence DS3 in main body of paper).

The basal Glenerie phosphate pebble bed marks initial transgression during deposition of the first post-Wallbridge depositional sequence (=Sequence DS1 of this paper). Continued transgression coupled with the onset of tectonically-induced foreland basin subsidence (Acadian Tectophase I) marks the change into the overlying Esopus Formation. Shallowing and siliciclastic progradation through the upper part of the Esopus Formation is truncated by apparent erosion of upper Esopus strata (see discussion above) below an unconformity at the base of the Schoharie Formation (=base of Sequence DS2 of this paper).

Turn around and proceed back west along NY Rt. 23.

49.8 0.2 intersection with Cauterskill Rd.; proceed ahead on Rte 23A.

49.9 0.1 Bridge over NYS Thruway.

50.0 0.1 Intersection with Old Kings Rd.- Greene Co. Rt. 47. Pull over to right and park. Walk carefully back along Rt. 23A to bridge over the Thruway.

Stop 3A. Schoharie and Onondaga Formations. Highway cuts along NYS Thruway and NY Rt. 23A visible from Rt. 23A bridge. Access to exposures along thruway only through written permission of the Thruway Authority.

Highway cuts along Rt. 23 and the Thruway show extensive exposures of the Lower Devonian Schoharie and lower part of the Middle Devonian Onondaga Formations. Buff-orange, brown, and gray colors characterize the Schoharie Formation along the exposure. Light-weathering limestones, with common yellowish-weathering cherts, are typical of the Onondaga Formation. Very long exposures of the Schoharie are visible along the Thruway to the north. To the left, immediately north of the bridge, a darker, banded interval is notable in the lower part of the exposure. This interval of siliceous shales in the lower part of the Aquetuck Member (seen at Stop 1) is correlatable to a calcareous shale interval at the same position at Kingston; as previously noted, it is also recognized in the middle part of the Needmore Formation in Pennsylvania (see body of paper). Overlying strata on the northwest side of the bridge belong dominantly to the Aquetuck Member.

On the opposite side of the highway north of the bridge, alternating bands of argillaceous, buff-weathering limestones and purer, more calcareous limestone layers characterize the Saugerties Member of the Schoharie Formation. Overlying coarse, crinoid-rich, non-cherty to chert-rich limestones represent the Edgecliff Member of the Onondaga Formation.

Walk back to intersection of Rt. 23A and Old Kings Rd.

Stop 3B. Schoharie and Onondaga Formations. Cuts along N.Y. Rte 23A at and near its intersection with Kings Highway, ~0.0-0.3 mi. west of the bridge over the New York State Thruway.

Outcrops to the east and west of and along Old Kings Road on the south side of Rt. 23A expose argillaceous limestones of the upper part of the Schoharie Formation (Aquetuck and Saugerties Members) and a thick section of limestones of the Onondaga Formation (~44-m thick; Edgecliff, Nedrow and Moorehouse Members). The Schoharie-Onondaga contact is exposed on the east side of Old Kings Road, south of Rt. 23A. Buff- and light gray-weathering limestones with thin quartz sand stringers (Saugerties Member) are gradationally overlain by light gray, clean limestones of the Onondaga Formation (Edgecliff Member). The lower part of the Onondaga section exposed along Old Kings Road (basal 2.0 m-thick non-cherty interval and overlying chert-rich strata) comprise the Edgecliff Member (ca 16 m-thick). The lower 2.0 m of relatively chert-free, coral, brachiopod, and crinoid-rich pack- and grainstones ("Jamesville Quarry facies" of Brett and Ver Straeten, 1994) are capped by a thick section of medium-grained, cherty, crinoidal packstones ("Clarence facies" of Brett and Ver Straeten, 1994). Similar alternations of cherty and chert-free

strata characterize the Onondaga Limestone in the Catskill area; chert-dominated facies are predominant locally, however.

The member level subdivisions of the Onondaga Limestone that are well-defined and easily recognized in central New York (see Brett and Ver Straeten, 1994) are less readily detectable in Hudson Valley outcrops. They can tentatively be separated by changes in chert and argillaceous content, and in bedding thickness.

A 6.6 m-thick covered interval, (which is exposed along Cauterskill Road, Mile 49.0 of this trip) at least in part represents the Nedrow Member. Overlying strata to the west along Rt. 23A comprise the Moorehouse Member. The coarse- to medium-grained, chert and crinoid-rich lithology of the Edgecliff Member below is replaced in the Moorehouse Member by finer-grained, cherty, wacke- to packstones characterized by brachiopod-dominated faunas. Calcite slickensides are common along bedding planes through the Moorehouse Member at Stop 3B; faulting, however, appears to be restricted along the bedding planes, associated with slippage of beds over each other during folding and overturn of the section.

The Onondaga Formation totals ~44 m at Stop 3B; uppermost strata of the formation are covered. Fifty meters of Onondaga strata are reported from outcrops along the NYS Thruway at Saugerties, 8.5 mi. to the south (Feldman, 1985); the top of the Saugerties section is thought to be close to the contact of the Onondaga Limestone and overlying clastics of the Union Springs Member (Stop 4, next locality).

A geologic overview of Localities 1-3 and 5 is summarized here (see Figure 3). Post-Wallbridge deposition of the Glenerie, Esopus, Schoharie, and Onondaga Formations represent a succession of, respectively, intrabasinal carbonates, extrabasinal siliciclastics, mixed siliciclastic and carbonate sediments, and a return to fully intrabasinal carbonates. These trends are associated with initial tectonic quiescence and carbonate-quartz arenite deposition above the Wallbridge Unconformity (Glenerie Formation), followed by: 1) a pulse of active tectonism in the Acadian Orogen, accompanied by volcanism and deposition of the Sprout Brook Bentonites, transport and deposition of fine- to medium grained siliciclastics, and subsidence of the foreland basin (Esopus Formation); 2) decreased tectonic activity and lowering of topographic relief in the mountain belt, associated with a decrease in siliciclastic input into the basin and deposition of calcareous shales and siltstones to argillaceous limestones (Schoharie Formation); and 3) a return to tectonic quiescence, marked by deposition of intrabasinal carbonates (Onondaga Formation). This succession of events 1-3 comprises Acadian Tectophase I as it is recorded in the foreland basin fill of eastern New York State.

Superimposed on these tectonically-related trends are three major depositional sequences (DS-1 to DS-3). These sequences appear to be dominantly controlled by eustatic sea level changes. The first sequence (DS 1) comprises the Glenerie and Esopus Formations. Initial deepening above the Wallbridge Unconformity continues upward through the Glenerie Limestone and lower to middle Esopus Shale (note: deepening during Esopus time is in part due to tectonically-induced subsidence of the basin). Upward shallowing through the upper middle to upper Esopus Formation is truncated by submarine erosion below a sub-Schoharie unconformity at the base of DS-2. Initial deepening through the Schoharie Formation culminates in deposition of widespread dark shaly strata represented at Catskill by an interval of dark, siliceous to cherty shales in the lower part of the Aquetuck Member. Overlying increasingly more calcareous strata are associated with a general shallowing-upward reaches a maximum at or near the Schoharie-Onondaga formational contact. Initial deepening at the base of the Onondaga Limestone marks the base of DS-3; transgression continued upward through the Edgecliff and Nedrow Members. Trends through the overlying Moorehouse Member indicate another major shallowing-upward trend during late highstand of DS-3. Transgression during the succeeding DS-4, which is dominantly represented by the overlying Union Springs Formation, was initiated during deposition of the upper part of the Moorehouse Member.

Return to the cars and proceed ahead west on NY Rt. 23A.

- 50.4 0.4 Cross over Kaaterskill Creek at Webber Bridge. Drive through the Bakoven Valley, with deeply eroded black shales of the Bakoven Member (Union Springs Formation) overlain by glacial sediments.
- 50.8 0.4 Turn left onto Underhill Rd., then turn around and return east of Rt. 23. Note outcrop of Stony Hollow Member (Union Springs Formation) at intersection; more outcrops of

Stony Hollow Member northward along Underhill Rd and at the crest of Rt. 23A to the southwest.

- 51.3 0.5 Recross Kaaterskill Creek and pull over and park on right side of Rt. 23. Walk back along the highway and down steep bank to the creek.

Stop 4. Onondaga Formation and Bakoven Member of the Union Springs Formation. East bank of Kaaterskill Creek, below the N.Y. Rt. 23A bridge. Type section of the Bakoven Member. PRIVATE PROPERTY.

The stream bed and east bank of Kaaterskill Creek expose the upper 0.85 m of the Onondaga Formation and the lower 27 m of the Bakoven Member of the Union Springs Formation. The abrupt change from shallow marine carbonates to basinal black shales seen in the outcrop mark the onset of Acadian Tectophase II.

Light-weathering, relatively fine-grained limestones with atrypid brachiopods, small rugose corals, and other forms characterize the Onondaga Formation at this locality. The upper 35 cm of limestone appears slightly darker in color and overlies a thin (1-cm-thick) K-bentonite. Key upper Onondaga marker beds (e.g., the Tioga B-OIN bentonite at the base of the Seneca Member) are not seen at this outcrop; it is not known whether the exposed strata belong to the Moorehouse or Seneca Members. A thicker but partially covered outcrop (2.9 m-thick) of upper Onondaga strata occurs immediately north of Rt. 23 east of the bridge.

The Onondaga-Union Springs formational contact is abrupt, and features numerous bored phosphate pebbles and well-defined burrows that are visible in the creek when the water level is low. The contact is overlain by a thin (ca 2-3 cm-thick), highly condensed bed of phosphatized fossil debris (e.g., crinoid ossicles) and fish bone and teeth (NOTE: Please do not collect from the bone bed-this is a classic section and type locality). The bed marks a major sediment-starved flooding surface associated with tectonic subsidence of the Appalachian foreland basin at the onset of Acadian Tectophase II of Ettensohn (1985). For further discussion of this contact, see Ver Straeten et al. (1994) and Lindemann and Feldman (1987).

Overlying rusty-weathering, fissile black shales of the Bakoven Member represent deposition of fine-grained, organic-rich sediments in the proximal subsiding trough of the Appalachian foreland basin. Anaerobic conditions on the sea floor prevented colonization by benthic faunas; only rare leiorhynchid brachiopods and pelagic (?) dacyroconariids (*Styliolina*) and rare cephalopods occur in the shales. Three thin intervals of more calcareous strata (styliolinid packstones) occur upward through the section. The poorly resistant shale features numerous tectonized intervals (small shear zones and mesoscopic folds; Marshak, 1990) and fault surfaces, especially above the Onondaga Limestone in the lower part of the Bakoven Member.

The low valley to the west is floored by the Bakoven Shale, and is partially infilled with Pleistocene sediments. Roadcuts on the west side of the valley exposes the Stony Hollow Member of the Union Springs Formation (not the Mount Marion Formation as reported by Murphy et al., 1980; Marshak, 1990; Cassie, 1990).

Return to cars and proceed ahead (east) on Rt. 23.

- 51.5 0.2 Pass overturned to steeply-dipping Onondaga Limestone of Stop 3B.
51.6 0.1 Turn right onto Old Kings Rd.-Greene Co. Rt. 47.
51.7 0.1 Schoharie Formation on left.
52.9 1.2 Esopus (?) Formation on right.
53.4 0.5 Catskill Escarpment and Hoogeberg Ridge visible in far and middle distance to right.
53.6 0.2 High Falls Rd. on right; road to scenic High Falls on Kaaterskill Creek, an excellent section of Otsego Member of the Mount Marion Formation.
54.1 0.5 Catskill Escarpment and Hoogeberg Ridge to right.
54.4 0.3 Onondaga Formation on right.
55.7 1.3 Onondaga exposure behind house on left.
56.0 0.3 Watch for "beefalo" in pasture to left of road.
56.3 0.3 Excellent view of Catskill Escarpment, with Kaaterskill Clove visible to right and back a little.

57.5	1.2	Katsbaan Reformed Dutch Church, built 1732. Stone walls dominantly made of blocks of Onondaga Limestone, except front wall (brown-green sandstones of Austin Glen Formation).
58.2	0.7	Intersection with Ulster Co. Rt. 34; continue straight.
58.35	0.15	<u>Turn left</u> onto NY Rt. 32 southbound. Outcrop of Schoharie- Onondaga contact on right after turn (a left turn at the intersection would pass outcrops of the Onondaga Limestone, the Stony Hollow Member (Union Springs Formation), the Otsego Member and undifferentiated upper Mount Marion Formation, and lowest strata of the Plattekill Formation.
58.7	0.35	Saugerties Member (Schoharie Formation) on left.
58.8	0.1	Schoharie-Onondaga formational contact on right.
59.0	0.2	Schoharie-Onondaga formational contact on right.
59.2	0.2	Excellent Mexican restaurant on left.
59.6	0.4	Onondaga Formation on left.
59.7	0.1	Site of 1994 Woodstock Festival on right.
59.9	0.2	<u>Turn left</u> to enter NYS Thruway/I-87 southbound.
59.95	0.05	Get card at toll booth.
60.0	0.05	<u>Merge</u> onto I-87.
60.1	0.05	Long outcrop of the Saugerties Member (Schoharie Fm.) and most complete section of the Onondaga Formation in the Hudson Valley (along both sides of I-87).
60.8	0.7	Large quarry of dark, shale-dominated Berne and Otsego Members (Mount Marion Fm.) on north shoulder of Mount Marion to right.
61.6	0.8	Crest of Mount Marion, highest hill in the Hoogeberg Ridge. Type section of Mount Marion occurs at south end of the ridge, in a high south bank of Plattekill Creek.
63.9	2.3	Outcrop of Stony Hollow Member (Union Springs Fm.) on left.
64.3	0.4	Bridge over Plattekill Creek; outcrop of upper Stony Hollow Member (Union Springs Fm.) or Cherry Valley Member (Mount Marion Fm.) in creek bed on right.
64.9	0.6	<u>Bear right</u> onto exit ramp at Ulster Service Area - <u>REST STOP</u> .
65.1	0.2	<u>Re-enter</u> I-87 South.
66.8	1.7	Middle part of Berne Member (Mount Marion Fm.).
67.1	0.3	Cross over Esopus Creek.
67.4	0.3	Cross under N.Y. Rt. 209.
67.5	0.1	Outcrops of Stony Hollow Member along I-87 for next 1.7 mi..
69.1	1.6	<u>Exit</u> from I-87/NYS Thruway at Exit 19-Kingston.
69.9	0.8	Pay toll at booth.
69.9	0.05	<u>Turn right</u> onto traffic circle.
70.0	0.05	<u>Fork right</u> onto N.Y Rt. 28 West.
70.1	0.1	Cross over NYS Thruway/I-87.
70.3	0.2	<u>Bear right</u> onto entrance ramp for U.S. Rt. 209 North.
70.4	0.1	<u>Merge</u> With Rt. 209 North.
70.5	0.1	Long outcrop of Stony Hollow Member, both sides of highway.
70.7	0.2	Hurley Member (type section; Union Springs Fm.) and Cherry Valley Member (Mount Marion Fm.) exposed on left.
70.9	0.2	lower part of continuous section of Berne Member (Mount Marion Fm.).
71.3	0.4	Berne and Otsego Members (Mount Marion Fm.) boundary ~2/3 height up bank on both sides.
71.9	0.6	Berne and Otsego Members in high bank on right.
72.2	0.3	Cross over Sawkill Road.
72.4	0.2	Cross over I-87/NYS Thruway.
72.7	0.3	Bridge over Esopus Creek.
73.5	0.8	Cross over NY Rt. 9W.
73.7	0.2	Outcrops of Esopus and Schoharie Formations on both sides of highway.
74.05	0.35	Outcrops of (in descending order) Glenerie, Port Ewen, Alsen, and Becraft Formations.

74.25	0.2	Outcrops of (in descending order) New Scotland, Kalkberg, Coeymans, and Manlius Formations.
74.5	0.25	<u>Exit right</u> to N.Y. Rt. 32.
74.7	0.2	<u>Turn right</u> onto Rt. 32 and proceed under bridge.
74.8	0.1	<u>Turn right</u> onto entrance ramp for Rtes. 199/209 West.
74.9	0.1	Outcrops of Ordovician Austin Glen Formation on both sides of entrance ramp. <u>Enter</u> Rtes. 199/209 and proceed along road in opposite direction to which you just drove.
75.6	0.7	<u>Pull off highway</u> and park on right.

Stop 5. Esopus and Schoharie Formations. Synclinal exposure on north side of Rtes. 199/209 east of Rt. 9W, north of Kingston.

Outcrops on the north and south side of Rtes. 199-209 expose upper strata of the upper member of the Esopus Formation and the unnamed, Aquetuck, and lower part of the Saugerties Members of the Schoharie Formation. The north outcrop is less weathered, and better displays various features of the strata.

16.3 m of the upper member of the Esopus Formation are exposed at Stop 5. These strata represent siltstones to fine-grained sandstones that cap the formation locally. The laminated unit at the base of the upper member seen in the Catskill region (Stops 1B & 2) is absent at this outcrop, but is partially exposed 0.4 mi. south at the north end of the Hudson Valley Mall (east of Rt. 9W). The contact between the laminated subunit and the overlying bioturbated siltstones to sandstones is sharp at the mall exposure. A continuous, almost complete section of the upper member is exposed along a railroad cut north of O'Reilly Street in southwestern Kingston (Stop #3 of Oliver et al., 1962), where the member totals 27.5+ m in thickness; the contact with the underlying middle member is covered below 5 m of the laminated subunit. The overlying siltstone-fine sandstone subunit totals 22.5 m in thickness, in contrast with 2.9 m at Stop 2 and 0.0 m at Stop 1 in the Catskill area.

The top surface of the Esopus Formation is well exposed on an eastward dipping bedding plane on the outcrop, where well-preserved *Zoophycos* and other trace fossils are exposed. The contact with the overlying unnamed member of the Schoharie Formation is sharp. Scattered quartz and phosphate pebbles are visible in the succeeding strata. The unnamed member at Kingston consists of bioturbated, calcareous, silty mudstones to argillaceous siltstones that total 32 m in thickness. This contrasts with ~45 m at the railroad cut in Kingston and 9.2 m (Stop 2) to 5.0 m (Stop 1) in the Catskill area. Subtle banding is visible in the unnamed member along the outcrop. The prominent black, siliceous marker bed of the unnamed member (black bed" of Oliver et al., 1962; 0.7 m-thick) is visible in the middle of the member along Rtes. 199-209.

The contact with the overlying Aquetuck Member (Schoharie Formation) is gradational. The highest white quartz pebbles in the upper part of the unnamed member were used by Oliver et al. (1962) to define the top of said member. These occur just above the uppermost prominent white band of limestone in the middle of the outcrop.

The siliceous/cherty nature of the overlying Aquetuck Member in the Catskill area and north is replaced by more calcareous strata in the Kingston area. The dark, siliceous, shaly interval near the base of the member is represented by a 3.2+ m-thick interval of blocky-weathering shales and thin limestone bands ~3 m above the base of the member. Prominent "coffee and cream" banding is well-developed in the Aquetuck Member as well as the other members of the Schoharie Formation along the outcrop. White nodular limestone bands in the Aquetuck Member are replaced upward by more continuous, tabular limestone beds in the overlying Saugerties Member.

Return to cars and proceed ahead on Rtes. 199/209.

77.0	1.4	<u>Recross</u> over NYS Thruway.
79.0	2.3	<u>Exit right</u> onto ramp for NY Rt. 28 West.
79.5	0.2	Merge with New York Rt. 28 West.
79.6	0.1	<u>Turn right</u> at stoplight onto Forest Hill Rd.
79.65	0.05	<u>Make an immediate right turn</u> onto City View Terrace.
79.47	0.05	<u>Park</u> along roadside before guard rail across from outcrop.

Stop 6. Bakoven and Stony Hollow Members of the Union Springs Formation. Cut along City View Terrace immediately east of N.Y. Rt. 28.

This roadcut and a subadjacent cut along Rt. 28 expose the upper 10 m of the Bakoven Member and the lower 23 m of the overlying Stony Hollow Member (Union Springs Formation). The Bakoven Shale at Kingston consists of ~100 m of fissile black shale (Rickard, 1989, Ver Straeten et al., in prep.). Similar to Stop 4, the member is nearly barren of fossils; styliolinids, cephalopods, and *Panenka* bivalves comprise the dominant forms found here. The Bakoven Member features numerous deformational structures similar to those at Stop 4, which include several zones of sheared and mesoscopically-folded shales below the overlying resistant strata of the Stony Hollow Member.

The Bakoven-Stony Hollow contact is gradational at Kingston. Calcareous, buff-weathering, thinly laminated to burrow-mottled lower Stony Hollow strata also feature a low diversity, anaerobic to dysaerobic fauna of dacroconariids and rare, small bivalves and brachiopods. Grain size increases upward through the Stony Hollow Member from laminated claystone-mudstone couplets in the lower part to fine-grained sandstones near the top of the unit. Biofacies upgrade from basinal anaerobic/dysaerobic dacroconariid-dominated faunas to fully aerobic brachiopod-crinoid-trilobite shallow marine assemblages through the Stony Hollow and Hurley Members. A dysaerobic, dacroconariid-leiorhynchid brachiopod-cephalopod assemblage in the overlying Cherry Valley Member indicates a return to deeper water conditions associated with transgression. The bottom of the Cherry Valley Member is a sequence boundary at the base of the Mount Marion Formation. The entire Stony Hollow and Hurley Members (type sections) and Cherry Valley Member (eastern clastic facies reference section) are exposed in the immediately adjacent area (see Griffing and Ver Straeten, 1991, p. 247-249).

Return to cars and proceed ahead on City View Terrace.

- | | | |
|-------|------|--|
| 79.8 | 0.1 | <u>Turn around</u> in parking lot of Potter Brothers Ski Shop and return to Forest Hill Rd. |
| 80.0 | 0.2 | <u>Turn left</u> onto Forest Hill Rd., then <u>turn left</u> onto Rt. 28 East at stoplight. |
| 80.3 | 0.3 | <u>Proceed into right lane.</u> |
| 80.4 | 0.1 | <u>Bear right</u> onto entrance ramp for Rt. 209 North. |
| 80.5 | 0.1 | <u>Merge</u> onto Rt. 209 North. |
| 82.7 | 2.2 | <u>Bear right</u> onto exit ramp for Sawkill Rd. |
| 82.75 | 0.05 | <u>Turn left</u> onto Sawkill Rd. |
| 83.05 | 0.3 | Cross Sawkill Creek. |
| 83.1 | 0.05 | <u>Fork right</u> onto Ulster Co. Rt. 31. |
| 83.15 | 0.05 | Berne Member on left. |
| 83.4 | 0.25 | Outcrop of Berne Member adjacent to outcrop along I-87/NYS Thruway (66.8 mi. of trip). |
| 83.6 | 0.2 | Berne Member on left. |
| 83.7 | 0.1 | <u>Turn left</u> into parking lot of Buzzanco's Greenhouses; <u>park and walk</u> to quarry exposure north of greenhouses. |

Stop 7. Berne and Otsego Members of the Mount Marion Formation. Abandoned quarry along Ulster Co. Rte 31, north of Buzzanco's Greenhouses. PRIVATE PROPERTY: contact Mr. Buzzanco at the greenhouses. Locality 3 of Ver Straeten, 1994.

The upper part of the Berne and lower part of the Otsego Members of the Mount Marion Formation (Depositional Sequence 5 of this paper) are exposed in this abandoned shale pit). ~55 m of dominantly dark gray mudstones and claystones crop out in the quarry. Cyclically-occurring shell beds, generally separated by 3-8 m of unfossiliferous mudstones, feature distinctive brachiopod-dominated faunas that in general increase in diversity upward through the interval exposed. These small-scale cycles represent individual parasequences. The strata are also punctuated by packages of thin sandstone beds; recent work shows that most of the shell beds and clusters of thin sandstones have been successfully correlated along the outcrop belt along much of the eastern New York outcrop belt (> 110 km, Kingston to Schoharie; Ver Straeten, 1994).

The base of the quarry lies ~85 m above the top of the Union Springs Formation. Regionally the contact of the Berne and Otsego Members is marked by the occurrence of a richly fossiliferous coral-

brachiopod biostrome (Halihan Hill Bed of Ver Straeten, 1994). This bed, which occurs ~12 m up in the quarry section, marks the first known occurrence of the classic fauna of the Middle Devonian Hamilton Group. Regional study indicates the bed represents initial flooding after a significant fall in relative sea level within Depositional Sequence 5; an abrupt contact of the Halihan Hill Bed with underlying claystones is due to local to regional submarine erosion of underlying sand-rich strata during sea level lowstand (Ver Straeten, 1994).

Overlying dark shale-dominated strata mark a return to deeper water settings associated with continued transgression. Low diversity shell beds, characterized by leiorhynchid and *Mediospirifer-Athyris* brachiopod associations characterize a 30 m-thick interval above the Halihan Hill Bed. Shell beds in the upper part of the quarry show an increase in diversity, apparently associated with a decrease in water depth. A package of thin sandstones at the top of the quarry are succeeded locally by a prominent, richly fossiliferous sandstone (up to 2.5 m-thick) that forms the caprock of waterfalls and ridges along the Hooageberg Escarpment in eastern New York (e.g., High Falls of Kaaterskill Creek, SW of Catskill).

Return to vehicles and return along Rt. 31 to Sawkill Rd.

- | | | |
|-------|------|---|
| 84.3 | 0.6 | <u>Turn right</u> onto Sawkill Rd. |
| 84.8 | 0.5 | Otsego Member on right. Exposures of middle Mount Marion strata for next 1.0 mi. |
| 85.2 | 0.4 | Cross over Sawkill Creek in village of Sawkill. |
| 85.5 | 0.3 | Hill Rd. on left; road to extensive turn-of-the-century flagstone quarries in upper part of the Mount Marion Formation. |
| 86.15 | 0.65 | Intricately decorated house on right side of Sawkill Rd. |
| 86.5 | 0.35 | Jockey Hill Rd. on left; road to more abandoned flagstone quarries. |
| 87.05 | 0.55 | <u>Turn left</u> onto Morey Hill Rd. |
| 87.5 | 0.45 | Massive sandstones of upper Mount Marion Formation on right. |
| 88.3 | 0.8 | Quarry in upper part of Mount Marion on right behind houses. |
| 88.7 | 0.4 | Roadside cuts in upper middle part of Mount Marion Formation for next 1.0 mi. |
| 89.3 | 0.6 | <u>Turn right</u> onto NY Rt. 28. |
| 89.9 | 0.6 | Upper part of Mount Marion Formation for next 0.3 mi. |
| 90.2 | 0.3 | <u>Turn right</u> into entrance to abandoned flagstone quarries; <u>park and walk</u> up roadway past concrete barriers. Extensive exposures throughout the quarries. this stop will concentrate on rocks along the entrance road and in the upper quarry behind abandoned buildings on quarried cliff. |

Stop 8. Upper part of the Mount Marion Formation. Abandoned flagstone quarry north of N.Y. Rt. 28 and east of Onteora Lake. PRIVATE PROPERTY.

This abandoned quarry features excellent exposures of nearshore marine to non-marine transitional deposits in the upper part of the Middle Devonian Mount Marion Formation. The upper strata of the formation are dominated by cyclic alternations of interbedded mudstone-sandstone facies and flaggy-bedded sandstone bodies on the order of 10 m total thickness. The coarser, sand-dominated strata may feature thin beds to lenses of quartz and chert pebble conglomerate. Normal marine faunas (brachiopods, bivalves, etc.) occur through most of the section; however, apparent plant-root traces in some beds indicate periodic subaerial exposure during deposition.

This locality was originally quarried around the turn of the century for bluestone flags that were used in New York City for sidewalks and curbstones; the quarry was reactivated during the 1970's for production of concrete aggregate (Banino, 1987). Two main levels were worked for quarry stone in the past. The main quarry face of the lower level exposes ~9 m of section, which is comprised of: a) a lower, 1.4 m-thick sandstone body; b) a middle 5.5 m-thick unit of interbedded dark gray mudstones and thin sandstones, which coarsens toward the top; and c) 2.0 m of flaggy-bedded sandstones at the top.

The upper level of the quarry features extensive lateral exposures of the flaggy-bedded sandstones. The top 0.5 m of the middle mudstone unit of the lower quarry face are exposed in the floor of the upper quarry, where scattered brachiopods (*Mucrospirifer* and *Devonochonetes coronatus*) and bivalves can be found. A large area of the upper quarry is floored by a 10 cm-thick quartz-rich sandstone with a symmetrically rippled top and common *Cruziana* trilobite traces on the underside of the bed.

The mudstone unit is overlain by conglomerate lenses and/or a pebbly, coarse sandstone bed at the base of the quarry face. The overlying interval (~1 m-thick) is characterized by thin- to medium-bedded sandstones and thin, black, fine-grained, organic- and pyrite-rich beds. These thin (~1-5 cm-thick), recessed interbeds contain abundant plant material, which include lenses of *in situ* root traces; yellow to white mineral crusts (Melanterite and Epsomite?) weather out of the crevices, probably associated with the decay of pyrite. Single valve rhynchonellid brachiopod shells also occur in some of the thin sandstones.

Overlying strata (ca 5 m-thick) are dominated by medium-bedded lithic arenites that appear planar- to cross-bedded. Pebbly lenses occur scattered through the section. Extensive bedding plane exposures of the flagstone interval are visible in the southern part of the quarries, where several unusual trace fossils have been found. A discontinuous but prominent crevice, ~halfway up the flaggy sandstone unit, is bounded by two lighter- and more smoothly-weathering sandstone beds. The upper bed (~40 cm-thick), which forms the caprock of the east wall, appears in side view to feature *in situ* plant root traces. Overlying sandstones are darker and flaggy bedded, similar to the beds below.

The top surface of the sandstone body is rippled (symmetric ripples) and is abruptly overlain by olive-gray colored, crumbly-weathering, sandy mudstones and thin sandstones. Relatively large *Camarotoechia* and *Mucrospirifer* brachiopods occur in the upper part of the sandstones (as densely packed lenses) and the overlying sandy mudstones. North of the quarry, in the woods, a relatively steeply-dipping body of thin, highly tough cross-bedded sandstones overlies the position of the olive-gray shales. It is not clear at present if this unit represents a submarine or subaerial channel-form deposit.

Return to the vehicles and return to Rt. 28.

- | | | |
|------|-----|---|
| 90.3 | 0.1 | <u>Turn right</u> onto Rt. 28 and continue westward toward the Catskill Mountains. |
| 90.4 | 0.1 | Upper Mount Marion strata. |
| 90.7 | 0.3 | Marine shales in uppermost Mount Marion in small quarry on right. |
| 90.8 | 0.1 | Top of marine section at quartz-rich sandstone at base of exposure on left hand side (= top of Mount Marion Fm.); overlying thick sandstones are in the succeeding Ashokan Formation. |
| 91.3 | 0.5 | Intersection with Zena (on right) and Rock (on left) Rds. |
| 91.7 | 0.4 | Exposure of Ashokan Formation on right side. |
| 92.2 | 0.5 | Intersection with NY Rt. 375 (to Woodstock); proceed straight on Rt. 28. |
| 92.6 | 0.4 | Outcrops of Plattekill Formation on left and right; lowest redbed exposures along Rt. 28. |
| 92.7 | 0.1 | Plattekill Formation on right. |
| 92.9 | 0.2 | <u>Pull off</u> to right side of road and <u>park</u> at outcrop. |

Stop 9. Plattekill Formation. Cuts along N.Y. Rt. 28, 0.7 mi. west of intersection with N.Y. Rt. 375.

This outcrop of red, green, and olive mudstones (lower part) and medium-gray, cross-bedded sandstones (upper part) of the Plattekill Formation are a gross generalization of facies that characterize the remainder of the Middle and Upper Devonian of the Catskill Front. The lower 5 m of green, green-mottled red, and red mudstones feature calcrete nodules, indicative of development of paleosols on subaerial flood plains. Plant fossils and root traces are visible throughout much of the section; distinctive weathering styles appear to characterize different paleosol units along the outcrop. Overlying olive-weathering mudstones (~1.5-3.0 m-thick) are erosively capped by ~4 m of planar cross-bedded, lithic arenites that represent deposition within a migrating fluvial channel.

Pedersen et al. (1976) reported a zone of deformation in a paleosol horizon ~2.5 above the base of the section. These structures on close examination appear to resemble separate, low angle, basin- or bowl-shaped structures ~0.5 to 1.0 m in diameter. They are similar to "pedogenic slickensides" described from redbeds of Silurian-, Devonian-, and Pennsylvanian-age by Gray and Nickelsen (1989). Slickenlines on the bottom-side of the structures visibly radiate outward and upward through 180° laterally on the outcrop. Pedogenic slickensides form in modern vertisols associated with seasonal wetting and drying of expansive clays in the B soil horizon (Gray and Nickelsen, 1989; Ciolkosz et al., 1979).

The succession of events for Stops 4 and 6-9 and overlying strata are summarized here (see Figure 4). Six major depositional sequences are recognized above (and including) the top of the Onondaga Limestone in New York State (see body of paper above). These sequences correspond to five formation-level subdivisions of the Hamilton Group (upper Onondaga and Union Springs Fms., Mount Marion/Oatka

Creek Fms., Skaneateles Fm., Ludlowville Fm., and the Moscow Fm., Sequences DS4-8, respectively) and the overlying Tully and Geneseo Formations (Sequence DS9). In the Catskill Front, marine rocks of Sequences DS4 and DS5 feature recognizable transgressive and highstand systems tracts (see body of paper). Strata of the overlying Sequences DS6-9 are fully non-marine; no sequence stratigraphic framework for these fluvial-dominated rocks has been proposed (however, see Bridge and Willis, 1994).

A larger-scale set of trends related to Acadian Tectophase II are superimposed onto Depositional Sequences DS4-9. Widespread, intrabasinal carbonate deposition during the late, quiescent stage of Acadian Tectophase I (Onondaga Limestone) is succeeded by a thick, wedge-shaped body of siliciclastics associated with Acadian Tectophase II (Hamilton Group). The progression of Tectophase II is characterized by: 1) onset of renewed tectonism accompanied by increased volcanism and deposition of the Tioga Bentonites, subsidence of a proximal foredeep (eastern NY) and uplift of a peripheral bulge (western NY), and transport/deposition of initially fine-grained black shale. This was followed by the onset of progradation of progressively coarser clastic marine sediments into eastern New York (Union Springs and Mount Marion Fms.); 2) decreased tectonism in the Acadian orogen and progradation of non-marine facies into eastern New York (Ashokan, Plattekill, and Manokill Formations). Decreasing siliciclastic input is indicated by a general upward increase in carbonate content in upper Hamilton Group strata in central to western New York (Ludlowville and Moscow Fms; this trend is analogous to general trends through the Schoharie Formation during Tectophase I); and 3) a return to tectonic quiescence and low relief of tectonic highlands in the late Middle Devonian, indicated by deposition of intrabasinal limestones (Tully Formation). Stages 1-3 comprise Acadian Tectophase II as it is developed in New York State. Onset of black shale deposition in the latest Middle Devonian (Geneseo Formation) marks the onset of Acadian Tectophase III.

END OF TRIP:

If traveling southbound, return east on Rt. 28 to I-87/NYS Thruway southbound.

If traveling north or east/returning to Schenectady, backtrack on Rt. 28 for 0.7 mi. and turn left onto Rt. 375. Follow Rt. 375 to T-intersection at Woodstock, then turn right onto NY Rt. 212 and follow to NYS Thruway northbound at Saugerties. Follow reverse directions of road log along Thruway north.

If traveling west, proceed west on Rt. 28 or follow directions back to Schenectady (above) and continue west on NYS Thruway.

ORDOVICIAN ROCKS IN THE MOHAWK VALLEY: GEOLOGIC SITES FOR EDUCATION OF HIGH SCHOOL AND COLLEGE STUDENTS

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INTRODUCTION

The Mohawk Valley has outstanding bedrock geology and the stratigraphy of the Middle Ordovician rocks reflects the tectonic events associated with the Taconic orogeny (Figures 1-4). One advantage of teaching in this area is the ability to integrate the local stratigraphic record into classes concerned with stratigraphy, sedimentology, depositional environments, tectonics, and other geologic studies (Garver, 1992). Most of the Mohawk Valley is underlain by Middle Ordovician rocks that were deposited prior to and during the collision of the Taconic arc. This collision is clearly reflected in changes in the stratigraphic record, as well as structural and metamorphic events farther to the east (see Kidd et al. [this volume] for the structural aspects of the frontal thrust; see Hollocher [this volume] for the metamorphic aspects across the Taconic Orogenic belt). These changes include: 1) a deepening upwards trend that reflects tectonic subsidence; 2) lateral facies changes that reflect the basin geometry; and 3) paleocurrent trends that ultimately define the nature of coarse clastic infilling during the orogeny. The immediate focus of this field trip is on the stratigraphic record of the Taconic orogeny in the Mohawk Valley. In our examination we will look at separate pieces of the stratigraphy in ascending stratigraphic order - we will be traveling upward through time. We will examine changes in depositional environments and how these changes can be inferred from the rock record. We will also examine the lateral variation in the stratigraphy to infer changes in the basin geometry. Finally, we will use paleocurrent indicators to infer the shape of the basin and the flow directions of sediment fill during the Taconic orogeny.

This trip has two goals. The first goal is to gain an understanding of how to interpret depositional environments and how the changes in depositional environments can be used to infer tectonic events. The second, and perhaps most important, goal is to discuss how these different geologic sites can be used for the education of students. I have little experience with primary and secondary students on the outcrop, but I do have considerable experience with undergraduate students in the field and some of the challenges encountered in this arena certainly apply to younger students. Of the five sites on this trip, I use four of them in my course "Stratigraphy and depositional environments of New York", which I teach at Union College (for a complete description of the course, see Garver, 1992). These sites are advantageous because they focus on the most dramatic changes in the stratigraphy and they are easily accessible - most are in public parks.

WHY SHOULD STUDENTS SEE ROCKS?

In my mind there are several reasons that students should see rocks, but it is important to bear in mind that different students should see rocks for different reasons. Perhaps the first reason is to capture their attention. There is little doubt that fossil and mineral collecting trips stir that imagination of many young (and old!) students. In this regard, just "collecting stuff" is fine - these trips may lead some students to further inquiry. The second reason for showing student rocks may center around simply telling them a story. A third reason to show students rocks is to have them understand scientific inquiry and the methodology of the geological sciences. Many would argue that this final reason is far too esoteric for the average elementary or even high school student - I disagree. Primary and secondary school students can benefit a great deal if the second and third reasons are combined. Much of what we discuss during this trip involves some rather simple concepts and ideas that one can draw on from their own experiences.

When I teach *Stratigraphy and depositional environments of New York*, I combine reasons two and three above. The entire course centers around understanding tectonics through the stratigraphic and sedimentologic record by conducting weekly research problems aimed at understanding different parts of the stratigraphy. An important aspect of this approach is to have students observe and collect their own data, and to have each field project build on previous work. This week-to-week unity provides the students with an intriguing motivation that, in my opinion,

In Garver, J.I., and Smith, J.A. (editors), Field Trips for the 67th annual meeting of the New York State Geological Association, Union College, Schenectady NY, 1995, p. 357-375.

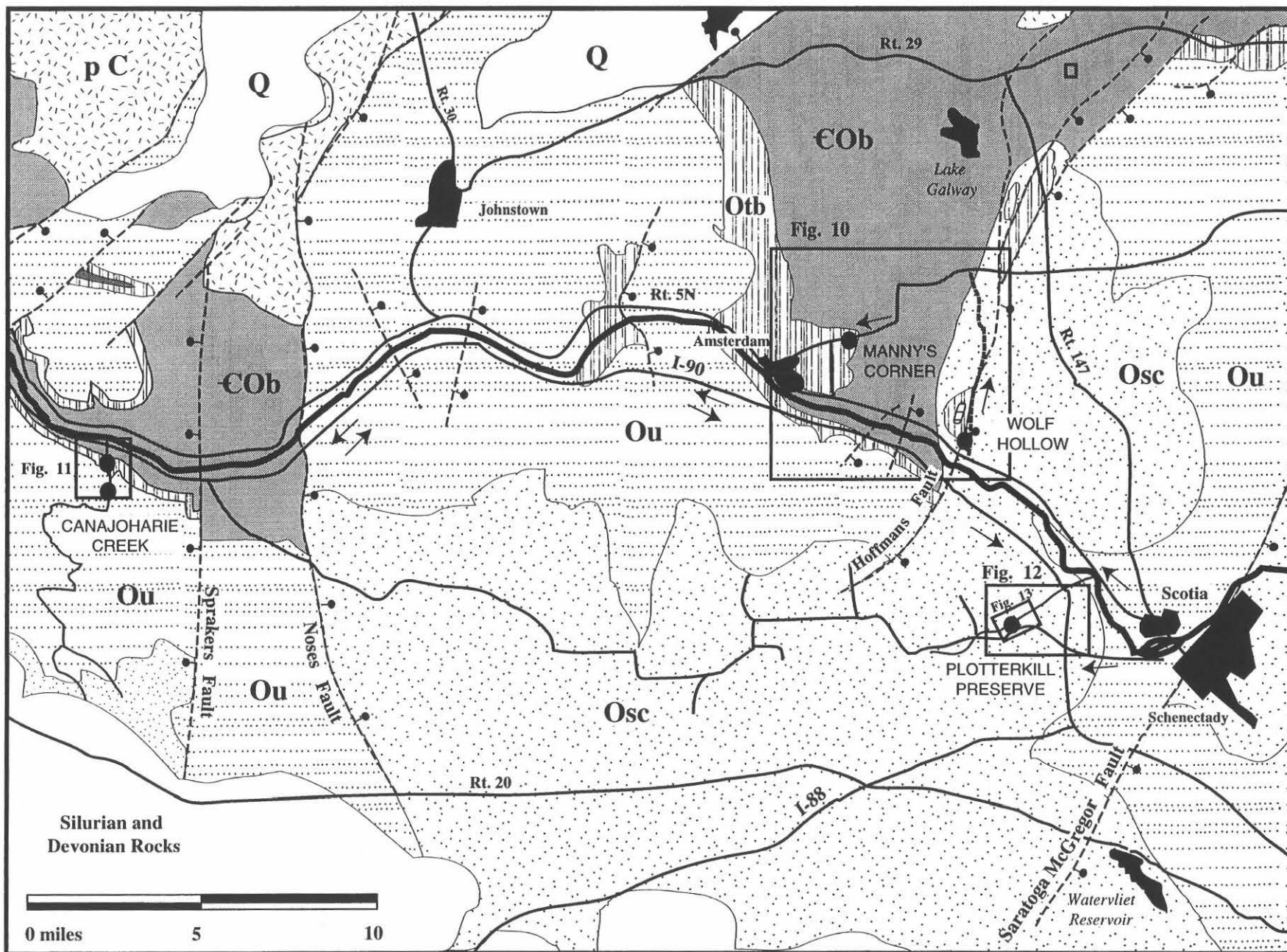


Figure 1: Simplified geologic map of the lower Mohawk Valley after Fisher et al., 1970, and Fisher 1980. Trip starts in Scotia, proceeds west on 5N to Wolf Hollow, north to Rt. 67 to exposures near Manny's Corners, then to Canajoharie Creek via Amsterdam and I-90. The last stop is in the Plotterkill Preserve near Schenectady. Normal faults shown with tick mark on the downthrown block. pC is Adirondack basement; COb is the Beekmantown Group; Otb is the Trenton and Black River groups; Ou is the Utica Shale; and Osc is the Schenectady Formation. Younger rocks (Silurian and Devonian) and cover (Quaternary) are not shaded.

appears to accelerate the learning process. I have students write a formal lab report for each project. This weekly report writing and rewriting dramatically improves writing skills and crystallizes their understanding of the scientific method.

The approach of this course is to present the students with weekly field and lab exercises that have a common goal of deciphering the tectonic evolution of eastern New York through the stratigraphic record. A specific format and a detailed explanation of each project help the students organize their thoughts, manipulate the data, and interpret the significance of the data. The need for well-exposed, easily accessible field sites is obviously important. The field projects are not show-and-tell field trips; they include the investigation and data collection in a single area or outcrop toward a goal that is clearly outlined in an accompanying project description. The purpose of the field projects is not only to teach methodology, but also to confront them with real scientific problems that they must try to solve. In many cases there is no correct answer. I emphasize that it is not the answer that is important, but how you arrived at your conclusions. In this regard, their reports are aimed at persuading the reader that their interpretation makes sense in the context of the field observations.

GEOLOGIC FRAMEWORK FOR THE MIDDLE ORDOVICIAN STRATIGRAPHY IN THE MOHAWK VALLEY

The Middle Ordovician rocks in the Mohawk valley are now generally interpreted to reflect the tectonic events associated with the Taconic Orogeny. The Taconic orogeny is inferred to be a period of Arc-Continent collision where the west-facing Taconic arc (the subduction zone dipped to the east) collided against the passive margin of North America. This trip is designed to highlight some of the most important changes in the stratigraphy and depositional environments of the units that record this collision.

Beekmantown Group

The oldest unit of concern for this discussion is the Cambro-Ordovician Beekmantown Group (Figures 2-5). This is a thick sequence of limestone and dolostone that was deposited in shallow marine to supratidal environments. Different formations of the Beekmantown Group are well exposed in the Mohawk Valley and their correlatives are present both north and south along the eastern seaboard. These rocks are dominantly Upper Cambrian to Lower Ordovician and are punctuated by several unconformities (Fisher, 1977). This time span represents some 41 Myr of shallow shelf sedimentation with no evidence of tectonic activity (~517 to 476 Ma - see Harland et al., 1990). The depositional environments and ultimately the stratigraphy are dominated by changes in global sea level (eustatic sea level) as opposed to tectonic effects. This sequence of rocks probably represents a *Passive Margin* (or Trailing Margin) sequence that rifted away from another continental mass long before these sediments accumulated. A modern setting that probably resembles the depositional setting of these ancient rocks would be the southern and eastern coast of Florida today where carbonates and very mature clastic rocks (well-rounded quartz sandstone) are deposited together in a variety of depositional environments.

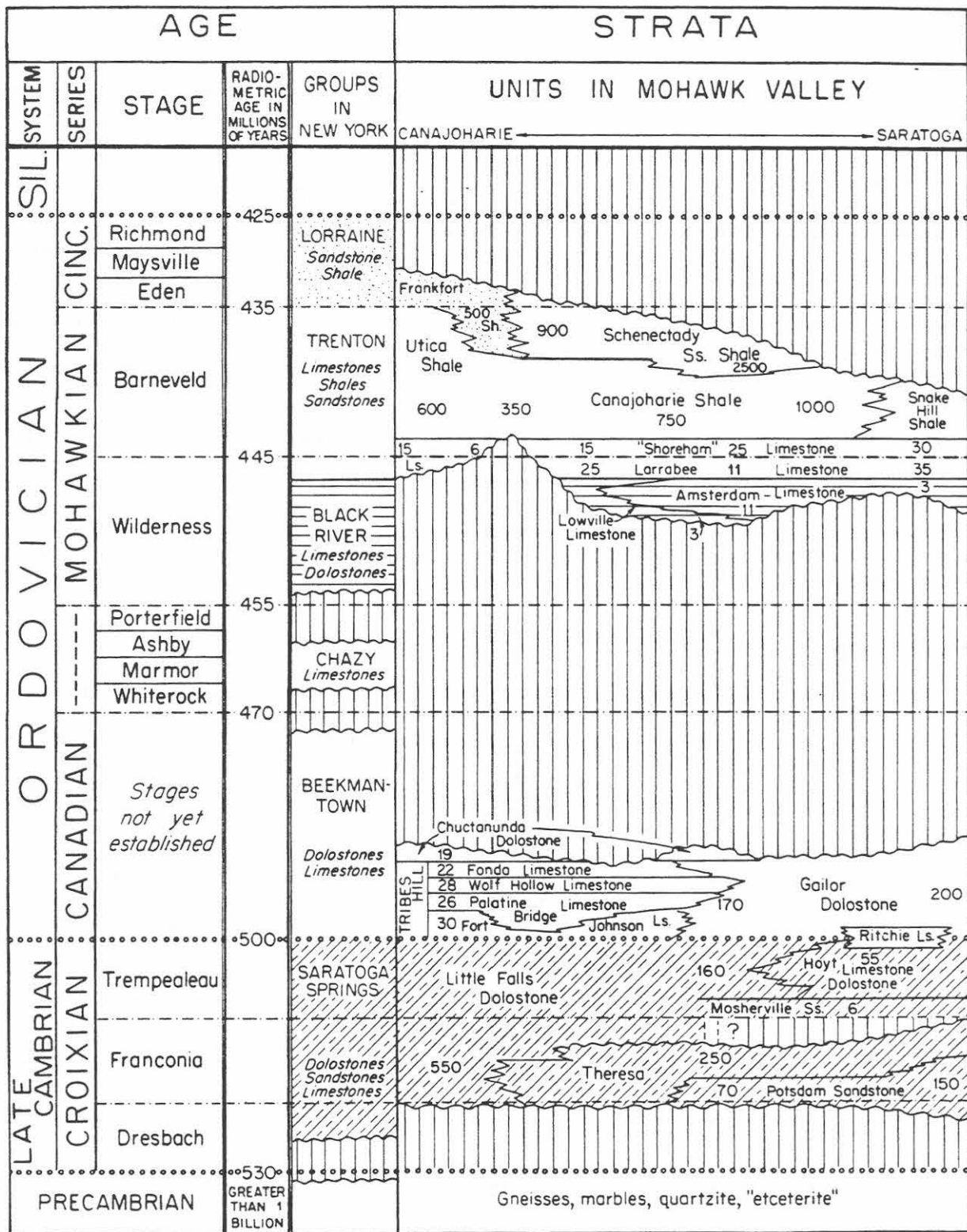
Middle Ordovician Limestones

Deposited above the upper units of the Beekmantown Group are several different limestone units - mainly the Black River Group and the "Trenton Limestones"¹ (Figures 2-5). The Black River Group contains the Lowville Limestone and the overlying Amsterdam Limestone (Fisher, 1977), both of which are generally fossiliferous and were probably deposited in a very shallow marine setting. We will use sedimentary structures and fossil content to make some inferences about their depositional setting.

The overlying limestones of the Trenton Group include, mainly, the Glens Falls Limestone². This unit is abundantly fossiliferous and generally contains both limestone beds and interbedded calcareous shale. The depositional setting of this unit is a deeper-water shelf as compared to the underlying limestones of the Black River Group. The general upward trend in this stratigraphic sequence is a profound deepening (transgression). In the field, we will discuss possible ways to estimate depth during deposition for each of these units; once we have depth estimates, we

¹ There is some confusing nomenclature surrounding the Middle Ordovician rocks in the Mohawk Valley. To many geologists the Trenton Group includes the Utica Shale, but to others it does not. For the ease of discussion, I will refer to the Trenton Limestones and the Utica Shale separately.

² Locally the Glens Falls Limestone can be subdivided into a lower unit, the Larrabee Limestone, and an upper unit, the Shoreham limestone (see Fisher, 1977; 1965a,b)



Note: All thicknesses shown in approximate feet.

Figure 2: Generalized lower Paleozoic stratigraphy of the Mohawk Valley (from Fisher, 1965a,b). Thickness of units given in feet.

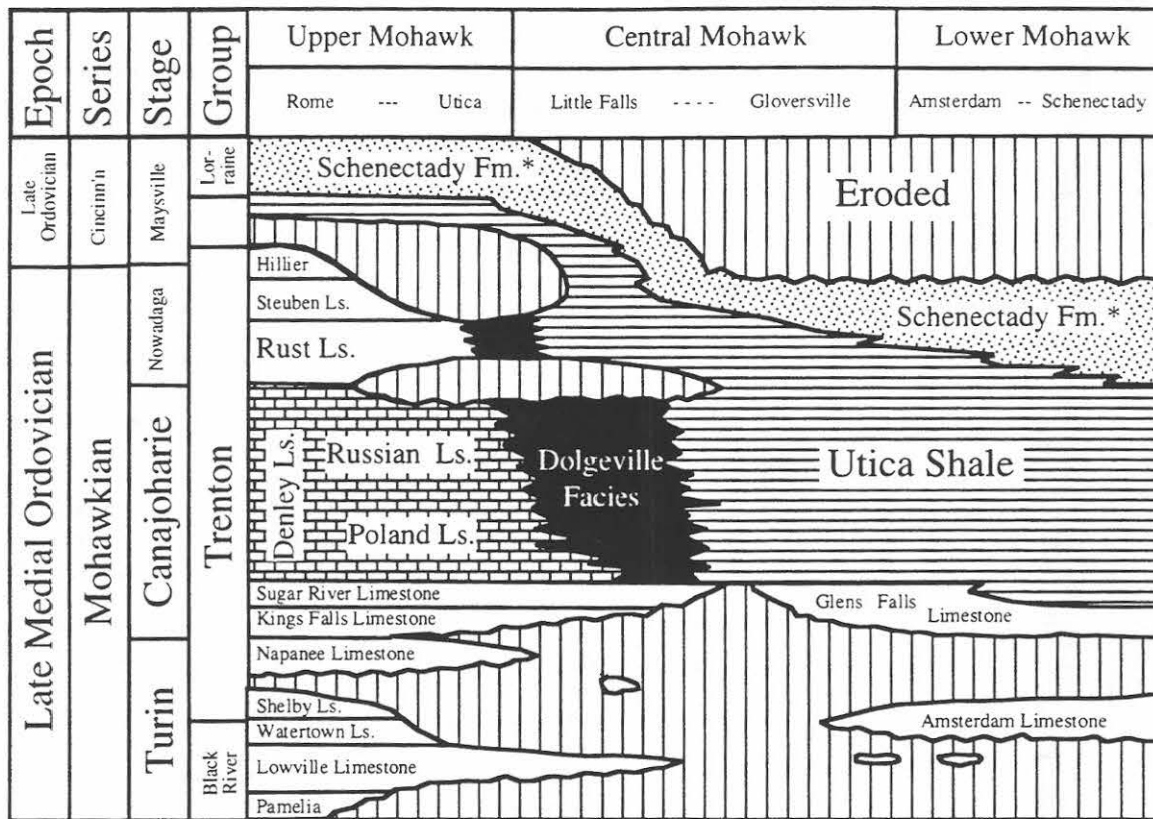


Figure 3: Generalized stratigraphic relationships in the Mohawk Valley showing biostratigraphic relationship between units (not thicknesses). Simplified from Fisher (1977). (*In this diagram the "Schenectady Formation" includes the Frankfort Formation.)

can make inferences about the influence of either sea level change (eustatic) or tectonic subsidence. I and others favor an interpretation in which the deepening trend is caused by tectonic flexure at the onset of the Taconic orogeny (see Baldwin, 1980).

Middle Ordovician Clastic rocks

The Utica Shale lies conformably above the transgressive limestones of the Trenton and Black River groups and reconstructions of depositional environments suggests that the transgression continued (Figures 2-5). The Utica Shale³ is a thick sequence of black shale that is locally fossiliferous and locally calcareous. Two aspects of the Utica Shale are noteworthy. First, it is extremely thick considering its age range (therefore the average accumulation rate must have been very high); it is about 300 meters thick in the Canajoharie area and it thickens to about 1000 m to the east of Schenectady. This rapid sedimentation rate implies that some tectonic mechanism was responsible for forcing the bottom of the basin downwards (as opposed to making sea level go up), allowing a thick sequence of sediments to be deposited. Second, it passes laterally into different units (Figure 5). Traced to the west, the formation becomes increasingly interbedded with limestone (Dolgeville facies) and much farther to the west it passes laterally into limestone with no shale (Denley Limestone; see Fig. 5). Laterally, to the east, it passes into interbedded shale and sandstone of the Schenectady Formation and its equivalents. These *Lateral Facies changes* are extremely important in helping us determine the geometry of the depositional basin in which these rocks were deposited.

³ This name has also been somewhat controversial. It is generally agreed that the Utica Shale, Canajoharie Shale, and Snake Hill Shale are more or less lithologically correlative, but are slightly different ages (see discussion in Kidd et al., this volume)

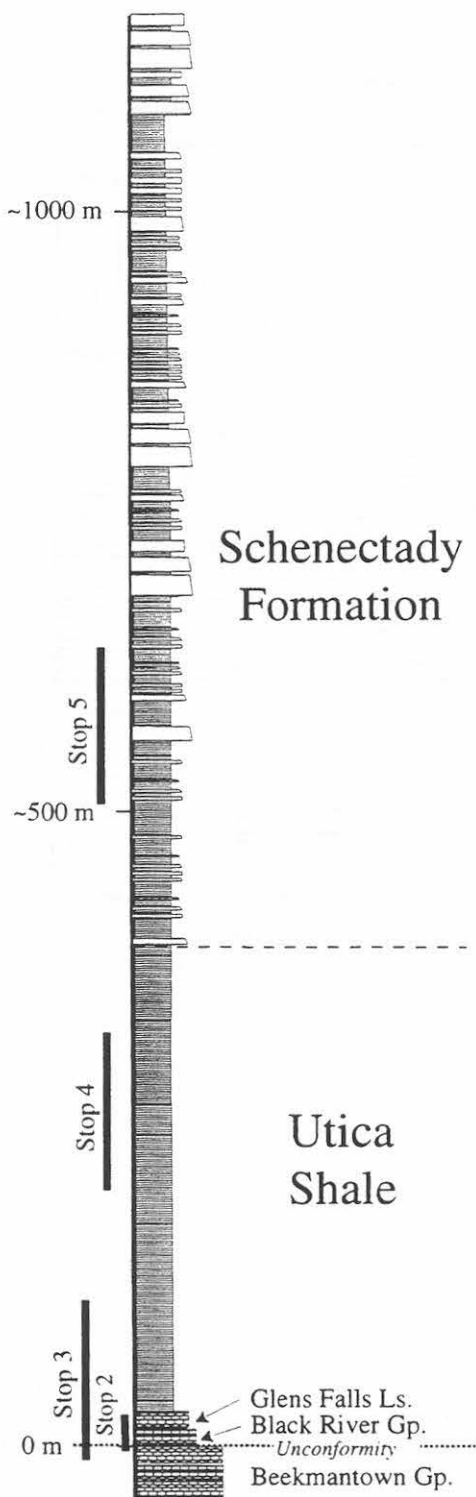


Figure 4: Generalized stratigraphic column for rocks seen on this field trip. Thicknesses are approximate (see Fisher, 1977). Approximate limit of the stratigraphy seen at each field trip stop is shown on the left-hand side.

The Schenectady Formation⁴ is a very thick sequence of interbedded sandstone and shale that is virtually unfossiliferous. The unit is very thick and is wedge-shaped, thickening dramatically from west to east (Fisher, 1977; see Figure 2). Unfortunately, because erosion has removed the upper beds of the Schenectady Formation, the true stratigraphic thickness of the Schenectady Formation is difficult to establish. In the Plotterkill area, the unit is greater than about 760 meters thick (Fisher, 1977). The Schenectady Formation is the foreland basin deposit that accumulated in an asymmetric basin in front of the advancing thrust complex associated with the Taconic arc. A *foreland basin* is a basin produced by flexure of the continental crust when a load (such as the Taconic arc) is placed on the edge of the crust (Figure 6). Perhaps the best way to envision the geometry of a foreland basin is to hold one end of a meter stick on the edge of a table and push down on the free end, flexing the ruler downward. The hand that is pushing the ruler down is a load -- similar to a thrust complex. The ruler (if thick and wooden) behaves very much like "rigid" continental lithosphere. We know, however, that it is not this simple. When the continental lithosphere was pushed down by the forward-advancing load, the upper part of the continental crust responded by breaking along normal faults (Figure 7). Bear in mind that the load that was responsible for flexure of the crust was not static. Instead, it moved progressively to the west so the region of maximum flexure (the deepest part of the basin) also moved progressively from east to west. The relationship between lithospheric flexure and concurrent normal faulting in the Mohawk Valley is discussed in Bradley and Kidd, 1991. In the field, we will discuss the offset amounts, timing, and effect that these faults had on Middle Ordovician sedimentation.

An important aspect of basin studies is *paleocurrent analysis*. A paleocurrent study involves determining the ancient flow directions (paleoflow) of sedimentary detritus in a basin. This sort of study is done to determine the position of the original source terrain, types of depositional environments, and lateral trends within a basin of deposition. If you work for an oil company, and you discover an oil-rich beach deposit, the direction and trend of that beach would be important information. Paleocurrent data from these beach sediments could help you make useful predictions about the lateral trends of this deposit. Many physical sedimentary structures are produced by traction currents and they commonly retain information about the direction of sediment transport. The sandstone beds in the Schenectady Formation represent turbidites that were deposited by fast-moving turbidity currents that moved downslope and were deposited in this part of the Mohawk Valley (Figures 7,8).

⁴ In this report, "Schenectady Formation" refers to the type Schenectady Formation as well as correlative units of interbedded sandstone and shale to the west and to the east (including deformed sequences). Notably, this would include the Frankfort (to the west) and parts of the Normanskill Formation (to the east).

Turbidity currents are sediment-laden currents that, because they are denser than water, flow downslope to the deeper parts of a sedimentary basin. When these currents reach the deepest part of the basin, or a break in slope such as is found at the base of the continental slope, they slow down and begin to deposit sediment. The decelerating current produces a very regular and systematic sequence of sedimentary structures that record the slowing of the current and the deposition of the sediment from the turbidity current (Walker, 1979). The deposition of a turbidite is very rapid (hours) and the downslope movement of the sediment may correspond to an earthquake, a storm, or any other mechanism capable of stirring up sands and silts in shallow water. Normally, deep-water environments receive only minor fine-grained sediment that rains down from the water column at extremely slow rates. It is common, therefore, for deep-water sediments to be composed of turbidites and interbedded shale with the former representing an instantaneous depositional event and the later representing the slow continuous accumulation of mud from the water column. Turbidites commonly have excellent paleocurrent indicators that can be used to reconstruct basin geometry.

In turbidite sequences, perhaps the most common **paleocurrent indicators** are found at the base of sandstone beds and represent the flow of the sand-rich turbidite over the muddy seafloor. Flutes and grooves are sole marks that commonly occur at the base of turbidite beds. Flutes are unidirectional sole marks that occur on the bottom of sandstone beds. They are caused by turbulence in the turbidity current striking the seafloor and moving forward at the same time. They look like inverted teaspoons with the sharp/steep side toward the source direction. Grooves are bidirectional sole marks. Grooves are elongate nearly straight marks on the base of sandstone beds. They are not asymmetric and they therefore simply indicate the orientation of the current but not the absolute direction. Paleocurrent indicators also occur in the main part of a turbidite bed. As turbidity currents slow and decelerate,

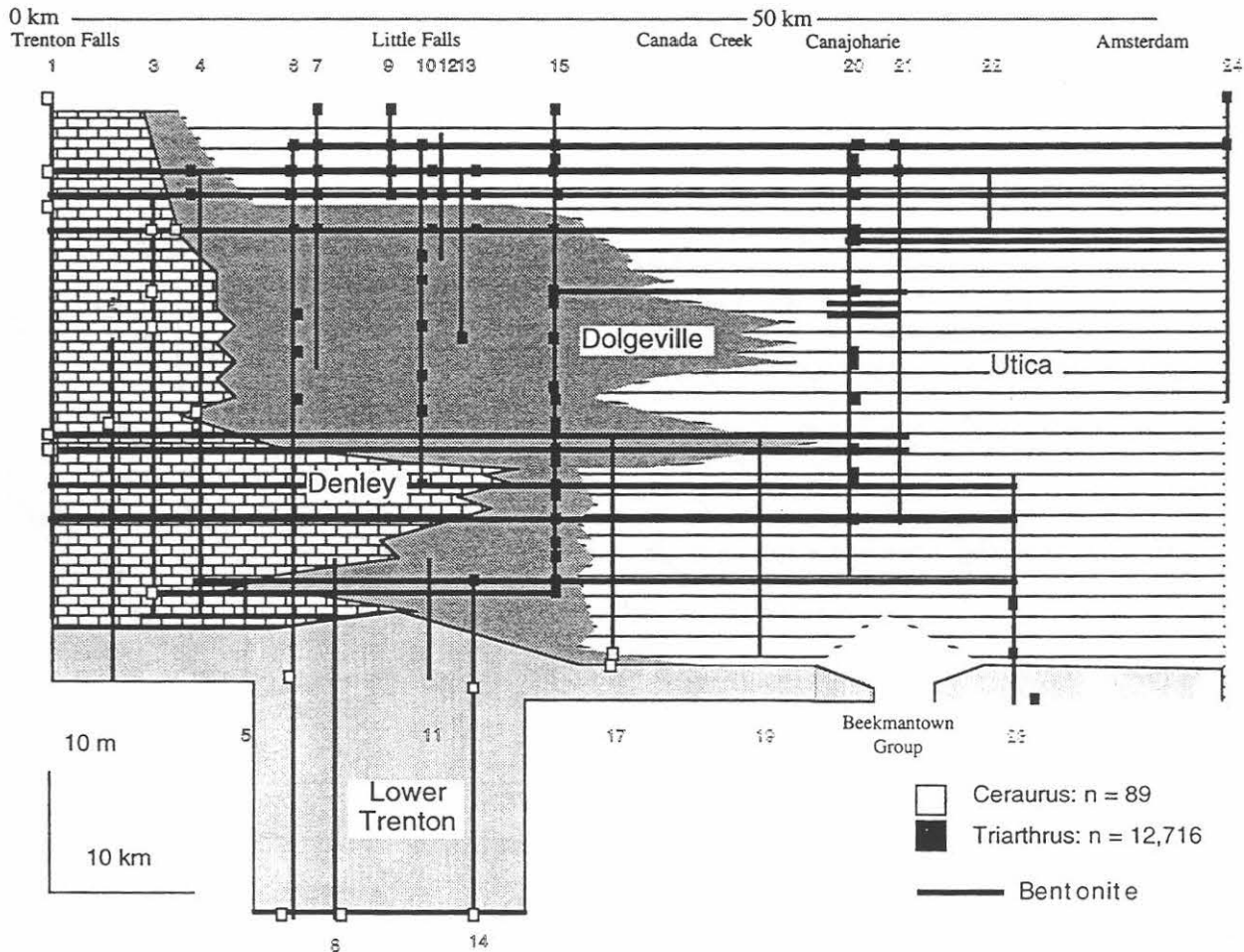


Figure 5: Relationship between bentonites (dark horizontal lines) and measured stratigraphic sections (vertical black lines) showing the distribution of *Cerurus* (open squares on the measured stratigraphic sections) and *Triarthrus* (filled squares). Modified from the Macintosh Hypercard stack by Ray Gildner (unpublished).

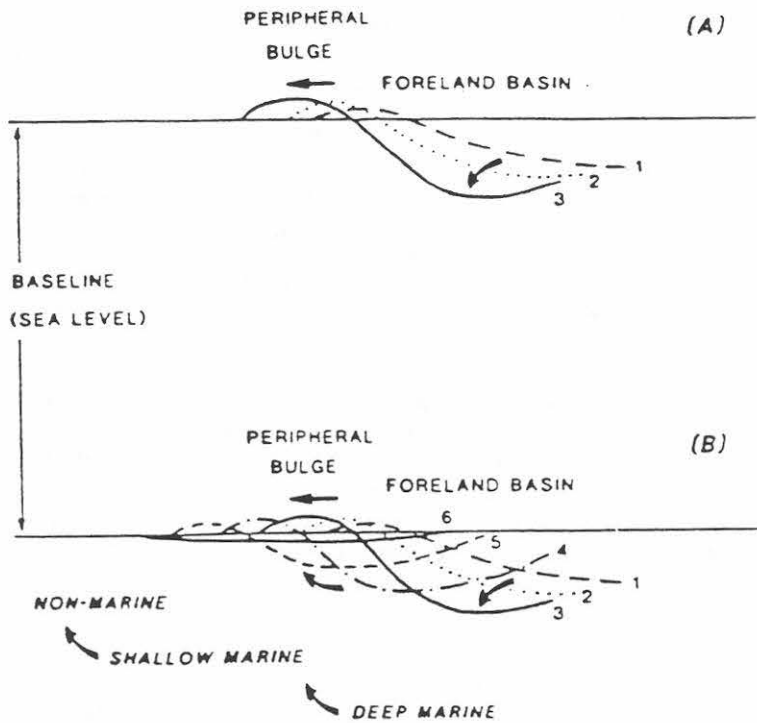


Figure 6: Development of lithospheric flexure due to thrust loading (from Pickering, 1977). Note that through time, the position of the maximum flexure (deep part of the basin) migrates forward with the advancing thrust complex [A]. Also note that at any one place that experiences this flexure, one can expect the environments to go from non-marine to shallow marine to deep marine. Locally, slight upwarping (the "peripheral bulge") can cause rocks to be uplifted above sea level.

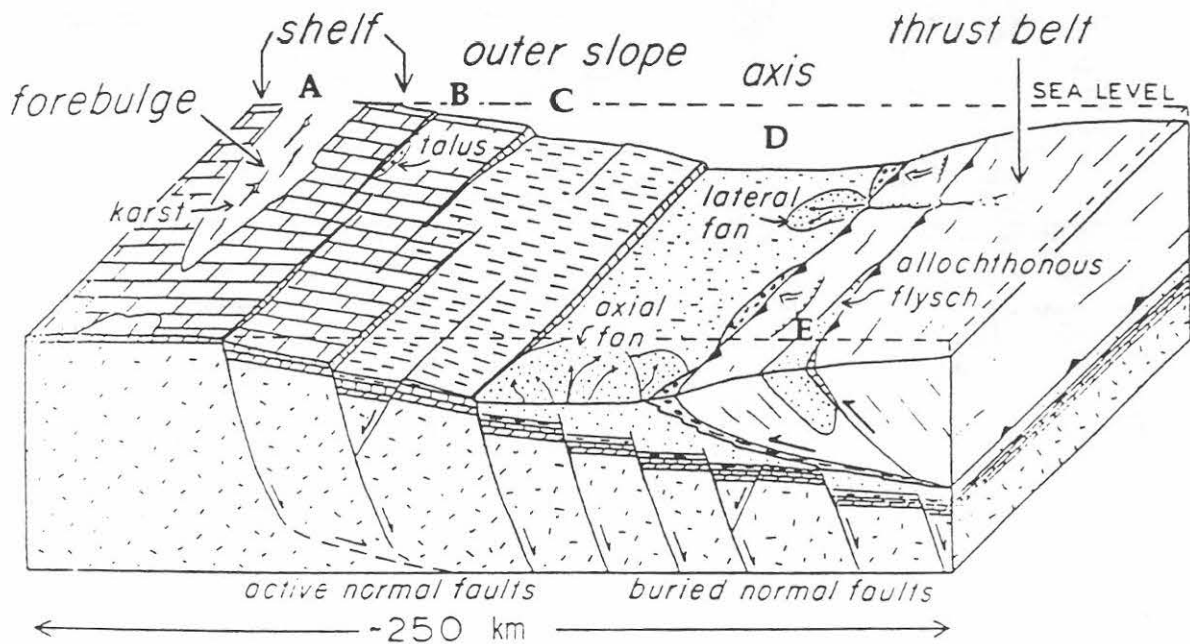


Figure 7: Relationship between basin infilling and normal faulting which is inferred to be concurrent with lithospheric flexure (slightly modified from Bradley and Kidd, 1991).

thus depositing their load, they produce planar laminated bedding. Parting lineation is a bidirectional current indicator that occurs on the split surface of parallel laminated sandstone beds. It occurs as a lineation parallel to the paleoflow direction and it is produced by mineral alignment during deposition. Eventually further slowing of the current produces rippling of the sand. At the top of a turbidite one can commonly observe ripple cross-lamination, which is a unidirectional current indicator. Unless exposures are excellent, however, it may be difficult to determine current direction. A cross-section of the ripples in a sandstone bed will display small (mm- to cm-scale) inclined foreset laminae that are inclined in the direction of ripple migration and current movement.

If you are interested in determining where the clastic came from, you would have to study the *Provenance* (source area) of the sediment. The provenance of the sandstones should give us a clue as to the composition of the highlands that eroded to fill the basin. In general, provenance studies of synorogenic clastic sediment are important because in other thrust belts the coarse-grained detritus is derived from the uplifted and eroding thrust complex and therefore the sandstone should provide a record of what collided. Thin sections of sandstones of the Schenectady Formation reveal that the sandstones are composed almost exclusively of quartz, feldspar, and sedimentary rock fragments (mainly fragments of limestone and sedimentary rocks and only minor metamorphic and volcanic rock fragments), and minor detrital mica. The grains of quartz are particularly interesting because many are well rounded multicycle grains. This observation suggests that the quartz grains are very mature texturally, and must have been reworked over and over in a stable tectonic setting - not typical of an active orogenic belt. Another important clue is the abundance of clasts of sedimentary rock fragments, which suggests that sedimentary rocks (including limestone) were common in the source area. Notable is the virtual lack of volcanic rock fragments. If the Taconic island arc collided with North America in the Middle Ordovician, then why aren't the sandstones rich in volcanic arc-derived detritus? The answer lies in what happens when an arc collides with the edge of a continent.

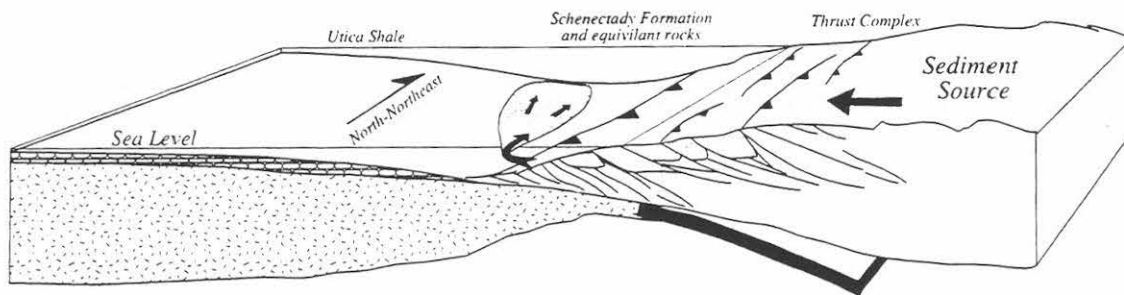


Figure 8: Large-scale relationship between flexure of the lithosphere and uplift of the thrust complex. The submarine fan drawn here emphasizes the axial transport of turbidites in the basin. From Garver and others (1996).

As we know from studies of modern arc-continent collisions (e.g. see Lundberg and Dorsey, 1988), the arc plows into the continental margin sediments (many of them deep-water) and imbricates them like a deck of cards in a rapidly growing thrust complex that separates the arc from the foreland basin (Figure 9). Because these continental margin sediments are the first to ride up onto the continental margin, they become rapidly uplifted and eroded, and the detritus then gets deposited in the adjacent foreland basin. Today, the thrust complex is well displayed in the Taconic mountains where deep-water (continental slope and continental rise) sediments have been internally imbricated and thrust over shallow water continental margin strata (Zen, 1961; Stanley and Ratcliff, 1985). The rocks in the Taconic allochthon are almost exclusively sedimentary rocks, including Cambrian and Ordovician units that are rich in well-rounded mature quartz sand (e.g. Poultney Formation, West Castleton Formation, and others) that was derived from contemporaneous units on the shelf (e.g. Theresa Formation) (Baldwin, 1983). The basement to the Taconic arc included ophiolites in many areas of the Appalachians. An ophiolite is a piece of ocean crust that includes basalt, gabbro, and ultramafic rocks. The ultramafic rocks are geochemically distinct because they contain far more chromium and nickel than any other rock common at the surface of the earth. Using the Cr and Ni geochemistry of the shale that is interbedded with the sandstone, Garver et al (1996) determined that ophiolites were not significant in the thrust complex that eroded to form the Schenectady Formation. They did find, however, that as one follows correlative sequences north to Newfoundland, the percentage of ophiolites in the source increased considerable. Similarly, they determined that sediments deposited very early in the collision reflect the nature of the Taconic arc, but that those deposited later in the collision are dominated by continental margin sediments.

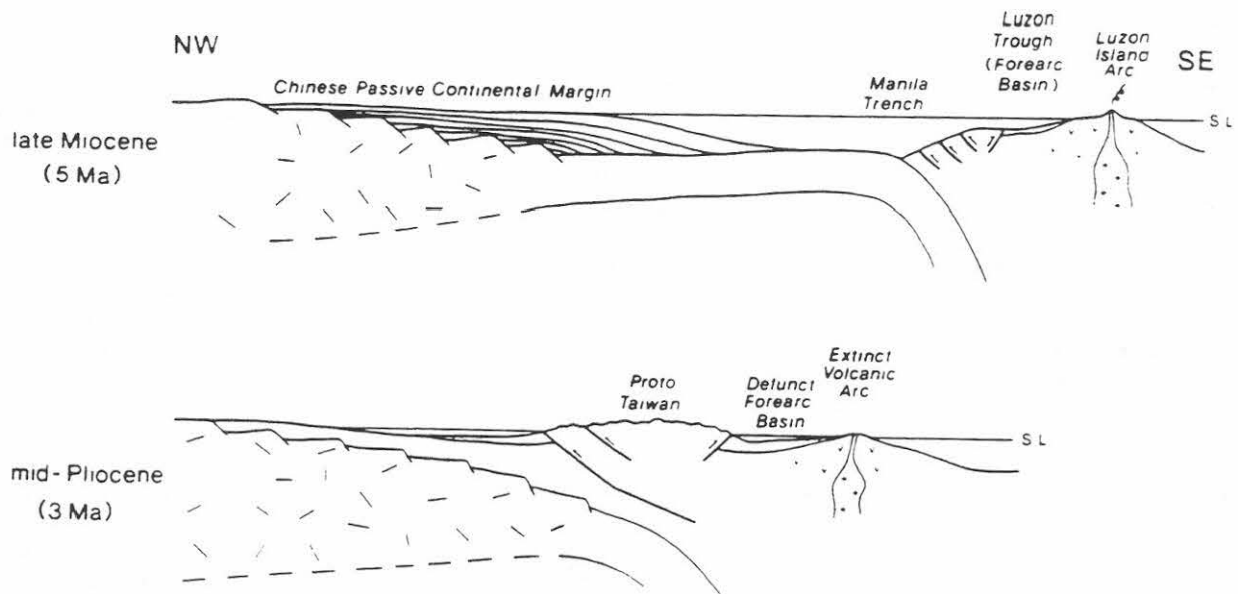


Figure 9: Inferred sequential development of the collision of Taiwan (from Lundberg and Dorsey, 1988). The west facing Luzon Arc is inferred to have collided with the passive margin of China only in the last 5 Million years. The inferred collision of the Taconic arc with the then passive margin of eastern North America was probably very similar. In the lower diagram (mid-Pliocene) note the development of "Proto-Taiwan" which is largely composed of imbricated sediments of the Chinese margin. Much of the sediment deposited in the foreland basin (to the west of Proto-Taiwan) is derived from the thrust complex and not from the Luzon arc - much the same scenario is envisioned for the collision of the Taconic arc.

As we examine the stratigraphy that records the collision of the Taconic arc, consider how these concepts of stratigraphy, sedimentology, depositional environments, and tectonics can be presented to students of all ages. I hope that the different experiences of the trip participants will come out as a lively and informative session that will help us all to be more effective in presenting rocks to students.

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ORDOVICIAN ROCKS IN THE MOHAWK VALLEY: GEOLOGIC SITES FOR EDUCATION OF HIGH SCHOOL AND COLLEGE STUDENTS

ROAD LOG

Road Log starts at the Intersection of Rt. 50 and Rt. 5N on Mohawk Ave in downtown Scotia.

- | | | |
|------|-----|---|
| 0.0 | 0.0 | Continue westbound on Mohawk Ave through Scotia. |
| 5.5 | 5.5 | Canal Park and Lock 9 to the left. |
| 7.6 | 0.0 | Turn Right at Hoffmans on Wolf Hollow Rd. |
| 8.25 | 0.0 | Bear left and continue on Wolf Hollow Rd. Note historic sign telling of the displacement on the Wolf Hollow Fault. |
| 8.4 | 0.0 | Pull off to the left at sharp curve in the road - use caution because it is difficult to see oncoming traffic. Stop 1 (optional). |

STOP 1: Hoffmans Fault

In Wolf Hollow a high-angle normal fault cuts the Schenectady Formation and older rocks (Figure 1 and Figure 10). This fault, which trends about 030 (north-northeast) is one of a family of high-angle normal faults that were probably active in the Middle Ordovician. Most of these faults dip to the east and the eastern block is downdropped - some notable exceptions occur farther to the west at the "Noses" (Figure 1). At this site one can see slightly tilted rocks of the Schenectady Formation (interbedded sandstone and shale) on the east side of the road. On the west side of the road are exposures of the Beekmantown Group - these are especially easy to see when the leaves are off the trees. Not visible from here are also exposures of the Trenton limestones and the Utica Shale, both resting above the Beekmantown Group. Using stratigraphic mismatch we can determine the apparent throw on the fault. Although the sign at the entrance to Wolf Hollow states that the fault has 1000 feet of displacement, is difficult to calculate the exact displacement because there are no clear stratigraphic markers in the Schenectady Formation. (See Locality 19 in Fisher, 1980). Using simple cross sections and reasonable stratigraphic thicknesses for these units, the apparent displacement is somewhere between 200 and 400 meters.

- | | | |
|------|-----|---|
| | | Continue north on Wolf Hollow Road. Note the excellent exposures of the Schenectady Formation on the east side of the road along the entire length of the Hollow. |
| 9.75 | 0.0 | West Glenville Road - note exposures of the very fossiliferous Trenton Group limestones (Glens Falls Limestone) on the left (west) side of the road). Continue straight. |
| 13.3 | 0.0 | Rt 67. Turn left (west). |
| 15.5 | 0.0 | Several outcrops of carbonates of the Beekmantown Group (Wolf Hollow Limestone which is Lower Ordovician) along the right (north) side of the road. |
| 17.0 | 0.0 | Pulloff in small overgrown drive on left side of road to old abandoned quarry. This is private property. Please call Cranesville Block Company (518-346-5749) for permission to enter the "Manny's Corner quarry". See Figure 10. |

STOP 2A: Quarry at Manny's Corner - Beekmantown , Black River, Trenton groups

In this quarry, the Cambrian to Lower Ordovician Beekmantown Group is disconformably overlain by Middle Ordovician carbonates of the Black River and Trenton Groups (see Locality 17 of Fisher, 1980). The floor of the quarry is composed of the Lower Ordovician Wolf Hollow Member of the Tribes Hill Formation. The Tribes Hill Formation is one of the upper units in the Beekmantown Group, a widespread and relatively uniform sequence of carbonates and dolomites found throughout eastern North America (for other stops in this unit see Friedman, 1972). Note that this unit is dolomitic, has algal mounds, and is replete with a single species of gastropod (*Ecculiomphalus*). The walls and nearby outcrops (north side of Rt. 67) are composed of the Lowville Limestone, the Amsterdam Limestone, and the Glens Falls Limestone (~Larrabee Limestone and the overlying Shoreham Limestone) all of which belong to the Black River Group and the Trenton Group (see Fisher, 1965; 1977). At this exposure, the unconformity between the Beekmantown Group and the overlying Middle Ordovician rocks is well displayed, and regional stratigraphic studies indicate that the upper part of the Beekmantown Group (the Lower Ordovician part) is missing. The Lowville Limestone is generally not fossiliferous, but it contains birdseye texture, a few vertical burrows, and mudcracks; locally it is dolomitic. The Amsterdam Limestone is irregularly bedded and the contact between the two units is marked by the first fossiliferous beds. These thin beds are very fossiliferous, with brachiopods, rugose coral, crinoids, and trilobites dominating the fauna. As you examine the outcrop from bot-

tom to top, note the distribution of coral.

For students in *Stratigraphy and depositional environments of New York*, this project involves measuring a short (10 m) stratigraphic section, in which they describe the fossil content of the units, and interpret the depositional environments of these shallow marine rocks. As we will see, the interpretation of the depositional environments is aided by carefully noting the sedimentary structures and faunal content of each rock unit. After examining the rock we will have a brief discussion about relative sea level changes and the role of tectonic activity. The change in faunal content upsection indicates deepening through time and most students use their newly acquired knowledge concerning sea level changes and basin subsidence to explain this apparent transgression. As we will discuss, the deepening in the stratigraphic section may be related to tectonic subsidence. The goal of this phase of the project is to relate depositional environments and the unconformity to regional or global events that might have produced this particular sequence.

- 17.1 0.0 Continue west on Rt 67.
 Pull off on right side of road to prominent outcrops on north side of Rt.67.

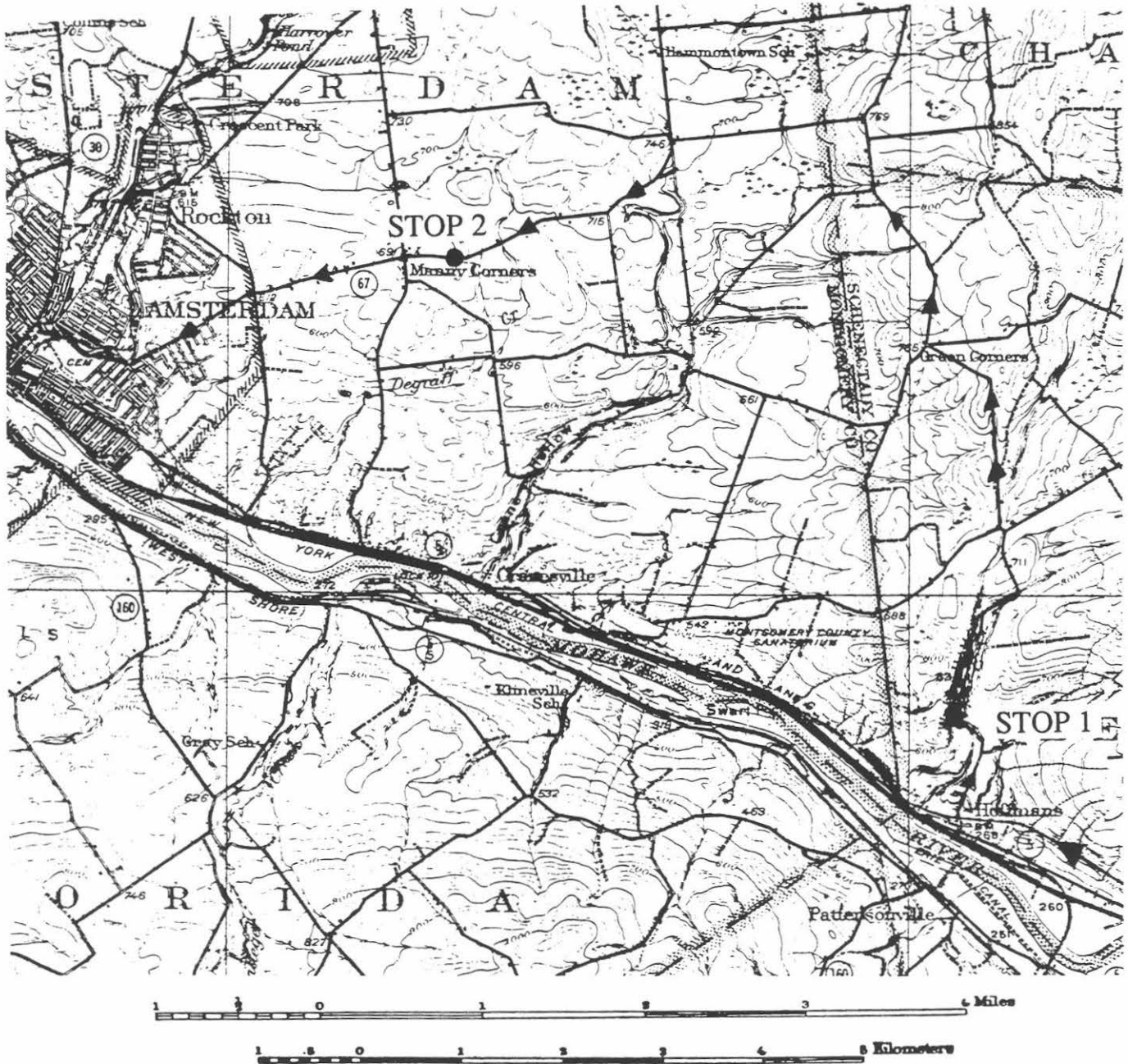


Figure 10: Topographic map showing the route and locations of Stop 1 (Wolf Hollow) and Stop 2 (Manny's Corners).

STOP 2B: Road cut across from Manny's Corner Quarry - Trenton Group Limestone

Exposed in the roadcuts along the north side of Rt. 67 are excellent exposures of the extremely fossiliferous Glens Falls Limestone. Note that these beds contain fossiliferous limestone (locally with scoured bases and graded beds) with interbedded calcareous shale which is also fossiliferous.

These rocks are replete with fossils; brachiopods, bryozoans, trilobites, crinoids, and others. Of particular note are two rather distinct trilobites. A small trilobite, *Cryptolithus tessellatus*, is common in the resistant limestone beds - they are typically the size of a fingernail. A much larger trilobite, *Isotelus gigas*, is common (unfortunately as fragments) in both the limestone bed and the calcareous shale. These are important index fossils. As we will see at Canajoharie Creek, geologists generally use fossils for either biostratigraphy or for reconstructing depositional environments. The concepts of biostratigraphic analysis are fundamental and are the basic tenet of the science of stratigraphy. As organisms evolve through time they change and new species eventually become extinct. By carefully studying the vertical distribution of distinct and common species, paleontologists have been able to determine fossil zones that are defined by the occurrence of a particular index fossil(s). *Cryptolithus tessellatus* is an index fossil for these beds. As we move upsection we will no longer see this fossil because it had become extinct. The distribution of index fossils and other more common and long ranging fossils is very closely related to depositional environments. (Good index fossils are found in many depositional environments). Some fossils can be used to determine the relative water depth. In the Mohawk Valley there have been very detailed studies of fossils with respect to water depth (see Cisnee and Chandler, 1982; Cisne et al., 1982). As we will discuss at the outcrop, the assemblage of fossils in this outcrop suggests deeper water than does the assemblage in the underlying Amsterdam Limestone. Additionally, the occurrence of fine-grained calcareous shale suggests deposition below the normal wave base (the effective depth to which fairweather waves affect the bottom sediments). In this regard the stratigraphic section is deeper than that at the quarry. A deepening trend such as this is referred to as a transgression.

- Continue west on Rt 67.
- 18.35 0.0 Enter the town of Amsterdam
- 19.2 0.0 Intersection - Stay on Rt. 67 by veering slightly to the right.
- 19.95 0.0 (small public park [to left] with very nice exposures of the dolostone of the Beekmantown Group.)
- 20.0 0.0 Left to Rt. 30 south, follow signs to Interstate 90.
- 21.0 0.0 Left lane entrance to Toll Plaza for Interstate 90. Take I-90 west to Utica, Buffalo, etc. While merging (at about 21.3), note the exceptional exposures of the Utica Shale across the interstate in the exit ramp for the eastbound traffic.
- 32.0 0.0 Excellent exposure of limestones of the Middle Ordovician Trenton Group in the opposite lane. Above this exposure (on Rt. 5S) is the conformably overlying Utica shale. At the far end of the road cut on I-90, the uppermost beds of the Beekmantown group are exposed. Once again, the Knox unconformity separates the Trenton Group and the Beekmantown Group.
- 35.6 0.0 Cross the Noses fault (see Figure 10). This fault belongs to the family of normal faults that cut Middle Ordovician and older rocks in the Mohawk Valley, and is therefore similar in timing to the Hoffmans Fault. This fault and others to the west bound an upthrown block (horst) that exposes some of the oldest rocks in the Mohawk Valley. Note that this horst was probably a topographic high during the Middle Ordovician - as can be seen from the regional geologic map (Figure 1), the Trenton Group was not deposited on this block. Elevation of the horst above sea level may explain this distribution of rock units.
- 35.8 0.0 Across I-90 and along the barely visible Rt 5S are prominent exposures of the Precambrian Basement and the unconformably overlying rocks of the Beekmantown Group. The contact between these rocks, which can be seen along Rt 5S, shows evidence of structural disruption, and the interpretation of the contact is complicated by this disruption. The Precambrian rocks are provisionally assigned to the Peck Lake Formation, which is a garnet biotite gneiss about 1.1 Ga (billion years old). Although this contact may be faulted, in this area the Beekmantown Group is known to unconformably overly the Precambrian Basement. For a discussion of this area see Locality 12 in Fisher, 1980.
- 37.3 0.0 For about 0.5 miles there are very steep cliffs of the Upper Cambrian Little Falls Dolostone on the left (south) side of I-90. This unit, which is part of the Beekmantown Group, is stratigraphically lower than all of the other exposures of the Beekmantown Group that we will see on this trip. In places this unit has vugs containing doubly terminated quartz crystals locally known as the "Herkimer Diamond".
- 41.0 0.0 Take Exit 29 - Canajoharie/Sharon Springs, Rt. 10. Pay toll (\$0.65 in 1995).

- 41.9 0.0 Right on 5S, enter town of Canajoharie.
- 42.0 0.0 Left onto Mitchell Street.
- 42.1 0.0 Cross Montgomery Street obliquely (slightly to left) and continue on Moyer St.
- 42.45 0.0 Turn right onto Floral Ave. (Opposite Barclay St.). Proceed to end.
- 42.6 0.0 Park at end of Floral Ave in dirt parking lot at the end of the street. Walk toward river and take foot path upriver for several hundred meters. Stop 3.

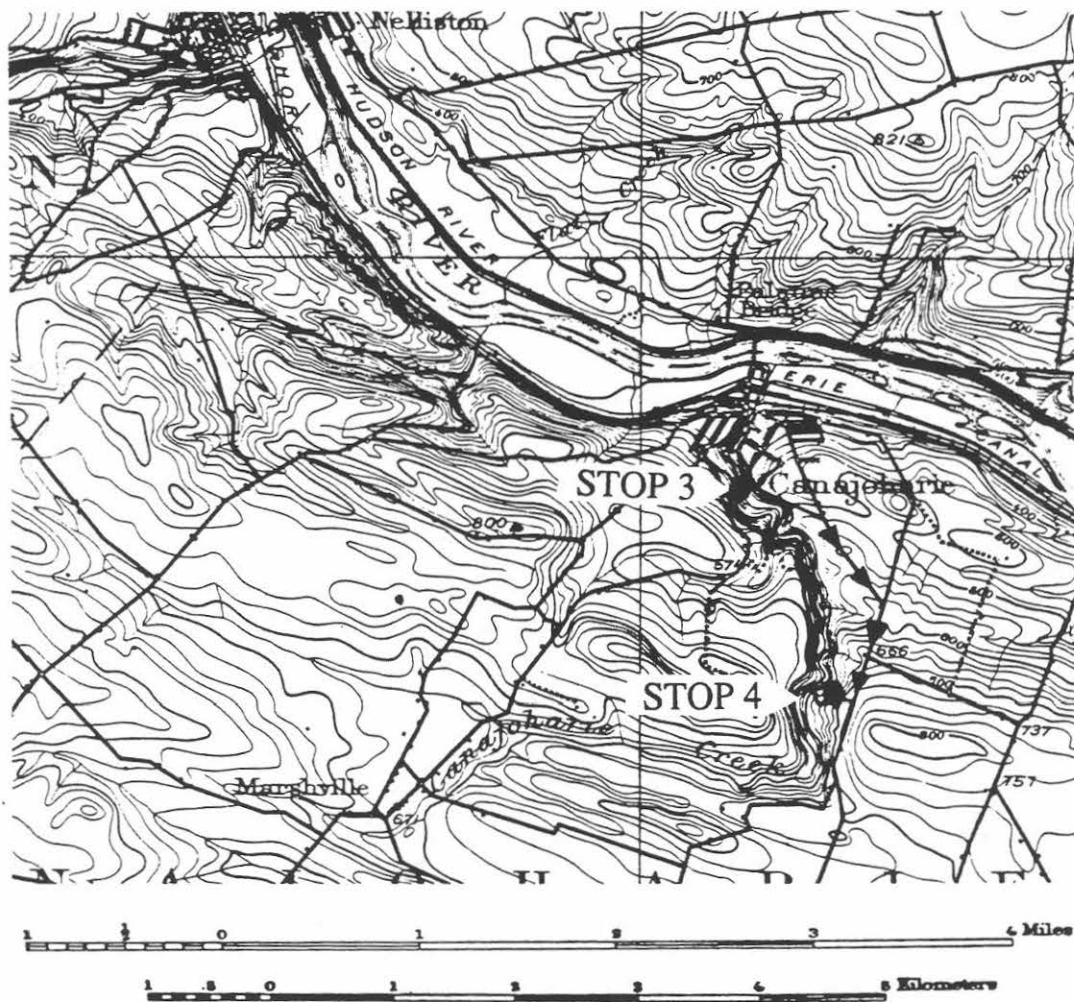


Figure 11: Topographic map showing the route and locations of Stop 3 (Canajoharie Creek on Floral Ave) and Stop 4 (Wintergreen Park).

STOP 3: Canajoharie Creek (Floral Avenue) - Beekmantown , Black River, Trenton groups

Similar to the exposures at Manny's Corners, limestone and shale of the Trenton Group disconformably overlie dolostone of the upper part of the Beekmantown Group in the river bed of Canajoharie Creek. This well-exposed and very fossiliferous stratigraphic section records important changes in the depositional history of the Mohawk Valley. Rocks in the Canajoharie Creek river bed include the Chuctanunda Creek Dolostone, which is part of the Beekmantown Group, and the disconformably overlying limestone and shale of the Trenton Group. Interbedded limestone and shale of the "lower Trenton Group" is conformably overlain by the Utica Shale, which is some 305 meters thick at this locality. All rocks of the Trenton Group were deposited during the Trentonian stage of the Middle- Late Ordovician (*circa* 448-458 Ma) and the rocks of the Chuctanunda Creek member of the Tribes Hill Formation were deposited during the Gascondian stage of the Early Ordovician (*circa* 505-488 Ma).

- Return to vehicles and drive down Floral Ave to Moyer St.
- 42.8 0.0 Turn right on Moyer St.
 - 43.8 0.0 Veer right on Carlise St.
 - 44.0 0.0 Keep right (straight) on Old Sharon Road.
 - 44.1 0.0 Turn right into the entrance of Wintergreen Park.
 - 44.2 0.0 Park at entrance gate in pulloff to the right. Walk down the park road to the Canajoharie Creek bed to exceptional exposures of the Middle Ordovician Utica Shale. (See Figure 11.)

STOP 4: Canajoharie Creek (Wintergreen Park) - Utica Shale

The second stop, which is near the top of the Utica Shale, can be accessed from Wintergreen Park (only open during the summer but it is possible to visit this locality yearround). Here the Utica Shale is a black, laminated and locally fossiliferous shale with minor thin (generally < 2-5 cm) seams of light gray and rusty weathering bentonite. The bentonites can be recognized easily because orange-red rusty streaks emanate from them. The rusty streaks are caused by the oxidation of pyrite which is common in the bentonitic seams. We will walk downstream to examine a fossiliferous section containing the trilobites *Triarthrus* and *Isotelus*, as well as brachiopods and graptolites. Locally, very large nautiloids can be found in this outcrop. For pioneering work on the Utica shale (slightly to the west) see Kay, 1953.

The changes in depositional environments and basin geometry during the deposition of these rocks are well recorded in both the stratigraphy and lateral and vertical distribution of fossils. As you can imagine, individual organisms (fossils) preferred different depositional settings and therefore they commonly are found in only certain facies. For example, a thick-shelled pelecypod may inhabit a high-energy niche in the shoreface while a thin-shelled brachiopod may have found its niche in a deep-water environment. Once preserved in the geologic record, these individuals may serve as useful indicators of the energy level or water depth during deposition of a particular unit. In addition, the distribution of such fossils may provide important information concerning the original configuration and deepening trends within a basin of deposition. (see Cisnee and Chandlee, 1982; Cisne et al., 1982). From the enormous database collected by John Cisne and his co-workers, Ray Gildner has developed a Hypercard stack (Macintosh-based) that shows the distribution and occurrence of 10,000 fossils in the database. In essence, one can see the places where an individual fossil was observed in many stratigraphic sections. The distribution of two depth-sensitive individuals are shown in Figure 5. Here you can see that the distributions of *Triarthrus* and *Cerarus* are very different; *Triarthrus* is restricted to the eastern (deeper) sections and *Cerarus* is present only in the western (shallower) sections. From this we can see that the basin deepens to the west.

The occurrence of the bentonite is extremely important (Figure 5). Bentonite is commonly interpreted to represent volcanic ash. Therefore not only can these thin beds serve as chronostratigraphic markers (one bed is the same age everywhere), but they can also tell us that volcanic activity occurred nearby. Although the stratigraphy records this dramatic subsidence, this volcanic ash is the first direct sample of the Taconic arc to the east.

- Return to vehicles and drive out of park and retrace the route to I-90.
- 44.3 0.0 Turn left to Old Sharon Road
 - 45.2 0.0 Turn left to Moyer St.
 - 46.0 0.0 Cut obliquely across (left) Montgomery St. to Mitchell
 - 46.1 0.0 Turn right onto Rt 5S.
 - 46.2 0.0 Turn left to I-90 entrance. Get ticket and proceed on I-90 eastbound (Albany)
 - 50.2 0.0 Again the Little Falls Dolostone is exposed to the right (south), but here you can also see a well exposed outcrop of the Precambrian basement across the Mohawk river on Rt. 5N. This exposure is also provisionally assigned to the Peck Lake Formation (circa 1.1 Ga) and is intruded by a dark-weathering basaltic dike that has been dated at about 700 Ma (see locality 12 in Fisher, 1980). Dikes of about this age are common in the Adirondacks.
 - 55.8 0.0 Exposure of Trenton Group resting unconformably on the Beekmantown Group. Look closely because the unconformity is at the very beginning of the outcrop (west side).
 - 72.2 0.0 Exposures of the Beekmantown Group on the right.
 - 72.6 0.0 Cross the approximate trace of the Wolf Hollow Fault.
 - 75.5 0.0 Excellent exposures of interbedded medium- to thin-bedded sandstones and shale of the Middle Ordovician Schenectady Formation. If observed from a distance, this turbidite sequence shows low-angle truncations that probably represent channeling on a submarine fan.
 - 76.2 0.0 Exposures of the Middle Ordovician Schenectady Formation.

- | | | |
|------|-----|--|
| 77.2 | 0.0 | Exit at Exit 26 - Schenectady, I-890. Pay toll (\$1.00 in 1995). Continue on I-890 east. |
| 80.3 | 0.0 | Take Exit 2 (Rt. 337 - Campbell Road). |
| 80.4 | 0.0 | Turn right on Rt. 337 south. |
| 80.9 | 0.0 | Turn right on Putnam Road. |
| 82.7 | 0.0 | Turn right on Rt. 159. |
| 85.0 | 0.0 | Turn right into the entrance/parking lot of the Plotterkill preserve. This is the Schenectady County Nature and Historic Preserve. Follow the Red trail (Figure 13) to the top of the waterfalls at the confluence of Rynex Creek and Plotterkill Creek. Stop 5. |

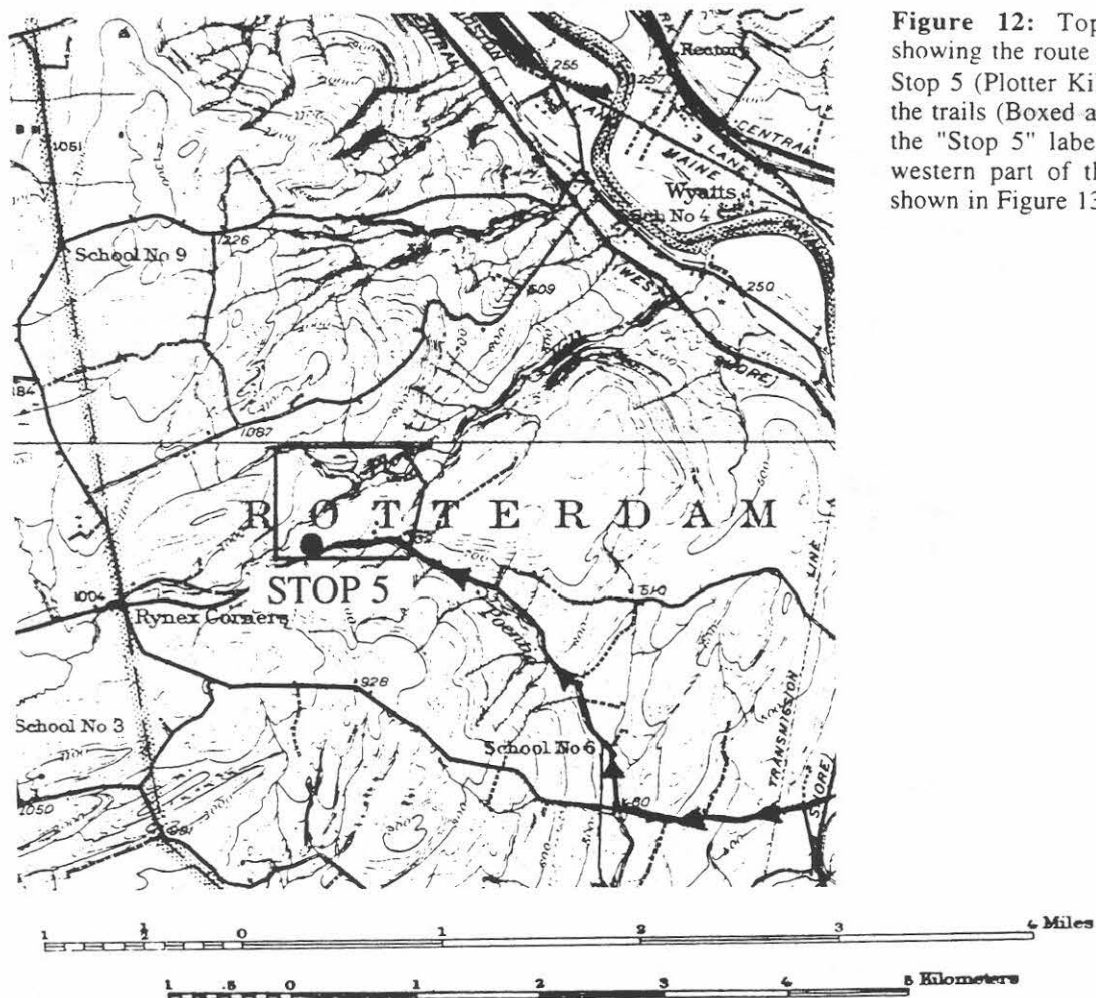


Figure 12: Topographic map showing the route and locations of Stop 5 (Plotter Kill). Location of the trails (Boxed area just north of the "Stop 5" label) in the southwestern part of the Preserve are shown in Figure 13.

STOP 5: Plotterkill Preserve - Schenectady Formation

The Plotterkill Preserve is a private preserve that surrounds the lower reaches of the Plotterkill stream which drains north into the Mohawk River (Figure 12 and 13). The stream cuts into and beautifully exposes sandstone and shale of the Middle Ordovician Schenectady Formation (see Locality 30 in Fisher, 1980). The Schenectady Formation is Mohawkian in age (upper Middle Ordovician). Most of the formation (and the Utica Shale) is restricted to the Nowadagan stage but the lower beds belong to the Canajoharian stage (i.e., the top of the Mohawkian). In general the Schenectady Formation is composed of interbedded light-brown- to buff-weathering, medium- to fine-grained sandstone with interbedded gray to black laminated shale. The sandstones are laminated and rippled and locally graded. They commonly have flutes and grooves on their bases and ripples on the tops. The unit is virtually unfossiliferous with exception of uncommon graptolites. These sandstones are interpreted to have been deposited in very deep water and to have been transported to the site of deposition by turbidity currents.

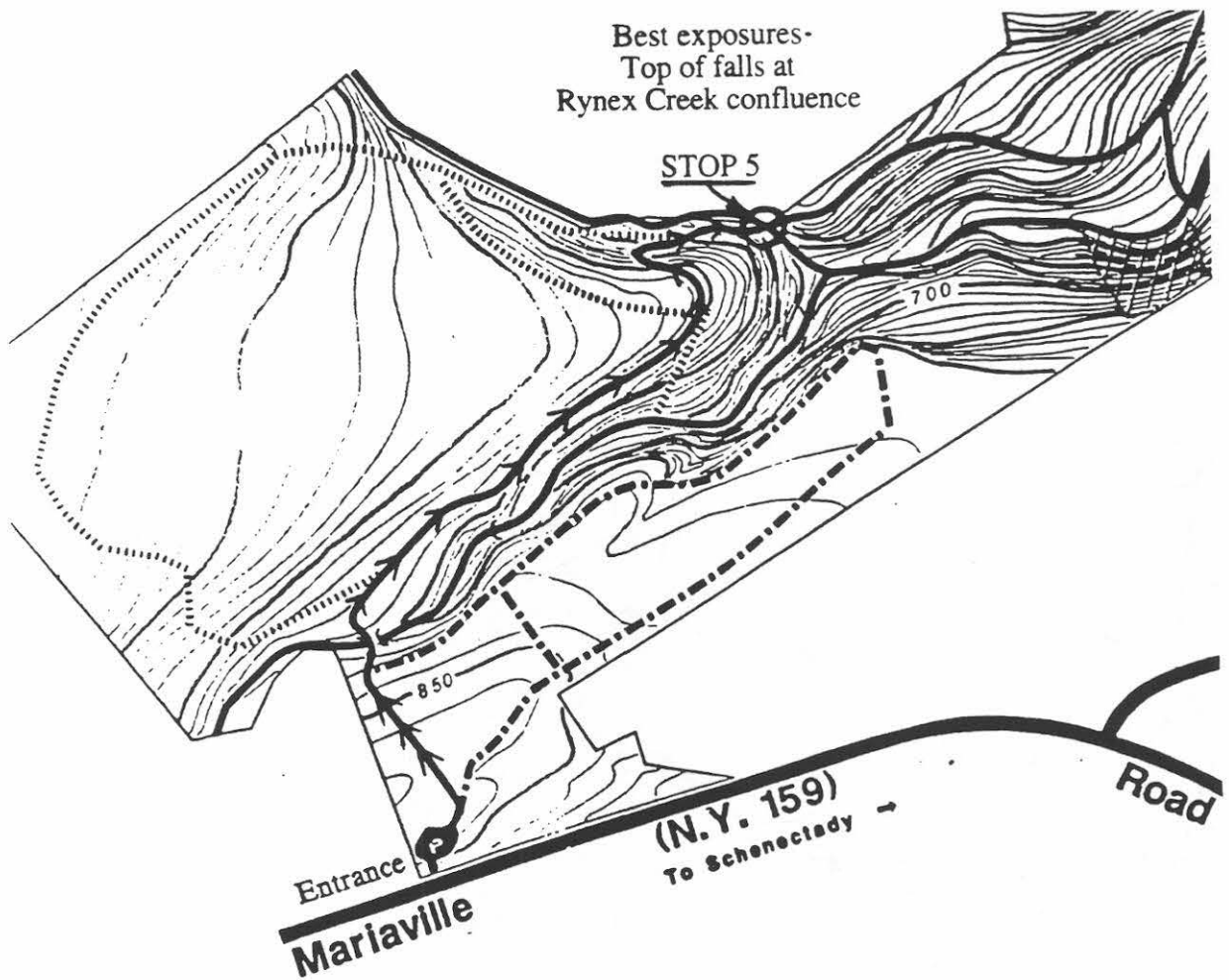


Figure 13: Trail map showing the route of the Red Trail which leads one to the confluence of Rynex Creek and Plotterkill Creek.

Although many geologists accept the notion that the clastic material had a source to the east, the paleocurrent data from this unit do not clearly support this contention. Using data collected from the Plotterkill Preserve and data from several other sites in the Mohawk Valley one can reconstruct the paleoflow directions within the basin during deposition. Virtually all of the paleocurrents trend to the north-northeast. As explained above, these current directions may reflect sedimentation on a submarine fan that formed axial to the principal structural trend of the foreland basin.

For students in *Stratigraphy and depositional environments of New York*, this project involves measuring, plotting, and interpreting paleocurrents and relating this information to concepts of basin infilling. In the field, students spend several hours measuring flutes, grooves, and ripple marks from relatively flat-lying sandstone beds with a Brunton compass. Measurements are then entered into computer data files and plotted using Macintosh software. In addition to the data set collected by the students (all the data from different groups are merged), they are given two additional data sets from geographically separated outcrops of the same unit.

The Schenectady Formation has very good paleocurrent indicators - mainly flute and grooves. Virtually all of the paleocurrents in rocks of the Schenectady Formation in the Plotterkill preserve indicate flow from the southwest to the northeast (flow direction to about 020 to 030°) (see Figure 14). If these sediments were derived from the Taconic Arc to the east, then why do the ancient flow directions tell us that the sandstone flowed from the southwest? The answer lies in a possible interpretation concerning how turbidites flow in a basin and the shape of the basin. Turbidites flow downslope. If they encounter the deepest part of the basin, they tend to turn to flow along the basin axis as opposed to flowing upslope on the other side. In this regard, we recognize that foreland basins can develop two different types of submarine fans. The first, called a lateral fan, is composed of turbidites that accumulate directly off the edge of a large source terrain. In this case, all of the paleocurrents tend to record flow directly away from the source. The second, called an axial fan, records flow parallel to the source and possible

parallel to the structures that influenced the source. This part of the Schenectady Formation was most likely deposited as an axial fan with a main feeder canyon somewhere to the south. It is interesting to note that the average paleocurrent direction in the Schenectady Formation is perpendicular to the direction of tectonic transport (thrusting direction) in the western Taconics (about 290 to 300° as determined at five localities). If thrusting in the Taconics was related to basin formation, then one would predict a foreland basin with a trough-like axis oriented at about 020-030° - exactly the orientation of the paleocurrents. A similar trend in sediment dispersal was recognized in correlative rocks in Pennsylvania, where the sandstones of the Martinsburg Formation indicate basin infilling by both lateral and axial submarine fans (McBride, 1962).

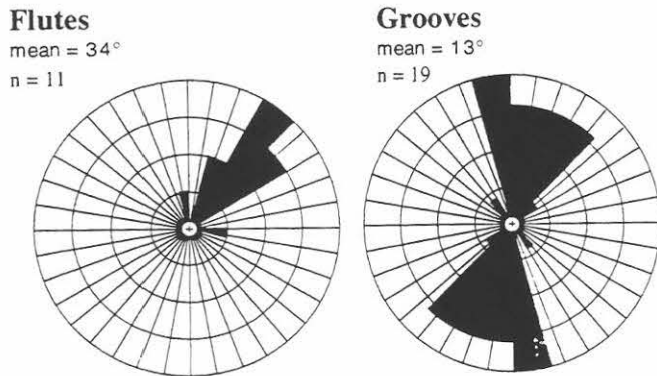


Figure 14: Paleocurrents from the Schenectady Formation in the Plotterkill Preserve (Stop 5). These polar histograms (or Rose diagrams) show the results of measurements taken in 1994 by the "Stratigraphy and depositional environments of New York" class at Union College. The diagrams can be read like a compass with north directly to the top, and east at 90°, etc. The flutes have a mean orientation of 34°, the grooves have a mean orientation of 13°; and together, if plotted all as bi-directional current indicators, they have a weighted mean of about 21°. From these data we can infer that the turbidites of the Schenectady Formation in the Plotterkill Preserve, were transported to the north-northeast. As discussed in the text, this orientation is inferred to be axial within a foreland basin.

In this area, the Schenectady Formation is underlain by the Utica Shale (i.e. Wintergreen Park), and together their thickness is nearly 1000 to 1800 meters. This shale/sandstone package thins gradually to the west and thickens dramatically to the front (east edge, near Troy) of the Taconic Range. The extreme thickness and the composition of these strata is an important sedimentation pattern. The thickness of strata preserved in a basin is generally a reflection of the relative depth or the amount of subsidence that occurred in the basin. This room for sediment is sometimes referred to as "accomodation space" because if there is not room for sediment (a basin), sediment tends to not accumulate. With this in mind, one could postulate that the basin was deepest, or experienced the most subsidence from here (Plotterkill) east to the Troy area where the unit thickens considerably. Shale that filled the basin has no obvious provenance or source area because the particles are so fine-grained. However, recent studies suggest that this shale reflects the changes in the composition of the thrust complex to the east (Garver et al., 1996). Sandstones contain coarse-clastic material that represents identifiable fragments of the source region. Because these and correlative sandstones have a distinct composition, workers have postulated that these sandstones had a source that lay to the east that was composed of mainly sedimentary rocks. Sandstones of the Schenectady Formation contain quartz, feldspar, and sedimentary rock fragments (with minor metamorphic and volcanic rock fragments), and a minor detrital mica.

The Schenectady Formation, therefore, was derived from the Taconic orogenic belt to the east - in this regard it is a synorogenic deposit. Paleocurrents indicate that the flow of sediment was to the north-northeast, which is exactly what we would predict if the structures presently preserved if the Taconic mountains produced a foreland basin. The sandstones and the shales were derived from the thrust complex and are therefore rich in sedimentary detritus and recycled grains - detritus directly from the Taconic arc is rare.

- Exit parking area and retrace route. Turn left out of parking lot and follow Rt. 159 eastbound.
- 87.3 0.0 Turn left on Putnam Road.
- 89.0 0.0 Turn left on Rt. 337 north.
- 89.5 0.0 End of trip. Take I-890 east for Schenectady, Scotia, Albany and all points east (take take I-90 east at the end of I-890). Take I-890 west for all points west of Schenectady, exit I-890 at the I-90 interchange. To reach I-88 to Binghamton, take I-890 West, then take I-90 East and then exit for I-88 south.

TREATMENT OF SURFACE AND GROUNDWATER: TOUR OF MOHAWK VIEW WATER TREATMENT PLANT

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INTRODUCTION

The Mohawk View Water Treatment Plant is unique in that it has three different raw water supplies: 1) a surface water reservoir in a suburban housing area with occasional algal blooms; 2) a ground water supply with high amounts of soluble iron and manganese; and, 3) a river supply with varying flows and water quality problems. The tour will show the wide variety of treatment methods used to deal with the water quality problems associated with these various raw water sources.

HISTORY OF THE LATHAM WATER DISTRICT

A public water supply was established in the Town of Colonie by the creation of the Latham Water District in the year 1929. When the District began operations, it served only a relatively small area around the present day Latham Circle. Over the years numerous extensions have been added until, at present, the Latham Water District serves most of the developed areas in the Town. When first created, the District relied on a few wells which pumped directly into the system.

As the service area expanded, the District began an almost constant search for new supplies in an effort to remain ahead of the increasing demands for water. A series of deep wells were developed throughout the Town, which also pumped directly into the system. Two of these original wells remain, useful only for emergency service. The 1950's saw the construction of the Stony Creek Reservoir Supply and the River Road Water Treatment Plant (RRWTP). In 1963, the site for the present day water supply complex was obtained and a 5.0 MGD well field was constructed adjacent to the river. This supply, completed in 1965, consisted of four wells, a 0.5 MG Clearwell, a 5 MGD High Lift Pumping Station, and a 30-inch transmission main to the distribution system. This supply was chlorinated and pumped directly into the Distribution System.

The Mohawk River flows along eight miles of the northerly boundary of the Town of Colonie. It is a great natural resource, draining an area of about 3,500 square miles. As the River is part of the Barge Canal system, the flow of water in the River is regulated by the New York State Department of Transportation. Average annual flow approximates two to three billion gallons of water per day. As early as 1963, the River was recognized as a future source of water for the Latham Water District. Various tests and studies of the River confirmed that it contained a water treatable by modern, conventional treatment processes.

In April 1966 the Town Board of Colonie, following the recommendation of O'Brien & Gere Consulting Engineers (1966), authorized what was called the Mohawk River Water Supply project, with an expenditure of \$5,000,000 for construction of a 15 MGD water supply from the Mohawk River. This additional supply of water was

In Garver, J.I., and Smith, J.A. (editors). Field Trips for the 67th annual meeting of the New York State Geological Association. Union College. Schenectady NY. 1995. p. 377-382.

planned to meet an estimated 1980 maximum daily demand of 26 MGD and to augment the existing Stony Creek Reservoir and Mohawk View well supplies.

Application was made to the New York State Water Resources Commission for permission to take 15 MGD from the River. After a public hearing, the Commission granted the taking by a decision dated October 4, 1966. The section of the River from which the Latham Water District draws water was initially classified "B" (swimming). The Commission reclassified this section of the River as Class "A" (water supply after treatment), effective February 26, 1970. The new class "A" rating compelled upstream wastewater discharges to upgrade their treatment plants.

The Latham Water District intake is located approximately midway between Crescent Dam and Lock No. 7. The impoundment from which the water is drawn has an estimated surface area of approximately three square miles and a volume of about five billion gallons. The elevation of the impoundment is controlled by the New York Power Authority at its Crescent Hydroelectric Generating Station. The New York State Department of Environmental Conservation maintains a water quality monitoring station at the Route 32 Bridge in Cohoes.

Overview of Latham Water District Water Supply

The major components of the Latham Water District Water Supply include the Intake Channel, Raw Water Pump Station, Mohawk View Water Treatment Plant (MVWTP), Clearwell storage tanks, the Mohawk View & River Road High Lift Pump Stations, and Lagoons for settling basin and backwash wastes. The water treatment plant consists of the Aeration Basin, Flocculation and Coagulation Basins, Settling Basins, Rapid Sand Filters, and chemical feed facilities. The well supply, completed in 1965, now consists of 5 wells, with a permitted withdrawal of 7.5 MGD. The well supply receives full treatment at the WTP. Although the RRWTP was abandoned in 1985, the Stony Creek Reservoir today still supplies over 5.1 million gallons of raw water (on an almost daily basis) to the MVWTP.

Intake Channel and Raw Water Pump Station

Water from the Mohawk River flows through a shore intake style Intake Channel. Water flows through a coarse bar rack and mechanized traveling water screen into the intake sump which makes up the lower level of the Raw Water Pumping Station. Suspended in this sump are four vertical turbine pumps which transmit raw water through 1900 feet of 30-inch pipe to the Aeration Basin. The four pumps are each nominally rated for five million gallons per day (MGD), thereby providing for 15 MGD capacity with any one unit out of service. A natural-gas-fired engine provides a 5.0 MGD emergency pumping capability. Chemical pre-treatment facilities were constructed in 1993 for metals oxidation and zebra mussel control.

Aeration Basin

The Aeration Basin is the point at which the River, Reservoir, and Well supplies enter the treatment process. The River and Reservoir supplies are combined in the yard piping upstream of the Aeration Basin. The Well Supply enters the Aeration Basin via an airbreak arrangement, preventing any back siphonage of raw water to the Wells.

The Aeration Basin functions primarily as a mix basin for the three possible raw water sources. Aeration and mixing are accomplished by eight floating electrically driven mechanical aerators, ensuring a high level of dissolved oxygen, providing oxidation of metals and taste-and-odor-producing compounds, and some air stripping of volatile compounds and other gases from the water. Under other operational modes employing powdered activated carbon, potassium permanganate, and/or chlorine dioxide, the Aeration Basin serves as a detention basin. Carbon is used to absorb organics, especially those causing taste, odor, and color. Chemical oxidants such as chlorine dioxide and/or potassium permanganate allow for oxidation of ferrous and manganous ions and some organics without the formation of trihalomethanes (THM). THM's are created with the interaction of chlorine with dissolved organics and forming Chloroform, etc.. The detention time in the basin at 15 MGD is 26 minutes.

Mohawk View Water Treatment Plant

The MVWTP includes facilities for rapid mixing, flocculation and coagulation, settling, and filtration of the water. The Plant has a nominal rated process capacity of 15 MGD and a hydraulic capacity of 40 MGD.

The first stage of treatment occurs before the water reaches MVWTP. Chemical pre-treatment for oxidation of iron and manganese along with taste and odor-producing compounds is practiced on all three raw water sources.

Contact times of up to 60 minutes occur before the water reaches the Aeration Basin. After the water flows through the Aeration Basin, various chemicals are introduced into the water in the Rapid Mix Chamber. Aluminum sulfate (or alum) and a polymer flocculant aid are fed to create a precipitate of aluminum hydroxide which coagulates and forms "floc" particles. These "floc" particles attract organic colloids, clay, algae, or any other particles which give the water some color or the appearance of being "dirty". The Rapid Mix Chamber is sized for 60 seconds detention time at 15 MGD.

Flow from the Rapid Mix Chamber is treated with chlorine dioxide for bacterial disinfection and oxidation as it is conducted into the Mix Basins. Gentle agitation from electrically driven vertical shaft flocculators help to form a denser, more settleable floc. Three 5 MGD dual compartmented basins provide two step flocculation. Total detention time in the flocculation basins at 15 MGD is 42 minutes.

Each Mix Basin discharges into a separate Settling Basin. Water enters the Settling Basin through orifices along the Dispersion Flume and leaves the basin through a similar arrangement in the Collection Flume. The basins are separated into two zones by a submerged half wall, and sludge is collected by two electrically driven mechanical rakes into hoppers in the bottom of the basins. The hoppers are hydraulically "vacuumed" off on regular intervals, and basins are drained and cleaned twice per year. Four hours detention time (at 15 MGD) is provided in the Settling Basins. This normally removes 90 to 99% of the particles in the water. The settled water is then delivered to the filters through the Settled Water Flume and Filter Influent piping.

Six dual media rapid sand filters perform the final step in clarification of the water. The filters are constructed of 22 inches of anthracite and 14 inches of sand on top of 10 inches of gravel. Filter bottoms are constructed with compound duplex vitrified tile blocks and each filter is provided with eight surface wash sweeps. Each filter has 595 square feet of surface area to process up to 2.5 MGD at a loading of 2.9 gallons per minute per square foot. The filters are backwashed when head loss exceeds 8 feet, when service run approaches 120 hours, or when effluent turbidity exceeds 0.35 NTU's. Filters are backwashed at a rate of 19 gallons per minute per square foot (16.5 MGD rate) until clean. Backwash water is a combination of water stored in a 0.8 MG backwash stand pipe (located to the rear of the WTP) and flow from the 30-inch finished water transmission main. A pressure regulating/altitude valve allows this tank to be refilled from a 12-inch connection to the transmission main before the water leaves the WTP site. For the first 30 minutes of a filter run after backwash, a polymer filter aid is added to condition the filter and enhance particle removal.

Final Treatment

After filtration, finished water is "post chlorinated" with chlorine for final disinfection. Corrosion control of the effluent flow is by pH adjustment with the addition of caustic soda (sodium hydroxide).

Clearwell Storage

Finished water from MVWTP flows into the Diversion Chamber. This concrete chamber splits the flow of water - approximately 70% flows downhill through a 36-inch pipeline to the MVHLPS Clearwells. The remaining 30% flows through a 12,000-foot-long 30 inch pipeline to the RRHLPS Clearwells. Two steel reservoirs are located behind the MVHLPS, a 0.5 MG tank constructed in 1965 and a 1.0 MG tank constructed as part of the Mohawk River Water Supply project. Three steel Clearwell tanks are located behind the RRHLPS, two 0.34 MG and one 0.5 MG reservoir.

High Lift Pumping Stations

Finished water is transmitted to the Distribution System by two high lift pump stations. The Mohawk View High Lift Pump Station (MVHLPS) has a rated capacity of 25 MGD. Five vertical turbine pumps are each powered by a 400 hp electric motor. Twin diesel generators provide emergency power.

The River Road High Lift Pump Station (RRHLPS) replaced the high lift pumping capabilities of the abandoned RRWTP, and has a rated capacity of 10 MGD. RRHLPS is equipped with two pumps rated for 2.5 MGD and one pump rated for 5.0 MGD. These vertical turbine pumps are each powered by a 400 hp electric motor. RRHLPS #3 has an engine-driven right-angle drive to provide emergency pumping capabilities.

Chemical Application

Chemicals may be fed in at a number of points, thereby providing flexibility in operation. Chemical feed points and chemicals which can be fed in are as follows:

<u>Reservoir</u>	Potassium permanganate
<u>Well Supply</u>	Potassium permanganate, chlorine dioxide
<u>Raw Water Pumping Station</u>	Chlorine dioxide, powdered activated carbon (PAC), potassium permanganate
<u>Rapid Mix Basin</u>	Alum, caustic soda, weighting agents, chlorine, chlorine dioxide, PAC, polymer
<u>Before Flocculation Basins</u>	Chlorine dioxide
<u>Before Filters</u>	Polymer, chlorine
<u>After Filters</u>	Chlorine, caustic soda

Chemical Handling and Feeding

Chemicals normally used in the treatment process are chlorine, aluminum sulfate, sodium hydroxide, powdered activated carbon, potassium permanganate, polymers, sodium chlorite, and chlorine dioxide. Bulk handling and storage equipment is provided for all except chlorine dioxide. The chlorine dioxide is generated on site by reacting sodium chlorite and chlorine.

Chlorine is purchased, stored and utilized in "Ton" containers each weighing approximately 3700 pounds. Electric-powered lifting equipment for handling the chlorine has been provided. Chlorine application is accomplished by standard feeders with duplication of the largest to ensure 100 percent standby.

Alum and caustic soda are purchased, stored, and fed in as liquids. Purchase is by truckload lots with storage tanks sized accordingly. Liquid alum is fed from day tanks while caustic is fed directly from storage by proportioning pumps. Volumetric and gravimetric feeders are available to feed bagged equivalent dry chemicals in the event of breakdown in the basic alum or caustic systems.

Powered activated carbon is purchased in bulk truckload lots in powder form. As it is being unloaded, the carbon is wetted and is stored, handled, and fed in slurry form. Bagged activated carbon can be fed in dry powder form via volumetric feeder in the event of failure of the normal system.

Instrumentation and Control System

The instrumentation and control system provides the Operator with indicators and records of all major WTP and Distribution System operations. From the master control console, the Operator can start and stop well pumps, raw water pumps, aerators, rapid mixer, flocculators, sludge collectors, and high lift pumps. Control valves at the inlet to the flocculation basins and at the filter discharge rate controllers are also operable from the master control console. Filter control consoles, one for every two filters, permit easy operation by manipulation of switches on the filter console. Filter backwashes are Operator-attended, and are initiated when effluent quality degrades to 0.35 NTU, service run reaches 120 hours, or if loss of head reaches 8.0 feet.

The instrumentation system utilizes pulse duration and 4-20 milliamp signals. Master control of the filter effluent rate of flow is based on the elevation of water in the plant. The Master Control matches the flow through the filters with the incoming flow of raw water (influent). Chemical feed rates are automatically varied in accordance with variations in influent flow rate.

Laboratory Facility

The MVWTP laboratory includes equipment for making physical, chemical, bacteriological and microscopic tests on the water. Pumped sampling lines deliver continuous samples to the laboratory from various stages of the treatment process. Frequent testing of the raw and finished water plus testing at different stages of treatment ensure that proper treatment is being accomplished to provide a safe and palatable water for the Latham Water District customers.

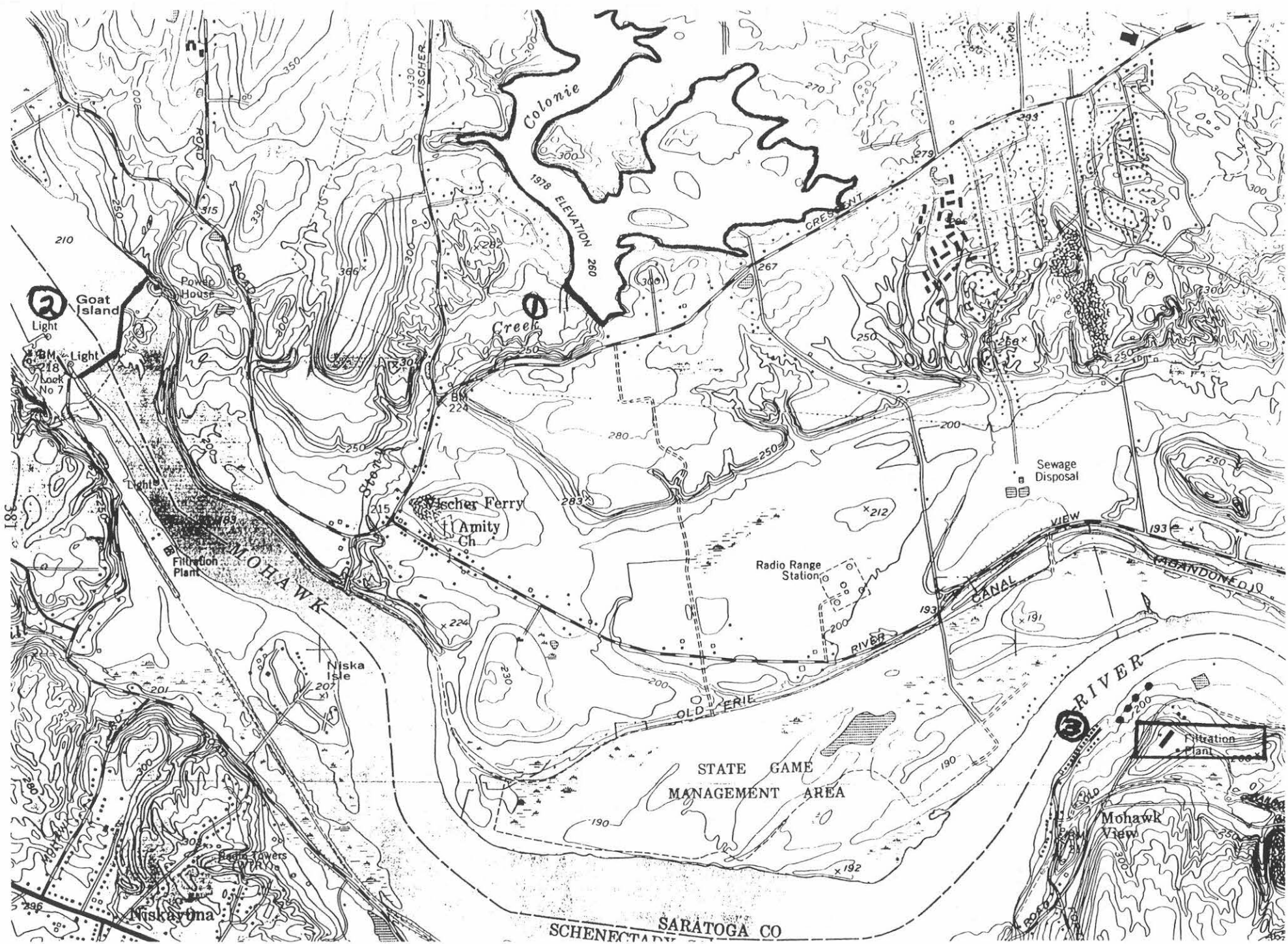


Figure 1. Location map showing: 1) the surface reservoir; 2) the Mohawk River; and, 3) the well field and Water Treatment Plant.

Waste Lagoons

Two lagoons provide gravity settling and thickening of process waste (filter backwash water, mix and settling basin waste flow, etc.) before the clarified supernatant water is discharged back into the Mohawk River. Discharge of this wastewater flow is governed by a New York State Department of Environmental Conservation (NYSDEC) State Pollution Discharge Elimination System (SPDES) permit. Original design provided for between 1 and 4 years of sludge storage before cleaning is required. The current cleaning interval is 1 year and is accomplished by conventional excavation equipment. Ramped access for vehicular traffic makes cleaning operations as easy as possible.

Future Expansion

Future expansion has been considered in design to permit the eventual doubling of the capacity of the MVWTP. Groundwater supplies adjacent to the river and in the glacial Colonie Channel have been explored. The Raw Water Pumping Station can be doubled in capacity by replacing present pumps and motors with larger units and duplicating the raw water transmission pipeline. Room is available for duplicating the Aeration Basin. Provision is made at the Rapid Mix Basin in the WTP to duplicate the piping to and from the mix basin. The number of flocculation basins, settling basins, and filters can be increased by addition of a single bay at the eastern end of the Filter Gallery or by construction of a completely new wing on the Western side of the WTP. There is ample space for constructing additional Clearwell Storage and improvements to the MVHLPS. Additional power supply and a second finished water transmission main to the Distribution System, required when total water demand from the site exceeds 25 MGD, were considered when laying out the present plant, pumping, and storage facilities.

ADDITIONAL SOURCES OF INFORMATION (NOT REFERRED TO IN TEXT)

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Simpson, E.S. 1949. Buried Preglacial Ground Water Channels in the Albany-Schenectady Area in New York. US Geological Survey and New York Water Power and Control Commission, Albany, NY, Bulletin GW-20A.

O'Brien & Gere Engineers, Inc. April 1991. Well Field Assessment, Mohawk View Site, Town of Colonie, 34 p.

Dineen, Robert, et.al. 1983. "Bedrock Topography and Glacial Deposits of the Colonie Channel Between Saratoga Lake and Coeymans, New York" with Waller, Roger "Ground-water Potential of Capital District Buried-Valley Deposits". New York State Museum Map and Chart Series #37, Albany, NY.

TRIP GUIDE

Leave Civil Engineering/Geology Building (Butterfield Hall) at Union College, exiting the Campus via Nott Street/VanVranken Street (northeast). Turn right at light onto Nott Street. At the second light, bear left onto Rosa Road (Ellis Hospital on right). Proceed for 1 mile until stop light on RT#146 (Balltown Road). Go straight (toward GE R&D Center) on River Road. Proceed 2.5 miles east, past GE R&D Center, on River Road, turning left at stop sign on to Rosendale Road. Follow Rosendale Road for 2 miles. Ignore turn to right at arrow onto Vly Road, and proceed straight on Old River Road. Proceed on Old River Road 2 miles until intersection of Forts Ferry Road. Proceed straight across intersection (River Road changes to Onderdonk Road after intersection). The treatment plant is on the right side immediately after the bicycle path crossing. Take access road into the parking lot (#3 on map).

Information as to access and possible tours of the plant can be answered by calling Robert Maswick (518-783-2774) at the Mohawk View Water Treatment Plant. A schematic of the plant will be made available at the Treatment Plant.

ENVIRONMENTAL TECHNOLOGY AND PRESERVATION: THE PINE BUSH, LANDFILLS, AND GROUNDWATER INTEGRITY

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West Sand Lake, NY 12196

INTRODUCTION

The active Albany Interim Landfill and the closed Albany City Landfill are located in the City of Albany, New York, in a region of stabilized dunes known as the Pine Bush (Figure 1). Surface deposits in the Pine Bush consist of highly permeable sands originally deposited by the paleo-Mohawk River as delta deposits during the high stand (maximum extent) of glacial Lake Albany. The sands constitute a large unconfined aquifer. Subsurface investigations have shown that the nearby unlined Albany City Landfill, which was opened in 1969, has contributed both organic and inorganic contaminants to the Pine Bush aquifer. The lined Albany Interim Landfill was designed in a manner intended to protect groundwater integrity in the surrounding Pine Bush region. Groundwater contamination on the scale of the area of the Albany City Landfill can be difficult to remediate. Remediation of smaller commercial sites with contaminated groundwater is a somewhat more tractable problem. This field trip provides an opportunity to visit the Pine Bush, the Albany Interim Landfill, the Albany City Landfill, and a small commercial site (Stewart's Shop #182) that is currently undergoing groundwater remediation.

THE PINE BUSH

The Pine Bush is a region of about 40 square miles of sand dunes and bogs covered by pitch pine and scrub oak forests located between Albany and Schenectady, New York. The Pine Bush occupies part of the townships of Colonie, Guilderland, and Rotterdam, and the western arm of the City of Albany. It is bounded by the Schenectady City Line and New York State Route 5 on the north, the Normans Kill on the west and south, and New York State Route 85, Osborn Road, and Albany-Shaker Road on the east (Figure 1).

The Pine Bush is a small segment of a dune field that extends from South Glens Falls to Delmar. The Mohawk River and the inlets and outlets to Saratoga and Round Lakes dissect this field. Dunes of windblown sand range from hundreds to thousands of feet long. The dunes lie on a nearly level surface known as the Lake Albany Plain which represents the bed of a glacial lake that existed 20,000 years ago (Dineen, 1975).

Bedrock Geology of the Pine Bush

Bedrock underlying the Pine Bush is dominated by Middle Ordovician shale and sandstone. The bedrock has been mapped as two units: the Schenectady Formation, which is mostly sandstone with some shale, and the Normanskill Formation, which is shale with some sandstone and chert (Fisher et al., 1977). The Schenectady Formation is jointed, locally faulted, and relatively flat-lying (see Kidd et al., this volume). The Normanskill Formation is highly faulted and folded. In general, the Schenectady Formation underlies hills and ridges and the Normanskill Formation underlies the preglacial valleys (Dineen, 1975).

In Garver, J.I., and Smith, J.A. (editors), Field Trips for the 67th annual meeting of the New York State Geological Association, Union College, Schenectady NY, 1995, p.

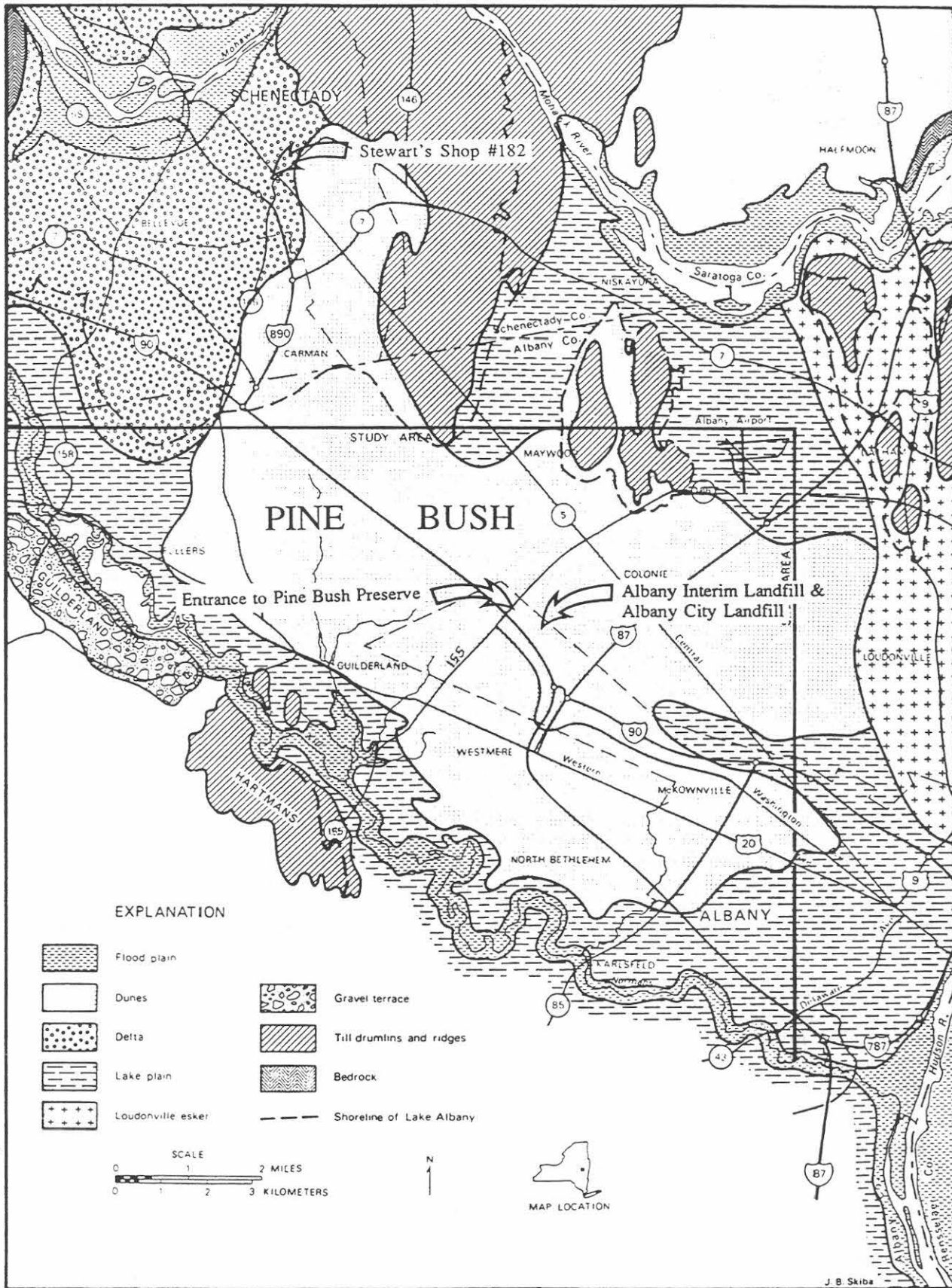


Figure 1. Generalized glacial geology map of the area between Albany and Schenectady. The Pine Bush is the area of dunes in the center of the map (modified from Dineen, 1975).

Prior to the last glaciation, major streams in the Capital District preferentially followed valleys in the less resistant shale; these streams included the preglacial Mohawk, Alplaus, and Colonie Rivers. The Mohawk and Alplaus Rivers met beneath present-day Guilderland; the Mohawk and Colonie Rivers met beneath the present site of the village of Karlsfeld and flowed south, west of the current Hudson River valley. The preglacial Mohawk and Colonie channels underlie the southern and eastern parts of the Pine Bush, respectively (Dineen, 1975; 1976; 1982). Because the Colonie channel has the flattest gradient of the tributaries as presently known, it may have been the major river draining the southeastern Adirondack region in Tertiary times (Isachsen, 1965). Figure 2, which is from Dineen (1976), shows the bedrock topography in the Pine Bush.

The buried bedrock surface in the Pine Bush has over 300 feet of relief. The bedrock surface with the highest gradient is toward the preglacial Mohawk Channel, which lies along the southern edge of the Pine Bush. The gradient eastward, towards the preglacial Colonie Channel, is gentler. A bedrock terrace with elevations of 275-300 ft above sea level (ASL) underlies the Pine Bush. The bedrock terrace is cut by tributaries to the Mohawk Channel at its western side and along its southeastern portion. The bedrock valleys acted as sedimentary basins that received thick accumulations of glacial sediment; the glacial sediments are thin on the bedrock terrace (Dineen, 1982).

The complex subsurface relationships of buried channels and overlying glacial deposits determine the local potential for groundwater under the Pine Bush (Dineen, 1976).

Glacial History of the Pine Bush

At its maximum extent about 20,000 years ago, the last major ice sheet covered all of eastern New York and New England north of Long Island and Staten Island. Ice thickness in the Capital District at this time may have exceeded 3,000 feet, while sea level stood about 350 feet below present sea level, exposing much of the continental shelf. Downwarping of the crust under the weight of glacial ice has been estimated at 1,000 ft in southern Québec and about 0 ft in New York City at the ice margin. Crustal uplift occurred during the glacial retreat as the weight of the ice decreased. In the Hudson-Champlain Lowland, a sequence of glacial lakes and marine invasion has been related to episodes of crustal uplift (LaFleur, 1976).

During 5,000 years of recession of the ice front, the terminus of the glacier withdrew from Long Island to Albany. Meltwater filled the Hudson Valley and the surrounding area with water to a level of 330 feet above present sea level, forming glacial Lake Albany, which extended from Glens Falls to Newburgh at this time. The entire Pine Bush region was submerged. During Lake Albany time, the glacial Mohawk River flowed through the Mohawk Valley and entered Lake Albany at Schenectady, depositing the extensive Schenectady delta (see Wall and LaFleur, this volume). There, delta deposits consisted mainly of cobbles and gravel. Glacial Mohawk River currents carried sand, silt, and clay further eastward into Lake Albany (Stoller, J. H., 1911; Dineen, 1975; Wall and LaFleur, this volume).

The draining of Lake Albany is generally attributed to post-glacial crustal uplift which favored a more active southward drainage by increasing regional gradient. The successor to Lake Albany was Lake Quaker Springs. Discharges through the Mohawk Valley of normal river flow alternating with catastrophic lake outbursts from central New York produced a series of channels beginning with the narrow valley now occupied by Ballston Lake and the Mourning Kill. Lowering of Lake Albany by 30 feet (to 300 feet above present sea level) to the Quaker Springs level exposed the Schenectady delta as a land mass, and required Mohawk drainage to flow northward around its western edge and through the Ballston Channel. An embayment of Lake Quaker Springs extended through the Pine Bush area where fine sands continued to accumulate in shallower water. Some winnowing of exposed Schenectady delta sands also occurred during this episode (LaFleur, 1976; Wall and LaFleur, this volume).

As the water level receded in Lake Quaker Springs, a temporary stillstand occurred near 190 feet. All of the Pine Bush stood high and dry during this episode. Wind deflation and incipient dune formation altered the original lake floor configuration. Catastrophic discharges through the Mohawk from a draining Lake Iroquois in central New York occupied the Ballston-Round Lake channel and may have initiated the modern course of the Mohawk from Alplaus to Cohoes (LaFleur, 1976; Wall and LaFleur, this volume).

With the recession of Lake Albany and the smaller lakes that succeeded it, the drained lake bed was exposed to the action of wind and streams which eroded channels into the exposed lake bottom. A cool, dry climate prevailed as the water and ice retreated farther north. The climate and the lack of subsoil on the lake floor inhibited development of vegetation. Wind action caused the sand to be eroded into finer particles and accumulate into dunes that currently characterize the Pine Bush. These dunes covered the old lakebed, and depressed areas around dunes developed into bogs (Dineen, 1975; Donahue, 1976).

Glacial Deposits in the Pine Bush

Seven units have been defined within the glacial deposits in the Pine Bush. Most of the glacial deposits are wedge-shaped masses that are draped against the bedrock terrace (Dineen, 1982). Figure 1 is a map of the glacial geology of the Pine Bush from Dineen (1975). A tabulation of glacial deposits in the Pine Bush is included as Figure 3 (Dineen, 1982). A generalized stratigraphic column for the glacial deposits is as follows (youngest deposits at the top, oldest at the bottom):

Dune sand
Lake Albany Sand
300-foot clay
Lake Albany silt and sand
Lake Albany Silt and Clay
Ice-contact Sand and Gravel
Till

Descriptions and interpretations are combined in the following introduction to each of the seven units, which are arranged in order from oldest to youngest to clarify the depositional sequence. The following is summarized from Dineen (1975 and 1982) and Donahue (1976).

1. Till

Till is a mixture of boulders, gravel, sand, silt, and clay. Generally, till overlies bedrock and underlies other units. Thickness ranges from 5 to 150 feet. Till is thickest in preglacial valleys of the Pine Bush (e.g., the valley of the preglacial Mohawk River) and thin-to-absent elsewhere. Till is thicker towards the northeast (in the vicinity of a drumlin field located just north of intersection of Rts. 5 and 155). Till was deposited directly beneath and on the glacier.

2. Ice-contact Sand and Gravel

Stratified sand and gravel deposited by the meltwater in contact with the glacier underlies much of the Pine Bush. Locally, this unit lies above the till and below the Lake Albany silt and clay. Exposures are few. Ice-contact sand and gravel partially fill the small tributary valleys that are cut into the bedrock terrace. This is a relatively thin unit that grades vertically and laterally into the basal Lake Albany silt and clay. Ice-contact sand and gravel (and the lower part of the Lake Albany silt and clay) were deposited in 150-foot-deep water of glacial Lake Albany between 0.5 and 2 miles of the retreating glacial terminus, at a time when the water elevation of Lake Albany was at 340 feet.

3. Lake Albany Silt and Clay

Very-fine-grained glaciolacustrine silt and clay grades downward into ice-contact sand and gravel and upward and laterally to Lake Albany silty sand and Lake Albany sand. This unit rarely extends above 260 ft ASL; it pinches out against the bedrock terrace, causing a "drape effect" on underlying bedrock. Local wedges and lenses of silty sand and sand are present. The lower part of the Lake Albany silt and clay (and the ice-contact sand and gravel) were deposited in 150-foot-deep water of glacial Lake Albany between 0.5 and 2 miles of the retreating glacial terminus, at a time when the water elevation of Lake Albany was at 340 feet.

4. Lake Albany Silty Sand

Lake Albany silty sand contains 10 to 50 percent silt. The unit grades upward and eastward into lake sand and thickens toward the south and southeast. The silty sand pinches out toward the east and is dominant above an elevation of 200 ft ASL. Several elongate valleys or depressions are present in the upper surface of the unit. Lake Albany silty sand was deposited offshore of a sand bar system in water that was 50 feet deep during successively lower water levels of Lake Albany (Dineen, 1982).

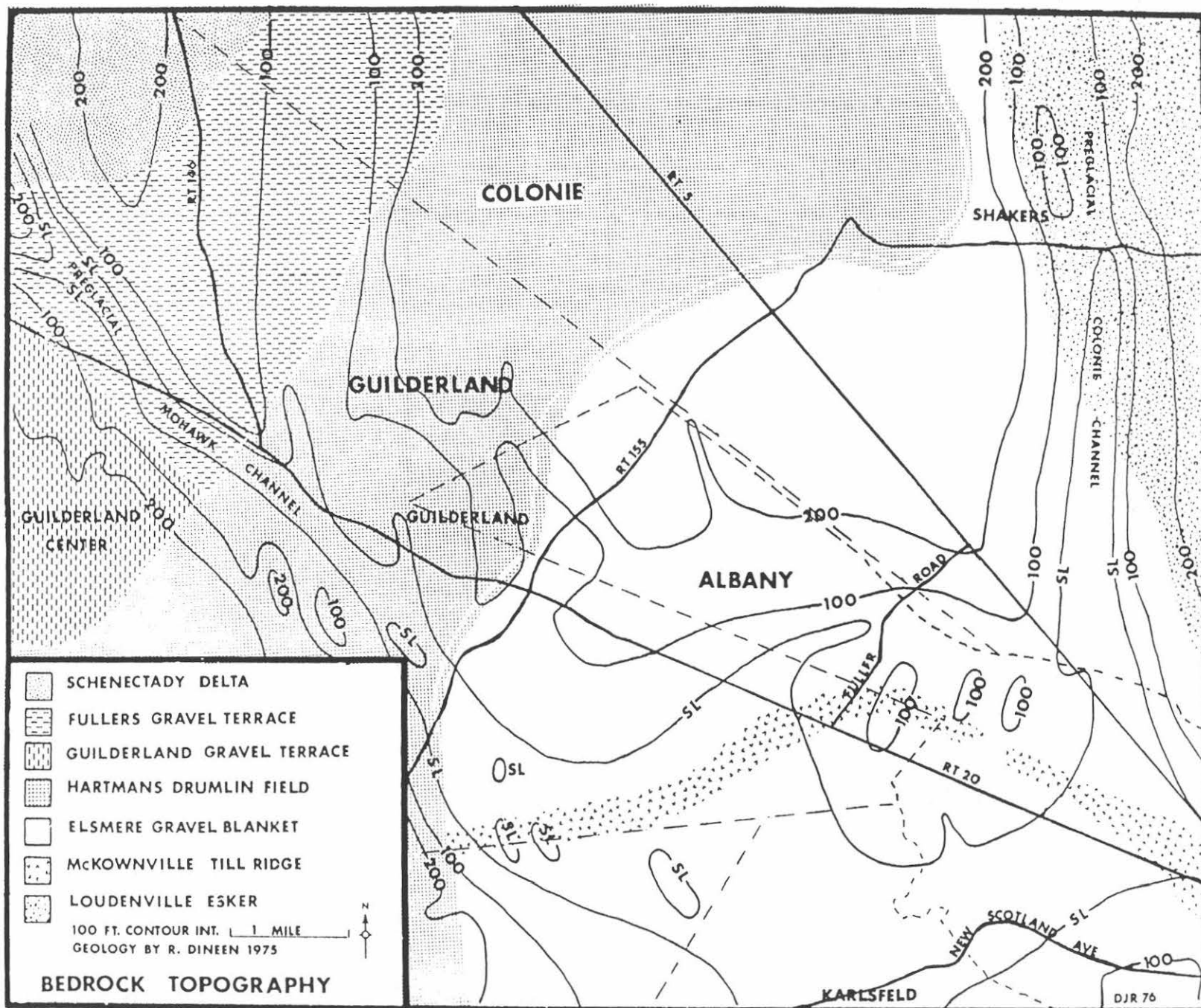


Figure 2. Bedrock topography of the Pine Bush. Note the locations of the Mohawk Channel at left and the Colonie Channel at right (from Dineen, 1976; used by permission).

AGE	UNIT	MAP SYMBOL	DESCRIPTION	SEDIMENTARY STRUCTURE	THICKNESS: METERS	NOTES	
RECENT	ARTIFICIAL FILL	AF	Heterogeneous, with concrete fragments, boulders, cobbles, sand, silt, clay, and organic matter. Color is variable.	Massive to thickly bedded.	0.3 to 15, 1 average	Permeable to impermeable, loose to very compact, can contain leachable materials and chemicals.	
	<i>Unconformity</i>						
HOLOCENE	DUNE SAND	QDS	Slightly silty, light yellow brown, fine to very fine sand, grains are subangular to subrounded.	Cross-bedded, > 15° dip	1.5 to 30, 6 average	Highly permeable, loose, weathered to 6m, readily eroded by wind and water.	
	<i>Disconformity</i>						
PLEISTOCENE	Woodfordian	LAKE ALBANY SAND	QLAS	Slightly silty, light yellow brown to light gray, subangular, medium to very fine sand.	Thin bedding, some ripple cross-laminae.	0.3 to 15, 1 average	Highly to moderately permeable (vertical permeability impeded where silt layers are present), loose to compact, unstable in steep slopes, easily eroded by water.
		300 FT. CLAY	QLAV	Silty, varved, light gray to light brown clay, siltier at top and bottom.	Varves and 0.3m beds	0.3 to 5, 1 average	Impermeable, plastic, unstable in slopes, spring lines form at upper contact with sand, is an aquiclude, impedes vertical water movement.
	<i>Disconformity to gradational contact</i>						
	LAKE ALBANY SILT & SAND	QLAM	Silty to very silty, light yellow brown to light gray sand.	Horizontal to ripple laminated.	0.3 to 15, 8 average	Moderately permeable, loose to compact, silt unstable in steep slopes, impedes water movement and "clouds" water, erodes easily by water.	
	<i>Disconformity to gradational contact</i>						
	LAKE ALBANY SILT & CLAY	QLAC	Varved, brown to gray silty clay and clayey silt, trace sandy beds.	Varves with ripple cross-laminated silt beds.	8 to 50, 15 average	Impermeable, plastic, some water movement along silt beds, flows or slumps in slopes > 12°, flows under load, is aquiclude.	
	<i>Unconformity to gradational contact</i>						
ICE-CONTACT SAND & GRAVEL	OI	Stratified brown to dark gray gravel to silty sand and gravel, tends to fine upward into silt.	Graded bedding is frequent, ripple-laminated (sand & silt).	0.3 to 3	Permeable, loose to compact, water is frequently under artesian pressure.		
<i>Unconformity to disconformity</i>							
TILL	OT	Dark gray to dark brown, bouldery, gravelly, sandy clay with few lenses of gravel.	Massive	1.5 to 15, 5 average	Impermeable, compact, sometimes permeable at base.		
<i>Unconformity</i>							
ORDOVICIAN	BEDROCK		Black shale and dark gray graywacke	Well bedded		Impermeable, and compact except where fractured.	

NEW YORK STATE GEOLOGICAL SURVEY Geology, R. Dineen Graphics, J. Skiba 12/80

Figure 3. Summary of glacial deposits in the Pine Bush (from Dineen, 1982).

5. Lake Albany Sand

Lake Albany sand consists of fine- to medium-grained sand with thin laminae of silt. Lake Albany sand is overlain by dune sand. Differentiation of lake and dune sand is based on vertical change from horizontal and ripple-laminated lake sand to dune sand with thick, high-angle (>30°) foreset beds. Lake Albany sand thickens towards the east. Thick wedges of lake sand were deposited offshore along a sand bar system, creating a lagoon where the 300-foot varved clay (below) was deposited. Silty sand (above) was deposited offshore of the sand bar system.

6. Lake Albany 300-foot Clay

The 3-to-6-foot-thick 300-foot clay layer is a varved clay deposit located within the upper section of the lake sand. The 300-foot clay has a distinct upper contact, which slopes downward toward the south and east. Several 200-to-500-foot gaps are present in the layer, coinciding with the depressions or valleys in the surface of the underlying silty sand. The gaps allow recharge to the lower units. The 300-foot varved clay was deposited in a lagoon protected by thick wedges of lake sand.

7. Dune Sand

Cross-bedded dune sand is the surficial deposit in about 80 percent of the Pine Bush. The steepest internal layers of the dunes (i.e., slipfaces) dip to the northeast to southeast, suggesting that the dune-forming winds were from the southwest or northwest. The lower contact of the dune sand is a thin, shale-granule lag-concentrate that overlies the lake sand. Particles of quartz dominate the dune sand. The dunes, which are between 5 and 50 feet high, form a discontinuous, irregular mantle over the lake sand. With a decrease of the level of glacial Lake Albany to 270 feet, wind and streams began to erode the exposed shore and sand plain sediments and the construction of large sand dunes began. Streams eroded valleys through the 300-foot varved clay, carrying sand to the near-shore zone of the lake, while additional varved clay was deposited in deeper water offshore. The exposed sand plain increased in width as the lake level continued to fall to 250 feet. Wave action formed a beach along the southern shore and large streams cut into the exposed lake floor. Dune formation continued until approximately 5,000 years ago. The cessation of dune formation has been attributed to dune stabilization by vegetation.

The Pine Bush Aquifer

The Pine Bush aquifer comprises the "silty sand", "lake sand", and "dune sand" units. These sands are widespread, thick, and permeable. The Pine Bush aquifer consists of very-fine to medium sand and ranges in thickness from 5 to 150 feet. Depth to the water table is 10 to 15 feet in most of the Pine Bush and rarely exceeds 20 feet. Hydraulic conductivity ranges from 65 to 70 feet per day. Finer grained silt and clay form the base of the aquifer, and discontinuous lenses of silt and clay are prevalent (Snaveley, 1983).

Recharge of the aquifer occurs near a groundwater divide that is located near the intersection of Rt. 155 and Washington Avenue Extension. Groundwater drains westward into the Hunger Kill, southward into the Kaikout Kill, and eastward into Patroon Creek. Silt and clay and the 300-ft varved clay are aquicludes that underlie the aquifer; they impede the vertical movement of water. Groundwater generally migrates downward along the top of the silt and clay until it enters a stream valley. The presence of the 300-ft clay causes perched water tables (Dineen, 1982), which increase the complexity of the hydrology of the Pine Bush aquifer.

Mean annual recharge in the Pine Bush, based on calculation of mean annual base flow, is estimated to be 12.7 inches. Precipitation is the only source of recharge, and approximately 38 percent of the annual precipitation recharges the Pine Bush aquifer, while the rest is lost to direct runoff to streams and evapotranspiration (Snaveley, 1983).

Computer modeling has been used to simulate drawdown by a pumping well that taps the center of the surficial aquifer. Results of the simulation indicated that a yield of 150 to 600 gallons per minute (216,000-864,000 gallons per day) could be obtained with maximum drawdown of 80 percent of the saturated thickness of the aquifer at hydraulic conductivity of 25, 50, or 100 feet per day (Snaveley, 1983).

Groundwater of the Pine Bush aquifer was sampled for phosphorus, nitrogen, and chloride by the U.S. Geological Survey in April 1979. These constituents were chosen because they are the most common chemical contaminants from septic-tank effluent and sewer-pipe leakage. Chloride was also of concern as an

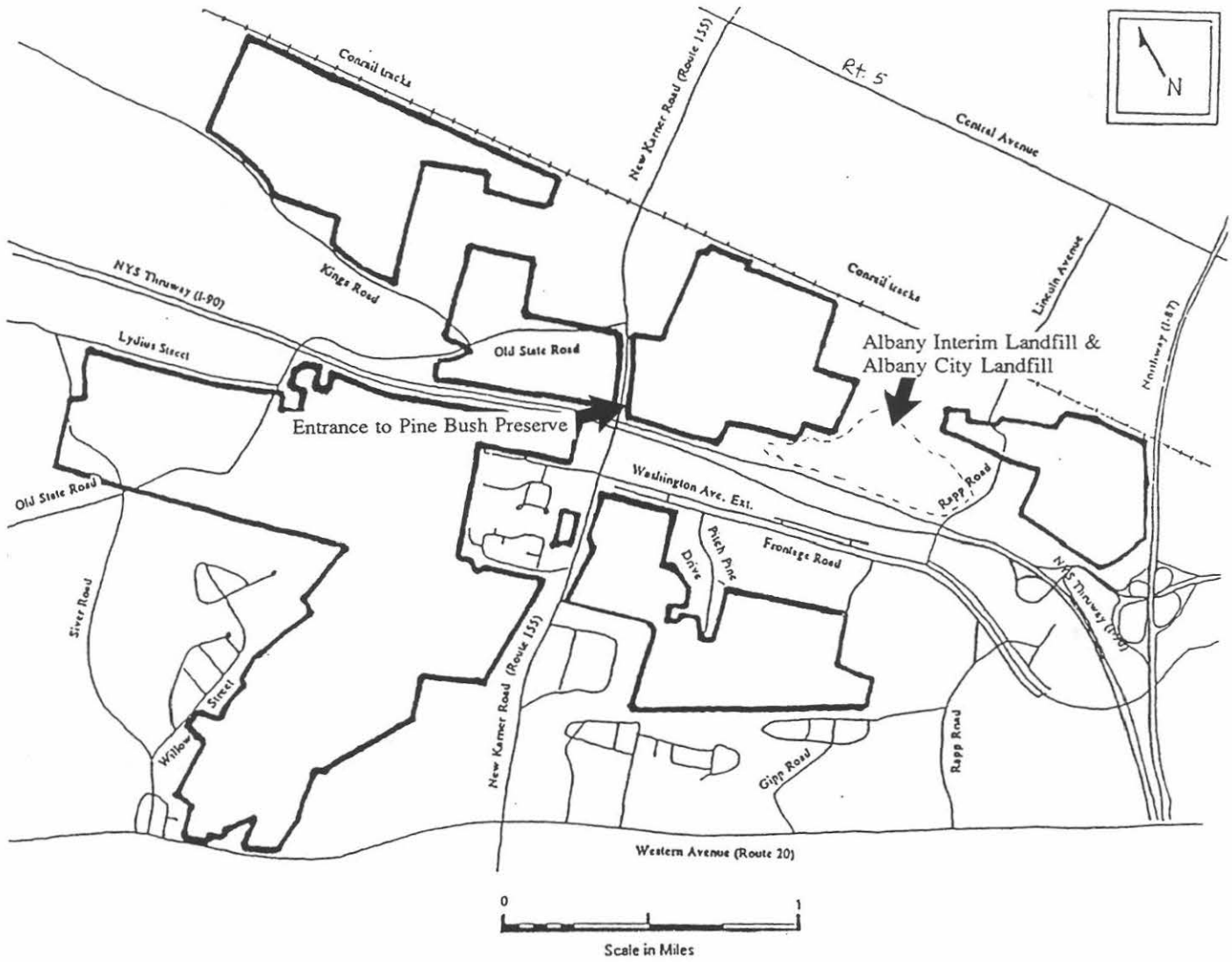


Figure 4. Boundaries of the Albany Pine Bush Preserve and approximate locations of the Albany Interim Landfill and the Albany City Landfill (modified from Albany Pine Bush Preserve Commission, 1995).

indicator of contamination from the nearby Albany City Landfill and from road salting on major roads and highways. The highest phosphorus concentrations were only slightly above the laboratory detection limit. The few high nitrogen concentrations were in shallow private wells near septic tanks. Chloride concentrations, however, ranged over two orders of magnitude (1.1 milligrams per liter, or mg/l, to 340 mg/l). Milligrams per liter are approximately equivalent to parts per million. Three of the thirty wells sampled had chloride concentrations that exceeded the 250-mg/l maximum level established by the State Sanitary Code in 1979. The highest chloride concentrations were generally in the central part of the Pine Bush near major highways. Because of this correlation, road salting was suspected as the source of the higher chloride concentrations. Wells that were located in the area farthest from the nearest major road and upgradient from the highway had the lowest chloride concentrations. Water from the deeper wells had higher chloride concentrations than water from the shallow wells, which was interpreted to suggest a chloride stratification in groundwater (Snaveley, 1983).

The sand deposits that are the primary source of shallow subsurface water in the Pine Bush have high permeability and contain relatively large quantities of water. Groundwater in the sand is readily contaminated by seepage from septic tanks and landfills, oil or chemical spills at construction and industrial sites, and from salt stockpiles and winter de-icing operations. Contamination may be especially severe where the water table is close to the surface (Dineen, 1975). Such a setting seems an unlikely location for a major municipal landfill.

The Albany Pine Bush Preserve

Approximately 2,000 acres of Pine Bush have been permanently protected through the cooperative efforts of The Nature Conservancy, the State of New York, and various local municipalities. Figure 4 shows the current boundaries of the Albany Pine Bush Preserve. In 1988, the New York State Legislature created the Albany Pine Bush Preserve Commission, which coordinates the management of protected lands within the Pine Bush (Albany Pine Bush Preserve Commission, 1995).

ALBANY CITY LANDFILL

The Albany City Landfill (the landfill) covers approximately 80 acres of the Pine Bush. The landfill is located on the north side of the New York State Thruway between Rapp Road, the Conrail railroad tracks, and the Pine Bush Preserve. The landfill has been capped and permanently closed since 1994.

The landfill is of a relatively simple design: it is unlined, it has no leachate collection system (Malcolm Pirnie, 1988), and it is believed to be a surface waste pile rather than an excavated pit refilled with waste (Engineering-Science, 1992). During its years of operation, the landfill was filled using conventional sanitary landfilling techniques of compaction and daily covering of waste. Methane and leachate were not extracted during operation. Methane extraction wells and automatic burners were installed in 1991 after methane migration toward the south (under the Thruway) was discovered.

The landfill began operations in August 1969. It was operated by a private company under contract to the City of Albany until 1974, when the Albany Department of Public Works assumed operational responsibility. At the time of the transfer of responsibility, weight scales were installed and the City began to keep records of the incoming waste. Until that time, the waste had been largely unrecorded and uninspected. It is suspected that during the 1969-74 period and potentially until the 1980s, numerous area industries had disposed of various amounts of industrial and hazardous wastes at the landfill (Malcolm Pirnie, 1988; Engineering-Science, 1992).

Beginning in 1983, a portion of the waste brought to the landfill was used as fuel for the ANSWERS Refuse-Derived Fuel (RDF) plant located on Sheridan Avenue in Albany. Fly ash from the incinerator at the ANSWERS RDF plant was brought to the landfill for disposal (Malcolm Pirnie, 1988). Between 1983 and 1987, approximately 203,265 tons of fly ash were disposed of at the landfill. In addition, approximately 343,143 tons of municipal and other waste were deposited at the landfill during the same period (Engineering-Science, 1992). The municipal and other wastes brought to the landfill for disposal included bypassed solid waste (waste not used as RDF), excess RDF, petroleum-contaminated soil,

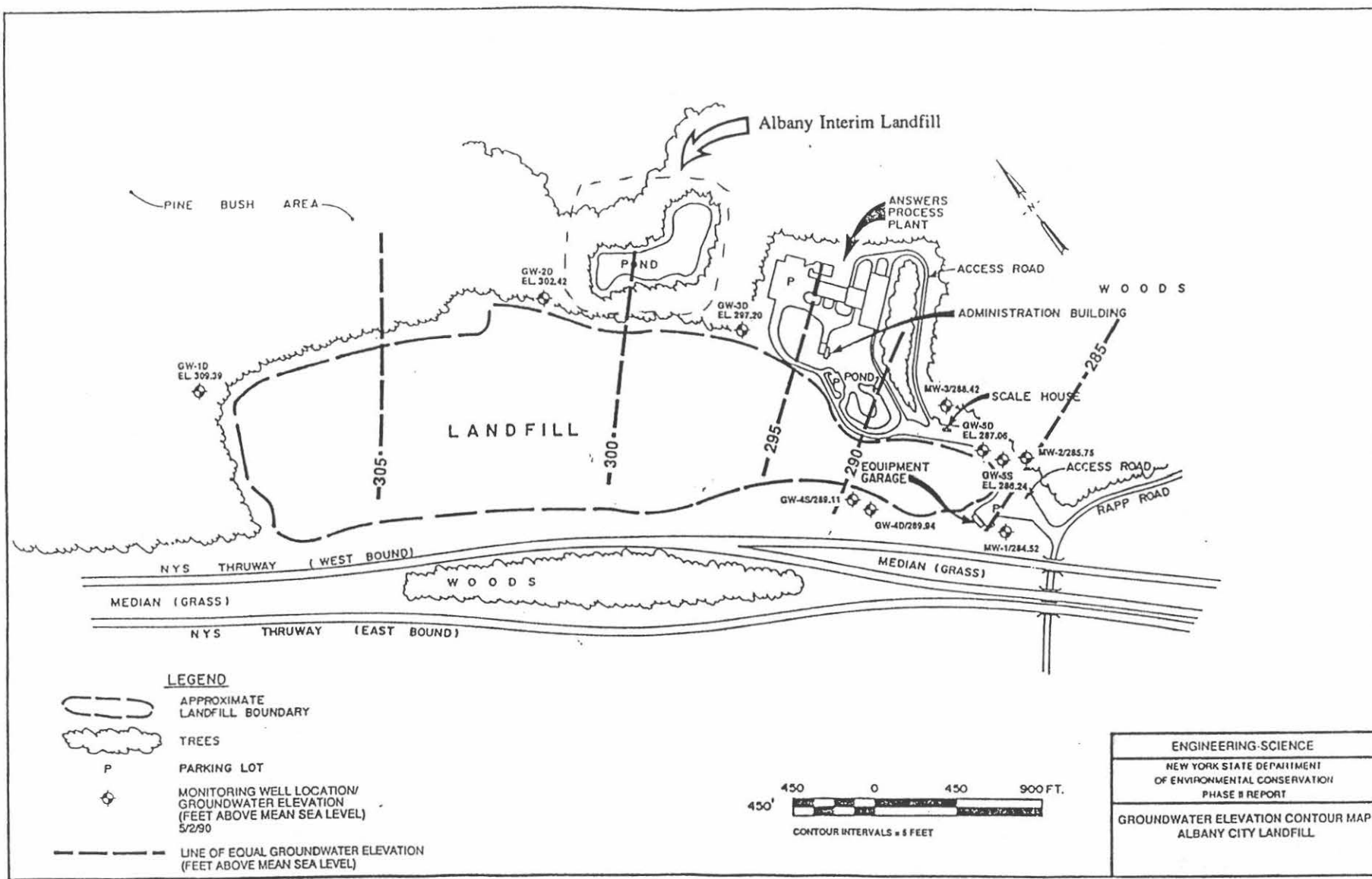


Figure 5. Albany City Landfill (circa 1990), monitoring well locations, groundwater elevations in wells (5/2/90), and groundwater elevation contour map (adapted from Engineering-Science, 1992). The approximate location of the Albany Interim Landfill is indicated by the dotted line.

picked material (unshreddable material removed during RDF processing), and ferrous material (Malcolm Pirnie, 1988), plus pharmaceutical wastes, iodine/soil wastes, and asbestos (Engineering-Science, 1992). The landfill began accepting petroleum-contaminated soil in 1987 (under the approval and recommendation of the New York State Department of Environmental Conservation, or NYSDEC), primarily for use as daily cover material; the landfill accepted approximately 14,000 tons of petroleum-contaminated soil in 1987 (Malcolm Pirnie, 1988).

Closure of the landfill began in 1982 under a phased approach. In 1985, the City of Albany signed a NYSDEC Order of Consent for the final closure of the landfill. In 1986, the City of Albany proposed an expansion project of the existing landfill to facilitate and expedite the closure process (Malcolm Pirnie, 1988). The expansion was approved by NYSDEC in 1989 (Engineering-Science, 1992), which resulted in the construction of the Albany Interim Landfill during the early 1990s.

At least three rounds of groundwater samples were collected at the landfill by Bender Hygienic Laboratories between 1981 and 1988. Several downgradient samples showed elevated values for a number of solid waste regulation baseline parameters, including iron, manganese, total phenols, and chloride (Engineering-Science, 1992).

The consulting firm of Malcolm Pirnie installed fourteen monitoring wells, including five deep-shallow pairs, in the location of the proposed expansion area in 1986. At that time, a fire-suppression pond occupied part of the area north of the landfill. Groundwater levels in the wells indicated groundwater flow from west to east and northeast across the site. The flow in the shallow groundwater appeared to be affected by surface topography such that a portion of the flow was toward the northeast (Malcolm Pirnie, 1988).

The Malcolm Pirnie wells were sampled in October 1986 and April 1987. Only the groundwater samples from two of the wells located closest to the landfill (MW-9D and MW-10S) were found to contain detectable concentrations of priority organic pollutants: 5 parts per billion (ppb) toluene in groundwater from MW-9D, 19 ppb benzene and 11 ppb ethylbenzene in groundwater from MW-10S. These three compounds are commonly found in petroleum products and solvents. Current New York state groundwater standards for these three compounds are 5 ppb, 0.7 ppb, and 5 ppb respectively (New York State Department of Environmental Conservation, Division of Water, 1993). At that time (1987), there were no NYS groundwater standards for toluene and ethylbenzene, although guidance values of 50 ppb had been established; the standard for benzene was then "non-detectable" (Malcolm Pirnie, 1988). Malcolm Pirnie concluded that the chemical analyses of samples taken from six of the monitoring wells did not show high concentrations of parameters often characteristic of landfill leachate plumes, but that higher than background levels of other parameters (e.g., iron, calcium, magnesium, specific conductance, and chlorides) in samples from MW-10S, coupled with relatively lower than background levels of the same parameters in downgradient well MW-4, suggested that those two wells may have delineated "the fringe of a weak plume from the existing landfill" (Malcolm Pirnie, 1988). In well designations, the suffix "D" indicates a deep well and the suffix "S" indicates as shallow well.

The April 1991 edition of the NYSDEC publication "Inactive Hazardous Waste Disposal Sites in New York State, Volume 4" listed the Albany City Landfill as inactive hazardous waste disposal site Number 401001. The landfill was given a classification code of 2a, which is a temporary classification assigned to sites that have inadequate and/or insufficient data for inclusion in any of the other classifications. The landfill was described as being located in the Pine Bush, as having allegedly received various amounts of industrial wastes from industries in Albany and Rensselaer Counties, and as having sandy soil and an existing high groundwater table. At that time, a groundwater monitoring program had been implemented. The City was implementing a phased closure under a consent order as a municipal landfill and closure was being negotiated. A Phase II (subsurface) investigation was in progress. The field work for the Phase II was complete and a Supplemental Search for documentation of disposal of hazardous waste was in progress. The Phase II draft report had been submitted for review (NYSDEC, 1991). The landfill was eventually (by the 1993 edition) removed from the NYSDEC listing of inactive hazardous waste disposal sites.

The Phase II investigation of the Albany City Landfill was completed by Engineering-Science for NYSDEC Division of Hazardous Waste Remediation. Field work for the investigation was performed

Well I.D.	Ground Surface Elevation (Feet*)	Top of PVC Well Pipe Elevation (Feet*)	Well Screen Interval Elevation (Feet*)	Water Level Data			
				Depth To Water Level (Feet**)	Water Level Elevation (Feet*)	Depth To Water Level (Feet**)	Water Level Elevation (Feet*)
GW-1D	348.5	351.04	307.5-297.5	41.6	309.44	41.65	309.39
GW-2D	320.8	323.40	295.8-275.8	20.85	302.55	20.98	302.42
GW-3D	301.3	303.33	266.3-246.3	5.95	297.38	6.13	297.20
GW-4D	299.5	301.79	238.5-218.5	11.7	290.09	11.85	289.94
GW-4S	299.6	301.36	277.6-257.6	10.0	291.36	12.25	289.11
GW-5D	292.0	294.11	231.0-211.0	7.0	287.11	7.05	287.06
GW-5S	292.1	294.09	278.1-258.1	7.7	286.39	7.85	286.24
MW-1	291.9	295.04	Unknown	--	--	10.52	284.52
MW-2	292.2	295.88	Unknown	--	--	10.13	285.75
MW-3	297.8	299.88	Unknown	--	--	11.46	288.42

Figure 6. Monitoring well information and water level data, Albany City Landfill (from Engineering-Science, 1992).

* Feet above Mean Sea Level

** Water level depth from top of PVC well pipe in feet

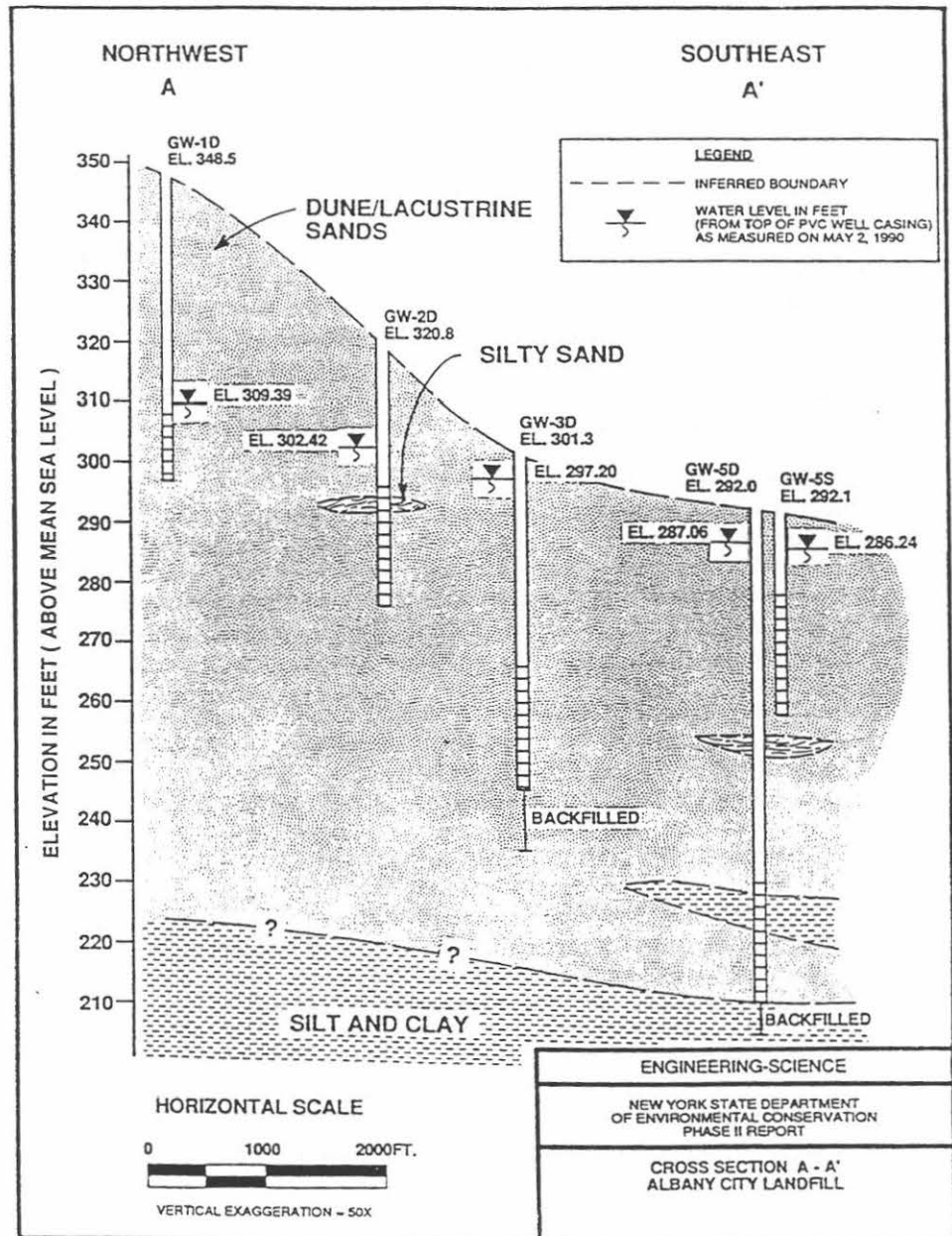


Figure 7. Cross-section through five monitoring wells at the Albany City Landfill showing subsurface materials, screened intervals, elevation of water table on 5/2/90, and decreasing elevation of water table from west to east (from Engineering-Science, 1992).

between September 1989 and May 1990. The investigation included completion of a geophysical survey using electromagnetic (EM) methods and installation of seven monitoring wells (Figure 5).

The geophysical survey was performed during September and October 1989. The geophysical survey indicated the potential presence of conductive contaminant plumes, and the monitoring wells were located so as to intercept the plumes. Groundwater, soil, and leachate samples were collected to determine whether hazardous substances or hazardous wastes were present at the landfill site (Engineering-Science, 1992).

The seven 2-inch diameter monitoring wells were installed into saturated sands of the Pine Bush aquifer to depths of 34 to 81 feet along the northern, eastern, and southeastern edges of the landfill in April 1990. The depths to water in the monitoring wells ranged from about 4 to 40 feet (Engineering-Science, 1992). Well depths, elevations, and other data are summarized in Figure 6. The data presented in Figure 6 and a cross-section showing the locations of well screens and the water table elevation in the wells (Figure 7) indicate that all of the wells were screened entirely below the top of the water table. The report indicates that the screened intervals were positioned to intercept contaminant plumes detected by the geophysical survey.

The upgradient well, GW-1D, was installed at the west end of the landfill. Wells GW-2D and GW-3D were installed on the north side of the landfill. Wells GW-5D, GW-5S, GW-4D, and GW-4S were installed at the east end of the landfill, the former two on the northeast side and the latter two on the south side. The suffix "D" indicates a deep well and the suffix "S" indicates as shallow well.

Based on the water levels measured in the seven wells, local groundwater flow was calculated to be mainly to the east and southeast (although elsewhere in the report, groundwater flow to the north is suggested). A vertically upward component of flow was inferred from the finding that the water levels measured in deep wells at the east end of the site (MW-4D, MW-5D) were higher than those in adjacent shallow wells (MW-4S, MW-5S).

Composite soil samples collected in the borings for the five deep ("D") wells were analyzed for organic compounds, metals, and cyanide. The analyses indicated low concentrations of toluene in two samples (GW-1D and GW-5D, 9 and 13 micrograms per kilogram, or $\mu\text{g}/\text{kg}$, respectively) and moderate concentrations of acetone in one sample (GW-2D, 170 $\mu\text{g}/\text{kg}$). Micrograms per kilogram are equivalent to parts per billion. Concentrations of metals did not exceed referenced naturally occurring ranges (Engineering-Science, 1992).

Groundwater samples were collected from the seven Engineering-Science wells and one well that had been installed previously (MW-1, Figure 5) in May 1990. The groundwater samples were analyzed for organic compounds, metals, and cyanide. Benzene was detected in the samples from GW-5S (5 micrograms per liter, or $\mu\text{g}/\text{l}$) and MW-1 (8 $\mu\text{g}/\text{l}$). Chlorobenzene was detected in the sample from GW-2D (140 $\mu\text{g}/\text{l}$). Micrograms per liter are approximately equivalent to parts per billion. Detectable concentrations of other organic compounds (e.g., acetone) were either below applicable standards or attributed to laboratory contamination, or both. Sixteen metals were detected, and downgradient concentrations consistently exceeded concentrations in the sample from the upgradient well (GW-1D) by more than three times, indicating an intermediate source (i.e., the landfill). Concentrations of seven metals (arsenic, chromium, iron, lead, magnesium, manganese, and sodium) exceeded applicable standards in one or more wells (the standard for iron was exceeded in all wells). The report of the Phase II investigation concluded that the landfill was adversely affecting groundwater quality in the Pine Bush Aquifer (Engineering-Science, 1992).

Three leachate samples that were collected north and south of the landfill showed some contamination by organic compounds and metals. Two organic compounds (chlorobenzene and 4-methylphenol) and five metals (iron, lead, manganese, sodium, and zinc) were detected at concentrations that exceeded NYS groundwater and surface water standards in one or more samples (Engineering-Science, 1992).

The Phase II Investigation report concluded that releases of organic compounds and metals to groundwater and leachate could be attributed to the landfill and that the data suggested that contaminants were migrating radially outward from the landfill, most notably to the north and east "in the direction of groundwater flow". The report stated that the "potential exists for off-site migration of contaminated

groundwater through the sands of the Pine Bush Aquifer” (Engineering-Science, 1992). The landfill was, however, subsequently removed from the NYSDEC list of inactive hazardous waste sites.

ALBANY INTERIM LANDFILL

The Albany Interim Landfill (AIL) covers an area of approximately 12 acres on the west side of Rapp Road in the City of Albany (Hansen, 1995). The AIL was constructed in the early 1990s with a double composite liner. The closed Greater Albany Landfill and the New York State Thruway lie to the south. A residential trailer park is located several hundred feet to the north. The Pine Bush Preserve borders the AIL to the west. Landfill service buildings and Rapp Road lie to the east. Figure 5 shows the approximate location of the AIL relative to the surrounding properties.

The AIL is operated by Landfill Technologies, Inc. (LTI) of West Sand Lake, New York. In 1991, a program of biostabilized solid waste placement was proposed at the AIL. The program began with a pilot project in 1989.

Biostabilized solid waste placement

Biostabilized solid waste placement is described in general terms by LTI literature (Hansen, 1995) as follows:

Solid waste to be landfilled is first shredded and moisture adjusted using mobile slow-speed shredding equipment. Such equipment is operated daily at the Albany, NY landfill. The prepared waste is placed in one hundred foot wide mats, six to eight feet high, and is immediately covered with Posi-Shell[®] synthetic cover. The length of the mat is continually extended as additional material is added to the mat's working face. Perforated ADS pipes provide aeration for a period of 60-90 days. Following this aerobic biostabilization period, the waste is compacted with standard landfill compaction equipment. A new stabilized layer is then placed above the compacted materials.

The process as performed by LTI is called In-Place StabilizationSM and has been granted U.S. patent 5265979; international patents are pending. Eight basic process steps are involved, as described in Hansen (1995) and summarized below:

1. Removal of recyclable material, specifically bulk metals (curbside recycling generally removes other recyclable materials).
2. Shredding by a mobile slow-speed shear mill with four-inch cutter spacing.
3. Moisture adjustment to 40-60% through addition of water by the dust suppression water spray system with which the shredder is fitted.
4. Placement of the shredded, moisture adjusted material in stabilization mats (100 feet wide and 8-10 feet high along the length of the landfill) by a track loader.
5. Covering with Posi-Shell[®] synthetic cover.

The Posi-Shell[®] synthetic cover consists of an aqueous slurry of 100% recycled material including a recycled pozzolon binder agent such as cement kiln dust and a minor quantity of recycled reinforcing fibers. Collected landfill leachate can be used as the aqueous liquid because the pozzolonic binder neutralizes odors and balances low pHs with its high pH. The Posi-Shell[®] cover is spray applied by mobile equipment capable of traversing all areas of a landfill. The Posi-Shell[®] material stiffens to a stucco-like consistency in a matter of hours and does not erode in subsequent rainstorms. The shell has sufficient pores to allow release of aeration gases, but is restrictive enough to maintain elevated temperatures immediately beneath its surface (necessary for extermination of fly and other insect larvae). The Posi-Shell[®] cover performs the functions of a soil cover during the

temporary biostabilization period (e.g., prevention of litter and fire, odor suppression, vector control).

6. Maintenance of the stabilization mats' interstitial oxygen and temperature conditions by addition of air.

Perforated pipes are placed within the mat at approximately 20-foot intervals. The pipes are manifolded to a high-volume, low-pressure blower that provides the air necessary to maintain aerobic conditions. Areas of higher temperature generally require more air to replace oxygen depleted by biological activity. Oxygen, temperature, and methane measurements are taken daily to determine appropriate aeration patterns.

7. Maintenance of the stabilization mats' interstitial moisture by addition of water.

The air blown into the stabilization mats is humidified by spraying small amounts of water in the blower discharge pipe; this water replaces evaporation losses.

8. Final compaction at the desired state of maturity.

The mat subsides to an average height of about 5 feet after 60-90 days of biostabilization. The mat is then compacted with standard landfill equipment to an average height of about 3 feet.

Petroleum-contaminated soil at the Albany Interim Landfill

The AIL accepts non-hazardous petroleum-contaminated soil for disposal. This soil is then used to construct roadways and driving surfaces within the landfill itself, because the stabilized waste and the Posi-Shell® cover are not suitable for vehicle traffic.

Methane and leachate at the Albany Interim Landfill

Methane is actively extracted from the AIL, in large part as an odor control measure. Three horizontal hoses run from the AIL to a burner on the west side of the AIL. Commercial use of the methane is being explored.

Leachate is actively collected and removed from the AIL. Leachate can then be either reinjected or routed to holding tanks and eventually to the sanitary sewer for disposal (provided it meets disposal criteria).

Conventional sanitary landfilling techniques

Solid waste disposal methods in the U.S. have evolved considerably during the twentieth century. The open or burning dump is generally a thing of the past, having been replaced by the sanitary landfill, so called primarily because of the "sanitary" manner in which waste is covered daily. The term "sanitary landfill" was apparently coined during the 1930s at a pioneering landfilling operation in Fresno, California, where the "cut and cover" or "trench" method of landfilling was first used in the U.S. (American Public Works Association, 1970).

Conventional sanitary landfilling procedures, as described in American Public Works Association (1970) and Sorg and Hickman (1970), consist of the following steps:

1. waste is dumped directly onto the landfill from collection trucks
2. waste is spread and compacted by a tractor
3. waste is covered with a 6-inch layer of compacted soil ("daily cover") before the end of the day
4. when the landfill space is full, the landfill is capped with a 2-to-3-foot layer of compacted soil

Conventional sanitary landfilling procedures differ from biostabilization procedures in several ways. Specifically, conventional sanitary landfilling involves no shredding or initial moistening of waste, no direct mechanical compaction after the daily cover has been applied, no aeration, no active methane or

leachate extraction, and no moisture adjustment. Before the advent of landfill liners, moistness was considered conducive to groundwater pollution (American Public Works Association, 1970).

Evaluation of biostabilization techniques

The fundamental criterion upon which the stabilized waste placement process at the AIL has been evaluated is the amount of waste that can be placed, which is evaluated in terms of effective density and airspace conservation. In general, LTI has found that 40% to 60% more waste can be placed in a given volume under stabilized placement techniques than under conventional landfilling techniques. Effective densities were found to be higher with the use of stabilized placement (Hansen, 1995).

Additional criteria have been used by LTI to evaluate the success and practicality of stabilized waste placement, including litter generation, odor emission, vector control, pathogen potential, and orderliness of operation. LTI found that the Posi-Shell[®] synthetic cover was highly effective in the control of litter after waste placement. Odor emissions were generally found to be a problem only if the mats were less than approximately 1000 feet from the receptors, particularly during warm weather. The Posi-Shell[®] was found to be a highly effective vector (e.g., rats, seagulls, flies) deterrent. Fly larvae were generally killed by the high temperatures in the stabilization mats. Rats and birds were not observed; only a few individual mice were observed. The average temperatures of 130-150°F recorded within the aerated mat were judged to be highly destructive to most human pathogens. The stabilized placement method received high ratings for orderliness from the project foreman or engineer and the landfill superintendent (Hansen, 1995). The additional costs for stabilized placement were estimated by Hansen (1995) at about \$12 to \$24 per ton.

STEWART'S SHOP #182

The volatile organic compounds benzene, ethylbenzene, and toluene, which were detected at relatively low concentrations in groundwater at the Albany City Landfill in 1990, are commonly found as groundwater contaminants at sites that have experienced subsurface gasoline spills. The volatile natures of these and many other compounds typically found in gasoline offer a means of remediating, or cleaning up, the contaminated groundwater by encouraging volatilization, or evaporation, of the compounds. One method of encouraging volatilization of compounds is to pump contaminated groundwater through an air stripper. This remediation method is being used at Stewart's Shop #182, a small site on South Brandywine Avenue in Schenectady, New York (Figure 8).

The following information is summarized from Passaretti Geological & Environmental Consultants, Inc. (1991).

Groundwater Study

Stewart's Shop #182, a convenience store and gas station, is located at 100 South Brandywine Avenue in Schenectady, New York. In 1991, three underground gasoline storage tanks and one underground waste oil tank were removed from the site as part of a tank upgrading program. Petroleum-contaminated soil was encountered during the tank removal. Approximately 250 cubic yards of soil were removed from the former tank pit for offsite disposal. New underground storage tanks were installed in a different part of the site. The site had reportedly been a Stop & Go Station prior to being a Stewart's Shop. The contiguous site to the northeast had also reportedly been a gas station. The NYSDEC representative requested completion of a groundwater study at the site.

Passaretti Geological & Environmental Consultants, Inc. (Passaretti) of Saratoga, New York were contracted to perform the groundwater study at the site. On October 30, 1991, three monitoring wells were installed at the site: one 4-inch diameter well (MW-1) and two 2-inch diameter wells (MW-2 and MW-3). Figure 8a is a copy of a site plan showing the layout of the site and the location of the first three wells and of seven wells installed at and around the site later. Groundwater was encountered during drilling at a depth of approximately 19 feet. The wells range in depth from 25 to 35 feet. Petroleum odors were detected in soils and groundwater from MW-1 and MW-3. No odors were detected in groundwater from MW-2. Data collected during a November 1991 site survey and corrected for the presence of free product discovered in MW-3 at that time indicated that groundwater flow was toward the southwest to west at a gradient of about 1%.

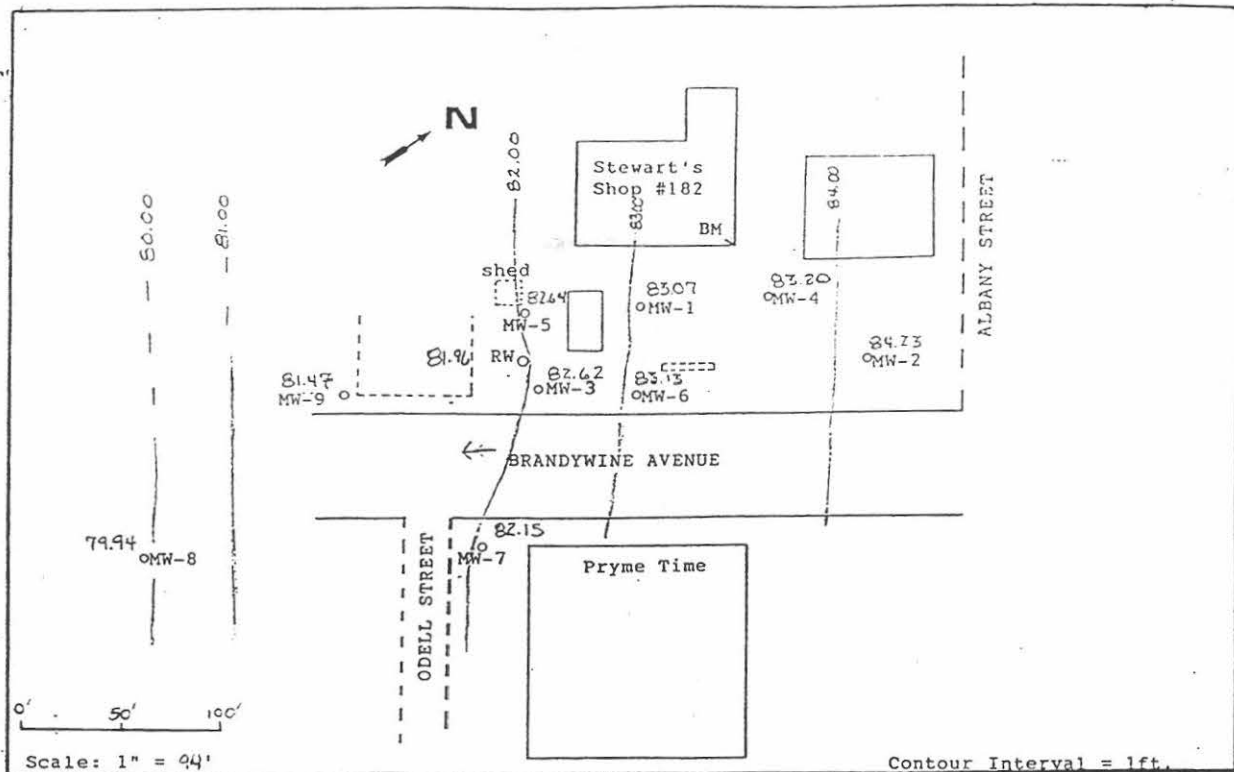


Figure 8a. Monitoring and recovery well locations, static groundwater elevations (5/25/93), and static groundwater contour map at Stewart's Shop #182, 100 South Brandywine Avenue, Schenectady, New York (adapted from Passaretti, 1994).

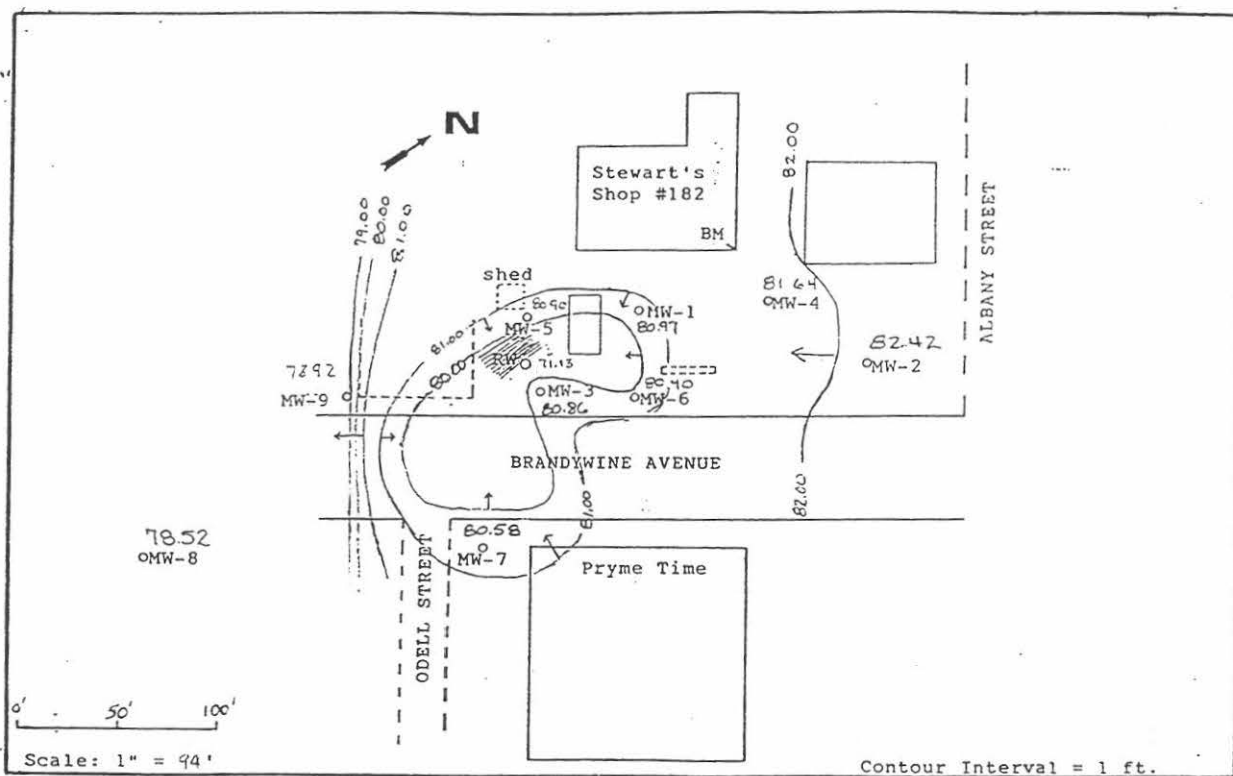


Figure 8b. Monitoring and recovery well locations, pumping groundwater elevations (11/19/93), and pumping groundwater contour map at Stewart's Shop #182, 100 South Brandywine Avenue, Schenectady, New York (adapted from Passaretti, 1994).

Groundwater samples were collected from the three wells on November 3, 1991. Analysis for volatile organic compounds indicated contamination by gasoline compounds of groundwater from MW-1 and MW-3. Total BTEX (benzene, toluene, ethylbenzene, and xylenes) concentrations in the samples from MW-1 and MW-3 were 903 and 27,581 ppb, respectively (excluding MTBE, a gasoline additive). [These are typically considered to be moderately high (MW-1) to high (MW-3) BTEX concentrations.] A trace amount (4.4 ppb) of the solvent tetrachloroethene was detected in the sample from MW-2.

Grab and split-spoon sampling during drilling indicated that the site is underlain by well-sorted, interbedded medium- and coarse-grained sands to a depth of about 20 feet (approximately to the water table), and below this by fine-grained sand with variable silt and clay content. These deposits are part of the Schenectady delta sediments.

Passaretti recommended installation of additional wells to define the plume of contamination and the extent of free product, bailing of free product from MW-3 and storage of the bailed product in a 55-gallon drum, remediation of groundwater with a pump and treat system, and remediation of remaining areas of contaminated soil with a soil vent system (following recovery of the free product). Passaretti recommended that the additional monitoring wells be installed with slotted screen from about 5 feet below the surface to below the water table so that they could serve both as groundwater monitoring points and as vertical vents. [The soil vent system was not installed.]

Site Remediation

The following information is summarized from Passaretti Geological & Environmental Consultants, Inc. (1994).

Between November 1991 and April 1992, seven additional wells were installed at and around the site to define the contaminant plume. In December 1991, three wells were installed onsite (MW-4, MW-5, and MW-6). In February 1992, three wells were installed offsite and downgradient of the site (MW-7, MW-8, and MW-9). In April 1992, a 40-foot, 4-inch diameter recovery well was installed onsite (RW). Figure 8a shows the locations of the original three wells and the seven additional wells, the static groundwater levels, and the calculated elevation of the static water table.

A remediation system was installed in June 1993, using the recovery well (RW) that was installed in April 1992. The soil vent system that was recommended by Passaretti in 1991 was not installed. The remediation system was described as follows in the 1993 Engineer Report prepared by Passaretti (1993):

A one-half horsepower, submersible pump pumps water from the recovery well into a 275 gallon separator tank. The tank contains a solitary baffle and a submersible pump. Volatilization of contaminants occurs across the separator. By decreasing the contaminant level in the water prior to its entering the shallow tray system, the need for carbon polishing is negated, as is the associated cost.

Currently the system treats 31 gallons of water per hour. Recent adjustments of the downhole electrical probes are intended to enhance this recovery rate. Draw down around the recovery well is excellent as indicated by the most recent groundwater contour data.

Figure 8b shows the locations of the ten wells, the groundwater levels during pumping, and the calculated elevation of the pumping water table. Note the drawdown of the top of the water table around the recovery well.

Between June and December 1993, approximately 98,000 gallons of groundwater were treated by the system. Analytical results for groundwater samples collected in the monitoring wells through November 1993 generally showed decreases in the total concentrations of volatile organic compounds detected in the samples (Passaretti, 1994).

Tabulations of readings of the water meter on the remediation system and calculated pumping rates through April 1995 show that as of April 12, 1995, 269,039 gallons of groundwater had been pumped into

the system. The average pumping rate was 14.47 gallons for the one-week period April 5-12 (Passaretti, 1995a).

A tabulation of groundwater analytical results for samples collected from the onsite wells, the offsite wells, and the remediation system through April 1995 shows mixed results. Improvement in groundwater quality can be seen in some, but not in all, wells (Passaretti, 1995b).

Our thanks to Mary Passaretti of Passaretti Geological & Environmental Consultants, Inc., for providing us with information about this site.

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**ENVIRONMENTAL TECHNOLOGY AND PRESERVATION:
THE PINE BUSH, LANDFILLS, AND GROUNDWATER INTEGRITY**

FIELD TRIP GUIDE

Meet in the Geology Department on the second floor of Butterfield Hall at Union College in Schenectady. We will begin with a slide show by David Hansen, P.E., of Landfill Technology, Inc. The slide show will provide an introduction to the design and operation of the Albany Interim Landfill.

After the slide show, we will leave Union College and take I-890 eastbound to the NYS Thruway (I-90) eastbound. We will leave the Thruway at Exit 24 (Albany). The road log starts at the Exit 24 toll booth.

Mileage

- 0.0 After exiting from the eastbound lanes of the Thruway, set the odometer to 0.0 in the toll booth at Exit 24 (Albany) of the NYS Thruway. Bear right out of the toll booth toward Exit 1S.
- 0.1 Take Exit 1S (Western Avenue/Route 20). You are now on the southernmost section of the Northway.
- 0.7 Exit right up the ramp that leads to Crossgates Mall Road.
- 0.9 Bear right through the light and yield sign onto Crossgates Mall Road.
- 1.05 Bear right at the sign for Washington Avenue Extension.
- 1.25 Turn left at the sign for Washington Avenue Extension Westbound. Continue on Washington Avenue Extension westbound to the intersection with Rapp Road.
- 2.1 Turn right onto Rapp Road at the light.
- 2.3 The access road for the Albany City Landfill, the Albany Interim Landfill, and the public solid waste tipping station is on the left. We will meet our guide at a prearranged location inside the landfill property.

STOP 1: Albany Interim Landfill

Access to the landfill is restricted. Call Landfill Technologies, Inc. (518-674-8694) in West Sand Lake, New York, to arrange a tour of the active and closed landfill areas.

Because the distance that we will drive inside the landfill property is unknown, we will reset the odometer to 0.0 at the intersection of the landfill access road and Rapp Road.

- 0.0 Set the odometer to 0.0 before turning onto Rapp Road from the landfill access road. Turn right onto Rapp Road.
- 0.2 Turn right onto Washington Avenue Extension westbound at the light.
- 1.4 Turn right onto Route 155 (New Karner Road) at the light. Turn right onto Route 155 (the sign says "east"). Almost immediately, you are crossing over the Thruway. Look at the right side of the road. A guardrail begins where the overpass ends. At the end of the guardrail is the beginning of an unpaved track that runs along the east (right) side of Route 155.
- 1.75 Pull off the right side of Route 155 onto the gravel shoulder and continue onto the unpaved track. Continue along the dirt track to the informal parking area marked by the "Albany Pine Bush Preserve" sign.

Mileage

- 1.85 Park near the entrance to the Albany Pine Bush Preserve marked by the sign and the wooden post-and-rail fence.

STOP 2: Albany Pine Bush Preserve

The Albany Pine Bush Preserve is open to the public. Sign in at the box near the entrance. We will walk generally east, pausing atop and between dunes to note the morphology and vegetation, eventually ending up at the west end of the closed Albany City Landfill. On the way, we will pass a monitoring well that is most likely GW-1D, the upgradient well installed by Engineering-Science in 1990. After looking at the landfills from this different perspective, we will make our way back through the Pine Bush to the vehicles.

Return to Route 155, either by continuing along the unpaved track to the entrance road for the State Employees Federal Credit Union or by backtracking. Head back to the Exit 24 toll booths.

- 1.9 Turn left onto Route 155.
- 2.35 Turn left onto Washington Avenue Extension eastbound at light.
- 4.2 Turn right at the sign for Crossgates Mall Road.
- 4.35 Turn left at stop sign onto Crossgates Mall Road.
- 4.7 Turn left at light onto the access ramp for I-87 (north). The signs indicate that this road leads to I-87 and I-90.
- 5.5 Take Exit 1W (to New York and Buffalo). Merge into westbound traffic on I-90 and continue west to the toll booths.

Reset the odometer to 0.0 in the toll booth at Exit 24 (Albany) of the NYS Thruway.

- 0.0 At the toll booth, get ticket, then leave toll booth and bear left onto I-90 westbound (to Buffalo).
- 5.4 Take Exit 25 to I-890 (Schenectady). Pay toll (\$0.20). Go west on I-890.
- 9.1 Take Exit 6 (Michigan Avenue). Turn right onto Brandywine Avenue at the stop sign at the end of the offramp.
- 9.4 Stewart's Shop #182 is on the left (west) side of Brandywine Avenue (#100) between the first and second traffic lights. Signs advertising Mobil gasoline are visible as you approach.

STOP 3: Brandywine Stewart's Shop

The site is a convenience store and gas station. The remediation shed is closed to the public. Please do not take up customer parking spaces or block customer access to the store or gas pumps.

End of trip.

GEOLOGY AND MINING HISTORY OF BARTON MINES CORPORATION, GORE MOUNTIAN MINE

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INTRODUCTION

This trip proceeds from Union College to the former Barton Garnet Corporation mine near the summit of Gore Mountain, New York. For 105 years, this was the site of the world's oldest continuously operating garnet mine and the country's second oldest continuously operating mine under one management.

The Barton Mines Corporation open pit mine is located at an elevation of about 800 m (2600 ft) on the north side of Gore Mountain (Fig.1). The community at the mine site is the highest self-sufficient community in New York State. It is 16 km (10 mi) from North Creek and 8 km (5 mi) from NY State Route 28 over a Company-built road that rises 91 m (300 ft) per mile. This road, like others in the vicinity, is surfaced with coarse mine tailings. About eleven families can live on the property. The community has its own water, power, and fire protection systems. On the property are the original mine buildings and the Highwinds Inn, built by Mr. C.R. Barton in 1933 as a family residence. The Inn is now privately leased from the Corporation and operates 10 months per year. It offers a four-bedroom lodge, a four star dining room, cross-country skiing and fantastic views of the Siamese Wilderness Area.

The garnet is used in coated abrasives, glass grinding, metal and glass polishing, and even to remove the red hulls from peanuts. Paint manufacturers add garnet to create non-skid surfaces and television makers use it to prepare the glass on the interior of color picture tubes prior to the application of the phosphors. Barton sells between 10,000 and 12,000 tons of technical-grade garnet abrasive annually. About 40% of the company's shipments are to foreign countries. All current U.S. production of technical-grade garnet is limited to the Barton Mines Corporation. The product is shipped world wide for use in coated abrasives and powder applications (Austin, 1993a,b).

Garnet has been designated as the official New York State gemstone. Barton produces no gem material but collectors are still able to find rough material of gem quality. Stones cut from Gore Mountain rough material generally fall into a one to five carat range. A small number of stones displaying asterism have been found. Garnets from this locality are a dark red color with a slight brownish tint. Special cutting schemes have been devised for this material in order to allow sufficient light into the stone.

HISTORY

The early history of the Barton garnet mine has been compiled by Moran (1956) and is paraphrased below. Mr. Henry Hudson Barton came to Boston from England in 1846 and worked as an apprentice to a Boston jeweler. While working there in the 1850's, Barton learned of a large supply of garnet located in the Adirondack Mountains. Subsequently, he moved to Philadelphia and married the daughter of a sandpaper manufacturer. Combining his knowledge of gem minerals and abrasives, he concluded that garnet would produce better quality sandpaper than that currently available. He was able to locate the source of the Adirondack garnet stones displayed at the Boston jewelry store years before. Barton procured samples of this garnet which he pulverized and graded. He then produced his first garnet-coated abrasive by hand. The sandpaper was tested in several woodworking shops near Philadelphia. It proved to be a superior product and Barton soon sold all he could produce.

In Garver, J.I., and Smith, J.A. (editors), Field Trips for the 67th annual meeting of the New York State Geological Association, Union College, Schenectady NY, 1995, p. 405-412.

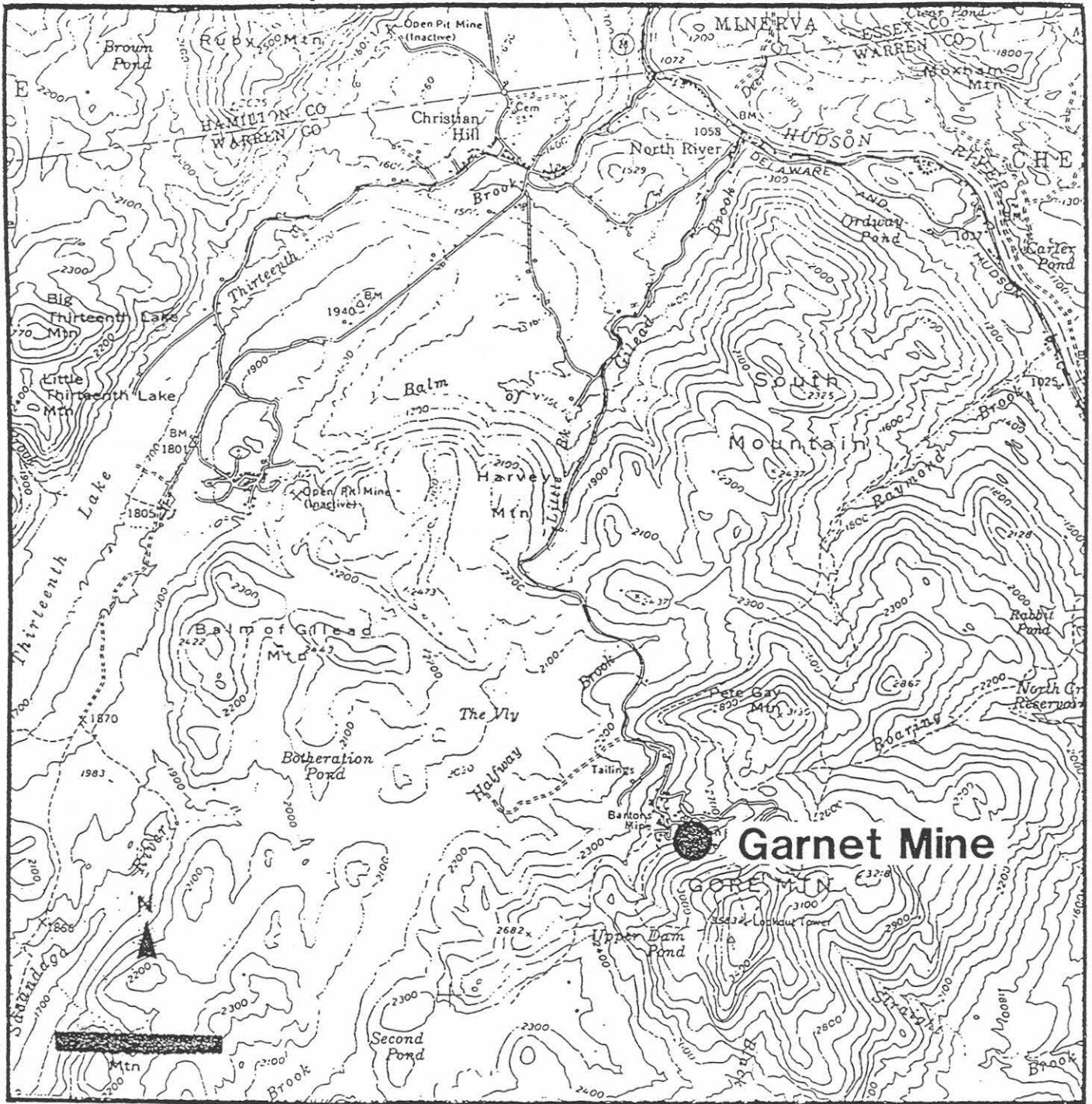


Figure 1. Topographic map (Thirteenth Lake 15' quadrangle) showing the location of the Barton garnet mine. Scale bar = 1.6 km (1 mi).

H.H. Barton began mining at Gore Mountain in 1878 and in 1887, bought the entire mountain from the State of New York. Early mining operations were entirely manual. The garnet was hand cobbled (*i.e.* separated from the waste rock) by small picking hammers and chisels. Due to the obstacles in moving the ore, the garnet was mined during the summer and stored on the mountain until winter. It was then taken by sleds down to the railroad siding at North Creek whence it was shipped to the Barton Sandpaper plant in Philadelphia for processing. The "modern" plant at Gore Mountain was constructed in 1924. Crushing, milling, and coarse grading was done at the mine site. In 1983, the Gore Mountain operation was closed down and mining was relocated to the Ruby Mountain site, approximately 6 km (4 mi) northeast, where it continues at present.

MINING AND MILLING

The mine at Gore Mountain is approximately one mile in length in an ENE-WSW direction. The ore body varies from 15 m (50 ft) to 122 m (400 ft) and is roughly vertical. Mining was conducted in benches of 9 m (30 ft) using standard drilling and blasting techniques. Oversized material was reduced with a two and one-half ton drop ball. The ore was processed through jaw and gyratory crushers to liberate the garnet and then concentrated in the mill on Gore Mountain. Garnet concentrate was further processed in a separate mill in North River at the base of the mountain. Separation of garnet was and is accomplished by a combination of concentrating methods including heavy media, magnetic, flotation, screening, tabling and air and water separation. Processes are interconnected and continuous or semi-continuous until a concentrate of 98% minimum garnet for all grades is achieved (Hight, 1983). Finished product ranges from 0.6 cm to 0.25 micron in size.

CHARACTERISTICS OF GORE MOUNTAIN GARNET

The garnet mined at Gore Mountain is a very high-quality abrasive. The garnets display a well-developed tectonic parting that, in hand specimen, looks like a very good cleavage. This parting is present at the micron scale. Consequently, the garnets fracture with chisel-like edges yielding superior cutting qualities. The garnet crystals are commonly 30 cm in diameter and rarely up to 1 m with an average diameter of 9 cm (Hight, 1983). The composition of the garnet is roughly 43% pyrope, 40% almandine, 14% grossular, 2% andradite, and 1% spessartine (Levin, 1950; Harben and Bates, 1990). Chemical zoning, where present, is very weak and variable (Luther, 1976). Typical chemical analyses of the garnet are presented in Table 1. Hardness of the garnet is between eight and nine and the average density is 3.95 gm/cm³.

Table 1. Electron Microprobe analyses of Gore Mt. garnet (almandine-pyrope) normalized to 8 cations and 12 anions.

Oxide Weight Percent	#29	#41
SiO ₂	39.43	39.58
Al ₂ O ₃	21.40	21.20
TiO ₂	0.05	0.10
FeO*	22.80	24.45
Fe ₂ O ₃ *	1.44	0.72
MgO	10.65	9.60
MnO	0.48	0.74
CaO	3.85	3.97
Na ₂ O	0.00	0.00
K ₂ O	0.00	0.00
Total	100.09	100.36

* Calculated by charge balance (Kelly and Petersen, 1993).

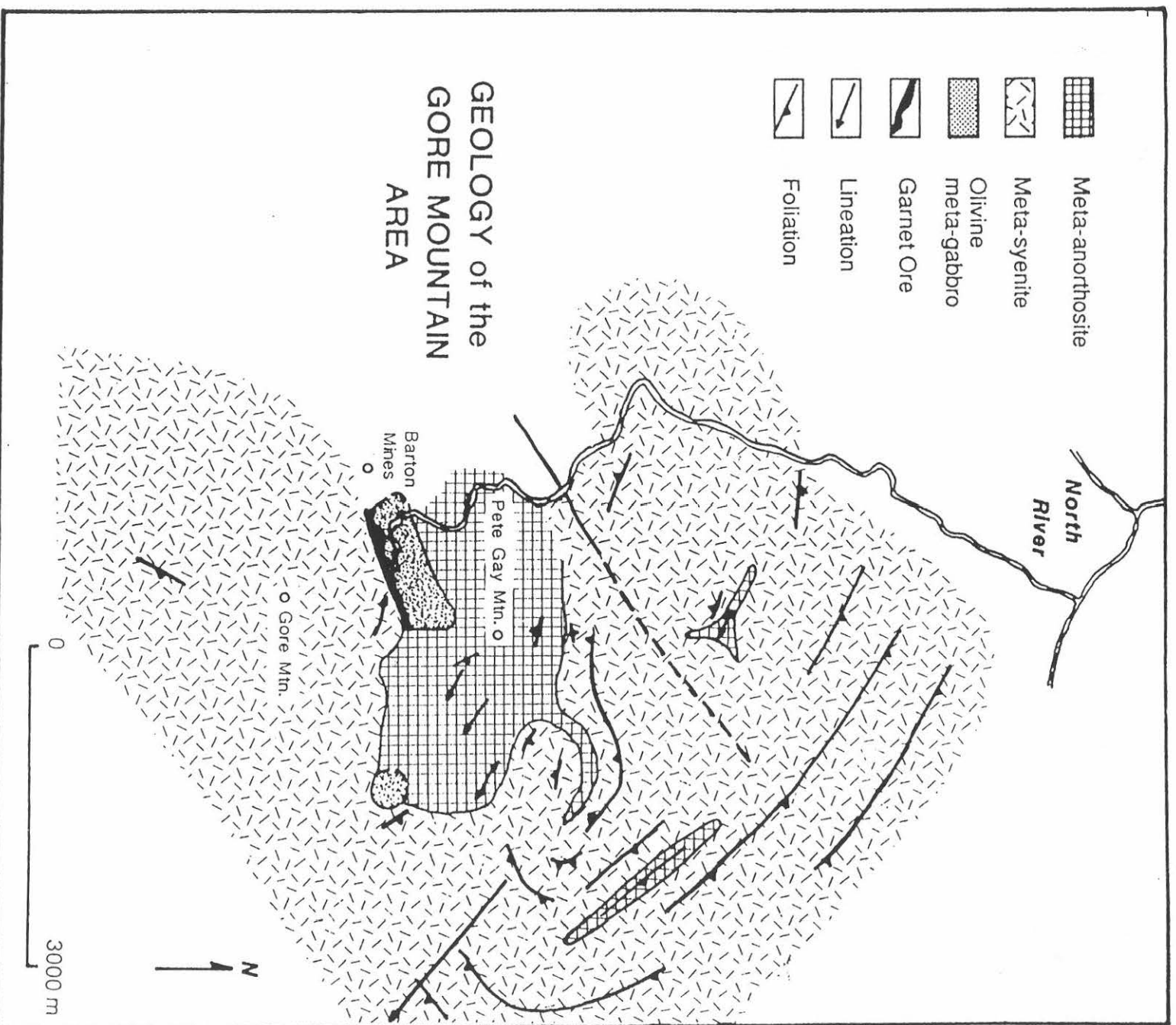


Figure 2. Geologic map of the vicinity of the Barton garnet mine (modified from Bartholomé, 1956).

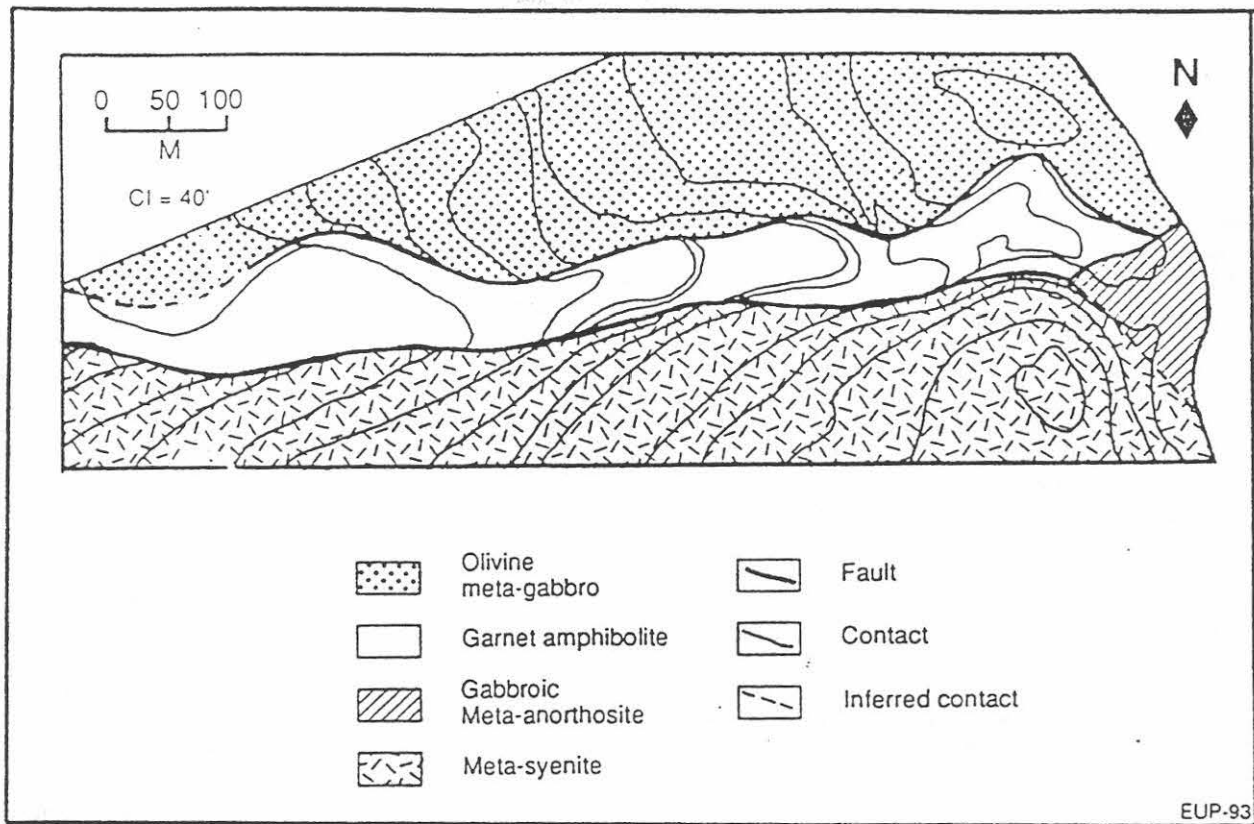


Figure 3. Geologic map of the Gore Mountain mine (modified from Goldblum and Hill, 1992).

GEOLOGY

The garnet mine is entirely hosted by a hornblende-rich garnet amphibolite unit along the southern margin of an olivine meta-gabbro body (Figs. 2,3). The garnet amphibolite grades into garnet-bearing gabbroic meta-anorthosite to the east. To the south the garnet amphibolite is in contact with a meta-syenite; a fault occurs parallel to this contact in places.

The olivine meta-gabbro bordering the ore zone is a granulite facies lithology with a relict subophitic texture. Preserved igneous features, faint igneous layering, and a xenolith of anorthosite have been reported in the meta-gabbro (Luther, 1976). Prior to metamorphism, the rock was composed of plagioclase, olivine, clinopyroxene and ilmenite. During metamorphism, coronas of orthopyroxene, clinopyroxene and garnet formed between the olivine and the plagioclase and coronas of biotite, hornblende and ilmenite formed between plagioclase and ilmenite (Whitney & McLelland, 1973, 1983). The contact between the olivine meta-gabbro and the garnet amphibolite ore zone is gradational through a narrow (1 to 3 m wide) transition zone. Garnet size increases dramatically across the transition zone from less than 1 mm in the olivine meta-gabbro, to 3 mm in the transition zone, to 50 to 350 mm in the amphibolite (Goldblum and Hill, 1992). This increase in garnet size coincides with a ten-fold increase in the size of hornblende and biotite, the disappearance of olivine, a decrease in modal clinopyroxene as it is replaced by hornblende, and a change from green spinel-included plagioclase to white inclusion-free plagioclase (Goldblum and Hill, 1992). Mineralogy in the garnet amphibolite ore zone is mainly hornblende, plagioclase and garnet with minor biotite, orthopyroxene, and various trace minerals. In both the olivine meta-gabbro and the garnet amphibolite, garnet content averages 13 modal percent, with a range of 5 to 20 modal percent (Luther, 1976; Hight, 1983; Goldblum, 1988). The garnet amphibolite unit is thought to be derived by retrograde metamorphism of the southern margin of the granulite facies olivine meta-gabbro. At the west end of the mine, a garnet hornblendite with little or no feldspar is locally present. This rock may represent original ultramafic layers in the gabbro (Whitney et al., 1989). In the more mafic portions of the ore body, the large garnet crystals are rimmed by hornblende up to several inches thick. Elsewhere, in less mafic ore, the rims contain plagioclase and orthopyroxene. Chemical analyses of the olivine meta-gabbro and garnet amphibolite show that the garnet ore was derived by retrograde isochemical metamorphism, except for an increase in the H_2O and fO_2 , of the olivine meta-gabbro (Table 2; Luther, 1976).

A strong, consistent lineation and weak planar fabric coincide with the zone of large garnet crystals and are an important feature of the garnet ore zone (Goldblum and Hill, 1992). The lineation is defined by parallel alignment of prismatic hornblende crystals, elongate segregations of felsic and mafic minerals, plagioclase pressure shadows, and rare elongate garnet. The foliation is defined by a slight flattening of the felsic and mafic aggregates.

ORIGIN OF GARNET

Although the garnet crystals in the ore zone at Gore Mountain are atypical in size, the modal amount of garnet is not unusually high for Adirondack garnet amphibolites. Garnet amphibolite that is texturally and mineralogically similar occurs elsewhere in the Adirondacks, usually on the margins of gabbroic rock bodies. The ore at the currently operating Barton Corporation mine at Ruby Mountain, for example, is of the same tenor but the garnets rarely are larger than 2.5 to 5 cm.

Petrologic studies (Buddington, 1939, 1952; Bartholome, 1956, 1960; Luther, 1976; Sharga, 1986; Goldblum, 1988; Goldblum and Hill, 1992) have agreed that the growth of the large garnets is related to a localized influx of water that caused the retrograde metamorphism along the margin of the granulite facies olivine meta-gabbro body. The Gore Mountain garnets are chemically homogeneous suggesting that (a) the garnets grew under conditions in which all chemical components were continuously available and the (b) temperature and pressure conditions were uniform during the period of garnet formation. A zone of high fH_2O along the southern margin of the original gabbro body may have enhanced diffusion and favored growth of very large garnets and thick hornblende rims at the expense of plagioclase and pyroxene. Luther (1976) speculates that physical and chemical conditions were favorable for the growth of garnet but poor for the nucleation of garnet so that the garnet crystals that did nucleate grew to large size. The presence of volatiles, particularly H_2O , promotes the growth of large crystals by aiding transport of components.

Recognition that the garnet ore body, retrograde metamorphism, and L-S deformation fabric all coincide with the southern margin of the olivine meta-gabbro body led Goldblum and Hill (1992) to hypothesize that the high fluid flow required for growth of large garnet crystals was the result of ductility contrast at a lithologic contact during high-temperature shear zone deformation. The olivine meta-gabbro is a granulite facies rock with a poorly developed foliation and little evidence for ductile deformation. In the transition zone between the olivine meta-gabbro and the garnet amphibolite, increased ductile deformation resulted in grain-size reduction of plagioclase and pyroxene. Microstructures in plagioclase in the transition zone indicate plastic deformation, and the concurrent modal increase in hornblende indicates an influx of fluid. Fabric development and hydration are most apparent in the garnet amphibolite of the ore zone. According to Goldblum and Hill (1992), the olivine meta-gabbro remained competent and initially deformed by brittle processes along its southern margin while the adjacent feldspar-rich meta-syenite and gabbroic meta-anorthosite deformed plastically during deformation at amphibolite facies conditions. Initial grain-size reduction by cataclasis along the margin of the meta-gabbro allowed hydration and retrograde metamorphism to produce the garnet amphibolite. As the hydrated ore body replaced the olivine meta-gabbro, ductile deformation mechanisms replaced cataclasis. During retrograde metamorphism, the garnet amphibolite was likely a high-strain zone of reaction-enhanced ductility. Eventually, metamorphic reactions apparently outpaced the rate of deformation and grain coarsening impeded ductile deformation processes (Goldblum and Hill, 1992).

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ROAD LOG

The Gore Mountain mine of the Barton Mines Corporation is generally open seasonally for visitors. However, it is suggested that visitors call in advance to be sure the mine is open. Arrangements for group tours at any time should be made through Barton Mines Corp., North Creek, NY 12853, (518) 251-2296. DO NOT attempt to visit this mine without permission.

This trip log begins at the intersection of US Route 90 (NYS Thruway) and US Route 87 (Northway) at Thruway Exit 24 (Northway Exit 2).

	Milage	Cumulative milage
Rt. 90/Rt. 87 Interchange	0	0
Travel north on Rt. 87 to Warrensburg, Exit 23	60.2	60.2
Go to end of exit ramp, turn left	0.3	60.5
Go to traffic light at NYS Rt. 9 north, turn right	0.1	60.6
Travel through Warrensburg on Rt. 9		
Traffic light, go straight	0.2	60.8
Traffic light at Rt. 418, go straight	0.6	61.4
Traffic light at fork, fork to right on Rt. 9	0.5	61.9
Travel Rt. 9 north to intersection with NYS Rt. 28,		
Turn left on Rt. 28 west	2.9	64.8
Blinker at Rt. 8 intersection, go straight	11	75.8
Intersection with Rt. 28N, go straight	6.6	82.4
Turn left on Barton Mine Road	4.6	87
Travel to end of Barton Mine Road	5	92

Note: the intersection of Barton Mine Road and Rt. 28 is marked by a small cluster of buildings. Among these are a Mom 'n' Pop general store with gas pumps and Jasco's mineral shop. On the east side of Rt. 28 facing south there is a sign opposite Barton Mine Road indicating the Barton Mine (Gore Mt.) mineral shop.

GROUNDWATER AQUIFERS OF THE UPPER MOHAWK VALLEY

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INTRODUCTION

Much of the groundwater used by the city of Schenectady and the several smaller communities in the near vicinity comes originally from the Mohawk River via gravel aquifers lying just below or at the present surface. In addition to the city itself, the towns of Niskayuna, Rotterdam Junction and the Scotia-Glenville area also use this major primary source (Figure 1). Of course, other smaller sources exist, including the bedrock sandstones and shales (the Middle Ordovician Schenectady Formation) and surficial but small perched sand and gravel deposits adequate for household supplies with volumes that range from 1.5 to perhaps as much as 10 gallons per minute for a reasonable period of time. The Mohawk River, however, provides millions of gallons of clean fresh water each day from gravel aquifers through which the less than desirable river water flows and is, as a result, cleaned to a remarkable degree.

The basic study of the area was done by Simpson (1949) and by Winslow *et al.* (1965) as well as several other authors. The present study, primarily in the Rotterdam Junction Aquifer (Figure 1), builds upon their work and corrects some basic misconceptions regarding gravel aquifers of this type. It was supported by the Schenectady Chemical Company, now Schenectady International, and was used in litigation designed to protect their own and the town of Rotterdam Junction's water supply. The matter became important because the gravels were being mined and the aquifer was in danger of total removal (Figure 2). The vice-president of the company at that time (Clinton P. Braidwood) provided a drill rig and crew for my exclusive use and hired me as a consultant to study the aquifer fully. Dr. Robert Yunick, the present vice-president, continued the effort and now that litigation has ended successfully, has given me permission to publish the study. The town of Rotterdam Junction now owns the property in question in that area and James Constantino, the present town supervisor, has given permission for our entry to that particular site and has permitted our visit.

GENERAL GEOLOGY

The gravel aquifers in the Schenectady, New York, area are located west of the city along the Mohawk River. They result from a series of glacial advances and withdrawals coupled with fairly complex erosion cycles intermixed with these glacial alternations. All of this lies on the eroded and well dissected preglacial surface of the Middle Ordovician Schenectady Formation consisting of thick-bedded, grey and brown sandstones overlain by a sequence of sandstones and dark grey to black shales followed by an upper series of grey to black shale beds. Each of these lithologies resembles the different members of the Normanskill Formation (Mount Merino and Austin Glen) but there are differences both in the character of the grains and in the included fragments common in the Austin Glen (see Kidd *et al.*, this volume). The Schenectady Formation differs from the other units in sufficient characteristics that it can easily be distinguished from them.

A lobe of the advancing ice turned from its southward course and flowed eastward down the Mohawk Valley. The advancing ice moved over the unconformity on the top of the then eroded Schenectady Formation, eroding it even more, and deepened the walls of that ancient Mohawk Valley. When the ice melted back and withdrew from the area, its fast moving outwash cut another new path in places, often producing deep but small gorges or waterfalls.

Excellent stream gravels were deposited along the course of that river bed but in the deeper areas farther east, glacial muds and grey silt and clay with shale fragments were deposited. These deeper locations may have been parts of an old river bed or merely places where erosion took place at an ice margin, producing deeper pools. As the stream velocity slowed, these finer materials were deposited. Therefore, an old stream valley

In Garver, J.I., and Smith, J.A. (editors). *Field Trips for the 67th annual meeting of the New York State Geological Association*. Union College, Schenectady NY, 1995. p. 413-425.

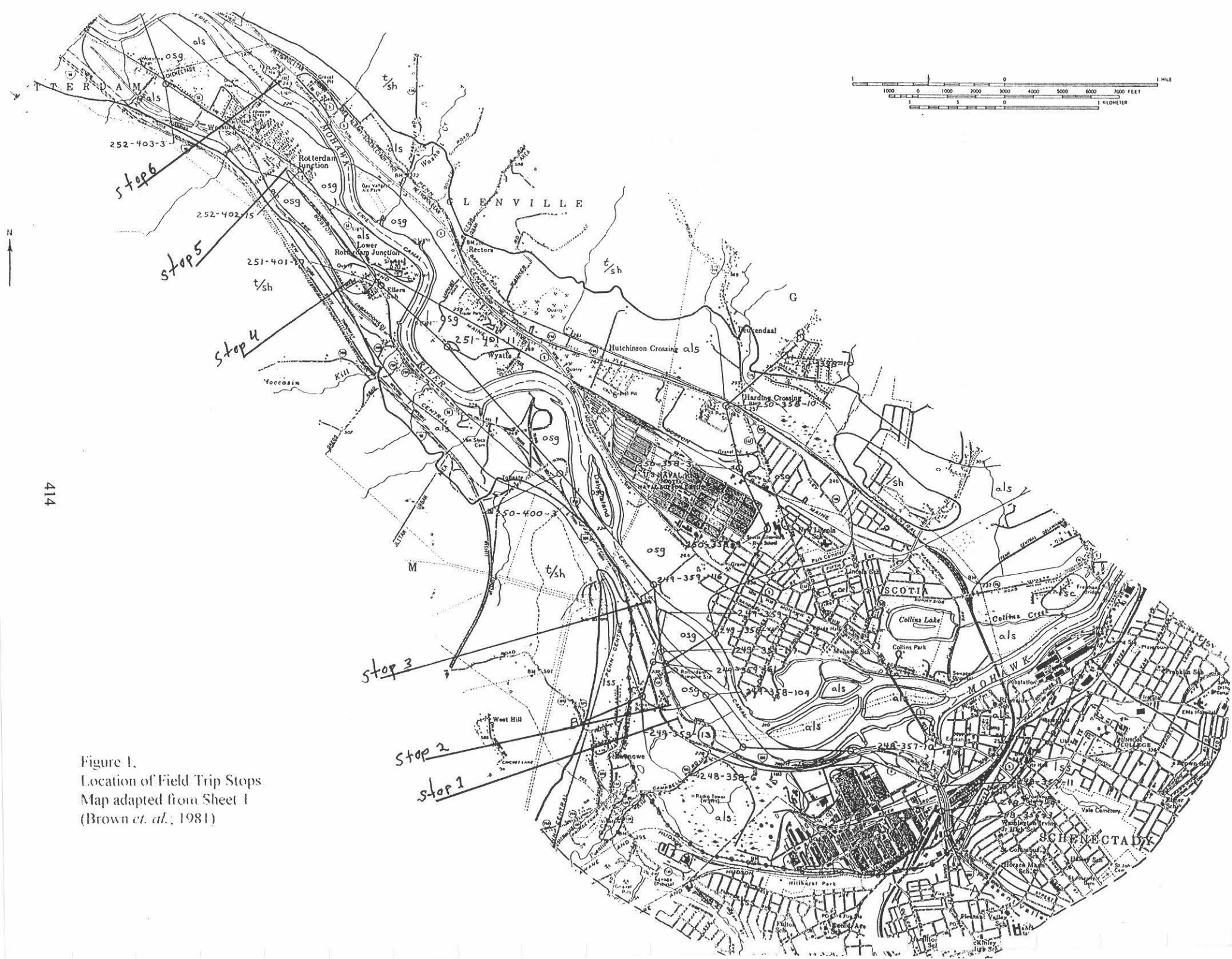


Figure 1.
 Location of Field Trip Stops
 Map adapted from Sheet 1
 (Brown *et. al.*, 1981)

south of, but parallel to, the present stream contains beds of economically useful and well-sorted gravels cut by thick deposits of the clay, silt and fragments. The thickness of the gravels in places is up to about 40 ft. and the thick clay and fine deposits are hundreds of feet thick (see the well logs in Table 1).

The last major event was a brief advance of the ice over the region and the subsequent deposition of a ground moraine of till (gravel, clay, etc.) up to 10 feet thick.

THE GROUNDWATER PICTURE

In no way does this report attempt to discuss the entire story of the groundwater of Schenectady County. That very large study was well done by Simpson (1949) and Winslow *et al.* (1965). However, neither of those studies had sufficient well data to describe the major aquifer and the prime source of water from the aquifer south of Rt. 5S in Lower Rotterdam Junction. This unit is closely related to the Schenectady water well supply which also supplies Rotterdam itself. It is the problem of the lower Rotterdam Junction aquifer that is of primary concern in this report. A discussion of the Schenectady-Rotterdam-Rotterdam Junction aquifer ties the general picture together. Our field trip will visit locations for these parts of the aquifer.

Each of the individual gravel "patches" represents a portion of an extensive series of gravels deposited in an earlier Mohawk River. Each was cut off from the other "patches" by meandering of the river and each became and is now a separate aquifer.

THE SCHENECTADY AQUIFER

These gravel deposits are located in an area known as the Great Flats and are found along the south bank of the Mohawk River over an area a few hundred feet wide and less than one-quarter of a mile in length. The L&M Motel lies on top of the easternmost extent of that aquifer. West of the Schenectady Rotterdam #5 Well Field is Lock #8, but it is Lock #7 which actually controls the level of the water in the River (called the Barge Canal) at that point. As will be shown, the locks on the Mohawk River actually control the flow of water into these gravels.

The quality of the Mohawk River water is dramatically changed after it enters the gravels, which have small amounts of sand and clay in the interstices. As a result, water pumped at the Schenectady and Rotterdam #5 Well Field is cleaned and requires only a small amount of chlorination and relatively minor treatment to make it palatable and highly potable. Of course, it is an area in delicate balance. On an early occasion during the construction of the motel, some contamination did in fact occur but it was rapidly ended and there appears to have been no problem since that time. Any construction or even contamination from the recently built mall on Campbell Rd. (Rotterdam Square Mall), or the recently closed gasoline station (Mobil) could also endanger these water supplies. Furthermore, the water level in the Mohawk River depends upon the lock levels for each of the aquifer portions of these gravels.

THE ROTTERDAM JUNCTION AQUIFER

By far the largest and most vulnerable of the gravel aquifers is the one that provides the water supply for the town of Rotterdam Junction and its main industry, Schenectady International. The length of this gravel aquifer is over 11,000 ft. and it varies from a few hundred to over a thousand feet in width. The gravel is typically a clean pea gravel, but also has a reasonable quantity of sand and minor clay. Thicknesses are quite variable but primarily the best gravels are as few as ten to fifteen to over forty feet thick in much of the length.

The most interesting part of the problem lies in the fact that this aquifer is also a valuable economic resource. Its first appearance was in the Kellam-Schaffer Pit which we will visit on Stop 4. Here gravel was mined for many years. Two wells in this area tap the aquifer and for years provided water for the function of the Rotterdam Junction Chemical Plant which had (and still has) need for a plentiful supply (at least 2000 gallons per minute) of clean water for its operation. Later, others purchased the pit and expanded the quarrying, leaving a large and deep pit (Figure 2). The question of the value of the gravel versus the value of the water produced

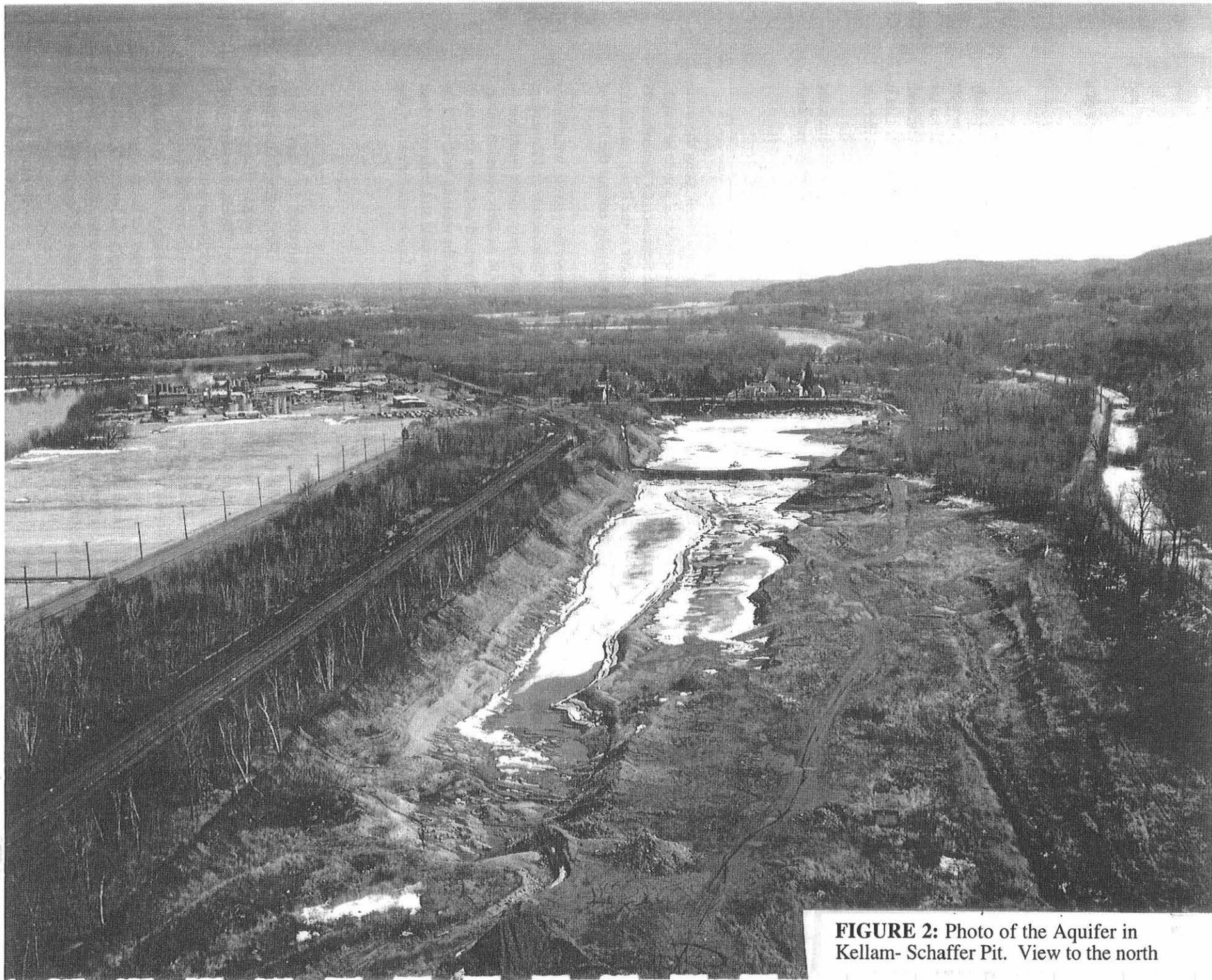


FIGURE 2: Photo of the Aquifer in Kellam- Schaffer Pit. View to the north

nearly thirty years of litigation, mining, periods in which mining was stopped legally, and continual study to learn everything we could about the aquifer in order to do all we could to prove its vulnerability and the probable loss of clean water with the loss of the aquifer to mining. Finally, thanks to Schenectady International, the aquifer area in the now much enlarged pit belongs to the town of Rotterdam Junction which plans to preserve it as parkland.

THE HYDROLOGIC REGIME

Very little recharge of the aquifer results from the surface except where the till cover and/or the gravel have been removed by mining. Originally, before mining, virtually no meteoric water passed through the highly clay-rich till at the surface overlying the aquifer. Now the surface of the aquifer is exposed and is constantly subject to potential contamination.

Recharge of this aquifer has been demonstrated to be from the Mohawk River west of Lock #9. For many years those who had earlier studied the aquifer believed recharge also came from the river east of Lower Rotterdam Junction and above Lock #8. Their argument presented some problems during litigation designed to protect the aquifer. As a result, however, wells were drilled in every possible location and direction east and north (as well as south and west, where possible) and well records fail to show the aquifer in any other position but that shown on Figures 1 & 3. A series of brief logs has been included at Table 1 which is keyed by H number (Hewitt number) to locations on the map (Figure 3) which demonstrates the recharge picture.

It is now very obvious that recharge occurs west of Lock #9 at the southward bend of the river. The water elevation of the river controls the water elevation in the aquifer. Records have been kept for over thirty years that demonstrate this correlation very closely. Within seven to ten days of the time Lock #9 is closed, the water at the old pit begins to rise. When the river level is lowered with the Lock #9 gates opened, the water level in the wells drops subsequently and in the same time interval. Were it not for the voluminous material and the need to consider space available in this guidebook, this material could easily have been included in this present report. Numerous wells have been drilled east of the old pit (Kellam and Schaffer) and east of Mabee Lane. No gravel is found at the horizon of the aquifer although a few minor gravels appear at shallow depths but show no hydraulic connection with the major aquifer. Some of these wells are also shown on the enclosed logs and are located in Figure 3 by H number (Hewitt number).

In addition to all of this, the river bank on the southward bend east of the pit area shows no sign of gravel and wells drilled all along that area show primarily upper-level gravels and deep silt and clay deposits some several hundred feet deep.

Brown *et al.* (1981) described the area in maps and cross-sections to show the geohydrology of the aquifer. In the area of the Rotterdam Junction aquifer, they used my data to produce their concept of this aquifer. Actually, our ideas appear to be quite similar but there are differences based on my complete well data. I have used their base map as my Figure 3 but have added a line marked with x's to show where we differ in concept. Therefore, my Figure 3 is adapted from their Sheet #1.

The aquifer is capable of very high hydraulic conductivity. Enormous volumes of water are passed through daily even at the lowest levels of the Mohawk River source. Schenectady International uses well over two million gallons per day and when one considers the entire area, including the Rotterdam Junction town wells and other users, it is probable that the aquifer could provide better than three million gallons per day with no appreciable loss of storage capacity. A rough estimate done by Prof. Carl George (Union College Biology Department) and Sandy Cardella, engineer, indicates yields and storage capacities even higher.

CONCLUSION

That the aquifer at Lower Rotterdam Junction has suffered from mining is quite clear from the aerial view in Figure 2. Even more obvious is that since the gravel has been mined so seriously, calcium carbonate deposits and surface clay settling prevent water from rising as high as it formerly did in the old pit. Fortunately,

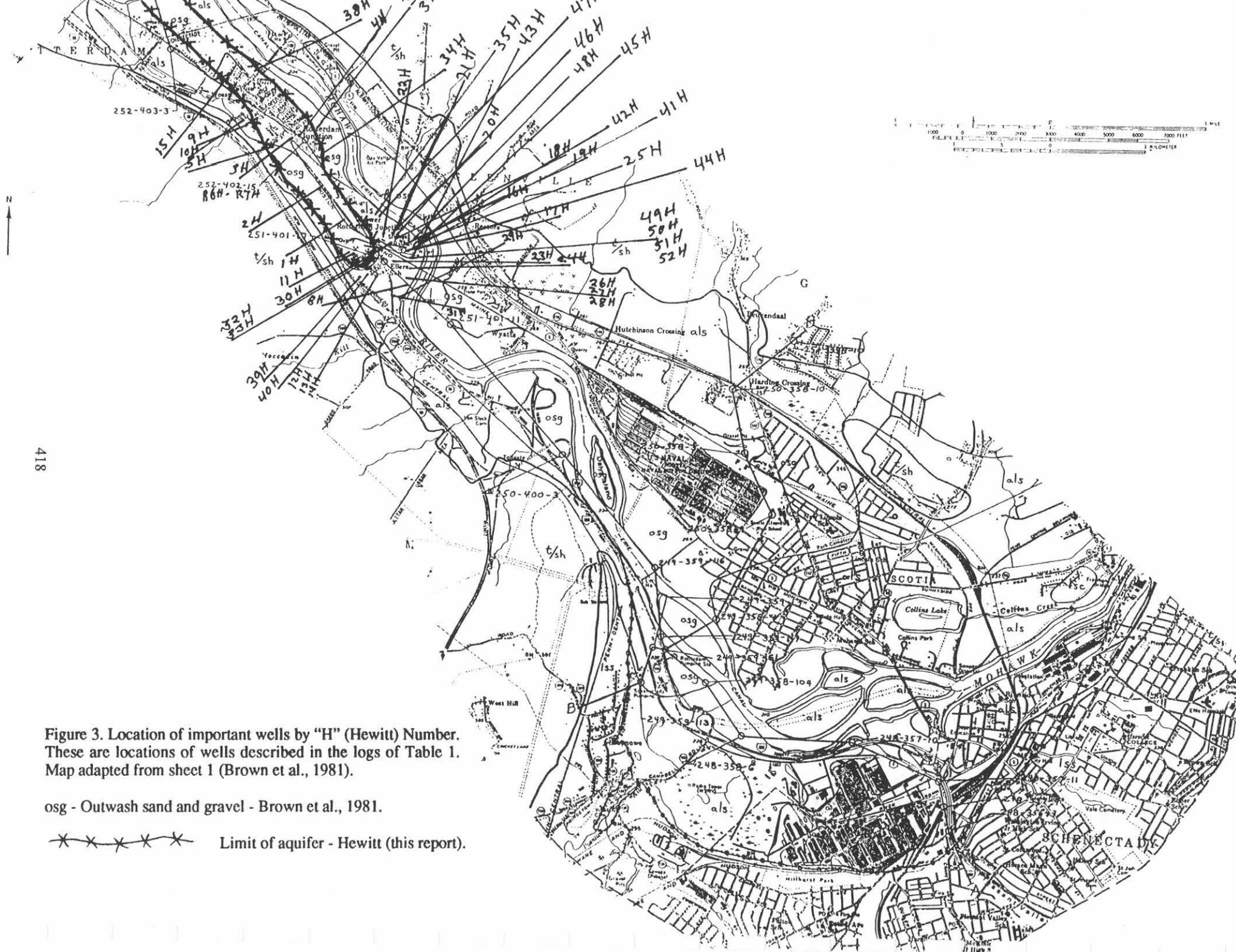


Figure 3. Location of important wells by "H" (Hewitt) Number. These are locations of wells described in the logs of Table 1. Map adapted from sheet 1 (Brown et al., 1981).

osg - Outwash sand and gravel - Brown et al., 1981.

---*---*---*---* Limit of aquifer - Hewitt (this report).

there remains much below surface storage and by intercepting the water in a more westerly area, the volumes are still great and transmissivity is still very high.

The aquifer further east, used by Schenectady and the town of Rotterdam, has not suffered in the same way and should provide its usual plentiful supply of clean water unless additional problems arise.

ACKNOWLEDGEMENTS

My gratitude to Schenectady International and all those associated with it, particularly Sandy Cardella, who assisted in every possible way to make this study successful and complete. There is no way to thank Sandy enough for his interest and his help. Even opponents who argued against certain conclusions are to be thanked. They literally and legally forced the study to go toward certainty and not merely scientific probability. They were actually of great help by making us do more than would normally be required.

TABLE 1: LIST OF WELLS AT ROTTERDAM JUNCTION BY "H" NUMBER (HEWITT NUMBER)

1H	Test Well	1-65		Dry hole
	Layne-New York Co.	Permeable bed	0 - 34"	Sand, gravel, boulders
			(19') - 53"	Sandy grey clay, gravel
			(8') - 61'	Grey clay
			(35') - 96'	Sandy grey clay
			(2') - 98'	Sand, gravel, clay
				98' Bedrock
2H	Layne-New York Co.	2-65		Water 35' 2"
			(16') - 51"	Sand, gravel
			(37') - 88'	Sand, grey clay
			- 88'	Bedrock
3H	Layne-New York Co.	3-65		Water 34'
			0 - 15'	Sand, gravel, clay
			(28') - 43'	Sand, gravel
			(24') - 67'	Sand, gravel, some clay
			(64') - 131'	Sandy, grey clay
			- 131'	Bedrock, sand, gravel on top (thin)
4H	Layne-New York Co.	4-65		Water 35'
			0 - 8'	Sand, gravel
			(6') - 14'	Hardpan clay, sand, gravel
			(56') - 72'	Sand, gravel
			(2') - 74'	Sandy clay, gravel
			(57') - 131'	Sandy grey clay
5H	Layne-New York Co.	5-65		
	Record enclosed		(6") - 6"	Top soil
			(13') - 13'	Sandy clay
			(8') - 21'	Sand, gravel & clay
			(24') - 45'	Sand & gravel
			(21') - 66'	Sand, gravel & boulders
			(16') - 82'	Sand, clay with gravel
			(2') - 84'	Soft rock
			- 84'	Hard rock
R6H		Stewart's	RJ 1957-1	Water 32.63'
	Get record from Stewart		(25')1 - 25'	Silty sand, gravel
			(11') - 36'	Clayey sand, gravel

by permission		(13.5)	- 49.5'	Silty sand, gravel
		(36.5')	- 86'	Sand, gravel (water bearing)
		(2')	- 88'	Hardpan
<u>R7H</u>	<u>Stewart's</u>	<u>RJ1947-1</u>	<u>Water 27.5'</u>	
			1' - 5'	Silty gravel
Get record from Stewart			(15') - 20'	Clayey sand
by permission			(20') - 40'	Gravel with clay
			(5') - 45'	Gravel, little clay
			(3') - 48'	Clean gravel
			(15') - 63'	Gravel (water bearing?)
			(2') - 65'	Clayey gravel
<u>8H B&M R.R.</u>	<u>(G.W. - 30)</u>			
				20 feet deep in gravel-dug 300 gpm
<u>9H R.J.</u>	<u>Water district #3 test 1 - south of Erie Canal at end of Iroquois St.</u>			
				Clay (see G.W. - 30) or Stewart or R.J. Dist. #3.
<u>10H</u>	<u>R.J. Test well for Water district #3 in clay, small yield - N side of Erie Canal.</u>			
				100+ ft. of clay
<u>11H</u>	<u>Schenectady International - 2 wells in quarry</u>		<u>Water 33'6"</u>	
			0 - 6'	Sand and gravel
			(2)6 - 8	Blue clay
			(9)8 - 17	Gravel, sand & clay
			(16)17 - 33	Gravel, sand & clay
			(7)33 - 40	Medium & large gravel
			(6.5)40 - 46.5	Large & medium gravel, boulders
			46.5	Bedrock
<u>12H)</u>			<u>98)</u>	<u>All shallow wells -note topography</u>
<u>13H)</u>		<u>Gravel</u>	<u>101)</u>	<u>of Simpson's Report.</u>
<u>14H</u>			<u>102)</u>	<u>G.W. - 30</u>
<u>15H</u>	<u>Layne-New York Co. 6-65</u>			<u>(Water)</u>
			(6") - 6"	Top soil
			(2.6") - 3'	Sandy brown clay
			(44) - 47'	Sand and gravel
			(4) - 51'	Dirty sand and gravel
			(11'9") - 62'9'	Grey sandy clay
<u>16H</u>	<u>Layne-New York</u>	<u>1-51</u>		<u>(No Water)</u>
			- 2'	Top soil
			(7) - 9'	Sandy clay
			(7) - 16'	Sandy clay & boulders
			(33) - 49'	Tough grey clay
			(56) - 105'	Sticky grey clay
<u>17H</u>	<u>Layne-New York</u>	<u>2-51</u>		<u>(No water)</u>
			- 2'	Top soil
			(18) - 20'	Sandy clay
			(2) - 22'	Sandy clay, little gravel & boulders
			(87) - 109'	Tough grey clay
<u>18H</u>	<u>Layne-New York</u>	<u>3-51</u>		<u>(No water)</u>
			(7) - 7'	Fill and boulders

		(12) - 19'	Sandy clay
		(14) - 33'	Tough grey clay
		(3) - 36'	Clay packed, little sand, gravel, boulders
		(64) - 100'	Sticky grey clay
19H	Layne-New York Co.	4-51	(No water)
		(18) - 18'	Sandy clay, little gravel & boulders
		(5) - 23'	Muddy sand, gravel, boulders-takes water
		(59) - 82'	Sticky grey clay
		(60) - 142'	Tough grey clay, little gravel
20H	Layne-New York Co.	6-51	(Some water)
N.G. pulled casing and filled hole. See comments 21H		(13) - 13'	Sandy clay, little gravel & boulders
		(18) - 31'	Dirty sand, gravel & boulders
		(1) - 32'	Muddy sand, gravel & boulders - takes water
		(2) - 34'	Coarse sand, gravel & boulders (water)
		(46) - 80'	Grey clay
		(10) - 90'	Sticky grey clay
21H	Layne-New York Co.	5-51	(Some water)
Record enclosed.		(9) - 9'	Sandy clay, little gravel, boulders
Ten ft. higher (all measurements)		(15) - 24'	Muddy sand, gravel & boulders (takes water)
than at S.C. wells. Not aquifer - relates to upper stratum 1# (dry) temporary water.		(11) - 35'	Coarse sand, gravel & boulders (water)
		(3) - 38'	Grey clay
22H	Layne - New York	7-51	(No water)
		(6) - 6'	Sandy clay, little gravel & boulders
		(20) - 26'	Muddy sand, gravel & boulders (takes water)
		(29) - 55'	Sticky grey clay
		(64) - 119'	Grey clay, little sand
		(19) - 138'	Grey clay, little gravel
		(5) - 143'	Clay, little shale & little gravel
		(2) - 145'	Broken rock
23H	Layne-New York Co.	8-51	(Some water)
N.G. pulled out		(8) - 8'	Sandy clay & boulders
		(19) - 27'	Muddy sand, gravel & boulders (takes water)
		(3) - 30'	Coarse sand, gravel, boulders (water)
		(13) - 43'	Sticky grey clay
24H	Layne-New York Co.	11-55	(No water)
		(29') - 29'	Fine brown sand & gravel
		(61') - 90'	Grey clay with sharp, black gravel
25H	Layne-New York Co.	12-55	(No water)
		(6) - 6'	Top soil
		(15) - 21'	Sand and gravel
		(36) - 57'	Grey clay
26H	Bradt home	285 ft. to bedrock	240'
See Mr. Bradt for record	All grey clay	285	
27H	Bradt home	85 ft. in grey clay	
See Mr. Bradt for record.			
28H	Bradt home	17 ft. well static level 10 ft.	
See Mr. Bradt for record.	(7 ft. of water) in gravel, Coarse with big boulders - not the aquifer.		

29H	Layne-New York Co.11a-51	(2'6" from surface) (Some water.)
	(5) - 5'	Top soil and clay
	(9) - 14'	River mud
Not aquifer	(6) - 20'	Dirty sand, gravel, boulders - water
Water at 2'6" - would not pump!	(2) - 22'	Muddy little tight sand, gravel & boulders
	(3) - 25'	Clay
30H	Layne-New York Co.10-51	(Water)(38'-46')
This well now dry. Had rapid drawdown.	(9) - 9'	Sandy clay, gravel & boulders
Note: material not aquifer.	(18) - 27'	Muddy sand, gravel & boulders (takes water)
	(11) - 38'	Dirty sand, gravel & boulders (takes water)
	(8) - 46'	Coarse sand, gravel & boulders (water)
	(4) - 50'	Clay
31H	Layne-New York Co. 9-51	
	(5) - 5'	Topsoil and clay
Static 9'. Pumping level 16'	(11) - 16'	River mud
Draw down 7' with G.P.M.	(6) - 22'	Muddy sand, gravel & boulders (water)
N.G.	(3) - 25'	Tough clay
32H	Layne-New York Co.14-55	(Water)
	(47) - 47'	Sand and gravel
	(30) - 7'	Clay
33H	Layne-New York 13-55	(No water)
	(29) - 29'	Yellow sand & gravel
	(31) - 60'	Grey silt & grey clay
34H	West End Orchard St.Hauserwas well north side of house.	
See owner for record.	260 ft. deep. All clay.	
35H	Layne-New York Co.Simpson G.W. - 30	(Some water)
	(24)0 - 24'	Sandy brown clay
	(2) - 26'	Muddy sand with wood
Not enough to pump!	(5) - 31'	Sand, gravel & clay
	(13) - 44'	Grey sand clay
	(13'6") - 57.6"	Grey clay with gravel
	- 57.6'	bedrock
36H	Mr. Zielinski30' in gravel (water)	
See Mr. Zielinski for record.	Shallow gravel (30')	
37H	Schultz	0 - 31' gravel (carbonated water!)
See Mr. Zielinski for record.	- 150'	clay, gravel below
38H	Applebee Well 90' in clay west side of Bridge St. at 5S	
	Hearsay record - cannot confirm, but fits other data.	
39H	Layne-New York Co. 8-65	Water (out at 15gpm)
	0 - 36'	Sand, gravel & clay
	36 - 42'	Grey clay
	- 42'	Bedrock
40H	Layne-New York Co. 9-65	
	(21) - 21'	Sandy clay with sand and gravel

		(5) - 26'	Brown sandy clay
		(17) - 43'	Soft grey clay with sand and gravel
		(2) - 45'	Sand, gravel with clay
		(1) - 46'	Soft rock
		- 46'	Hard rock
41H	Layne-New York Co. 10-64		Water (out at 7 gpm)
		(16) - 16'	Sandy silty clay
		(16) - 31'	Sand, gravel, boulders with clay
		(7) - 39'	Silty clay with sand
		(30) - 42'	Black shale
42H	Layne-New York Co. 11-65		Some water
		(19)0 - 19'	Yellow sandy clay with sand
		(8) - 27'	Yellow sand, gravel & clay streaks
		(4) - 31'	Grey clay with sand & gravel
		(18) - 49'	Grey sticky clay
43H	Layne-New York Co. 12-65		(No water)
		(14') - 14'	Brown sandy clay, sand, gravel, boulders
		(7') - 21'	Grey clay with sand & gravel
		(24') - 45'	Sticky grey clay
44H	Stewart	1-80	
		0 - 29	Brown sand, gravel (yellow!), some silt, trace clay, few gravel pieces
		29 - 50	Grey clay with silt and fine sand
		50 - 55	Silt with fine sand and clay
		55 - 60	Grey clay with silt and fine sand
		60 - 80	Grey clay
		80 - 85	Grey clay with silt and fine sand
		85 - 110	Grey clay
		110 - 120	Grey clay with silt and fine sand
		120 - 133	Grey clay
		133 - 137	Till - compact - dense fine sand, silt, clay - rock fragments
		137 - 165	Med. -fine sand - silt - trace clay, few pieces gravel
		165 - 180	Grey clay with fine sand - silt - few small gravel pieces
		180	Gas - clay and shale fragments - bedrock
45H	Stewart	2-80	
		0 - 10	Sand - coarse, yellow, medium-fine with silt, some pieces fine gravel, stone
		10 - 17	Sand - coarse, medium-fine with silt odor - chemical unrelated to aquifer on this alone and too high.
		17 - 20	Grey clay
46H	Stewart	3-80	
		0 - 10	Brown sand - gravel - some silt (yellow?)
		10 - 15	Brown sand -silt - some fine gravel (yellow?)
		15 - 20	Brown fine sand - silt (yellow?)
		20 - 25	Brown fine sand with silt - some fine gravel (yellow?), clay 24-25 ft.
		25 - 40	Grey clay, silt - some fine sand
		40 - 50	Grey clay

		50 - 60	Grey silt, clay fine sand
		60 - 70	Grey clay
		70 - 85	Sandy grey clay
		85 - 100	Grey clay
47H	Stewart	4-80	
		0 - 10	Brown sand - silt- some fine gravel (yellow?)
		10 - 15	Brn. sand - cs., med., fine silt, pieces of cobbl.
		15 - 20	Brown fine - medium sand, some coarse and few pieces of large cobble, silt
		20 - 25	Brown fn. sand, some cs. & med. sand with silt
		25 - 30	Brown cs. sand - with med. & fine gravel -silt
		30 - 35	Brown sand - coarse, medium, fine with silt
		35 - 37	Brown clay - sand, silt, trace grey clay
		37 - 65	Grey till(?) clay, sand, silt, stone
		65 - 75	Grey clay
		75 - 80	Grey till (?) clay, sand, silt, stone
		80 - 83	Grey clay
		83 - 90	Lumps of grey clay with sand silt stone (till ?)
		90 - 95	Grey clay
		95 - 105	Fine-med. sand, silt, clay; some stone
48H	Stewart		
		0 - 29	Brown (yellow?) sandy loan, sand - silt -fine gravel, cobbles
		29 - 32	Brown fine sand (yellow?)
		32 - 40	Grey clay, silt, sand with some stones
		40 - 55	Silt - sand - grey clay
		55 - 70	Silt - sand - clay - some stone (till)
		70 - 100	Grey clay
49H	Layne-New York Co.	1-66	No water
		0 - 8	Fill
		8 - 12	Clay and gravel
		12 - 64	Sand and gravel
		64 - 132	Clay with streaks of gravel
		133	Rock
50H	Layne-New York Co.	2-66	Not pumpable water
		0 - 14	Fill
		14 - 22	Brown clay and gravel
		22 - 48	Sand and gravel
		48 - 51	Brown clay
		52	Rock
51H	Layne-New York Co.	3-66	Not pumpable water
		0 - 12	Fill
		12 - 21	Brown sandy clay
		21 - 46	Sand and gravel
		46 - 49	Clay
		49 - 52	Soft rock at 52
52H	Layne-New York Co.	4-66	Dry hole
		0 - 9	Fill
		9 - 11	Brown clay and gravel
		22 - 15	Sand and gravel
		15 - 17	Rock

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- Simpson, E.S. 1952. The Ground-water resources of Schenectady County, N.Y., New York Water Power and Control Comm. Bull. GW30.
- Winslow, J.D.; Stewart, H.G., Jr.; Johnston, R.H.; and Erain, L.J. 1965. Ground-water Resources of Eastern Schenectady County, New York; Bull. 57, Water Resources Comm., State of New York Conservation Department.

ROAD LOG

Part of this Road Log is being prepared at a time when construction has altered or impeded travel. The mileage accuracy cannot be precise since road intersection changes may be made by the time the Field Trip is made. Enough data are given to make a successful trip, however. The trip leaves from the west side of the Ramada Inn parking lot near the Automatic Car Wash, Nott St., Schenectady, NY..

Cumulative Miles

- Left on Nott St.
- .1 Traffic light - Left (S) on Erie Blvd.. Move to right side of Erie Blvd. for right turn.
- .6 Right turn on State St. at traffic light. Move to left lane for left turn.
- 1.8 Left turn at light. Follow signs for Interstate 890 to 90 or Rt. 5S toward Thruway. Bear to right at Schenectady County Community College. (General Electric Co. plant on left.) Go West on Interstate 890.
- 2.8 Mohawk River on right. Take exit for Rice Rd. L&M motel ahead on right.
- 3.1 **STOP 1** at L&M Motel. Discuss aquifer location.
- 3.3 Schenectady Water Pumping Station on left. **STOP 2**. Continue discussion of aquifer location. This is source of Schenectady City Water. Note Rotterdam Town Wells #5 District is in same well field. Discuss the aquifer itself and probable environmental concerns.
- 3.5 Go west 0.2 miles on Rice Road to Lock #8. **STOP 3** Park. Discuss source of water here.
- 4.0 Return on Rice Rd. to Rt. 890 (90). Turn right (west) toward NYS thruway entrance. Stay on Interstate 890 (90) to Rt. 5S. Do Not take entrance ramp to the Thruway (Interstate 90) - continue straight (left lane) to Rt. 5S.
- 7.8 Entering Lower Rotterdam Junction. Note bedrock on this road is Schenectady formation.
- 8.2 Mabee Lane. Turn left. Enter gate at fence on right at south corner. Park. **STOP 4** at old Kellam-Schaffer pit. Discuss Rotterdam Junction Aquifer. Note groundwater elevation differences. Note surface till. Note gravel and clay and calcium carbonate blocking pores. Discuss pumping situation and excavation problems. Return to cars.
- 8.4 Turn left onto Rt. 5S. Note Schenectady International Plant at NE location.
- 9.3 #3 Pump House on left side of Rt. 5S opposite Post Office. **STOP 5**. Discuss pumpage of well and the well development. Continue aquifer discussion. Return to cars. Continue west on Rt. 5S.
- 9.8 Turn right (N) on Bridge St. (Rt. 103)
- 10.1 Turn right into Canal Park. **STOP 6**. Discuss Lock #9 and Lock System in general. Field trip ends.

To return to Rt. 5S, leave park; turn left on Rt. 5S to Interstate 890 or 90 to Thruway east or west.