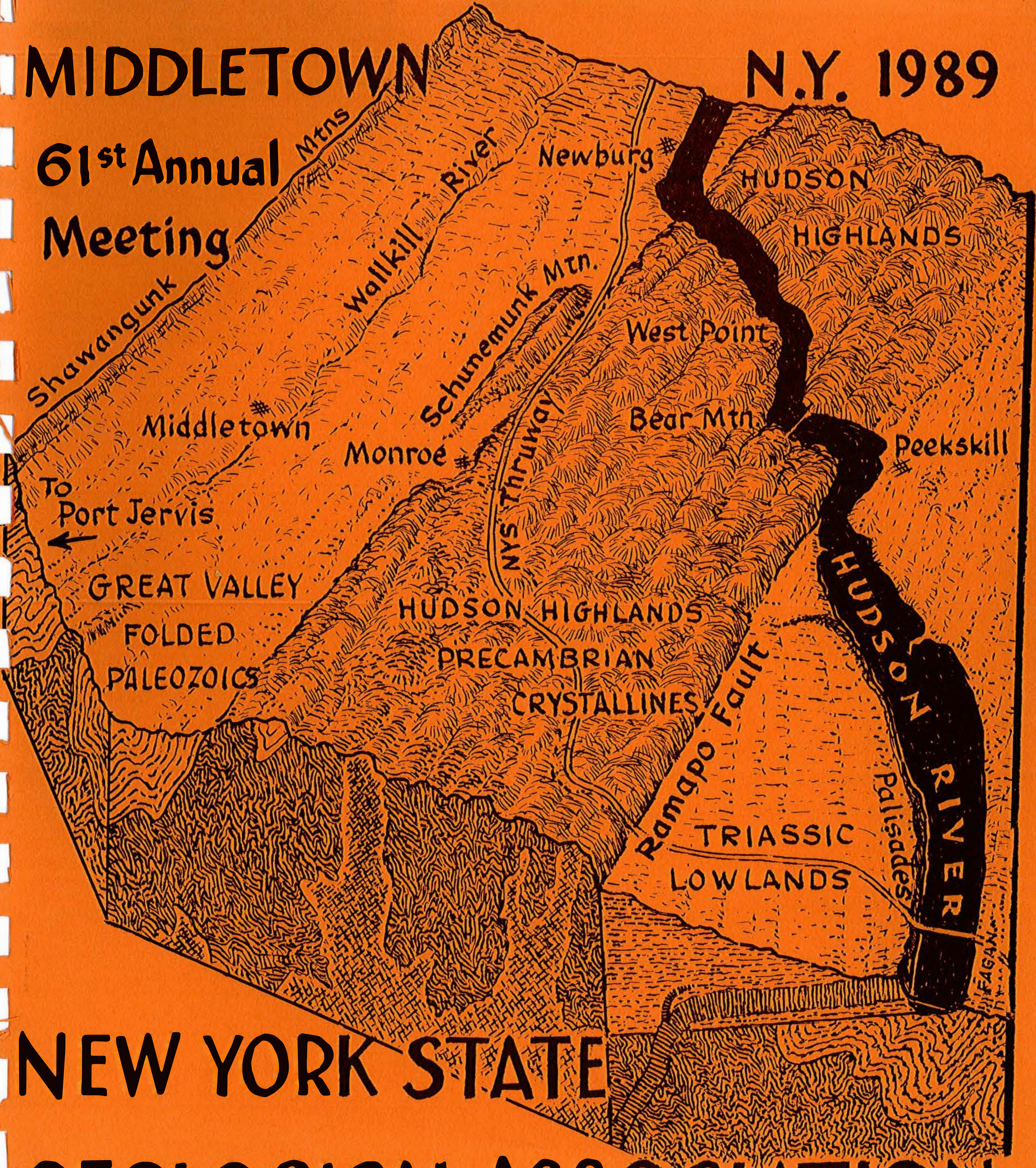
FIELD TRIP GUIDEBOOK

MIDDLETOWN N.Y. 1989

61st Annual Meeting



NEW YORK STATE GEOLOGICAL ASSOCIATION

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61st Annual Meeting

October 13 - 15, 1989

Middletown, New York

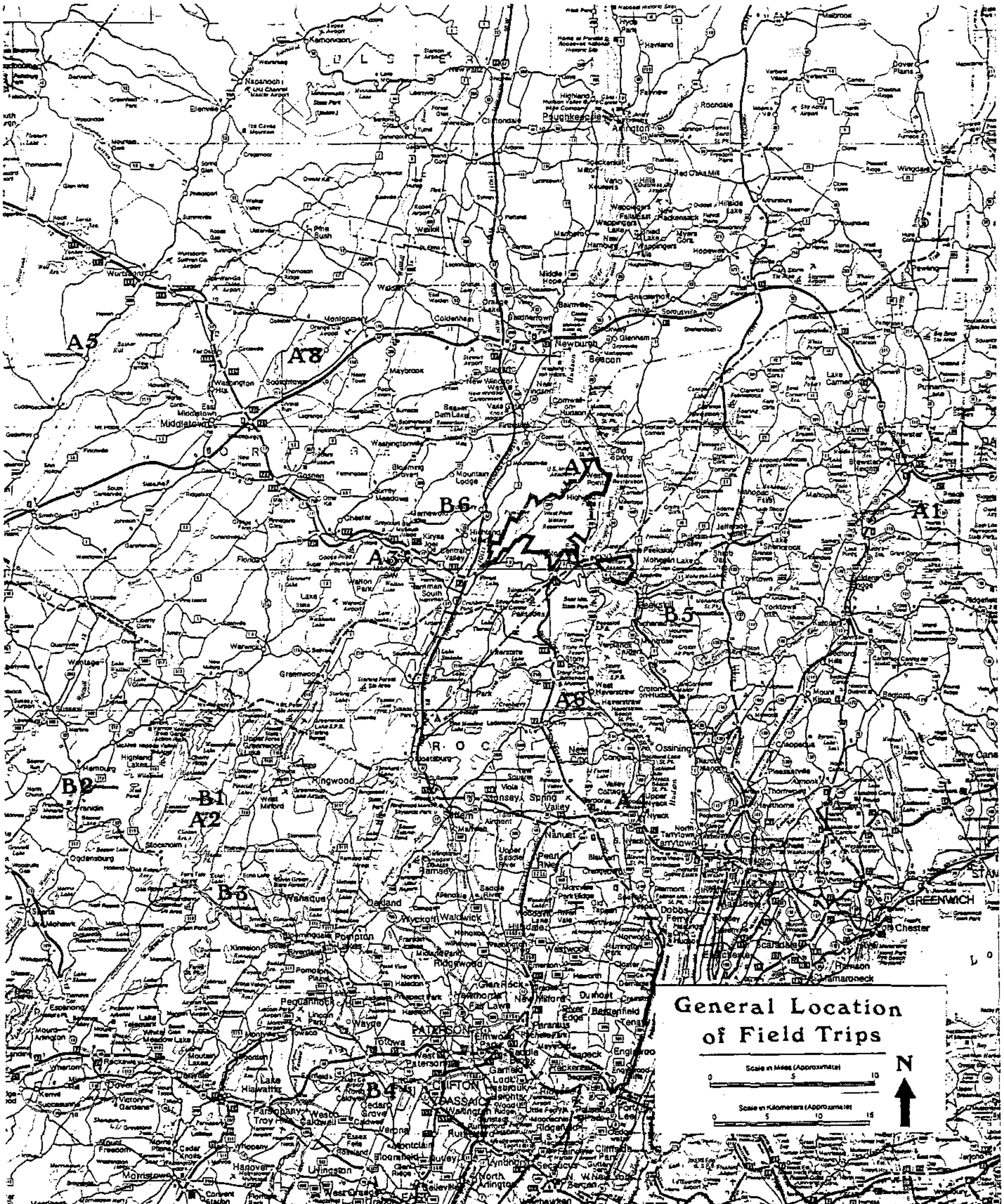
Field Trip Guidebook

Edited by

Dennis Weiss
Department of Earth and Planetary Sciences
City College of CUNY

Hosted by

Larry O'Brien
Department of Science and Engineering
Orange County Community College


General Location
of Field Trips
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Scale in Kilometers (Approximate)
N
A1
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B6

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PREFACE

The diversity of the geology of the Lower Hudson Valley and environs is reflected in the breadth of field trips contained in this guidebook for the 61st Annual Meeting of the New York State Geological Association. No where in New York can one view, in an area of comparable size, such a wide variety of rocks, structures, and sediments representing more than one billion years of Earth history. Evidence of major events in the Earth's history and development ranging from periods of extreme deformation, to times of extensive sedimentation, to the breakup of continents, to the covering of the land by vast ice sheets are contained in the area's geologic sequences. In addition, modern problems in applied geology such as seismic risk and slope stability are also of particular concern.

Several of the field trips retrace the paths their authors led us on at previous meetings of the Association. Howard and Elizabeth Jaffe; Gordon Connally, Les Sirkin, and Don Cadwell; and Bob Finks update information presented at various meetings over the past twenty-nine years. It is with nostalgia that I remember that these were the first trips of the Association that I attended as a student. Moreover, some of you may recognize the art work of Jack Fagan which adorns the cover the guidebook. New views are presented by Pam Brock on the stratigraphy of the Manhattan Prong; by Tony Prave, Moses Alcala, and Jack Epstein on the nature of sedimentation in the middle and late Silurian; and by Alan Kafka, Margie Winslow, and Noel Barstow on the neotectonics of the Lower Hudson Valley region. Clearly, there is continuing interest in the Green Pond Outlier. Larry Malizzi and Alec Gates; Jim Mitchell and Randy Forsythe; and Bob Finks and Michael Raffoni present their respective interpretations of that area's geologic development. Triassic/Jurassic intrusive and extrusive igneous activity is examined by Jeff Steiner and John Puffer, respectively. Over the years the area of the Lower Hudson Valley has been extensively developed and the surficial deposits of the area disturbed. Despite this added difficulty, Les Sirkin, Don Cadwell, and Gordon Connally have been able to complete new studies of the Pleistocene geology of eastern Putnam County and Westchester County. As growth and development of the region have continued, the expansion of its highways has been accompanied by problems of slope stability. Clay Bolton's field trip will examine this concern which, in recent years, has led to catastrophy in several instances. Looking back into prehistory, this region has been occupied by man for over 6,000 years. Phil La Porta's presentation on chert, relates its stratigraphic occurrence to regional archeologic studies. Finally, as our concerns for science education increase, Bill Tucci and Bob Kalin in their "Geologic Climb of Schunnemunk Mountain" present a useful "hands on" teaching exercise for Earth Science teachers.

In closing, I would like to thank the authors of each trip for their effort and cooperation. Without them this guidebook would not be the professional reference and source of information it is. In addition, my gratitude goes to Jack Fagan for finding the time to do the cover of the guidebook, Jim Olmstead and Fred Wolfe for their wisdom and guidance, Larry O'Brien for his uplifting and cheerful support during the editing process, and finally to my wife Susan for patiently enduring the piles of manuscripts and edited texts left all over the house.

Dennis Weiss, Editor
October, 1989

STRATIGRAPHY OF THE NORTHEASTERN MANHATTAN PRONG, PEACH LAKE QUADRANGLE, NEW YORK - CONNECTICUT

PAMELA CHASE BROCK
Queens College and the Graduate Center of the City University of New York

INTRODUCTION

The purpose of this trip is to consider the stratigraphy of the Manhattan Prong along the New York-Connecticut border, in the Peach Lake quadrangle (Fig. 1). Recent mapping in this area has suggested that a major stratigraphic revision is necessary. Overlying the Grenvillian rocks of the Fordham Gneiss and below the magnesian marbles of the Cambro-Ordovician Inwood Marble, a traceable suite of metasedimentary and metavolcanic rocks appears to exist. This suite, here informally named the "Ned Mountain Formation", includes the K-feldspar rich quartzites typical of the "Lowerre Quartzite"; but it also contains a variety of other lithologies, all of which have previously been assigned to either the Fordham Gneiss or to post-Lowerre Paleozoic units. Ways in which rocks of the "Ned Mountain Formation" can be distinguished from other units will be pointed out on the trip and in the text.

Recognition of this new, expanded, Late Precambrian to Early Cambrian unit will have important implications both for the rift-to-drift stage of geological history and (by its map scale distribution) for Paleozoic structures in the Manhattan Prong.

GEOLOGICAL SETTING

Rocks of the Manhattan Prong consist of Grenville basement gneisses and overlying Cambro-Ordovician strata. In the Croton Fall-Peach Lake area of New York, the Prong was multiply deformed and metamorphosed at K-feldspar-sillimanite grade during the Taconian orogeny. Later, probably during the Carboniferous, it was cut by shear zones and its K-feldspar-sillimanite assemblages locally retrograded (Brock and others, 1985; Brock & Brock, 1985a & b).

The generally accepted interpretation of the Prong's stratigraphy, familiar from the works of Hall (1968, 1979), Ratcliffe and Knowles (1969), and many others, consists of four major units (Table 1). The structurally lowest unit, Fordham Gneiss, is generally interpreted as a Grenvillian basement complex. Overlying it, the Cambrian Lowerre Quartzite, Cambro-Ordovician Inwood Marble, lower Middle Ordovician Walloomsac Schist (Manhattan A of Hall), and allochthonous Cambrian Manhattan Schist (Hall, 1968, 1979) comprise the accepted post-Grenville succession of the Manhattan Prong. Cameron's Line, which marks the boundary between the Manhattan Prong and overthrust Hartland terrane, nicks the northeastern border of the Peach Lake quadrangle (Fig. 2; Table 1).

The most compelling lines of evidence for the existence of the expanded Late Precambrian-Cambrian sequence of the "Ned Mountain Formation" come from a) the persistence of distinct lithologies in between the granulites of the Fordham Gneiss and the magnesian marbles of the Inwood (Fig. 2); b) traceability of rock units within this intervening sequence (Fig. 4); and c) truncation of units of the Fordham Gneiss against the base of this sequence (Fig. 2).

DESCRIPTION OF ESTABLISHED UNITS: FORDHAM GNEISS, INWOOD MARBLE, WALLOOMSAC SCHIST, MANHATTAN SCHIST, AND HARTLAND FORMATION

Fordham Gneiss. Each of the six units recognized in the Fordham Gneiss of the Peach Lake quadrangle is at least locally in contact with the "Ned Mountain Formation". Mineralogies of these units reflect their history of Grenvillian granulite-facies metamorphism, variably recrystallized at somewhat lower grade during the Taconian orogeny. These six units are:

Yf_I. Leuco- to mesocratic, medium-grained gneiss, with 3 mm- to 8 mm-

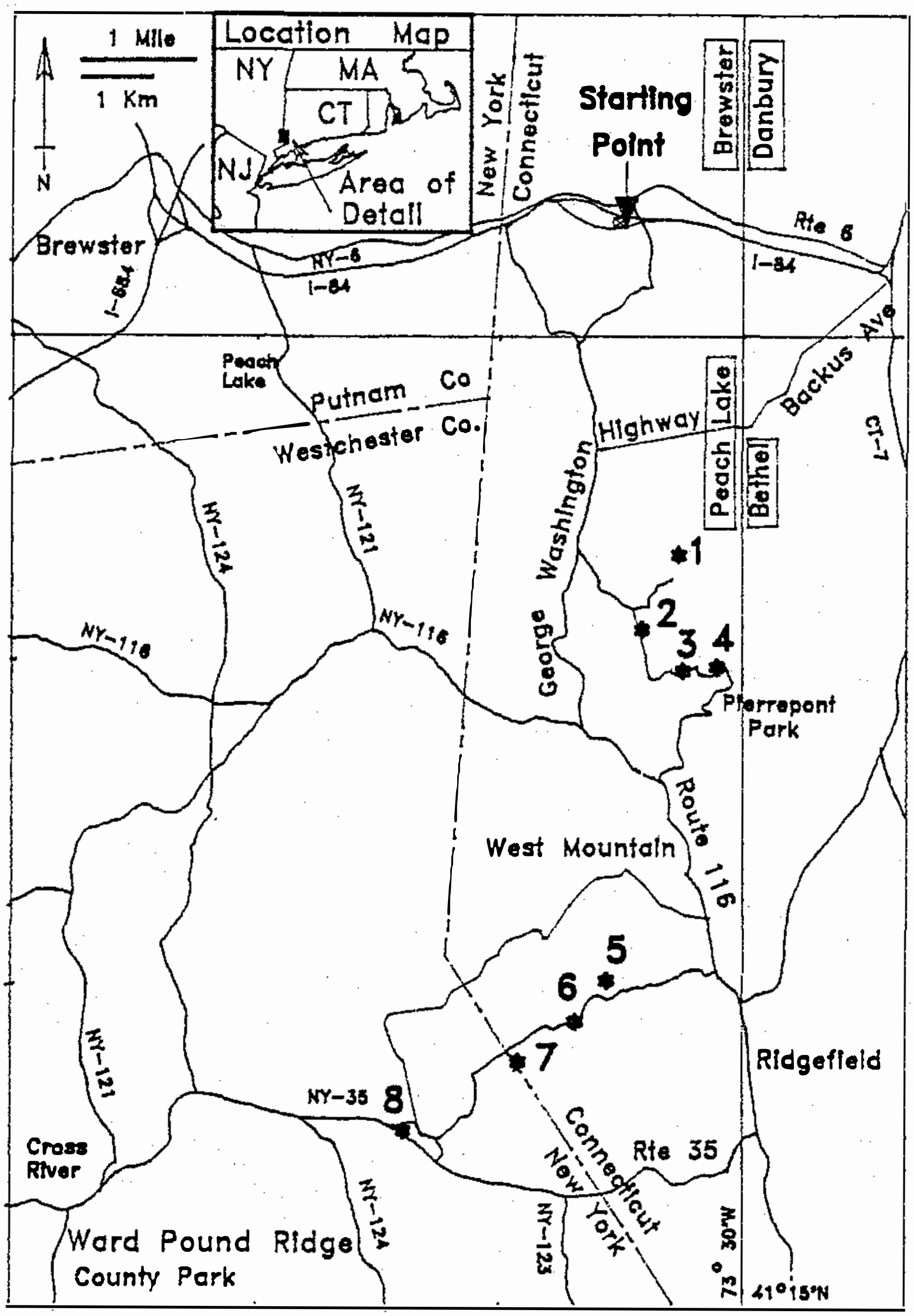
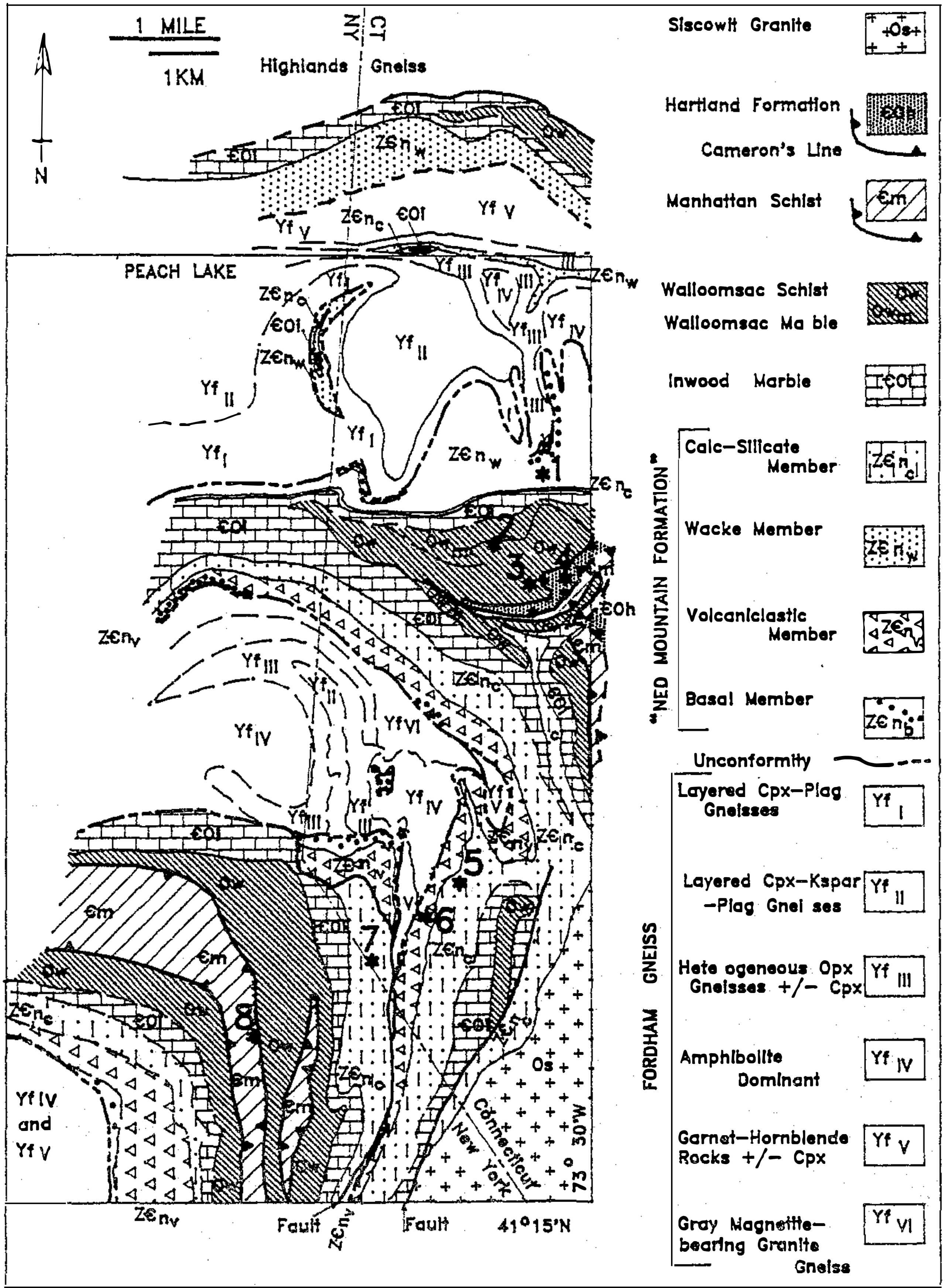


Figure 1. Location map showing field trip stops. Peach Lake & surrounding quadrangles are labeled.

Figure 2. (opposite) Geologic map of parts of the Peach Lake and Brewster quadrangles.


1 MILE
1KM
N
NY
CT
Highlands Gneiss
PEACH LAKE
Fault
Fault
41°15'N
73° 30'W
Connecticut
New York
Siscowit Granite
Hartland Formation
Cameron's Line
Manhattan Schist
Walloomsac Schist
Walloomsac Ma ble
Inwood Marble
"NED MOUNTAIN FORMATION"
Calc-Silicate Member
Wacke Member
Volcaniclastic Member
Basal Member
Unconformity
FORDHAM GNEISS
Layered Cpx-Plag Gneisses
Layered Cpx-Kspar -Plag Gnei ses
Hete ogeneous Opx Gneisses +/- Cpx
Amphibolite Dominant
Garnet-Hornblende Rocks +/- Cpx
Gray Magnetite-bearing Granite Gneiss
Yf IV and Yf V

thick discontinuous streaks and broader spaced (0.5-2.0 m) compositional layering. Mesocratic to leucocratic layers are plagioclase-quartz-hornblende-biotite±clinopyroxene; the amphibolites are hornblende-rich. Minerals show little planar parallelism.

Yf_{II}. Leuco- to mesocratic quartz-2-feldspar-hornblende-biotite±clinopyroxene streaky gneiss, again with amphibolitic interlayers. This unit strongly resembles Yf_{I} and is distinguished mainly by its K-feldspar content.

Yf_{III}. Interlayered coarse-grained, orthopyroxene-bearing gneisses (Stop 1A), consisting of a) light-colored quartz-plagioclase-orthopyroxene-hornblende-biotite±K-feldspar±garnet gneiss, with dark orthopyroxene-hornblende-plagioclase layers; b) mesocratic quartz-plagioclase-orthopyroxene-clinopyroxene-hornblende±biotite gneiss; and amphibolite-facies equivalents, remetamorphosed during the Taconian orogeny. Relicts of orthopyroxene embayed by cummingtonite are often found in these rocks. The unit is recognized at amphibolite facies in part by its characteristic compositional layering, with layers 1 to 5 cm thick. At amphibolite facies, these layers consist of a) quartz-plagioclase-hornblende-biotite-garnet±relict orthopyroxene, with amphibolite and biotite-rich interlayers, and b) quartz-plagioclase-hornblende-biotite±clinopyroxene layers. The amphibolite-facies gneisses are much better foliated than the granulite-facies rocks, due to the increase in biotite and decrease in pyroxene content.

Yf_{IV}. Amphibolite and amphibole-plagioclase-quartz-biotite gneisses, with minor pyroxene. This unit is well-layered but appears to be much more uniform and melanocratic than Yf_{III}. These rocks are medium-grained and equigranular.

Yf_{V}. Melanocratic to leucocratic, coarse-grained hornblende-garnet-plagioclase-clinopyroxene±orthopyroxene and quartz-plagioclase-hornblende-biotite gneisses (Stop 1B). At peak metamorphic grade, these rocks often have a red-and-green appearance due to intergrowth of garnet and clinopyroxene. Hornblende appears to be (at least in part) the product of Taconian metamorphism. Associated with the garnet-clinopyroxene gneiss in this unit is a rusty-staining, medium-grained, quartz-plagioclase-biotite-garnet-graphite granofels. Amphibolites are also common and the unit may in fact grade into Yf_{IV}. Some coarse-textured hornblende-clinopyroxene-biotite-plagioclase-quartz gneisses are interpreted as amphibolite-facies equivalents.

Granites of probable Grenville and Taconian ages are found with all the above units. Some of the Grenvillian granites are hornblende-bearing. One granite is large enough to be considered as a separate unit:

Yf_{VI}. Quartz-plagioclase-K-feldspar-biotite-magnetite±hornblende granitic gneiss. The gneiss contains amphibolitic layers. This grey granitic gneiss is remarkable for its modal abundance (up to 5%) of 1 to 3 mm clots of magnetite.

Many of the Fordham units share certain features that help distinguish them from post-Grenvillian rocks; this is especially true where effects of Taconian deformation and metamorphism are minimal. In such areas, Fordham rocks are typically medium- to coarse-grained (most grains 2 mm to 15 mm in diameter) and well-layered. They occasionally have a well-developed linear fabric, but are poorly foliated. They have relatively little biotite and proportionally large amounts of pyroxene. Feldspars often appear waxy and somewhat greenish. Equant and tabular grain shapes predominate, suggesting extensive grain growth after Grenvillian deformation. Post-Grenvillian rocks of similar compositions, in contrast, are finer-grained, better-foliated, contain more biotite, and have white, grey, or pink feldspar. A tendency towards mineral segregation and a high incidence of ribbon quartz (as well as of flattened grains of minerals like garnet) contributes to the pronounced fabric characteristic of post-Grenvillian rocks in the area.

Metamorphic grade, as reflected by mineralogy, can help distinguish Fordham and younger rocks. Grenvillian metamorphism attained higher grade than any known in the Paleozoic rocks of the area: coexistence of hypersthene and K-feldspar has been documented in the Fordham (Brock and Brock, 1983), and the garnet-clinopyroxene-plagioclase assemblages of Yf_V are distinctive. For many bulk compositions, however, mineralogies of the Fordham and younger rocks can be similar.

Taconian metamorphism was at least K-feldspar-sillimanite grade throughout the Peach Lake quadrangle. But within the Peach Lake quadrangle, an early Taconian metamorphic gradient has recently become evident. Peak assemblages generally increase in grade towards the northeast; the highest grade was reached in the vicinity of Pierrepont Park, Connecticut (Fig. 1). For example, many mafic units of Paleozoic age contain cummingtonite-hornblende-garnet-biotite (orthopyroxene-absent) assemblages in the southern Peach Lake quadrangle. Hypersthene-hornblende-biotite is common in Paleozoic rocks in the northern area, where peak Taconian assemblages do not appear to contain cummingtonite. Hartland and Walloomsac lithologies include two-pyroxene-bearing amphibolites in the northeastern Peach Lake quadrangle (Stop 4). This distribution of mineral assemblages is consistent with a transition from barely K-feldspar-sillimanite grade upwards into granulite facies (Hollocher, 1985). In the peak metamorphic zone, late Taconian K-feldspar-sillimanite assemblages overprint cordierite-garnet-sillimanite grade rocks (Stop 3), and sparse periclase has been found in magnesian marbles (Stop 1E). Assemblages in marbles, pelites and mafic rocks all apparently reflect the early Taconian regional gradient.

But even in the zone of Taconian metamorphism, Fordham and younger rocks remain texturally distinct. The orthopyroxene-bearing amphibolites of Paleozoic age are fine-grained, well-foliated, and in hand specimen resemble their cummingtonite-bearing counterparts much more than the orthopyroxene gneisses of the Fordham (Yf_{III}). Petrographic characteristics, combined with stratigraphic relationships, permit Grenvillian and younger rocks (including the "Ned Mountain Formation") to be confidently distinguished in most cases.

Paleozoic Strata: Inwood, Walloomsac, Manhattan, and Hartland. The distinctive magnesian lithologies of the Cambro-Ordovician Inwood Marble (Stop 1E) form an invaluable stratigraphic marker (Fig. 4 & Table 1). This unit is correlated with dolomites of the Wappinger Group and is inferred to have been deposited in a stable shelf environment. The marble of the Inwood is often calcitic in the Peach Lake quadrangle; forsterite+calcite coexist, and only low-silica Inwood retains dolomite (Fig. 3). The marbles are white-weathering where calc-silicate minerals are minor; calcite-clinopyroxene-forsterite-phlogopite marbles appear greenish and calcite-dolomite-forsterite-phlogopite marble weathers buff. Opaque minerals are generally very minor (well under 1%), but, rarely, graphite is abundant. Phlogopite-rich and quartzite interlayers occur. The Inwood has not been subdivided here, because a relative scarcity of outcrop makes it difficult to determine its internal stratigraphy.

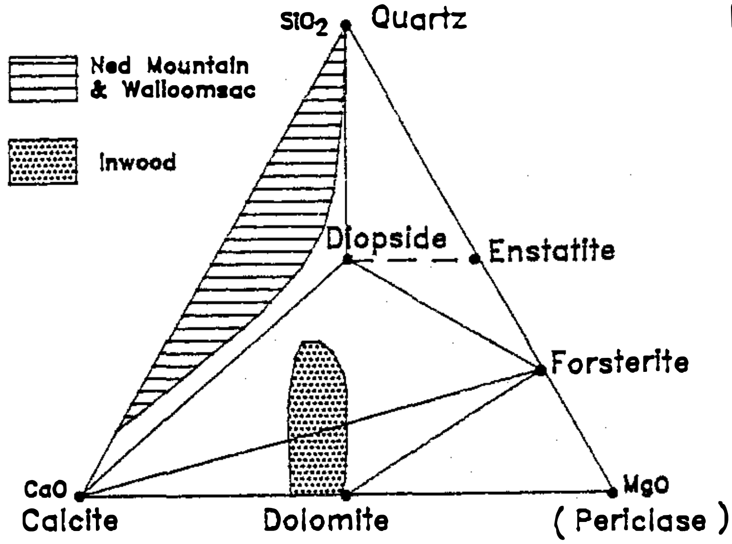


Figure 3.

CaO–MgO–SiO_2 plot showing compositional differences within the Paleozoic marbles and calcsilicates. K–feldspar–sillimanite grade assemblages are illustrated; at higher grade in the northern Peach Lake quad, dolomite begins to break down to calcite + periclase

In the area of peak Taconian metamorphism, Mg-Al spinel partly replaced by phlogopite is sometimes found; and dolomite has begun to dissociate according to the reaction

$$CaMg(CO_3)_2 = MgO + CaCO_3 + CO_2$$
$$\text{dolomite} = \text{periclase} + \text{calcite} + \text{carbon dioxide}$$

Decomposition of dolomite requires extremely high temperatures and is inhibited by high partial pressures of CO_2. The presence of periclase and spinel in the Inwood suggest Taconian peak temperatures >800^{o}C and relatively low pressures (Winkler, 1974).

Unconformably overlying the Inwood, the lower Middle Ordovician Walloomsac Schist (Table 1) contains a basal marble and calc-silicate unit (Ow_m). Ow_m can be distinguished from Inwood in a number of ways. Graphite is common and the unit is sulfidic and rusty-weathering. Walloomsac marbles are much less magnesian than the Inwood (Fig. 3); typical mineral assemblages include calcite-clinopyroxene-quartz-plagioclase-K-feldspar. In Ow_m, carbonate is less abundant and silicates generally more abundant than in the Inwood Marble. The rocks often have a green-and-white speckled appearance due to the abundant green clinopyroxene grains.

Walloomsac marble (Stop 2) is associated with Walloomsac granofelses and schists (O_w). Walloomsac granofelses are fine-grained rocks containing quartz-plagioclase-biotite-garnet±graphite, and little (1-2%) or no K-feldspar (Stop 2). The granofelses tend to break in a slabby fashion. Walloomsac schists usually contain quartz-plagioclase-K-feldspar-garnet-sillimanite-biotite±graphite, though cordierite is present at peak metamorphic grade (Stop 3). Minor cummingtonite-bearing amphibolites also occur in the formation. Bedding is well-defined in the calcareous and sandy parts of the Walloomsac, and poorly defined in the thicker-bedded schists. The schists and granofelses, like the calc-silicates, are usually sulfidic (containing pyrrhotite and pyrite) and rusty-weathering.

Manhattan Schist (Stop 8) is dominantly quartz-plagioclase-K-feldspar-biotite-sillimanite-garnet schist. In contrast with the more sulfidic Walloomsac, ilmenite and magnetite are the major opaque Fe minerals. Hornblende

TABLE I. Stratigraphic Correlation between Peach Lake and Surrounding Areas

	Drake (1969,1984) Delaware Valley N.J.-PA.	Prucha et al. (1968) Northeastern Manhattan Prong	This Report Northeastern Manhattan Prong	Hall (1968, 1979) Southeastern Manhattan Prong	Fisher (1977) Knopf (1962) Dutchess County, N.Y.
Structural Position of Overthrust Units			Hartland Formation Manhattan Schist	Manhattan Schist, Members B & C	
Middle Ordovician	Martinsburg Formation Jacksonburg Limestone	Manhattan Schist	Walloomsac Schist Marble	Manhattan Schist, Member A Marble	Walloomsac Formation Balmville Limestone
Early Ordovician	Beekmantown Group	Inwood Marble	Inwood Marble	Inwood Marble E D	Stockbridge Group: Copake Limestone, Rochdale Limestone, Halcyon Lake
Cambrian	Allentown Dolomite Leithsville Formation			C B A	Wappinger Group: Briarcliff Dolomite, Pine Plains Formation, Stissing Dolomite
	Hardyston Quartzite		"Ned Mountain Formation": Calc-silicate Member, Wacke Member, Volcaniclastic Member, Basal Member	Lowerre Quartzite	Poughquag Quartzite
Proterozoic Z	Chestnut Hill Formation			? Yonkers Gneiss ?	
Proterozoic Y	Gneisses of the Reading Prong	Fordham Gneiss	Fordham Gneiss	Fordham Gneiss	Gneiss

amphibolite is commonly interlayered with the pelites; garnet-rich granofelsic layers up to 60 cm thick are scattered throughout. This unit is thought to be of Cambrian age, a correlative of the Hoosac Formation (Rodgers, 1985) allochthonously emplaced over the Middle Ordovician Walloomsac (Table 1) (Hall, 1968).

Hartland Formation (Stop 4), undivided here, consists of hornblende-orthopyroxene-plagioclase amphibolites, thick-bedded quartz-plagioclase-biotite granofelses, and rhythmically bedded schist-and-quartz-feldspar granofels. Cameron's Line marks the early (D_1) Taconian thrust along which the Cambro-Ordovician Hartland Formation was emplaced against the rocks of the Manhattan Prong (Fig. 2 & Table 1).

THE "NED MOUNTAIN FORMATION": LITHOLOGIES AND STRATIGRAPHY

The lower limit of the "Ned Mountain Formation" is defined by the post-Grenville unconformity along which the units of the Fordham Gneiss are truncated. The upper limit is set by the magnesian marbles of the Inwood, which are succeeded at structurally higher positions by Walloomsac, Manhattan, and Hartland lithologies. Between these two limits, a complex suite of rocks with its own distinctive stratigraphy can be traced (Figs. 2 & 4).

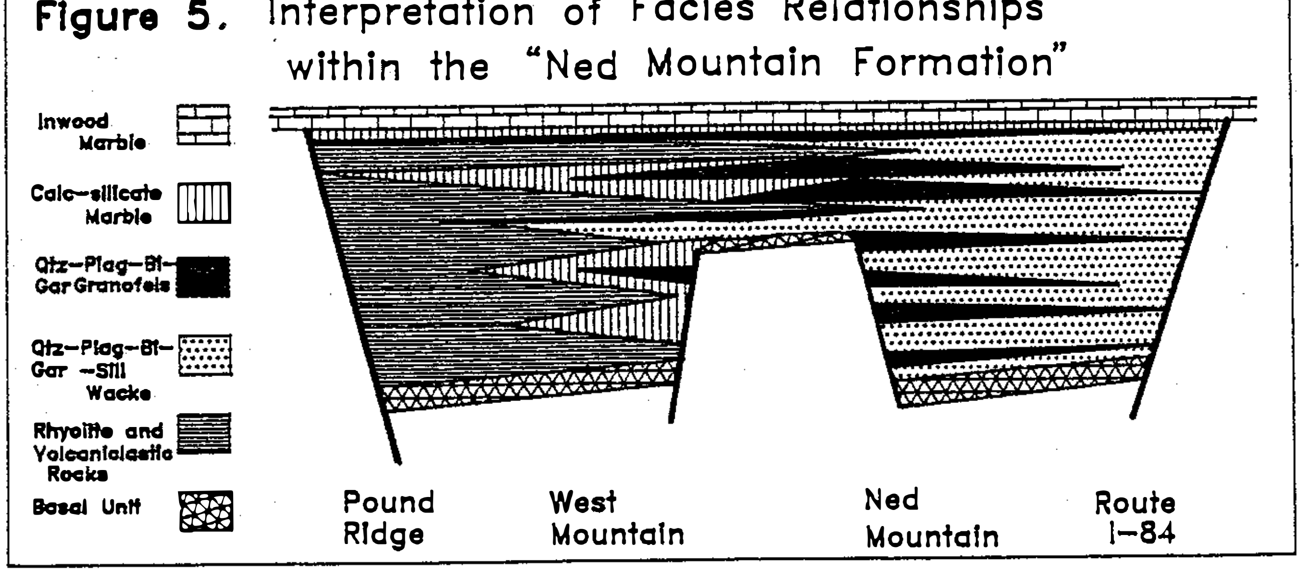
Figure 4. FACIES RELATIONSHIPS WITHIN THE "NED MOUNTAIN FORMATION"

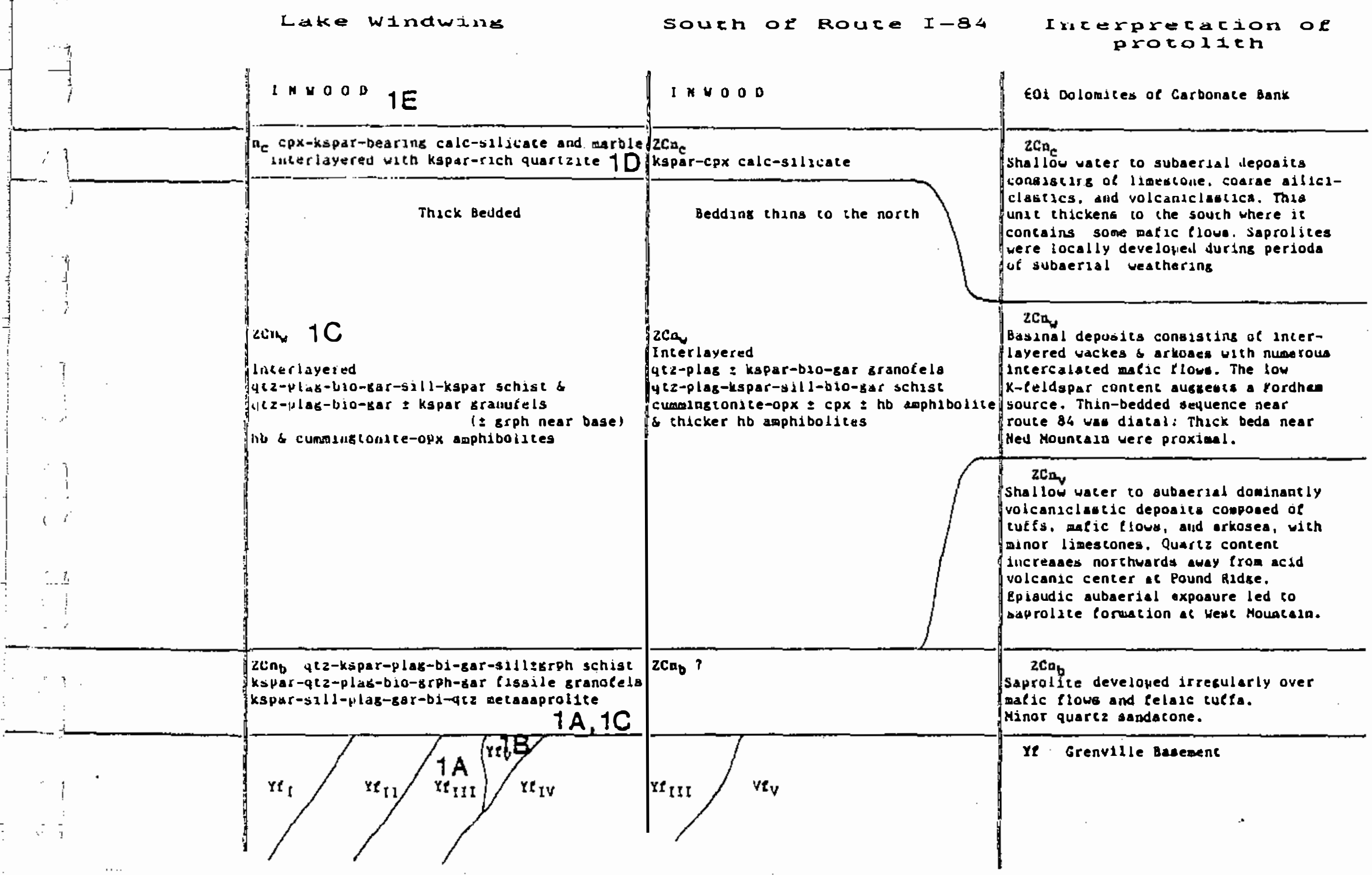
	Pound Ridge	West Mountain: South and West	West Mountain: North and East
€Oi	I N W O O D	I N W O O D	
ZCn_c <30m thick in north	ZCn_c qtz-plag-bio ± grph fissile granofels minor qtz-plag-kspar-gar-sill schist interlayered kapar dominant & quartzose granofels hb-cpx amphibolite, and cummingtonite amphibolite	ZCn_c increasing carbonate content **5, 7** interlayered mostly thin-bedded kspar-dominant & quartzose granofels (+plag±redBio±gar±grph) cpx-bearing calc-silicates with marble & hb amphibolite interlayers qtz-plag-red Bio-gar ± grph fissile granofels kspar-sill-bio-gar ± grph ± qtz metasaprolite and minor garnetiferous quartzite qtz-plag-bio-gar-sill schist	
ZCn_v		- - - Gradational Boundary - - -	
ZCn_v	ZCn_v hb-cpx ?calc-sil Pound Ridge granitic gneiss Hypersolvus Granite	ZCn_v **6** mostly thin-bedded feldspathic & quartzose granofels minor cpx-bearing layers thin layers rich in bio, amph & plag hb amphibolite thick amphibolite near the base kspar decreases ----------------> qtz increases	kspar-sill-gar ± qtz ± grph metasaprolite qtz-kspar-plag-bio & qtz-bio-plag-kspar granofels hb amphibolite
ZCn_b <50m thick	ZCn_b cumm-hb-gar-bio-grph fissile granofels qtz-kspar-sill-gar schist	ZCn_b k-spar- & gar-bearing quartzite cummingtonite amphibolite kspar-sill-gar-bio metasaprolite kspar-qtz-sill-gar-bio ± plag schist	cummingtonite-opx-plag-grph fissile granofels kspar-sill-gar-bio-grph±qtz metasaprolite plag-bio-gar-qtz-graph
Yf	Yf_{IV} and Yf_V	Yf_{III} Yf_{II} Yf_{III} Yf_V Yf_{IV}	Yf_{VI} Yf_{IV} Yf_V

Abbreviations: bio = biotite; redBio = red biotite;
cpx = clinopyroxene; opx = orthopyroxene
gar = garnet; grph = graphite; hb = bornblende;
kspar = potash feldspar; opx = orthopyrnxene; qtz = quartz;
plag = plagioclase; redBio = red biotite; sill = sillimanite

The large numbers indicate the stops at which the unit can be seen.

Figure 5. Interpretation of Facies Relationships within the "Ned Mountain Formation"





This intervening unit, "Ned Mountain Formation", contains a wide variety of lithologies. It appears to be the product of a highly variable and rapidly evolving depositional environment. Significant facies changes exist within the Peach Lake area (see Figs. 4 & 5). The "Ned Mountain" has been provisionally subdivided into four members, each of which is lithically variable.

<u>"Ned Mountain Formation", Basal Member</u>. The basal unit ($Z\epsilon n_b$) contains several rock types, some quartz-rich, others quartz-poor. One common lithology is a sulfidic, rusty-weathering, graphite-bearing, fissile, biotite-rich granofels (Stop 1A). Plagioclase is the dominant feldspar in this lithology, though K-feldspar is locally important. The biotite-rich granofelses sometimes contain minor cummingtonite (or orthopyroxene at higher grade). Graphite has been found with cummingtonite-garnet or orthopyroxene-garnet in several localities. Cummingtonite-bearing amphibolites are also found locally in the Basal Member of the "Ned Mountain Formation"; they contain 5%-10% quartz, in addition to plagioclase-hornblende-biotite-cummingtonite. The quartz content of cummingtonite-bearing granofelses is variable, generally ranging from 20% to 60%. The graphitic, cummingtonite-bearing granofelses are tentatively interpreted as organic-rich lacustrine sediments derived from weathered basalts.

The remaining lithologies of the Basal Member are K-feldspar-rich granofels, quartzite with minor garnet±K-feldspar±plagioclase, and distinctive, aluminous rocks low in quartz but rich in K-feldspar. The K-feldspar-rich granofelses contain quartz-K-feldspar-biotite-plagioclase. They are similar to the K-feldspar-rich rocks of the overlying Volcaniclastic Member of the "Ned Mountain Formation", and are interpreted as somewhat weathered, rhyolitic ash. The low-quartz rocks contain assemblages of K-feldspar-garnet-biotite-sillimanite±quartz±plagioclase. The low silica, high alumina, and high potash content of these rocks lead them to be interpreted as metasaprolites derived from and/or mixed with K-feldspar-rich volcaniclastic material. Some metasaprolites contain magnetite, suggesting deposition in a relatively oxidizing (subaerial?) environment; others, which contain graphite and iron sulfides, may consist of material washed into nearby, stagnant lakes.

To summarize, the Basal Member is interpreted as the product of variably weathered rhyolitic and basaltic rocks. Some units, particularly the amphibolites, may have undergone little alteration at the surface; others, like the quartz-rich cummingtonite-bearing lithologies and quartz-feldspar granofels, may have been arenaceous sediments derived from volcanic rocks. The sandy sediments may have been deposited in lakes, where they and saprolitic material washed in from adjacent highlands could mix with organic material. At one locality a thin-bedded quartzite (>80% quartz) contains thin laminae of biotite-garnet-cummingtonite and a thicker (>1.5 cm) quartz-free layer of K-feldspar-biotite-garnet-sillimanite-magnetite. An amphibolite adjacent to the quartzite may represent the source of the mafic volcanic debris. The rock records deposition of quartz-rich sands, saprolite, and mafic volcanic debris at close-spaced intervals.

The $Z\epsilon n_b$ member has been found overlying each of the Fordham lithologies recognized in the Peach Lake area (Figs. 2 & 4). Possible discontinuity of the unit can be explained by topographic relief on the post-Grenvillian erosional surface (Fig. 5). The $Z\epsilon n_b$ unit is generally reminiscent of the Chestnut Hill Formation of Drake (1984) in the Reading Prong of Pennsylvania. The Chestnut Hill Formation occupies the same stratigraphic position as the "Ned Mountain Formation" (Table 1): it overlies the Grenvillian basement, and underlies the Cambro-Ordovician dolomite sequence. It consists of arkose, ferruginous quartzite, metarhyolite, and metasaprolite (Drake, 1984). Drake (1969) identified the metasaprolite on the basis of its chemistry. Composed of only 42.7% SiO_2 and 0.16% CaO, in contrast to 20.2% Al_2O_3 and 7.9% K_2O, its composition is similar to that of saprolite derived from volcanic rocks of the Catoctin Formation in the Blue Ridge (Reed, 1955). The "low-quartz", K-feldspar-rich aluminous rocks of $Z\epsilon n_b$ probably have a similar origin.

Rocks immediately overlying the Basal Member of the "Ned Mountain Formation" are divided into two units: the Volcaniclastic Member ($Z\epsilon n_v$) in the south, and Wacke Member ($Z\epsilon n_w$) in the northern Peach Lake quadrangle. Lithologies in these two members seems to represent environments changing from subaerial volcaniclastic to a distal, basinal setting (Figs. 4 & 5).

<u>"Ned Mountain Formation", Volcaniclastic Member</u> ($Z\epsilon n_v$). The Volcaniclastic Member is composed of K-feldspar-plagioclase-quartz-biotite granitic gneiss, feldspathic granofelses and hornblende±clinopyroxene amphibolites. The granitic gneiss is coarse- to fine-grained, highly feldspathic (K-feldspar+plagioclase ~70%), and very leucocratic; mafic minerals, principally biotite, are usually <5% of its modal composition. Rocks of this description are interpreted as metarhyolites. The granofelses are medium- to fine-

grained, richer in quartz (ranging up to ~60%), and contain biotite-rich beds. Compositional layering, which ranges from a few mm to a couple of cm thick, is characteristic of the granofels. This layering is not compatible with a strictly igneous protolith. K-feldspar content of the granofels varies, often being ~55%. These rocks are thought to be volcaniclastic, with their relative quartz and biotite enrichment due to weathering of rhyolitic material or mixture with other sediments. The granofelses are hornblende±clinopyroxene bearing near the top of the $Z\epsilon n_v$, and contain lenses and beds of calc-silicate (Stop 6). At West Mountain, the Volcaniclastic Member grades upwards into the clinopyroxene-bearing arkoses, marble, and biotite granofelses of the Calc-silicate Member.

Granitic gneiss is thickest at Ward Pound Ridge County Park (Figs. 1 & 4) in the southern Peach Lake quadrangle. This is the "Pound Ridge Granitic Gneiss" of Scotford (1956). In the southern West Mountain area, on the other hand, the Volcaniclastic Member is dominated by granofelses and amphibolite. On the north side of West Mountain, $Z\epsilon n_v$ contains hornblende amphibolites, graphite-bearing metasaprolite (similar to that in $Z\epsilon n_b$), and quartz-rich quartz-plagioclase-biotite-K-feldspar granofelses. Rhyolitic volcaniclastic material appears to be less abundant and/or more altered in this region. A dike of one-feldspar, mesoperthitic, hypersolvus granite that lies within the microcline-rich granitic gneiss of Pound Ridge is interpreted as a feeder to a volcanic vent. Pound Ridge is inferred to have been the regional center of rhyolitic volcanic activity. North of Pound Ridge, granofelses of the Volcaniclastic Member appear to reflect increasing degrees of weathering and transportation.

The Yonkers gneiss of southern Westchester County (Table 1) is lithically similar to the "Pound Ridge granitic gneiss" and occupies a similar stratigraphic position. Hall (1979) noted truncation of units of the Fordham under the Yonkers, and suggested that the Yonkers might represent felsic volcanics or arkoses lying over an unconformity. Geochronological studies on both granitic gneisses support their assignment to Late Precambrian-Early Cambrian time. The Pound Ridge has a Rb-Sr whole-rock age of 579±21 Ma (Mose and Hayes, 1975)[1]. Rb-Sr whole-rock studies on the Yonkers Granitic Gneiss have yielded ages of 563±30 Ma (Long, 1969) and 530±43 Ma (Mose, 1981). Zircon from the Yonkers gives a nearly concordant U-Pb age around 515 Ma (Grauert and Hall, 1974). Grauert and Hall (1974) found Middle Proterozoic inherited components in zircons from the Fordham, Manhattan, and Lowerre. Only zircons from the Yonkers were free of Grenvillian contamination. This indicates an igneous origin for the Yonkers Gneiss during Late Precambrian to Cambrian time.

"Ned Mountain Formation", Calc-silicate Member. The Volcaniclastic Member grades upwards into the Calc-silicate Member of the "Ned Mountain Formation" (Stops 5 & 7) ($Z\epsilon n_c$, Fig. 4). This unit contains a mostly thin-bedded (0.5 to 10 cm) sequence of clinopyroxene-bearing granofelses and calc-silicates, calcareous marble, K-feldspar-rich granofels, quartz-plagioclase-biotite±K-feldspar±garnet±graphite granofelses, occasional hornblende and cummingtonite

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1. All Rb-Sr ages have been recalculated using a decay constant of 1.42×10^{-11} yr^{-1}.

amphibolites, a few saprolitic (low-quartz, K-feldspar-rich, sillimanite-bearing) layers, and minor quartz-plagioclase-biotite-garnet-sillimanite±K-feldspar metawackes. The K-feldspar-rich granofelses are indistinguishable from the granofelses of the underlying Volcaniclastic Member, and are thought to have a similar origin. Lithologies seem to grade continuously from rhyolitic towards calc-silicate (K-feldspar-rich, variable carbonate and clinopyroxene content), from rhyolite towards wacke (with increasing quartz and biotite, decreasing K-feldspar, ± graphite, garnet, and sillimanite) and from rhyolite towards saprolite. Thick calcite, low-Mg marbles (Fig. 4) lie near the top of this unit in the West Mountain area. Tourmaline is an occasional accessory in the granofelses, calc-silicates, and low-quartz rocks of $Z\epsilon n_c$, and is rare in every other stratigraphic unit in the region.

The Calc-silicate Member is thickest at West Mountain (possibly over 300 meters); it thins dramatically to the north and appears to be less than 30 meters thick where it overlies the Wacke Member ($Z\epsilon n_w$) of the "Ned Mountain Formation" (Fig. 4). In the southernmost area, at Pound Ridge, $Z\epsilon n_c$ seems to contains only the K-feldspar-rich and quartz-plagioclase-biotite-garnet-graphite granofelses, amphibolite, and metawacke lithologies. Calc-silicate there appears limited to clinopyroxene-rich amphibolites. In the central area, at West Mountain, all lithologies are present and most are intimately interbedded. In the northern Peach Lake quadrangle, the thin $Z\epsilon n_c$ member contains K-feldspar-rich quartzite interlayered with clinopyroxene-K-feldspar calc-silicate. Rocks identical to the sillimanite-bearing metawacke, quartz-plagioclase-biotite-garnet granofels and amphibolite seen in $Z\epsilon n_c$ at West Mountain and Pound Ridge are restricted to the underlying Wacke Member in the northern Peach Lake quadrangle. Thus, the Wacke Member of the "Ned Mountain Formation" appears to be correlative with both the Volcaniclastic and much of the Calc-silicate Members of the southern Peach Lake quadrangle (Figs. 4 & 5).

<u>"Ned Mountain Formation", Wacke Member</u> ($Z\epsilon n_w$). The Wacke Member of the "Ned Mountain Formation" (Stop 1C) overlies the Basal Member ($Z\epsilon n_b$) in the northern Peach Lake quadrangle (Figs. 2 & 4). It contains metawackes of quartz-plagioclase-K-feldspar-biotite-garnet-sillimanite, quartz-plagioclase-biotite-garnet±K-feldspar granofelses, and both hornblende and cummingtonite (+hornblende±orthopyroxene) amphibolites. It is locally graphitic near the base, where it overlies graphitic granofelses and metasaprolite of $Z\epsilon n_b$ (Stop 1C). Although it appears to be partly correlative with the Volcaniclastic Member of the "Ned Mountain", it seems to contain relatively little rhyolitic material. K-feldspar is minor or absent in most granofels layers, though some beds are K-feldspar rich. This is consistent with the northward decrease in unweathered rhyolitic material already observed within the Volcaniclastic Member.

While the Calc-silicate Member and the Wacke Member contain similar metawackes and granofelses, proportions of these rock types are very different. Granofelses are far more abundant in the Calc-silicate Member than metawackes, while the reverse is true in the Wacke Member. The two units appear to be shallow- and deeper-water equivalents: the shallow-water to subaerial $Z\epsilon n_c$ contains interbedded calc-silicates, granofels, and saprolite, while the $Z\epsilon n_w$ consists of basin-type sediments. Only after a deep basin had been filled, perhaps, could a shallow-water environment be established in the northern Peach Lake quadrangle and the thin calc-silicates of $Z\epsilon n_c$ be deposited. The much higher volcanic input in the "Ned Mountain Formation" of the southern Peach Lake area may account for the longer period of shallow-water deposition there.

The Manhattan Schist and the Wacke Member may in fact be similar in age and analogous in origin. Hall (1968, 1979) long held that the Manhattan (units B and C) was a Cambrian unit allochthonously emplaced over Ordovician Walloomsac Formation (Hall's Manhattan A). The Manhattan has been correlated with the Cambrian Hoosac Formation (Rodgers, 1985), and the Hoosac, in turn, has been interpreted as rift-related deposits over Grenville basement (Stanley and Ratcliffe, 1985). The Wacke Member appears to have had a similar origin (Figs. 4 & 5). The Wacke Member, like portions of the Hoosac and unlike the Manhattan, is autochthonous: it is still in place over the basement on which it (and the Basal Member) were originally deposited.

STATUS OF THE LOWERRE QUARTZITE

The "Lowerre Quartzite" is the only unit of the Manhattan Prong previously recognized between Fordham Gneiss and Inwood Marble. Most lithologies of the "Ned Mountain", (e.g., the amphibolites, calc-silicates, quartz-plagioclase-biotite-garnet granofelses, metasaprolites, and metawackes), have all been previously assigned to either Fordham, Manhattan or Inwood (Prucha and others, 1968). Usually only the feldspathic sandstones or rare quartzite have been designated "Lowerre", although Hall and his students (for example, Jackson and Hall, 1982) included some schistose rocks. The sporadic nature, K-feldspar richness, and conformity of the "Lowerre" quartzites with underlying "Fordham" lithologies have been sources of confusion, and caused many workers in past years to deny the existence of a post-Fordham, pre-Inwood stratigraphic unit in the Manhattan Prong. Fluhr and Bird (1939) found that a pebbly quartzite seemed to grade into a sheared granite. Berkey (1907), Prucha (1956) and Scotford (1956) argued that the "Lowerre" was merely a sheared or quartz-rich portion of the Fordham. Prucha (1956), working in the Peach Lake region, concluded that the Fordham and Inwood were stratigraphically continuous. On the other hand, many workers (e.g., Norton and Giese, 1957; Norton, 1959; Hall, 1968; Alavi, 1976) found occasional outcrops of quartz-rich sandstone below the Inwood Marble.

Recognition that, at least in the Peach Lake area, the "Lowerre Quartzite" is not a basal sandstone unconformably overlying Grenvillian Fordham allows a resolution of this controversy. The diversity of rock types that belong to the "Ned Mountain Formation", and the fact that many at least superficially resemble other stratigraphic units, suggests that the unit very likely has gone unrecognized in other parts of the Manhattan Prong.

Placing the rocks of the "Lowerre Quartzite" within the expanded context of the "Ned Mountain Formation" makes their K-feldspar richness understandable. Previous workers have remarked on the enigmatic abundance of K-feldspar in correlative Cambrian sandstones like the Cheshire and Poughquag Quartzites. In the Lincoln area of Vermont, Tauvers (1982) found that while older rift deposits (Pinnacle Formation) contain abundant plagioclase and appear to be derived from local basement, the overlying Cheshire Quartzite contains K-feldspar. He suggested a change to a "cratonal" source for these sediments. Aaron (1969) noted that the (unmetamorphosed) Hardyston Quartzite of New Jersey-Pennsylvania is rich in fresh K-feldspar and surprisingly poor in plagioclase. Plagioclase predominates over K-feldspar in Grenville gneisses of the Reading Prong (Aaron, 1969), as it does in the Manhattan Prong. The erosion of K-feldspar rich (crystal?) tuffs and associated felsic volcanic edifices of the Volcaniclastic Member of the "Ned Mountain Formation" provides a source for the K-feldspar so characteristic of the quartzites underlying the passive margin dolomite sequence.

INFERENCES FROM THE "NED MOUNTAIN FORMATION": RIFT-STAGE GEOLOGICAL DEVELOPMENT OF THE NORTHEASTERN MANHATTAN PRONG

The varied lithologies of the "Ned Mountain Formation" present a picture of a dynamic and rapidly evolving depositional environment. Together with the bimodal nature of its volcanic rocks, the abundance of clastic material, abruptness of facies changes, and its stratigraphic position (underlying the stable shelf deposits of the Inwood Marble), the "Ned Mountain Formation" appears to be the product of a rift environment (Fig. 5). The "Ned Mountain" resembles the late Proterozoic to Early Cambrian Chilhowee Group of Virginia in containing both rhyolitic volcaniclastics and basalt flows (Simpson and Eriksson, 1989). But the "Ned Mountain", like the Hoosac and Pinnacle Formations of New England, (Stanley and Ratcliffe, 1985) also contains marble, particularly near the top of the rift sequence.

The post-Grenvillian depositional history of the northern Manhattan Prong can be summarized this way:

1. As rifting is initiated, a Late Precambrian erosional surface forms over the Fordham Gneiss. Volcanic activity begins. Thin rhyolitic and basaltic beds are laid down. Aluminous saprolite develops in higher elevations; weathered volcanic debris and saprolite is washed into stagnant lakes ($Z\epsilon n_b$).

2. A relatively deep basin develops in the northern Peach Lake region. It receives immature sediments from eroded Fordham basement and basaltic flows. An acid volcanic center is established at Pound Ridge, resulting in thick rhyolite deposits. In the intervening area, shallow-water weathered volcaniclastic sediments are deposited ($Z\epsilon n_v$ and $Z\epsilon n_w$).

3. While deep-water deposition continues in the northern Peach Lake quadrangle, in the south episodic shallow-water deposition and subaerial exposure occurs. An interbedded sequence consisting of limestone, arkose, volcanic ash, and minor wacke ($Z\epsilon n_c$) is laid down on top of $Z\epsilon n_v$. Eventually, the deep-water basin is filled, and the shallow-water calc-silicates and arkoses of $Z\epsilon n_c$ overlie $Z\epsilon n_w$ in the north as well.

4. A stable shelf environment is established. Algal dolomites (which, when metamorphosed, produce magnesian marbles) are (conformably?) deposited over the $Z\epsilon n_c$ member.

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ROAD LOG FOR THE STRATIGRAPHY OF THE NORTHEASTERN MANHATTAN PRONG, PEACH LAKE QUADRANGLE, NEW YORK - CONNECTICUT

The Starting Point for the field trip is the Rest Area on east-bound Route I-84 located 4 miles east of the intersection of I-84 and I-684 (Fig. 1). If you are approaching from the east (on the west-bound lane of I-84), then you need to overshoot and to do a U-turn at the Sawmill Road exit in order to reach the Rest Area.

The locations of the stops are shown on several maps (Figs. 1, 2, 6, & 7) and the schematic cross section (Fig. 4).

Cumulative Mileage	Mileage from last point	Route Description
0.0	0.0	Start from Rest Area. Log starts from the entrance to the service road. Take service road onto east-bound route I-84.
2.3	2.3	Turn off right onto Route 7 south.
2.7	0.4	Turn off right to Park Ave. exit.
3.1	0.4	Turn left (southwest) at traffic light onto Backus Ave.
3.8	0.7	Traffic light. Continue straight.
4.2	0.4	Stop sign. Turn right (west) onto Miry Brook Road.
4.7	0.5	Complex intersection. Go straight ahead -- bearing slightly LEFT onto George Washington Highway.
6.0	1.3	Stop sign. Turn left (south) onto continuation of George Washington Highway.
6.9	0.9	Bear left at Y-junction onto Old Stagecoach Road. Road becomes Bennetts Farm Road. Sharp turns.
7.7	0.8	Bear left (east) on Bennetts Farm Road.
7.9	0.2	Turn left (north) on INCONSPICUOUS little road called Sky Top Drive into a housing development.
8.0	0.1	Turn left at Y on a nameless road. Continue past "No Outlet" sign.
8.1	0.1	Turn right (east) onto North Shore Drive.
8.3	0.2	Park at the end of the road on the north shore of Lake Windwing.

STOP 1. STRATIGRAPHIC SECTION FROM FORDHAM GNEISS, THROUGH "NED MOUNTAIN FORMATION" ($Z\epsilon n_b$, $Z\epsilon n_w$, AND $Z\epsilon n_c$), AND INTO INWOOD MARBLE

This stop entails a 1.6 kilometer hike, only part of it on trails, and some fairly steep slopes. There are several stations, so the stop will require some hours.

Traverse northwards (Fig. 6) up the hill, over the crest, and down onto the east-west trail (passing metawackes and granofelses of the $Z\epsilon n_w$ member as you climb). Turn and go east on the trail until you encounter a T-junction with a north-south trail. Follow this trail north for 0.6 km to a bridge over a northeast flowing steam. Turn west. Go 30 meters off the trail to the base of the hill slope. Station 1A.

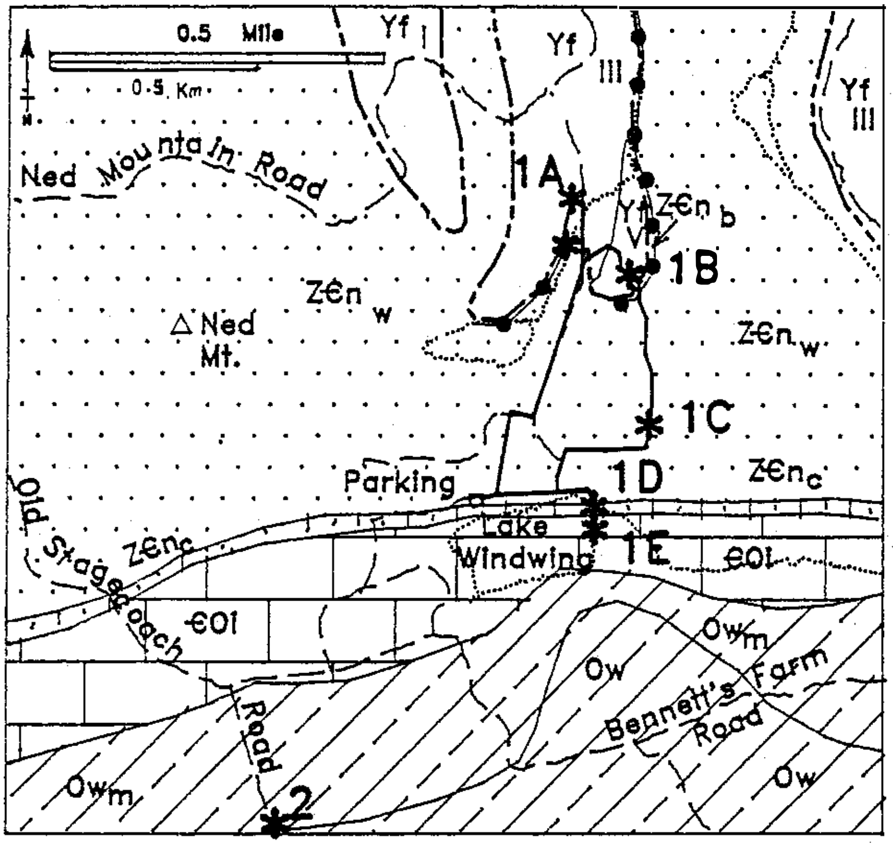

Figure 6.

Stops 1 and 2. Traverse is shown by heavy line. Roads are dashed. Streams and lakes are dotted. Other symbols are as in Figure 2.

<u>Station 1A.</u> <u>Fordham Gneiss and overlying Basal Member of the "Ned Mountain Formation"</u>

The Yf_{III} member of the Fordham Gneiss is exposed on this hill. It consists of well-layered and compositionally heterogeneous coarse-grained, orthopyroxene-bearing gneisses. The scale of compositional layering ranges from about 2 cm to a meter. Some layers are highly leucocratic, quartz-plagioclase-orthopyroxene-biotite-clinopyroxene±K-feldspar; dark layers are generally hornblende-rich. On close inspection, pitted and rust-stained orthopyroxenes can be seen on the weathered surface of the outcrop.

Here the Fordham largely retains its characteristic, poorly foliated Grenvillian texture. Locally, lineation defined by mafic minerals is evident. The effects of Taconian metamorphism can be seen in thin section: many orthopyroxenes are surrounded and embayed by cummingtonite and/or hornblende. Some pyroxene-bearing granite is present along the hill slope.

A number of small shear zones cut discontinuously across the gneisses. These seem to be related to an episode of post-Taconian shearing seen elsewhere in the northern Manhattan Prong. Age constraints from these areas suggest a Carboniferous age for the post-Taconian deformation (Brock and others, 1985; Brock and Brock, 1985 a & b).

After examining the Fordham, follow south along the cliff base for 60 to 100 meters to see the Basal Member of the "Ned Mountain Formation".

Here the Basal Member of the "Ned Mountain Formation" is a fissile, rusty-staining, graphitic granofels. The rocks are medium- to fine-grained, and contain quartz-K-feldspar-plagioclase-biotite-graphite±garnet. The tough Fordham gneisses form the crown of the hill, and the softer granofelses lie along its southern flank.

Biotite in the graphitic granofelses is concentrated into close-spaced (though discontinuous) streaks; as a result, the rocks tend to break into thin folia. This habit, and their K-feldspar-richness, distinguish the "Ned Mountain" granofelses from similar rocks in the Fordham and the Walloomsac. The graphitic, rusty-weathering granofels of the Fordham unit Yf_V contains no K-feldspar and little biotite. Walloomsac granofelses are typically quartz-plagioclase-biotite-graphite rocks containing little if any K-feldspar and breaking in a slabby fashion.

Fordham lithologies are truncated against the "Ned Mountain Formation". Where $Z\epsilon n_b$ is first seen, it is in contact with heterogeneous, layered gneisses; following the unconformity ~30 meters southwest, largely leucocratic, homogeneous Fordham gneisses are against it; after another hundred meters large amphibolites lie at the unconformity; still further southwest, leuco- to mesocratic gneisses are in contact with the graphitic granofelses of $Z\epsilon n_b$. Fordham gneisses adjacent to the unconformity tend to show the effects of Taconian metamorphism and deformation more than they do elsewhere. They are richer in biotite and have a better-developed foliation.

After finishing with this station, cross back to the east over the old dam. (The dam stands above the stream and swamp to the south of the hill.) Retrace your steps along the trail southwards for approximately 100 meters to the junction with a branch trail that climbs up to the east. Follow it east for ~30 meters (Fig. 6). Then turn south off it to the outcrops of Station 1B.

Station 1B. Fordham Gneiss, Unit Yf_V

Here Unit Yf_V of the Fordham consists of well-layered, medium to coarse-grained, melanocratic to leucocratic gneisses. Plagioclase-quartz-hornblende-biotite-pyroxene±cummingtonite gneisses and amphibolite are present in the first outcrop. Rocks still showing the distinctive assemblage of YF_V, gar-net-clinopyroxene-plagioclase±orthopyroxene, can be seen in the outcrop further from the path. Partial replacement of this assemblage by hornblende (and of orthopyroxene by cummingtonite) is thought to have occurred during Taconian metamorphism. The assemblage has not been found in any rocks of post-Grenville age.

Next, proceed southeastwards through the bush looking for outcrops. When you reach the north-south trail marked with yellow dots, follow it southwards to the crest of the scarp overlooking Lake Windwing. En route you will see a good section through the metawackes, pelites, hornblende amphibolites and cummingtonite amphibolites of the $Z\epsilon n_w$ member of the "Ned Mountain Formation".

Station 1C. Wacke Member of the "Ned Mountain Formation"

As you walk, you will first pass outcrops of coarse-grained K-feldspar-plagioclase-sillimanite-garnet-biotite-quartz±graphite schist of the Basal Member. The rocks are garnet-rich and contain prominent sillimanite trains; the contrast between the sillimanite-biotite-garnet and feldspathic segregations produces a pin-striped appearance. Metawackes of the Wacke Member, slightly higher in the section, have a similar pin-striped appearance but contain more quartz and less K-feldspar than the Basal Member.

A number of medium-grained granofelsic layers are present along the trail. These range from about 10 to 60 cm thick, and contain quartz-plagioclase-biotite-garnet±K-feldspar. Fine layering is often visible on close inspection.

Near the scarp several amphibolites are interlayered with the metawackes and granofelses. These include slightly greyish hornblende-cummingtonite-orthopyroxene bearing amphibolites and blacker orthopyroxene-free hornblende amphibolites.

The thick "pinstriped" metawackes, thin granofelses and hornblende amphibolites of the "Ned Mountain Formation" Wacke Member at this locality make it strongly resemble the Manhattan Schist. But in addition to the distinctive stratigraphic position of the "Ned Mountain", it can be distinguished from the Manhattan in several ways: a) while granofelses of the Manhattan contain little or no K-feldspar, those in the "Ned Mountain" in some places have substantial amounts; b) the Wacke Member is locally graphitic near its base, while the Manhattan contains no graphite; c) cummingtonite-bearing amphibolites have not yet been found in the Manhattan Schist; and d) bedding of the granofelsic layers in the Wacke Member thins towards the north, while no geographical variation is apparent in Manhattan Schist. At this locality, granofelses in the Wacke Member have similar thicknesses to those in Manhattan Schist, mostly between 5 cm and 1 meter. But about three kilometers to the north, granofels and hornblende-orthopyroxene amphibolite beds are both typically 1.5 cm to 8 cm thick, and 4.5 km to the north, many beds are ~1 cm thick. This thinning is interpreted to reflect a southerly provenance for the sediments of the Wacke Member (Figs. 4 and 5).

Getting down the scarp is somewhat tricky. Good exposures of migmatized metawackes of $Z\epsilon n_w$ are present in the cliffs below, but for safety, it is better to follow the trail along the crest until it turns down. From the bottom of the slope, go to the spillway of Lake Windwing (at the lake's eastern end).

<u>Station 1D.</u> <u>Calc-silicates, Arkosic Sandstones, and Pelite of the "Ned Mountain Formation, $Z\epsilon n_c$ member</u>

K-feldspar-rich quartzite and clinopyroxene-K-feldspar bearing calc-silicate are interbedded on the south side of the spillway. Beds range from 5 to 50 cm thick. A biotite-rich pelite is present on either side of the spillway. This outcrop displays the thin northern variant of the Calc-silicate Member, which overlies the Wacke Member of the "Ned Mountain Formation". Shearing and pegmatite intrusion thought to be of Carboniferous age has resulted in local development of muscovite in the pelites and tremolite in the calc-silicate.

Go 15 meters south. Scattered outcrops of white marble: Station 1E (Fig. 6).

Station 1E. Inwood Marble

The Inwood here consists of well-bedded, white and buff weathering calcitic marble. The outcrop at the lakeside contains a bed of clean quartzite. Buff-weathering marbles contain the assemblage calcite-forsterite-dolomite-phlogopite±periclase whereas white marbles contain calcite-dolomite-phlogopite±periclase. Although quite rich in calcite, these are magnesian marbles typical of the Inwood.

The Inwood contains more carbonate and less silica and K_2O than does the Calc-Silicate Member. It is also much less diverse lithologically. Marbles of the Calc-silicate Member of "Ned Mountain Formation" and Inwood Marble can be distinguished by their mineral assemblages (Fig. 3). Forsterite and phlogopite (occasionally spinel), ± clinopyroxene are typical constituents of the more magnesian, originally dolomitic Inwood. Clinopyroxene and K-feldspar are characteristic of both the low-magnesium Calc-silicate Member and Walloomsac marble.

Walk west 0.3 km on the trail along the north shore of Lake Windwing to the starting point of the traverse.

8.3	0.0	Leave Lake Windwing. Retrace route.
8.5	0.2	Stop sign. Bear right.
8.7	0.2	Turn right (west) onto Bennetts Farm Road.
8.8	0.1	Turn left (south) onto Old Stagecoach Road.
9.1	0.3	Pull off to right and park.

STOP 2. OUTCROPS OF WALLOOMSAC SCHIST AND MARBLE

Here (Fig. 6) we see Walloomsac marble that was unconformably laid down over the Inwood (Table 1). Small outcrops of Walloomsac marble lie on either side of the road. These consist of well-bedded, rusty-stained, graphitic marble; they contain calcite-quartz-plagioclase-K-feldspar-clinopyroxene ±phlogopite. Some coarse, late tremolite is present. Walloomsac marbles somewhat resemble calc-silicate rocks of the "Ned Mountain Formation", but differ from "Ned Mountain" in their distinctive rusty staining and in the abundance of graphite.

A few meters south on the west side of the road is a small outcrop of Walloomsac Schist. The Walloomsac here consists of a rusty-weathering, graphitic, quartz-plagioclase-biotite-garnet schistose granofels. K-feldspar-rich granofelses, typical of the Calc-silicate Member of the "Ned Mountain Formation", are never found in the Walloomsac.

9.1	0.0	Continue south. The road swings round to the east and becomes Aspen Ledges.
9.8	0.7	Park on Fox Drive on right. Walk east round the curve to see road cut of Walloomsac Schist.

STOP 3. CORDIERITE-BEARING WALLOOMSAC SCHIST

Slabby-breaking, rusty-staining, quartz-plagioclase-garnet-K-feldspar-sillimanite-cordierite graphitic schistose gneiss of the Walloomsac Schist outcrops here. More pelitic than at previous stop, the Walloomsac contains cordierite-sillimanite-garnet assemblages overprinted by K-feldspar-sillimanite-biotite. In thin section, garnets are rimmed by sillimanite against cordierite. The cordierites are rust-stained and obviously relict.

9.8	0.0	Continue east on Aspen Ledges.
10.1	0.3	Turn right (south) onto Bob Lane. The road swings to the east. (In Hartland Fm.)
10.4	0.3	Pull off on left beside road cut.

STOP 4. HARTLAND FORMATION AMPHIBOLITE AND GRANOFELS

Between Stop 3 and here, you have crossed Cameron's Line, the boundary between the rocks of the Manhattan Prong and the Hartland terrane (Fig. 2). Manhattan Schist, which elsewhere structurally overlies the Walloomsac, was cut out along this part of Cameron's Line. The Hartland was emplaced over the rocks of the Manhattan Prong early in the Taconian orogeny, and was later deformed along with them. At this outcrop, quartz-plagioclase-biotite granofels of the Hartland Formation is interlayered with hornblende-orthopyroxene$\pm$clinopyroxene amphibolite. Bedding ranges from 5 cm to 50 cm in thickness. The Hartland is distinguished from the Manhattan by its abundant amphibolites, its non-rusty, relatively garnet-poor granofelses, and its rhythmically interbedded schist and granofels sequences (meta-turbidites) not seen here.

A granite showing graphic intergrowth of quartz and K-feldspar and large books of biotite is also present at this outcrop. It contains no primary muscovite and is undeformed. It is thought to be late Taconian in age.

This completes a section from the Fordham Gneiss to the structurally highest rocks in the northern Peach Lake quadrangle. Next, we will examine a section in the southern part of the quadrangle.

10.4	0.0	Continuing east
10.5	0.1	Turn right (south) onto Knollwood Drive.
10.7	0.2	Forced right/bear right onto Twixt Hills Road.
10.9	0.2	Turn left (south) onto Pierrepont Drive. Down around hairpin bends. Road swings west past Lake Naraneka and is here called Barlow Mountain Road.
11.5	0.6	Turn left (south) at stop sign. Still on Barlow Mt. Road. (Right branch is Ledges Road.)
11.9	0.4	Turn right (west) at T-Junction. Still on Barlow Mt. Road. (Left branch is North Street.)
12.2	0.3	Turn left (south) onto Route 116. Tackora Trail.
13.8	1.6	Turn right (west) onto Saw Mill Hill Road.
14.4	0.6	Bear right (west) onto Rampoo Road.
14.8	0.4	Bear right (west) onto Oreneca Road.
15.0	0.2	Turn right (northeast) onto Sharp Hill Road.
15.2	0.2	Park at end of road.

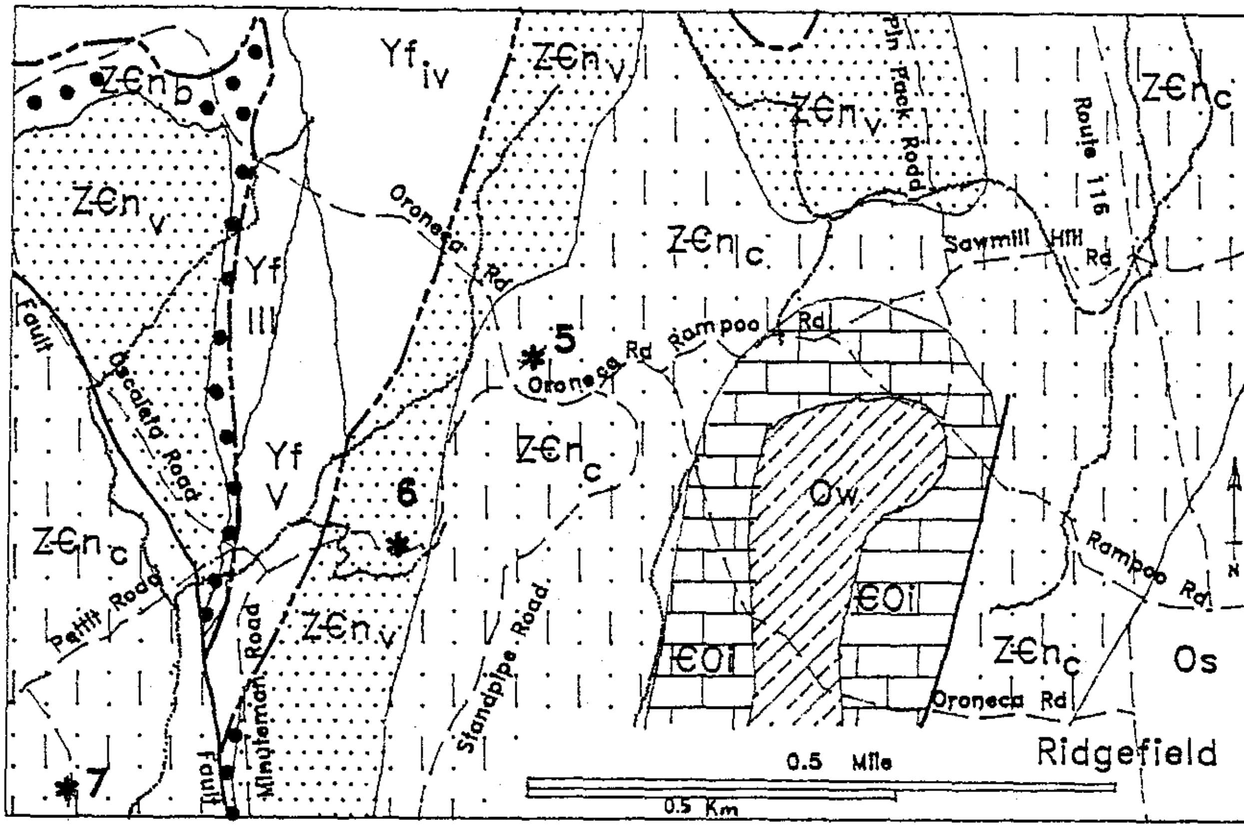

Figure 7. Stops 5 to 7. Streams are dotted; roads are dashed. Other symbols as in Figure 2.

STOP 5. CALC-SILICATE MEMBER OF THE "NED MOUNTAIN FORMATION", SOUTHERN SECTION

These outcrops are on private property. If you come by yourself, ask permission from the owners before looking. Outcrops are located at the end of the road and on the road's north side, by the driveway going north up the hill. Beware of poison ivy.

Inwood Marble lies in the bottom of the valley you have just driven across (Fig. 7). You are on the east limb of an F_2 antiform, and will continue going down stratigraphically as you proceed west. Here in the southern portion of the Peach Lake quadrangle (Fig. 1), the Calc-silicate Member of the "Ned Mountain Formation" is much thicker than at Lake Windwing. At this locality, white-and-green clinopyroxene-bearing calc-silicates and marbles are interlayered with greyish, laminated, quartz-K-feldspar-plagioclase-biotite granofelses and hornblende amphibolites. Retrogressive tremolite and epidote is locally present in calc-silicate layers. The characteristic ribbing on the weathered surface of the calc-silicates reflects variations in bed-by-bed carbonate content. Some granofelsic layers are clinopyroxene-hornblende-bearing, but rich in K-feldspar whereas others are plagioclase- and biotite-rich. Modal K-feldspar in the granofelses varies from 20% to 70%, while quartz ranges up to about 60%. The granofelses show a faintly developed linear fabric.

NOTES

STRUCTURE AND PETROLOGY OF THE PRECAMBRIAN ALLOCHTHON, AUTOCHTHON AND PALEOZOIC SEDIMENTS OF THE MONROE AREA, NEW YORK[1]

HOWARD W. JAFFE
University of Massachusetts
Amherst, Massachusetts

ELIZABETH B. JAFFE
Amherst, Massachusetts

INTRODUCTION

The area covered by this trip lies in the northern part of the Monroe 7.5 minute quadrangle, New York, and consists of a folded and faulted complex of autochthonous Precambrian gneisses, Lower Cambrian through Middle Devonian sediments and allochthonous Precambrian gneisses. Geologic maps covering the trip area have been published by Ries (1897), Fisher, et. al. (1961), and Jaffe and Jaffe (1962 and 1973). Unpublished maps prepared by Colony and by Kothe (Ph.D. thesis, Cornell Univ.) undoubtedly contain valuable information but are not available for study. Recent workers in adjacent areas include Dodd (1965), Helenek (1971) and Frimpter (1967), all in the Precambrian autochthon, and Boucot (1959) and Southard (1960) in the stratigraphy and paleontology of the Paleozoic sediments. The work of Colony (1933), largely unpublished, is impressive.

An attempt to unravel the complex structural history of the region has suggested the following sequence of events:

1) Deposition in the Precambrian of a series of calcareous, siliceous, and pelitic sediments and basic volcanics of the flysch facies in a eugeosynclinal; folding and metamorphism involving complete recrystallization to granulite facies gneiss assemblages which characterize the Precambrian autochthon (Jaffe and Jaffe, 1962; Dodd, 1965). Foliation in the autochthon trends northeast and is generally vertical or dips steeply to the east, with overturning west; fold axes most often plunge gently northeast. The metamorphic foliation appears essentially Precambrian in origin. The present Precambrian allochton was deposited and recrystallized at about the same time as the autochthon; recrystallization took place about 1100 million years ago. The sediments of the allochthon are graphitic, siliceous, calcareous and pelitic and appear to represent a clastic wedge (molasse) deposited in a reducing environment, possibly to the east. Graphitic gneisses are absent from the autochthon of the Monroe quadrangle, although they do occur in the Popolopen Lake quadrangle to the east (Dodd, 1965).

2) After extensive erosion, the Lower Cambrian Poughquag conglomerate, arkose, and quartzite were deposited unconformably on the Precambrian autochthon. As in most of the Hudson Highlands, the Poughquag has been sporadically preserved and here occurs only in the buttressed area northeast of Block 2 (Fig. 2). The Poughquag dips gently to the north.

[1]This article is reproduced (with minor editorial changes) from the New York State Geological Association Guidebooks of 1962 and 1967. The log has been modified to reflect road changes in the last 22 years.

3) Deposition of the Cambro-Ordovician Wappinger Formation, which in this area consists entirely of dolomite. In Block 2 (Fig. 2), it also dips gently to the north. In Block 3 it is moderately to strongly folded along a northeast trend. In Block 5 it outcrops between the Ordovician shales and the Precambrian of the Goose Pond klippe in a northeast striking band that dips west. In about this same attitude it underlies the Precambrian Museum Village klippe in Block 9.

4) Intrusion of lamprophyre dikes into the northwest-trending tension fractures in Precambrian and overlying Cambro-Ordovician rocks (Jaffe and Jaffe, 1962). These dikes have been found only in Blocks 1, 2, and 3.

5) Deposition of the Hudson River shales (Middle? Ordovician) over the entire area. This was followed by either:

   a) gentle folding, followed by erosion, or

   b) upfaulting of the Wappinger dolomite against the shales.

6) Overthrusting of the Precambrian allochton as a nappe from the east, most probably during the Taconic orogeny. Evidence for thrusting is:

   a) GEOMAGNETIC. The Precambrian autochthon everywhere shows a strong positive anomaly, whereas the klippen show none and can therefore be no more than 500 - 600 feet thick (R.W. Bromery, personal communication). The relief of Goose Pond Mountain is of this magnitude. Bull Mine Mountain, which contains magnetite deposits, does show a positive anomaly.

   b) GEOLOGIC. The Precambrian of Bull Mine Mountain is perched on Ordovician shale; the Museum Village klippe can be seen to rest on Wappinger dolomite. At the base of Goose Pond Mountain is a fault breccia; such a zone also exists on Bull Mine Mountain at the contact of the shales and the gneiss.

   Near the klippen, the Ordovician shales are always more strongly folded than elsewhere in the area, and are often overturned.

   c) PETROGRAPHIC. Quartz pebbles and grains, commonly optically continuous except for strain, have length-width ratios up to 17:1. The texture of the klippe gneisses is consistently more deformed and cataclastic than that of the Precambrian autochthon.

7) Folding of the nappe along a northeast axis, followed by erosion, leaving (a) synclinal remnant(s) along the fold axis, extending from Goose Pond Mountain to Snake Hill, near Newburgh, and beyond, to just west of Balmville. The klippen must once have formed such a single line as would be left by an eroded downfold; the alignment from Bull Mine to Snake Hill is too perfect to be a coincidence, and Goose Pond Mountain is on strike with a klippe west of Balmville, a town just north of Newburgh.

8) N 75 W cross-faulting along the Quickway (N.Y. 17) with the north block moving east. This accounts for the present displacement of the Bull Mine - Snake Hill line of klippen from the Goose Pond klippe (The Museum

Village klippe has been rotated from this line by a later fault presumably Triassic.) Apparent displacement is about one mile. The upper calcareous feldspathic quartzite member of the Hudson River shales, which outcrops on Lazy Hill, is essentially absent north of the Quickway fault. If its displacement north of the fault is the same as that of the klippen, its present position north of the fault is somewhere under the western talus slope of Schunemunk Mountain. Except for rotated Blocks 8 and 9, Silurian and younger formations line up on strike across the Quickway fault. The major lateral movement along this fault must therefore have been pre-Silurian.

9) After an erosion interval, the Shawangunk conglomerate and orthoquartzite (Lower to Middle Silurian) were deposited unconformably on the older rocks.

10) Deposition of Lower Devonian sediments.

11) Convincing evidence for the Acadian orogeny in the area is lacking. Such an event might account for pebble-stretching in the Shawangunk conglomerate, and for slight additional east-west movement along the Quickway cross-fault.

12) Deposition of the Cornwall shales and Bellvale graywackes in the Middle Devonian.

13) Appalachian folding. In the course of this folding, the relatively thin and brittle Shawangunk beds broke into detached plates which were thrust over the more yielding shales above and below. This thrusting produced the fluting parallel to the dip of the beds and the marked stretching of the Shawangunk pebbles in both the a and b fabric axes. Pebble beds in the thicker and more massive Bellvale graywackes show far less shattering and stretching of their pebbles.

14) Following the Appalachian revolution, the area was uplifted and has remained positive. During the Triassic orogeny, faulting, partly with and partly across the grain of the country, reactivated old faults and produced a complicated pattern of tilted and rotated, up-faulted and downfaulted blocks (Figs 2 and 2A):

a) Block 1 (Fig. 2) was uplifted along a N 33 E fault to form the Ramapo Mountains.

b) Block 2, which includes Poughquag quartzite and Wappinger dolomite nestled in the curve of the Precambrian massif and dipping gently north, was uplifted. Block 2 is truncated to the north by an eastward continuation of the Quickway fault, as is shown by geomagnetic evidence (R.W. Bromery, personal communication; Henderson, 1962). Block 2 is in fault contact with the younger sediments of Blocks 3 and 4.

c) Block 3 is a graben about 1500 feet wide at the south end of the map, and perhaps one or two miles wide at the north end of Block 2. In this graben, the Wappinger dolomite is moderately to steeply folded on a northeast axis.

d) Block 4 was downfaulted relative to Blocks 2 and 5, and upfaulted relative to Blocks 6, 8, and 9. Anomalous northeast dips at the north end of Block 4 may be the result of drag during faulting.

e) Block 5 was uplifted relative to Block 4, but downthrown relative to the klippen north of the Thruway fault. The Precambrian of Goose Pond Mountain outcrops from an elevation of 480 ft upward; the Museum Village klippe rests on dolomite at about 600 ft. The shale-gneiss contact on Bull Mine Mountain was observed at about 840 ft.

f) The main mass of Schunemunk Mountain, Block 6, is downthrown along northeast-southwest faults on both sides. It must also be considerably downthrown relative to its continuation to the south (Block 4), which has a moderate positive geomagnetic anomaly indicating that the basement is not very far down. The syncline's east limb is truncated to the south.

g) Block 7, in which the Esopus dips about 25° N and under which the basement anomaly is absent, probably was tilted to the north during the uplift of Block 2 and the sinking of Block 6.

h) Block 8, where the synclinal axis of Schunemunk swings to the north-south, has been rotated counterclockwise.

i) The Museum Village klippe and Bull Mine Mountain (Block 9), with the dolomite beneath, have also been rotated counterclockwise as a single block.

j) The Shawangunk sliver, Block 10 between Blocks 8 and 9 has been ground, thrust, rotated and crumpled during the rotation of the blocks.

Interrelations of the faults are highly problematic, partly owing to the masking effects of Pleistocene glaciation.

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ROAD LOG FOR STRUCTURE AND PETROLOGY OF THE PRECAMBRIAN ALLOCHTHON AND PALEOZOIC SEDIMENTS OF THE MONROE AREA, NEW YORK

CUMULATIVE MILEAGE	MILES FROM LAST POINT	ROUTE DESCRIPTION
0.0	0.0	This trip starts at intersection (traffic light) of NY 17, 32 and US 6 in Central Valley, NY. Proceed west on Niniger Road.
0.6	0.6	Niniger Road becomes Dunderberg Road.
1.9	1.3	Left at T-intersection with County Route 105
2.4	0.5	Bear right at fork to Spring Street.
3.4	1.0	Bear right onto North Main Street.

4.1	0.7	Bear right onto to Oreco Terrace at the intersection of N.Y. 208, Orange and Rockland Road, and Oreco Terrace. Park on right side of road.

STOP 1. ORECO TERRACE: BELLVALE GRAYWACKE

The outcrop is on the right side of Oreco Terrace just above the intersection of N.Y.208, Orange and Rockland Road, and Oreco Terrace. Here, the Bellvale graywacke of the Hamilton Group (Middle Devonian) is exposed in a section approximately 220 feet thick. It consists of 20 - 40 foot beds of dark blue-gray, green-gray, or gray, fine- to medium-grained (0.08 - 0.25mm average grain size) lithic arenite or graywacke, rhythmically interbedded with thin beds of dark green-gray to blue-gray shale. A representative modal composition of the graywacke follows:

Mode Of Bellvale Graywacke

Detritals:		
	quartz	20%
	oligoclase	2
	shale phyllite	44?
	siltstone chert	15
	greenstone	1
Matrix:		
	clay, sericite, chlorite, Mn oxide	15?
Metamorphic:		
	chlorite	2
	muscovite	1
		100%

Texturally, the rock consists of angular, elongated slivers of detrital quartz and predominantly phyllitic rock fragments (0.08 - 0.25mm) set in a fine matrix of sericitic muscovite, clay, and chlorite. It is often difficult to distinguish smeared-out phyllitic fragments from balled-up micaceous matrix, both of which frequently blend or flow together. Depending upon the uncertainty of the matrix content, or the classification used, the Bellvale is either a low-rank graywacke (Krynine, 1948), a subgraywacke or a lithic graywacke (Pettijohn, 1957) or a graywacke (Folk, 1954). Larger bent grains of chlorite and muscovite in the matrix, are here interpreted to have grown from fine matrix material marking the beginning of the chlorite zone - greenschist facies of regional metamorphism imprinted during the Appalachian orogeny.

In general, the Bellvale graywackes tend to show rhythmic interbedding with shale, with graded bedding low in the section and strong current crossbedding higher in the section. Occasional brachiopods are found low in the section; plant fossils are found higher up. Both features suggest gradation from marine to non-marine depositional environment. The provenance was a low-rank metamorphic or sedimentary terrane.

The shattered outcrop at Oreco Terrace is at the southwest corner of Rotated Block 8 (Fig. 2) and lies near the intersection of four directions of faulting. Attitudes of prominent structures at the outcrop are tabulated:

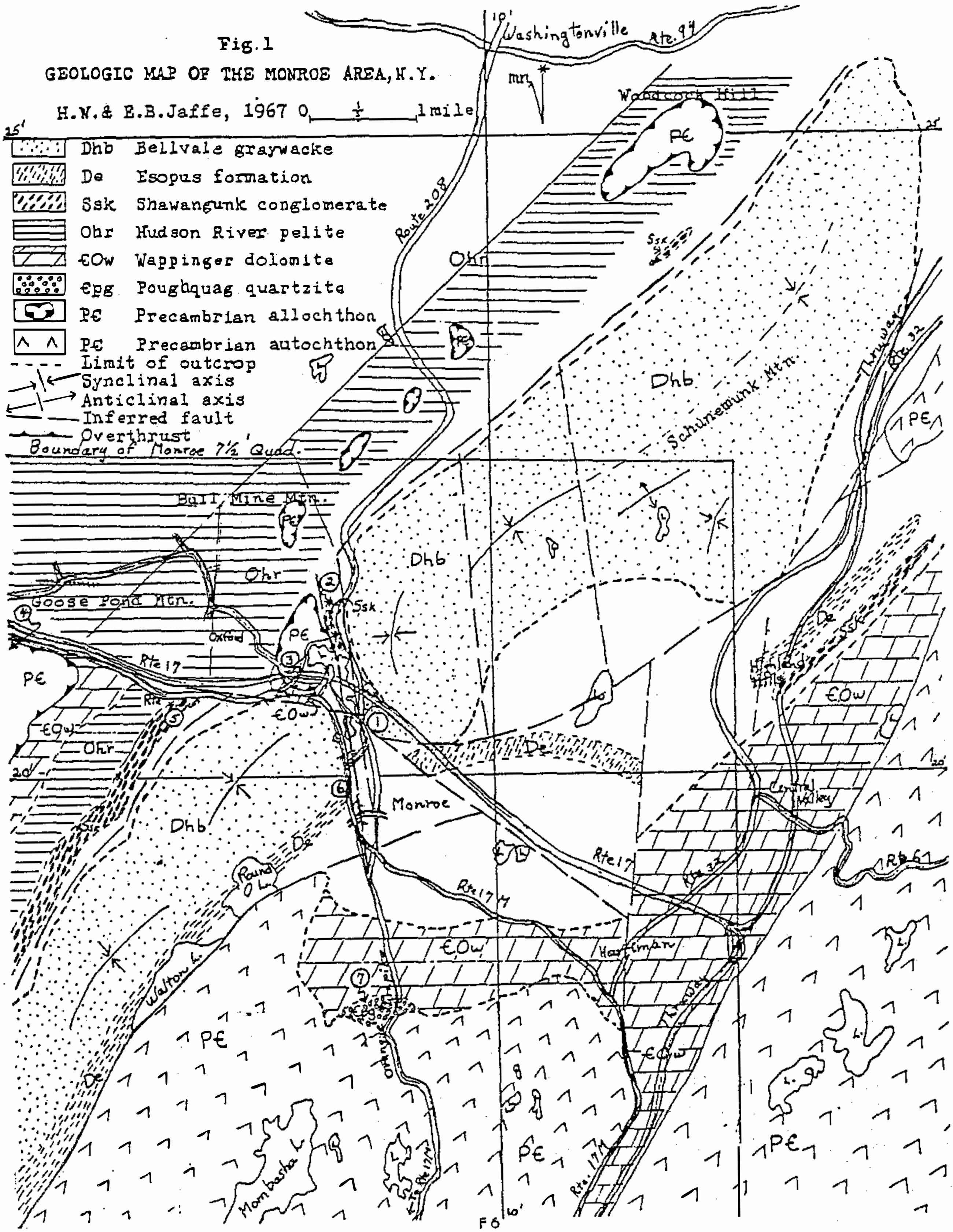

Fig. 1

GEOLOGIC MAP OF THE MONROE AREA, N.Y.

H.W. & E.B. Jaffe, 1967

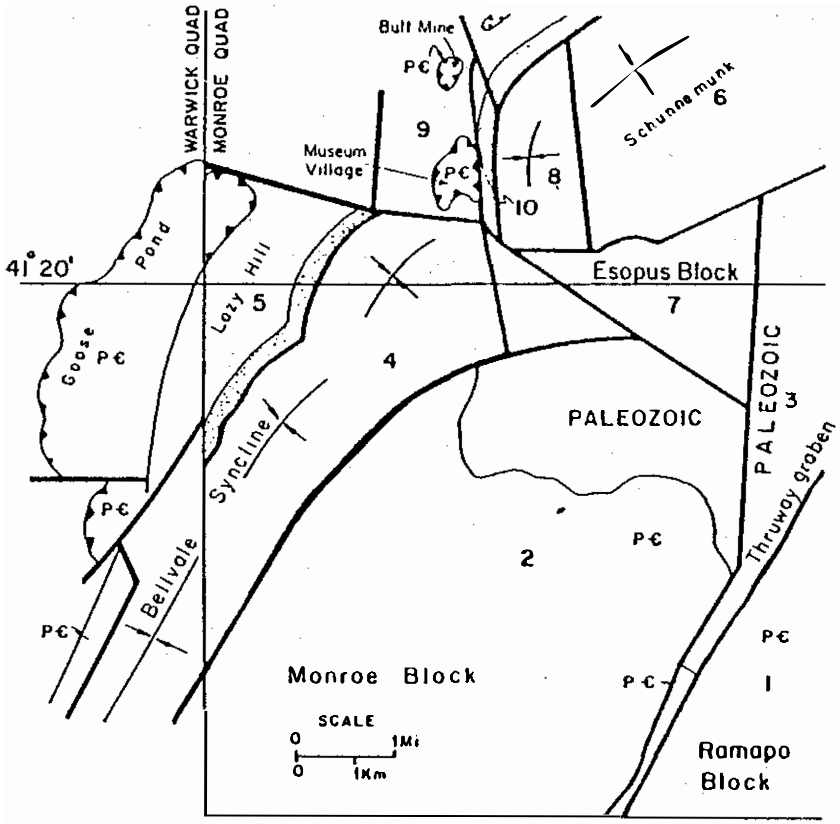

FIGURE 2. Principal fault blocks of the Monroe quadrangle, and their extension in the Warwick quadrangle (geology of the Warwick quadrangle after Offield, 1967). Stippled area is the Shawangunk.

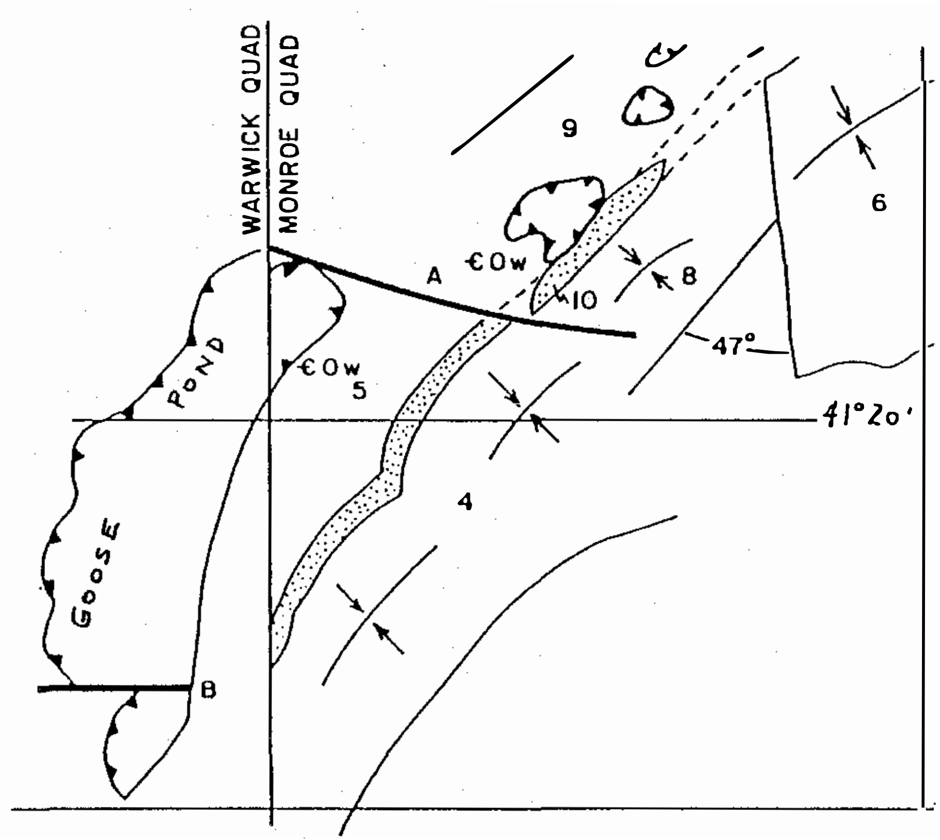

FIGURE 2A. Geologic map of parts of the Monroe and Warwick quadrangles as they may have appeared before the rotation of Blocks 8, 9, and 10 (Warwick quadrangle modified after Offield, 1967).

	Strike	Dip	Plunge
Bedding	N 67 W	30NE	----
Cleavage	N 50 E	37SE	----
Fault	N 13 E	Steep W	
Slickensides			45 N
Fault	N 38 E	Steep W	
Slickensides			60N
Fault	N 30 W	52SW	
Slickensides			39W

5.3	1.2	Drive north about 1.2 mi on N.Y. 208 and turn left at Golf Range sign onto a small road for about 300 feet. Park at the T-intersection and walk 300 feet along dirt road to the left to outcrops on the left side of the road.

STOP 2: GOLF RANGE: SHAWANGUNK CONGLOMERATE

The Shawangunk conglomerate and quartzite (Greenpond conglomerate unit) of Lower to Middle Silurian age occurs in a series of small outcrops extending northeastward along the western edge of Schunemunk Mountain syncline. At this stop the Shawangunk forms a small, relatively inconspicuous topographic knob as contrasted with its occurrence near Stop 5, where the same unit forms the spine of the steep, southeast-facing escarpment of Lazy Hill. At the present stop the Shawangunk consists of about 75% buff pebble conglomerate intercalated with 25% fine-grained green-gray quartzite.

The conglomerate consists of white pebbles of milky vein quartz (averaging 15 - 40mm in length) in a matrix of finer pebbles and grains of rounded quartz, all cemented by secondary silica and buff-orange-red ferric oxides. Occasional pebbles of white orthoclase are present as are black pebbles consisting of green tourmaline and quartz. The color of the weathered surface of the outcrop varies from pink (hematite) to yellow-brown (goethite) or black (manganese oxide dendrites) with some green contributed by lichens.

The pebbles, obviously well-rounded when deposited on the Late Ordovician erosion surface, have taken on a secondary angularity and elongation due to stretching, crowding, rotation, and slippage in the bedding planes produced during Paleozoic orogenies. Most of the pebbles show maximum elongation parallel to the fold axis (b-fabric axis). Many of the pebbles have been corrugated and a large number are cracked and sliced parallel to the b-c fabric plane. Bedding surfaces are slickensided, fluted, and warped parallel to the a-axis (down-dip). The fine-grained, gray-green interbedded quartzite is composed of quartz and minor orthoclase cemented by authigenic quartz, muscovite and chlorite, and Fe-oxides.

The elongation and shattering of the pebbles here at Lazy Hill to the south-west greatly exceeds that observed in pebble beds in Lower and Middle Devonian rocks (Connelly conglomerate of Oriskany age, and Bellvale graywacke, respectively) in this area, suggesting that the Shawangunk was involved in an additional deformation episode of possible pre-Oriskany age.

A pre-Acadian, Silurian deformation period in New York was reported by Megathlin (1939) and discussed by Kay (1942). The sporadic outcrops of Shawangunk quartzite to the north along the western limb of the Schunemunk

Mountain syncline are all heavily silicified and sheared, again much more so than quartzites of Oriskany and Bellvale age.

The conglomerate and quartzite outcrop is S-shaped, with the attitudes of the bedding and a cross-fault as follows:

	STRIKE	DIP
Bedding, Cgl., North end of hill	N 8 E	77E
Bedding, Cgl., North center of hill	N 59 W	70NE
Bedding, Cgl., Center of hill	N 27 W	50NE
Cross-fault	N 77 W	90
Bedding, Qtz., South end of hill	N 2 W	60E

The cross-fault displaces the stratigraphically higher Shawangunk quartzite member to the west, putting it on strike with the lower conglomerate member of the Shawangunk Formation. On first viewing the outcrop, the alternating pebble beds and fine feldspathic quartzites lead one to suspect that the beds are overturned. On close inspection this does not appear to be the case.

5.6	0.3	Return south on N.Y. 208 for 0.3 mi turning right on to Museum Village Road.
6.1	0.5	After 0.5 mi turn right on Old Mansion Road and drive to the far edge of the outcrop.
6.3	0.2	Park near the house on the right.

STOP 3. MUSEUM VILLAGE KLIPPE: ALLOCHTHONOUS PRECAMBRIAN LEUCOGNEISS RESTING ON CAMBRO-ORDOVICIAN WAPPINGER DOLOMITE

From the Old Mansion Road cut in the Precambrian Museum Village klippe, look south across the N 75°W-trending Quickway crossfault. The ridge due south is the Bellvale synclinal extension of Schunemunk Mountain. The next ridge to the west is Lazy Hill, held up by Shawangunk conglomerate, and offset from the Golf Range Shawangunk conglomerate of Stop 2. To the west of Lazy Hill, the next prominent ridge is the Goose Pond Precambrian klippe, formerly continuous with the Museum Village klippe and now offset about 2 miles along the Quickway cross fault.

Museum Village klippe is a thin, synclinal, saucer-shaped slice of gray-white albite-quartz-microperthite leucogneiss that has survived five or six orogenies. The rock of the allochthon were deposited as a clastic wedge (molasse) in a reducing environment, perhaps as long as 1500 million years ago; they were folded and metamorphosed to the sillimanite-almandine-orthoclase metamorphic grade about 1100 million years ago; thrust from the east in Taconic or Late Ordovician time; refolded and faulted in Taconic time; possibly again in Acadian time; refolded and faulted in Appalachian time; and finally shattered by Triassic block faulting and associated block rotation. In outcrop, the leucogneiss is heavily shattered and slickensided with lineations often running in three directions at a given place. Quartz grains and pebbles are stretched into thin corrugated tongues and sheets showing elongations of 15:1 and 20:1 parallel to the b- and a- fabric axes. Over most of the outcrop, biotite and garnet are extensively retrograded to chlorite, and abundant calcite veinlets cross at all angles.

Towards the central and western part of the cut there occur occasional thin layers rich in fresh biotite and uncommonly coarse laths (not needles) of fresh blue-gray sillimanite that have survived the complex orogenic history. As none of the Cambro-Ordovician or younger rocks in the area show any metamorphic grade higher than chlorite zone metamorphism, the sillimanite is assumed Precambrian in age, and its preservation in large fresh grains in an otherwise extensively retrograded outcrop is remarkable.

Modal analyses of samples collected along an east-west traverse across the Museum Village klippe follows:

Mineral	East Side 36	Sample Number Center 769-Si	Sample Number Center 769-N	West Side 769-W
Microperthite	49.5%	37.0%	22.5%	68.8%
Albite (An 0-5)	20.2	-	-	3.3
Andesine (An 32)	-	36.0	32.4	-
Quartz	22.2	2.0	35.1	21.5
Chlorite	5.8	-	-	2.5
Biotite	-	8.0	6.0	-
Almandine-pyrope	0.1	2.0	-	-
Graphite	-	3.0	-	-
Sericite	0.7	-	0.3	-
Sillimanite	-	12.0	3.7	-
Tourmaline	-	+	+	-
Calcite	+	-	-	3.4
Apatite	0.7	+	+	+
Zircon	0.1	-	-	+
Ilmenite	0.7	+	+	+
Pyrite	+	+	-	0.5
Sphene	+	-	-	+
	100.0%	100.0%	100.0%	100.0%

The presence of albite in the most retrograded gneisses of this and the other allochthonous Precambrian blocks studied, fairly consistently suggests that it may be a retrograded mineral after an originally more calcic plagioclase. The K-feldspar in all of these rocks is microperthite (usually microcline microperthite) and is indicative of a temperature of Precambrian metamorphism of the order of 600°C. The presence of two plagioclases, one as free grains and the other exsolved in microcline microperthite, is characteristic of many of the granitic gneisses of both the autochthon and the allochthons. In some of the allochthonous leucogneisses, the microperthite and the albite tend to occur in separate bands which may reflect original compositional differences.

Return to the cars and cautiously descend the hill to the left to the Quickway (N.Y. 17-West/U.S. 6). Beware of high speed traffic and stay close to the outcrop which parallels the highway. Walk to the extreme west edge of the roadcut where the Precambrian leucogneiss rests in overthrust contact on the light gray Cambro-Ordovician Wappinger dolomite. Note the occasional flat-lying, slippery fracture planes of the contact.

The metamorphic layering in the allochthon dips to the east whereas the dolomite bedding dips predominantly west.

The average foliation of the leucogneiss is N33°E, 20°SE. The average attitude of the dolomite beds is N22°W, 35°W. Prominent faults in the klippe parallel the metamorphic layering and trend N22°E, 20°SE. The fault contact of the dolomite and the klippe is irregular and has the same general attitude. Both the dolomite and the klippe are cut by vertical faults trending N27°W.

		Drive west along Old Mansion Road observing the flat topography superposed on the Hudson River pelites. These are more gently folded with increasing distance from the Precambrian allochthon.
7.1	0.8	At T-intersection turn right on to Oxford Road (County Route 51).
7.6	0.5	Turn left onto Greycourt Road.
8.5	0.9	Greycourt Road becomes Oxford Road.
9.6	1.1	Bear left to stop sign and turn left onto Greycourt Road.
9.9	0.3	Greycourt Road becomes Leigh Avenue and continues past Hudson River pelites to intersection with NY 17M (stop light).
10.5	0.6	Make hard left turn onto NY 17M east.
11.6	1.1	Pull off N.Y. 17M just west of the ridge of Goose Pond Mountain (Goose Pond west) into parking area on right just beyond entrance to NY 6/17 East.

STOP 4. GREEN POND MOUNTAIN: PRECAMBRIAN ALLOCHTHON OVER OVERTURNED HUDSON RIVER PELITE.

Observe black, fissile Hudson River shales (Middle? Ordovician) in the road cut at the northwest edge of the Precambrian allochthon Careful observation will show that the attitude of the bedding and cleavage is N78°E; the bedding dips 70°S and the cleavage 40°S. This indicates that the outcrop is on the limb of an overturned fold with the synclinal axis north. The shales both here and at Bull Mine Mountain klippe to the north are all wildly folded and overturned close to the overriding Precambrian allochthons.

Many of the Hudson River black shales are calcareous, and consist of fine laminae (0.05 - 0.1mm) of dolomitic silt or mud (marl) intercalated rhythmically (occasionally cross-bedded) with carbonaceous shale. An estimated thin-section mode of a representative Hudson River "shale" follows:

Mineral	Calcareous laminae	Carbonaceous laminae
Detritals and matrix		
dolomite, calcite	30%	5%
quartz	45	25
plagioclase, microcline	10	+
mica, clay, chlorite	15	50
carbonaceous matter, graphite	-	15
pyrite	-	+
Metamorphic:		
chlorite, muscovite, biotite	-	5
	100%	100%

			Walk about 100 - 150 feet into the woods southwest of the shale outcrop on N.Y. 17M.

A moss-covered rubble of shale, dolomite, and "limonitized" fault breccia indicate where the covered contact has been crossed. The edge of the Precambrian allochthon of Goose Pond Mountain is found about 130 feet south from the road, and the first rock found in place is a graphitic calcareous quartzite of which sample No. 527 is representative.

12.0	0.4	Return to the cars and drive 0.4 mi east on N.Y. 17M, stopping at a white albite-quartz-microcline microperthite leucogneiss (sample No. 20), in places graphitic, biotitic or rarely, garnetiferous.

Note the extreme elongation and smearing out of quartz pebbles and grains similar to that seen at the Museum Village outcrop of Stop 3. Further to the east, the rocks become increasingly calcareous (sample No. 788). At the extreme eastern edge, prehnitized calc-silicate leucogneiss is interlayered with some amphibolite, the latter of probable basic volcanic origin. Modes of representative rock types of the Goose Pond allochthon follow:

Modes Of Gneisses Of The Goose Pond Mountain Allochthon

	East Side	Center		West Side
Sample No.	529	788	20	527
Mineral				
microperthite	- %	10.0%	42.2%	2.6%
albite (An 0-5)	65.6	-	27.7	-
oligoclase (An 20)	-	49.0	-	-
quartz	16.8	5.0	29.0	81.7
biotite	-	2.5	-	-
chlorite	1.7	-	-	0.3
sericite	7.2	-	0.8	-
graphite	-	-	+	4.2
actinolite	0.2	-	-	3.0
brown hornblende	0.2	33.0	-	-
diopside	1.3	+	-	8.0
sphene	-	-	-	0.2
apatite	+	0.3	0.1	+
ilmenite	-	0.2	0.2	-
prehnite	7.0	-	-	-
zircon	+	+	+	+
	100.0%	100.0%	100.0%	100.0%

Here, as in the Museum Village klippe of Stop 2, the greatest amount of retrograding occurs at the eastern and western margins of the allochthon with some fresh rocks occurring near the center. If the klippen are indeed synclinal saucers the presently exposed centers of the masses would lie at a further distance from the sole of the thrust and would be expected to show less alteration. It should be emphasized that the Precambrian autochthonous gneisses in the southern part of the Monroe quadrangle (Jaffe and Jaffe, 1962) are not comparably retrograded except near Triassic border faults.

13.6	1.6	Continue east on N.Y. 17M about 1.6 mi. and turn right (south) on Bull Mill Road for about 0.2 miles.
13.8	0.2	Turn right at dirt road and park.

STOP 5. LAZY HILL - SHAWANGUNK QUARTZITE

Walk 0.2 miles west crossing buried northeast-trending fault contact between the Shawangunk ridge of Lazy Hill rising steeply ahead and the Bellvale ridge of Durland Hill to the rear. Walk to the north nose of Lazy Hill (permission of the owners, the Durlands, is necessary) where a large outcrop of Shawangunk quartzite is exposed. The rock is a thin-bedded, pink, buff and white orthoquartzite consisting of:

Mode of Shawangunk Orthoquartzite, Durland Property

Quartz	93%
Chert	5
Sericite, Chlorite	2
Zircon	+
Hematite	+
Goethite	+
Pyrite	+
Green tourmaline	+
	100%

The quartz grains are well-rounded, moderately elongated, well-sorted (average diameter, 0.75mm), and the rock is very tightly cemented. Each grain of quartz is cemented to another by authigenic quartz overgrown in optical continuity with the detrital cores. Undulatory extinction due to deformation passes through the core and overgrowth of each grain. The Lazy Hill ridgetop to the south (not visited) is formed of a coarse white pebble conglomerate interbedded with white orthoquartzite (occasionally ripple-marked) and grades eastward to a red arkosic conglomerate below the ridge-top. Quartz pebbles and orthoclase pebbles are strongly elongated (3:1 and 4:1) and heavily shattered and veined in both the red arkosic and the white conglomerate.

14.0	0.2	Return to N.Y. 17M and turn right (south).
15.9	1.9	Drive south to the Monroe Bowl-O-Fun parking lot.

STOP 6. MONROE BOWL-O-FUN: CONNELLY (ORISKANY) - ESOPUS CONTACT)

The outcrop at Monroe Bowl-O-Fun originally was described by Jaffe and Jaffe, (1967) to consist of about 300 feet of the Esopus Formation underlain at the rear of the cut by red and white pebble conglomerate, the Connelly Conglomerate of Oriskany age. Most of the exposure has been covered by a railroad tie retaining wall. Here, a small outcrop of the Connelly is still exposed to the far right near the gasoline service station, and was described by Jaffe and Jaffe (1967) to consist weathered yellow, "limonitic" conglomerate (3 ft or more), succeeded by white to buff, pebble-bearing orthoquartzite (5 ft), which is in turn overlain by bright red hematitic quartzite (10 ft). The pebbles in the Connelly conglomerate are of white, round to slightly elongated quartz, averaging 1-2 mm in maximum dimension. The Connelly is disconformably overlain by a lowermost member of the Esopus Formation, recognized by Southard (1960). A small exposure of the Esopus

Formation is exposed at the far left of the retaining wall. The attitude of both the Connelly and the Esopus at their contact was described to be N68°E, 45°N. A heavily slickensided fault surface, trending N53°, 60°SE cuts across the Connelly beds and presumably also cuts the overlying Esopus Formation.

The lowermost member of the Esopus, at its base, consists of fissile, blue-gray siltstones which weather to brown and orange on cleavage surfaces. Many of the rocks are marked with Taonurus cauda-galli on bedding planes. At this outcrop the authors have collected a remarkable fauna including a specimen of the giant trilobite, Coronura myrmecophorus, not previously reported from the Esopus Formation. According to D.W. Fisher, New York State paleontologist who identified the specimen, it has previously been reported from the Schoharie and Onondaga Formations. The specimen was donated to the N.Y. State Museum collection.

The lowermost member is also relatively rich in conulariids, none of which have yet been identified. Other fauna include the brachiopods: Leptocaelia flabellites, Schuchertella sp., Acrospirifer macrothyris as well as some chonetid and orbiculoid genera. Platystomid and loxonemid gastropods, rugose corals and a dalmanitid trilobite were also collected by the authors.

The lowermost member grades into the black, poorly fossiliferous Lower Mudstone member which in turn grades into a purple sandstone at the north end of the 350 foot exposure. The sandstone is presumably the lower part of the Highland Mills member of the Esopus Formation. The fauna of the lowermost member at the Bowl-O-Fun appears to differ significantly from that of the Highland Mills member of the Esopus Formation found at Bakertown and Highland Mills (described by Boucot, 1959). The fauna should receive some serious study by specialists before the outcrop is demolished by new construction.

16.6	0.7	Drive south on N.Y. 17M to the second traffic light and turn onto Stage Road.
17.4	0.8	Bear right at stop sign on to the Orange Turnpike
18.0	0.6	Park on the side of the road opposite development.

STOP 7. ORANGE TURNPIKE: POUGHQUAG QUARTZITE (LOWER CAMBRIAN)

Walk 0.16 miles due west over hilltop to the edge of a cliff formed by a 10 foot section of the Poughquag Formation (Lower Cambrian). The section consists of alternating 2 inch to 2 foot thick beds of ferruginous orthoquartzite, conglomerate, and arkose, striking N75°W and dipping 8°N, overlying the vertically dipping Precambrian autochthon with marked angular unconformity. This represents original sedimentary onlap with gentle warping or folding in subsequent geologic time.

Apparently the Precambrian Monroe Massif (Block 2) was sufficiently rigid throughout the Paleozoic to prevent the deformation of the overlapping embayment of Poughquag quartzite and Wappinger dolomite. This is indicated by both the gentle warping observed and also by the relative sphericity of the quartz pebbles in various Poughquag beds.

Several of the beds are feldspathic, a feature uncommon in the Poughquag of the Poughkeepsie Quadrangle (Gordon, 1911). One such bed at the Monroe outcrop is a conglomeratic arkose which is a true high rank arkose in the sense of Krynine (1948). A remarkable textural feature of this rock is the abundance of authigenic feldspar (microcline?) which is the principal cementing medium in sample No. 466. The specimen consists of 1-2m quartz and microcline pebbles (all very round) lying in a matrix of authigenic microcline (?) cement which is clear in appearance. Some sawtooth or hacksaw terminations on the detrital microcline cores (Edelman and Doeglas, 1931) indicate that interstratal solution has taken place after deposition, presumable _in situ_. The authigenic feldspar overgrowths show only weak twinning when grown around detrital cores showing strongly developed microcline twinning. A mode of such rock is as follows:

Mode Of Lower Cambrian Poughquag Conglomerate Arkose
Specimen No. 466

Microcline	47.4%
Microcline microperthite	+
Quartz	48.6
Albite-oligoclase	+
Muscovite	+
Rutile, Anatase, Tourmaline (green + brown)	0.5
Zircon	0.5
Hematite } Mn oxides}	3.0
	100.0%

The mineralogical composition of the Poughquag at Monroe leaves little doubt that it was derived from erosion of the granitic gneisses it overlies.

On the return walk to the road, stops may be made at exposures of post-Wappinger lamprophyre dikes which the authors believe to be of late Ordovician age (Jaffe and Jaffe, 1962). The authors have studied the dikes in considerable detail and would suggest a possible age of intrusion similar to that of the ultramafic intrusion of the Cortland Complex at Stony Point, New York (Ratcliffe, 1967). The Cortland Complex has been dated by Long and Kulp (1962) at 435 million years by K/A isotopic ratios obtained on biotite from the complex, a date close to the accepted Ordovician-Silurian boundary. An age of 398±17 m.y. was obtained from Ar^{40}/K^{40} ratios on the amphibole (kaersutite) phenocrysts in one of these dikes (Jaffe and Jaffe, 1973). This age may be a bit low because of argon loss during a regional Paleozoic reheating. These ages were run some years before development of more refined Ar^{40}/Ar^{39} methods now in use.

Thus, lamprophyre dikes of deep-seated (mantle-derived?) origin are not restricted to Mesozoic and Tertiary ages. They may range from Precambrian to Recent (Zartman, _et al_., 1967).

Stops 8, 9, 10, 11, and 12 (Fig. 3) are located in the granulite facies gneisses of the authochthonous block, here called the Monroe crystalline massif of Precambrian (Proterozoic) age. For reemphasis, we note that the gneisses of the authochthon do not show the pervasive retrograde alteration and extensive mineral stretching that characterize the rocks of the allochthon.

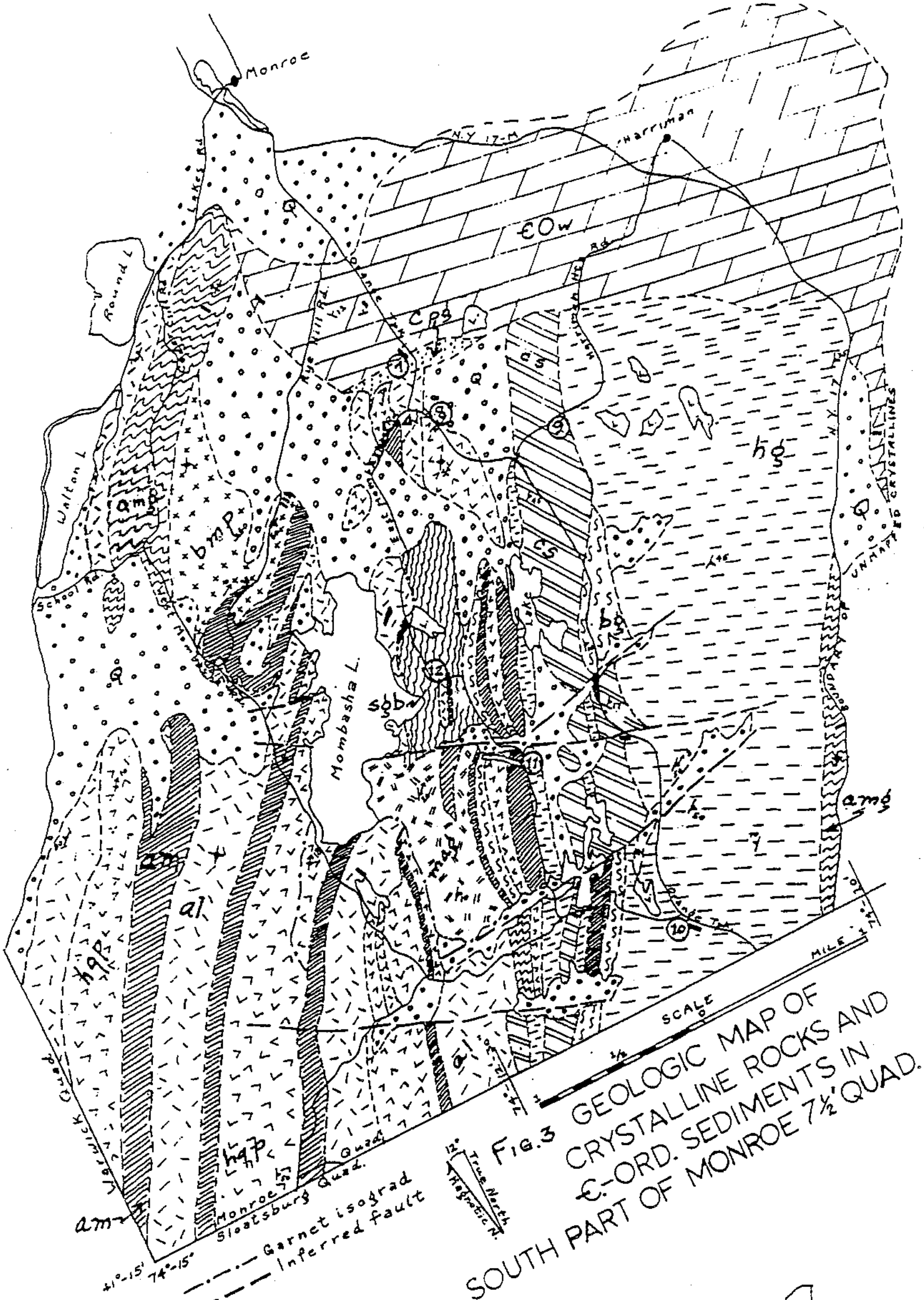

FIG. 3 GEOLOGIC MAP OF CRYSTALLINE ROCKS AND €-ORD. SEDIMENTS IN SOUTH PART OF MONROE 7½' QUAD.

H.W. & E.B. JAFFE
MAY, 1962

	Quaternary	Q	alluvial and glacial cover
Paleozoic	Cambrian - Ordovician		dikes of lamprophyre, leucophyre, diabase.
Paleozoic	Cambrian - Ordovician	€Ow	Wappinger Group, dolostone, minor shale
Paleozoic	Cambrian	€pg	Poughquag Formation : quartzite, arkose, conglomerate

Gneisses of Monroe and Ramapo Blocks

Proterozoic	hg	hornblende (ferrohastingsite) granite gneiss, and pegmatite
Proterozoic	am	amphibolite, pyribolite
Proterozoic	amg	interlayered amphibolite and granodiorite
Proterozoic	cs	calc - silicate gneiss
Proterozoic	hqp	biotite - hypersthene - quartz - plagioclase gneiss
Proterozoic	sgb	sillimanite - cordierite - almandine - biotite - quartz - feldspar gneiss
Proterozoic	bg	biotite gneiss
Proterozoic	bmp	biotite - mesoperthite - gneiss and albite - oligoclase - leucogranite
Proterozoic	hk	biotite - hypersthene - K- feldspar - quartz - plagioclase gneiss
Proterozoic	mag	magnetite - garnet - andesine gneiss
Proterozoic	al	microcline alaskite (leucogranite); contacts with all other Proterozoic units may be intrusive

The prevalence of hypersthene and total absence of muscovite from folded gneisses of the Monroe crystalline massif verify that regional metamorphism took place in the granulite facies. The omnipresent coexistence of hypersthene with augite in mafic rocks; hypersthene-K-feldspar association in granitic gneisses (charnockite); sillimanite-K-feldspar assemblages in pelitic gneisses; Fe-rich cordierite-sillimanite-K-feldspar-almandine-tourmaline assemblages, also in pelitic gneisses; and copious exsolution of pigeonite in host hypersthene collectively indicate that metamorphism occurred at T = 700 - 800°C, at P = 2 - 4 kbars (7-14 km depth) (Jaffe and Jaffe, 1973).

18.4	0.4	Drive southeast on the Orange Turnpike, stopping at the road cut on the west side of the road.

STOP 8. ORANGE TURNPIKE: QUARTZ-OLIGOCLASE GNEISS

Leaving Stop 7, the Orange Turnpike turns southwest and crosses the concealed unconformable Cambrian-Precambrian contact. The quasi-horizontal Poughquag and Wappinger beds overlie vertically dipping Precambrian gneisses, permitting the delineation of a major unconformity, unfortunately not exposed in the area. Beyond the Lipalian interval, the first rock encountered is a fine-grained pink alaskite composed mainly of 1-3 mm microcline-microperthite and quartz, minor sericitized albite-oligoclase and an occasional flake of biotite. Within 100 ft to the south, the pink alaskite grades through a narrow zone of coarse biotite-microperthite-oligoclase granite and granodiorite into a medium-grained (2 mm), gray, essentially massive hypersthene quartz diorite gneiss which forms the bulk of the outcrop. It is variously called quartz oligoclase gneiss or enderbite by other workers. It contains: oligoclase 70%, quartz about 25% and hypersthene, biotite, magnetite and chlorite about 5%. In thin section, quartz is not uniformly distributed but rather forms long tongues which embay adjoining oligoclase grains; these are well-twinned and antiperthitic. On top of the outcrop, observe several lenses of biotite-hornblende-hypersthene-labradorite (An55) pyribolite infolded in the quartz diorite gneiss. Foliation measured on the pyribolite is N35° to 50°E, with a dip close to 90°. Slickensided joint faces strike N24° to 65°W.

About 0.3 mi south (not a scheduled stop) the quartz diorite gneiss darkens in color, the quartz content drops markedly, and the rock grades to an augite diorite gneiss interlayered with hornblende-hypersthene-andesine pyribolite. Where quartz becomes locally abundant, it embays and replaces both plagioclase and the ferromagnesian minerals.

Hypersthene-quartz-oligoclase gneiss (quartz diorite gneiss) is thus formed from the metamorphic reconstitution of pyribolite accompanied by the introduction of silica and small amounts of potash. These constituents could logically be derived from anatectic granitic liquids derived from the fractional melting of sedimentary precursors.

18.9	0.5	Continue south on the Orange Turnpike and take the first left turn onto Harriman Heights Road.
19.3	0.4	Outcrop is on south side of road. Park on north side of road at entrance to Our Lady's Rosary Garden Gift Shop.

STOP 9. HARRIMAN HEIGHTS ROAD: CALC-SILICATE MIGMATITE

A fresh roadcut exposes a dark gray, green, and pink banded migmatite. The gray rock is a calc-silicate paragneiss composed of quartz, microcline, bytownite (An80), augite, green epidote (pistacite), dark brown sphene (titanite), zircon, apatite, and magnetite. The pink bands consist mainly of quartz and microcline or microcline microperthite. Other samples of this migmatite contain almost pure anorthite (An95). Epidote, common throughout the region in calc-silicate units, evidently formed by retrograde metamorphic alteration or exchange of Ca, Al, Si, and Fe in anorthite+augite, in a wet oxygenated environment.

22.4	3.1	Return the 0.4 mi to Orange Turnpike and continue south to the junction with Bramertown Road entering from the west. Park and walk 0.1 mi. south on Orange Turnpike to the outcrop.

STOP 10: ORANGE TURNPIKE: CAMPTONITE DIKE CUTTING HORNBLENDE GRANITE GNEISS

The outcrop on the west side of the road shows a 20-25 ft dike of lamprophyre called camptonite forming the low pavement of the outcrop. It strikes N38°W and dips 86°NE. The dike contains green hornblende phenocrysts up to 1/4 in lying in a matrix of albite laths, 0.2 x 0.4 mm, which, in turn enclose granular epidote. In these rocks, it appears that deuteric alteration of an initially more calcic (intermediate) plagioclase has resulted in the growth of albite+epidote. The camptonite contains: albite - 53.3%, hornblende - 21.5%, epidote - 12.6%, chlorite - 7.3%, and apatite+opaque+calcite+quartz+k-feldspar - 5.3%.

The country rock intruded by the dike is the hornblende granite gneiss that makes up about one-quarter of the volume of the Monroe crystalline massif and is also widespread in the other areas that constitute the Hudson Highlands. In the Monroe block, these granitic gneisses are uncommonly iron-rich and the hornblendes are ferrohastingsites with 100Fe/(Fe+Mg) = 86 that coexist with Ti- and Fe-rich biotites in which 100Fe/(Fe+Mg) = 90, ratios found in minerals at this outcrop. Here, the granite gneiss contains occasional schlieren of very biotite-rich rock.

23.2	0.8	Turn west onto Bramertown Road, and then turn right, north, on first paved road (East Mombasha Road). The road follows the contact of granite gneiss (east) and pyribolite (west).
24.1	0.9	Outcrop is on the west side of the road at a sharp bend to the west.

STOP 11. EAST MOMBASHA ROAD: LEUCOPHYRE DIKE INTRUDING AMPHIBOLITE

Stop 11 shows a 16 ft thick granodiorite leucophyre dike which strikes N48°W, cross-cutting the foliation of the surrounding amphibolite which strikes 57°E and dips 20°S. Note the large wedge of amphibolite in the center of the dike and the occasional pink K-feldspar-quartz bands in the amphibolite. This dike is unique in the Monroe quadrangle, and perhaps in the Highlands. It shows sparse pink phenocrysts of oligoclase, quartz, less microcline and biotite lying in a dark gray matrix, which is again porphyritic on a microscopic scale. The second generation of microphenocrysts is made up of square to rhombic zoned potash feldspar and laths of albite-oligoclase. These lie in a very fine granophyric groundmass made up of feldspar, quartz, mica, chlorite, "limonite" and manganese oxide. The extremely fine-grained oxides form megascopic crenulated black streaks which give the dike a distinct flow layering in parts of the outcrop. A mode was not obtained because of the fine nature of the matrix. X-ray data on a powdered sample indicate that oligoclase>quartz>microcline, hence the dike is of granodioritic composition. East of the road, the dike is not found, and may be cut off by a north-south fault; if so, the dike is very old. An outcrop of the same rock was found 0.25 mi to the west cutting migmatite; it may be an extension of the same dike.

25.1	1.0	Continue north on East Mombasha Road stopping at a dark, mica-rich gneiss just north of Stop Number 12 on the map (Fig. 3).

STOP 12. EAST MOMBASHA ROAD: PELITIC PARAGNEISS

This is a tightly folded, crenulated pelitic paragneiss in the sillimanite-K-feldspar zone of metamorphism. The outcrop consists of thin bands of gray biotite-microcline-labradorite-quartz gneiss intercalated with bands rich in orthoclase cryptoperthite (anorthoclase) and quartz. Abundant garnet (almandine-pyrope), Fe-rich cordierite, dark blue-green tourmaline and prismatic sillimanite are developed along the interfaces of the biotitic and alaskitic layers. Biotite is, under the microscope, intensely pleochroic from "paprika-red" to almost colorless and is undoubtedly rich in Ti as well as Fe. Sillimanite and tourmaline lie in the foliation planes with their long axes parallel to the fold axes. Cordierite, not recognized when the 1962 NYSGA Guidebook was written, is abundant in parts of the outcrop and can be recognized in some places by its characteristic blue or purplish blue color imparted to hand specimens. Under the microscope, it shows both polysynthetic twinning and sector-zoning or twinning; alteration to fibrous pinite plus the twinning make it difficult to identify and it may be mistaken for altered plagioclase; characteristic yellow, bulls-eye halos are virtually absent. Thin sections cut across foliation planes show numerous square cross-sections of sillimanite needles enclosed in cordierite. The relatively Fe-rich nature of this cordierite, $100Fe/(Fe+Mg) = 25$ is useful in limiting the pressure at which the high-temperature regional metamorphism occurred; it is 2-4 kbars with water vapor present but not necessarily saturating the pore spaces. This, plus the hypersthene and sillimanite-K-feldspar parageneses fix the parameters of regional metamorphism in this part of the Hudson Highlands at $T = 700 - 800^{\circ}C$ and $P = 2 - 4$ kbars.

26.4	1.3	Continue north on East Mombasha Road. Turn left onto the Orange Turnpike.
27.2	0.8	Orange Turnpike becomes Still Road.
28.0	0.8	Continue to traffic light at NY 17M. Continuing north past intersection Still Road becomes Freeland Street.
28.6	0.6	Freeland Street becomes County Route 105.
29.0	0.4	Turn right at intersection with Dunderburg Road.
30.9	1.9	Intersection of NY 6, 17, and 32. End of trip.

NOTES

THE GREEN POND OUTLIER OF NEW YORK AND NEW JERSEY; A MESOSTRUCTURAL LABORATORY

JAMES P. MITCHELL,
The Pennington School, Pennington, NJ 08534

RANDALL D. FORSYTHE
Department of Geography and Earth Science, UNCC, Charlotte, NC 28227

INTRODUCTION

Exposures of the folded Silurian and Devonian strata of the Green Pond outlier within the Reading Prong of New Jersey and New York (Fig. 1a.) contain abundant small scale deformation features. These include: slickenfiber faults, en echelon vein arrays, two sets of disjunctive cleavage, and small scale folds. This trip visits a series of outcrops within the outlier that permit a detailed study of the interrelationships and origin of these mesostructures.

Students of structural geology are commonly introduced to small scale (meso) structures classed as being either brittle or ductile. In the Earth, however, brittle deformation is not necessarily restricted to the upper "seismogenic" layer, and ductile deformation likewise is not restricted to the deeper medium to high grade metamorphic levels. Folds that form under moderate confining pressure and low temperatures, often reveals a confusing array of both ductile and brittle mesostructures. The manner in which these ductile and brittle features are related is one of the interesting topics that can be examined on this trip.

The trip also provides students an opportunity to understand some of the dynamic or stress conditions that operate during folding. Published work on folded strata (for discussion see Ramsay, 1967) has generally focussed on the geometric development of folds and its strain-related cleavages or foliations. The array of semi-brittle features that formed along with the cleavage in the Green Pond outlier permits an appreciation of the forces or stresses responsible for folding.

THE GREEN POND OUTLIER

Paleozoic formations in the outlier. Figure 1b shows the principal distribution of the units within the outlier. For simplicity, the ten Silurian and Devonian formations (stratigraphy shown in Fig. 2) are grouped into three lithostratigraphic units (III, IV, and V). These rest with angular discordance over Middle Proterozoic rocks (unit I) in some places but over a thin veneer of Cambrian and Ordovician rocks (unit II) in others (Barnett, 1976; Herman and Mitchell, in press). The 2,400 meters of Silurian and Devonian rocks, that makeup the bulk of the outlier (see Fig. 3) have strong inter-, and intraformational competency differences. Unit III, the Green Pond Formation (Silurian), is equivalent to the Schawangunk Conglomerate of the Valley and Ridge Province (Lewis and Kummel, 1915), and is composed of

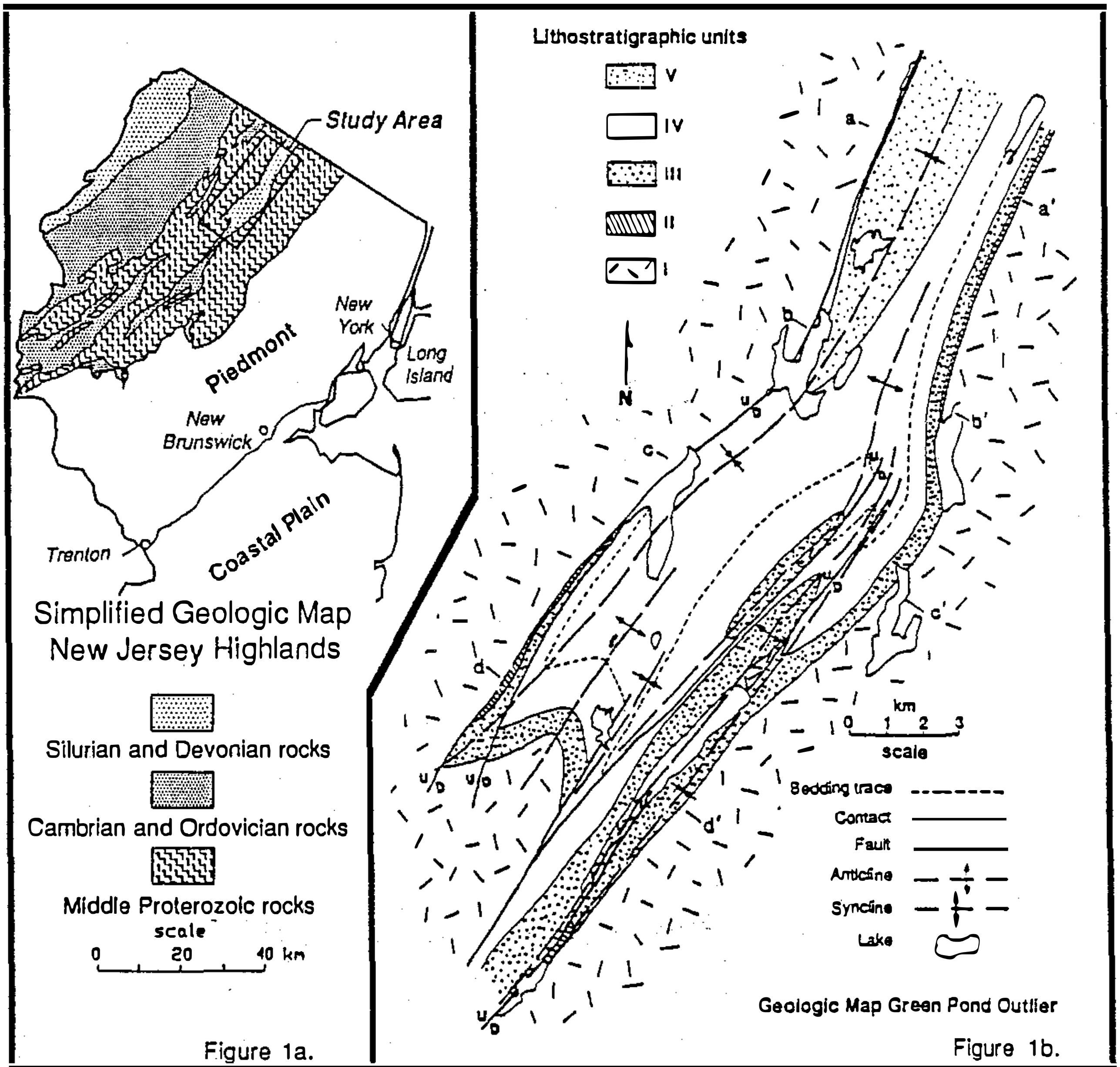

Figure 1a. Figure 1b.

interbedded purple-red conglomerate, cross bedded sandstone and minor siltstone units. The conglomerate and sandstone units are well indurated. They have highly sutured clasts and a siliceous and hematitic cement. The central portions of the sequence, unit IV, contain some of the less competent units, such as the Longwood (Sil.) and Cornwall Shales (Dev.). These units have the best development of disjunctive cleavage, but are not as well exposed in the outlier. The Middle Devonian Skunnemunk Formation (Catskill delta equivalent), is a ridge former, and like the basal Green Pond Formation is also well indurated, though the former tends to break around the grains when fractured.

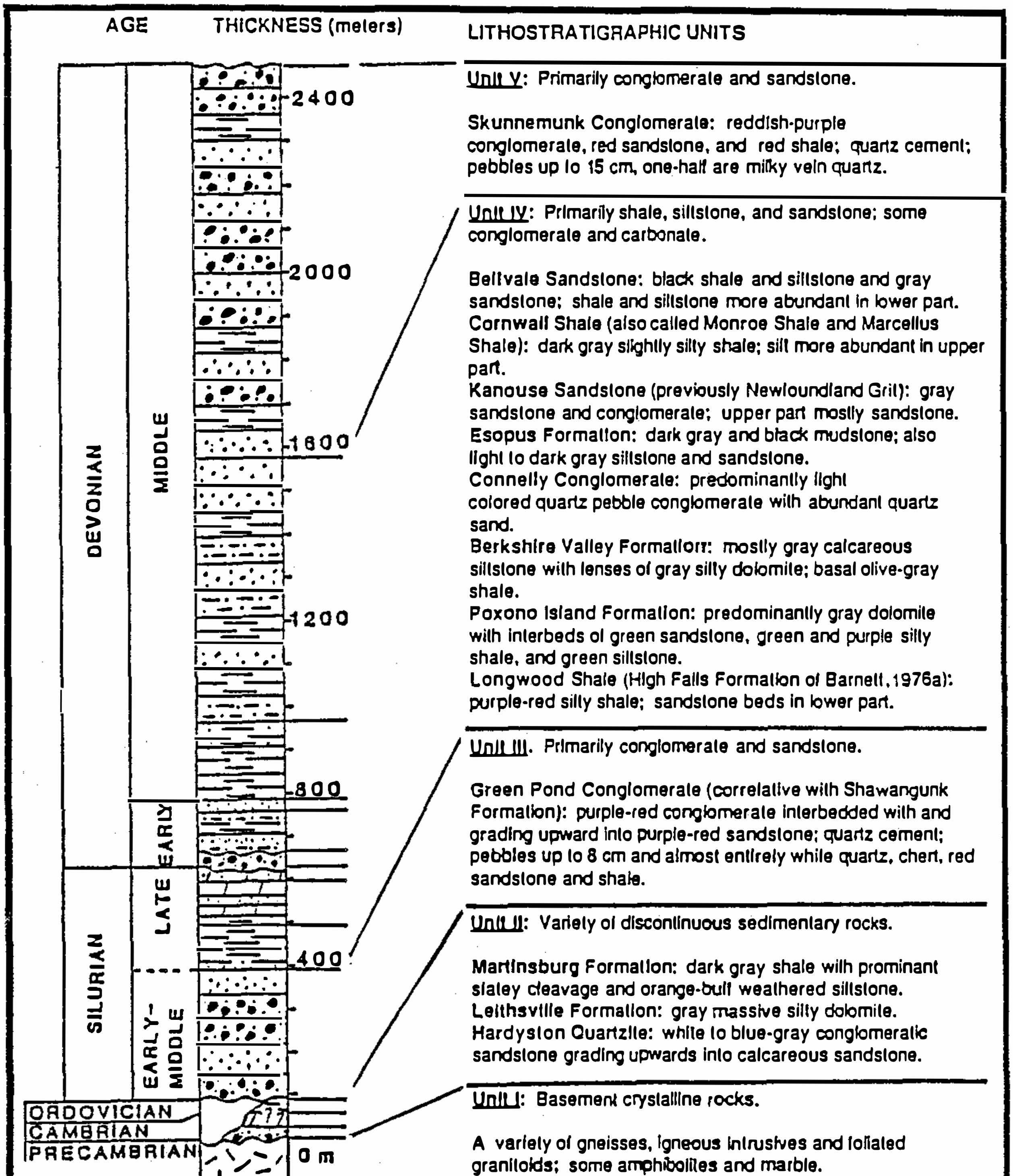

Figure 2. Green Pond Outlier Stratigraphy

<u>Structure of the Green Pond outlier.</u> The outlier is defined by a series of northeast trending fold structures that have been down-faulted within the Reading Prong. (Fig. 1b,4a). The large scale folds defined by the more competent units (III & V) and perhaps mechanically controlled by the basal contact with the underlying Proterozoic crystalline rocks have relatively planar limbs and sharp well defined hinges suggestive of a kink fold geometry. Within these large scale structures occur smaller amplitude open to tight folds that have curved parallel geometries. Most hinges plunge gently to the

northeast but others are doubly plunging. In New York State the outlier is defined by one large tightly folded asymmetric syncline with a steeply dipping to slightly overturned eastern limb. It is fault bounded to the northwest and southeast. Southwestwards in New Jersey the outlier's structural form expands to include three smaller amplitude folds. Each of the fold structures is fault-bounded, with the exception of the eastern margin of the outlier in our study area where the Green Pond Formation is thought to rest in depositional contact with both lower Paleozoic and Proterozoic rocks (Kummel and Weller, 1902; Barnett, 1976). Faults trend parallel to the axial traces of folds and tend to be associated with the eastern limb of anticlines. Fault attitudes and displacement histories have been variably interpreted over the years (Kummel and Weller, 1902, Lewis and Kummel, 1910 & 1915; Ratcliffe, 1980). However, the most recent mapping (Herman and Mitchell, in press) has identified high angle reverse faulting in the central part of the outlier and conjugate reverse faults to the southwest around Dover, New Jersey.

MESOSTRUCTURES

Small scale structures, believed to have been developed in addition to the large scale folding and faulting of the outlier are: two sets of cleavages, small scale folds, en echelon vein arrays (EVA's) and slickenfiber faults (SFF's) (Mitchell and Forsythe, 1988).

Cleavages. Within the outlier cleavages are best developed in the more argillaceous formations such as the Longwood and Cornwall Shales. In most localities a slaty to spaced cleavage can be observed in the shaly and silty units that is subparallel to the axial surfaces of the local fold structures. However, in addition to the axially planar cleavage, several localities reveal a cleavage of slaty, pressure solution, or crenulation type, that trends obliquely across the fold structures with attitudes ranging from N 60 E to E-W. In a few places (e.g., Stop 3) both cleavages are found together and document an interesting shift or noncoaxial character in the directions of 'finite' shortening accompaning their development (discussed below).

En echelon vein arrays (EVA's). (terminology of Beach, 1975) These are also known by some as en echelon tension gashes (Shainin, 1950; Dalziel and Stirwalt, 1975). They are staggered sets of veins that are aligned in one common plane. Examples in Figure 3b, show two of their common geometric and dynamic configurations. Unless they occur in conjugate sets or with other brittle features, it is not obvious whether they are of a specific type. The production of an EVA in a rock is the distortional equivalent to a dilating zone of shear. It is, however, produced by the progressive cracking and infilling of the veins during shearing. Thus these 'shear zones' are regarded as a class of "brittle-ductile" features (Ramsay, 1980b). In the Green Pond outlier, they are most commonly developed in the indurated sandstone and conglomerate layers of units III(Green Pond Formation) and V(Skunnemunk Conglomerate).

Slickenfiber faults (SFF's). (terminology of Wise et al., 1984) These also are features that, in part, form by crack propagation (Durney and Ramsay, 1973; Ramsay, 1980). These are regarded as a subset or special class of faults within which fibrous minerals grow and infill the spaces opened up within the

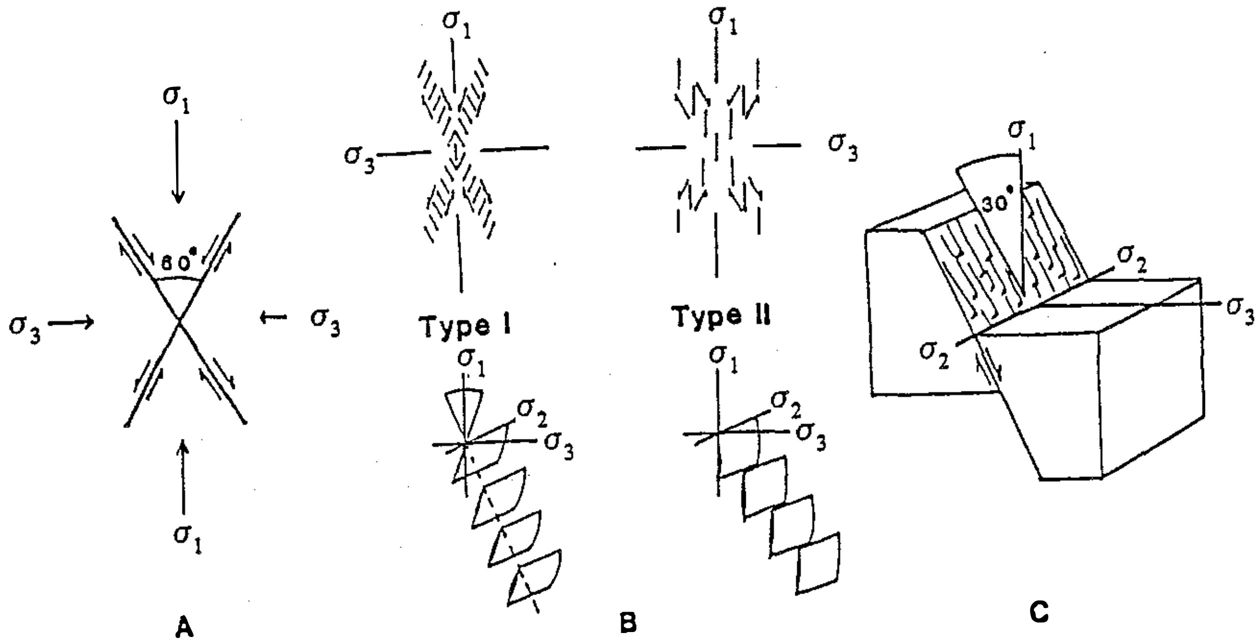

Figure 3. A. general relations between shear failure planes and principal stresses
B. relations between two types of EVA and principal stresses, after Beach, 1975
C. relations between slickenfiber faults and principal stresses.

fault zone during faulting. The fibers tend to connect formerly adjacent points on the opposite walls of the fault. Slickenfiber faults, like the EVA's are developed in the more competent units of the Green Pond outlier. Relations suggest that they develop instead of an EVA due to pre-existing surfaces of weakness in the rock being appropriately oriented for slip. Cross-bedding in the conglomerates and sandstones is commonly a plane of weakness along which SFF's were developed.

STRAIN AND PALEOSTRESS FIELDS

There are important distinctions to be made between the indicators of strain and paleostress in rocks.

Strain Marker. Generally most strain markers are found representative of a finite or long term product of distortion. Cleavage, for example, is commonly thought to be a plane of finite flattening disposed normal to the maximum axis of finite shortening (Ramsay, 1967). Viewed as such, it is a physical product of a protracted history of deformation that can tell us little about the specific flow or distortions occuring at a given instant. Thus strain markers in general provide a restricted or limited view of the deformation history.

Paleostress markers. Most paleostress indicators on the other hand are more like snap shots of the conditions of deformation that existed at various times during the deformation. In general both EVA's and SFF's contain geometric attributes which can be modeled after brittle features observed in the laboratory and which by analogy permit the inference of the orientation (but not magnitude) of ancient stress fields.

Paleostress determinations. Assessments of paleostresses are made with confidence when EVA or SFF features are present in conjugate sets. Such sets have formed the basis of many previous studies (Roering, 1968; Hancock, 1972; Beach, 1975; and Rickard and Rixon, 1983). In each case the direction of maximum compressive stress and incremental axis of greatest shortening bisects the acute angle between intersecting shear zones (Fig. 3). This condition is true for both biaxial and triaxial strains (Hancock, 1985; see Fig. 2). Two types of conjugate EVA geometry (Fig. 6b) have been described by Beach, 1975 (also see Roering, 1968; and Rickard and Rixon, 1983). In type 1 geometry, not all veins are parallel, rather, the veins of one shear parallel the orientation of the complimentary conjugate shear. Consequently in type 1 geometry the bisector of the conjugate EVA zones also bisects the acute angle between each vein and its enveloping shear. Also, a single large vein typically occurs at the intersection of the two shears which parallels the conjugate shears' bisector (Rickard and Rixon, 1983). In type 2 geometry (Fig. 3b) the veins of both shears parallel the conjugate shears' bisector.

In the field, the easiest case for paleodynamic determinations is for conjugate shears. For isolated arrays or SFF features additional considerations are necessary. In the Green Pond outlier, Mitchell and Forsythe (1988) made the following simplifying assumptions:

1) For each **EVA** the maximum principal stress axis bisects the acute angle between an individual vein and the plane containing the vein array for which the individual is a set member (Fig. 3b, type 1). This assumption appears valid since the bulk of our conjugate EVA's are type 1 of Beach (1975).

2) For each **SFF** an approximation of the maximum principal stress is assumed to lie about 30 degrees from the shear surface and in a plane containing the normal to the plane and the fiber lineation observed on the surface of the fault (Fig. 3c). Since two solutions are possible the sense of displacement must be determined from the overlap of the incremental fibers (Durney and Ramsay, 1973) for a unique solution. The 30^{o} angle is measured from the fault surface in a sense compatible with the sense of displacement (e.g. clockwise for right lateral, and counterclockwise for left lateral displacements).

3) For each **EVA** and **SFF** the intermediate principal stress is assumed to lie in the plane of the shear zone and perpendicular to the displacement direction. For EVA's it also parallels the intersection made by each vein with its enclosing array.

DEFORMATION OF THE GREEN POND OUTLIER

Paleostress trends. The results and interpretations of a regional investigation of SFF and EVA from the central portion of the Green Pond outlier were presented by Mitchell and Forsythe (1988). The study included 130 paleostress determinations from SFF's and 200 from EVA's. Approximately 80 EVA's were from conjugate sets. Most of the observations came from exposures of the Green Pond Formation, but 4 localities were located in the Middle Devonian Skunnemunk Formation. The data were grouped into 17 localities (Fig. 4a) that covered 4 southeast dipping fold limbs, 3 northwest dipping fold limbs and 2 hinge zones. The data are reproduced on stereonets in Figure 4b and on rose diagrams for each locality in Figure 5 for SFF and EVA data. The plots document a number of the general findings. First, most SFF'S and EVA'S record vertical S_1 - S_3 planes with S_2 horizontal (S_1, S_2,

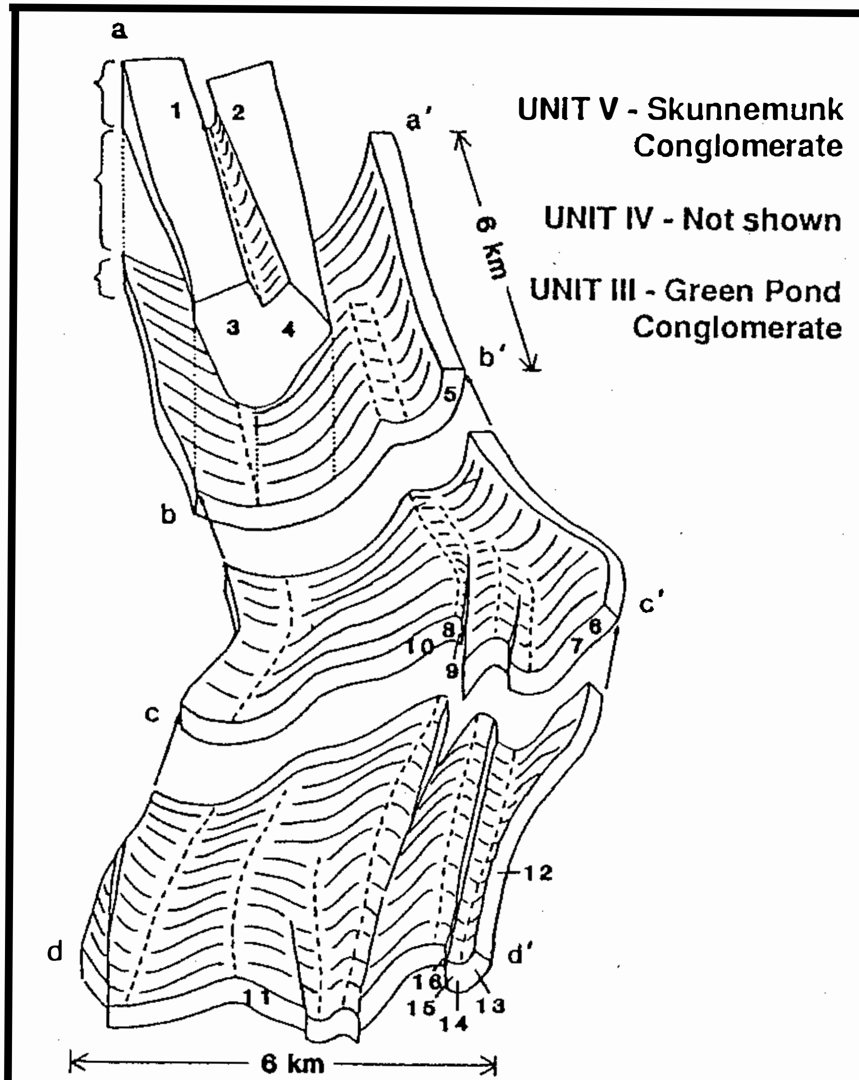

Figure 4a. Perspective drawing of the outlier with data sites

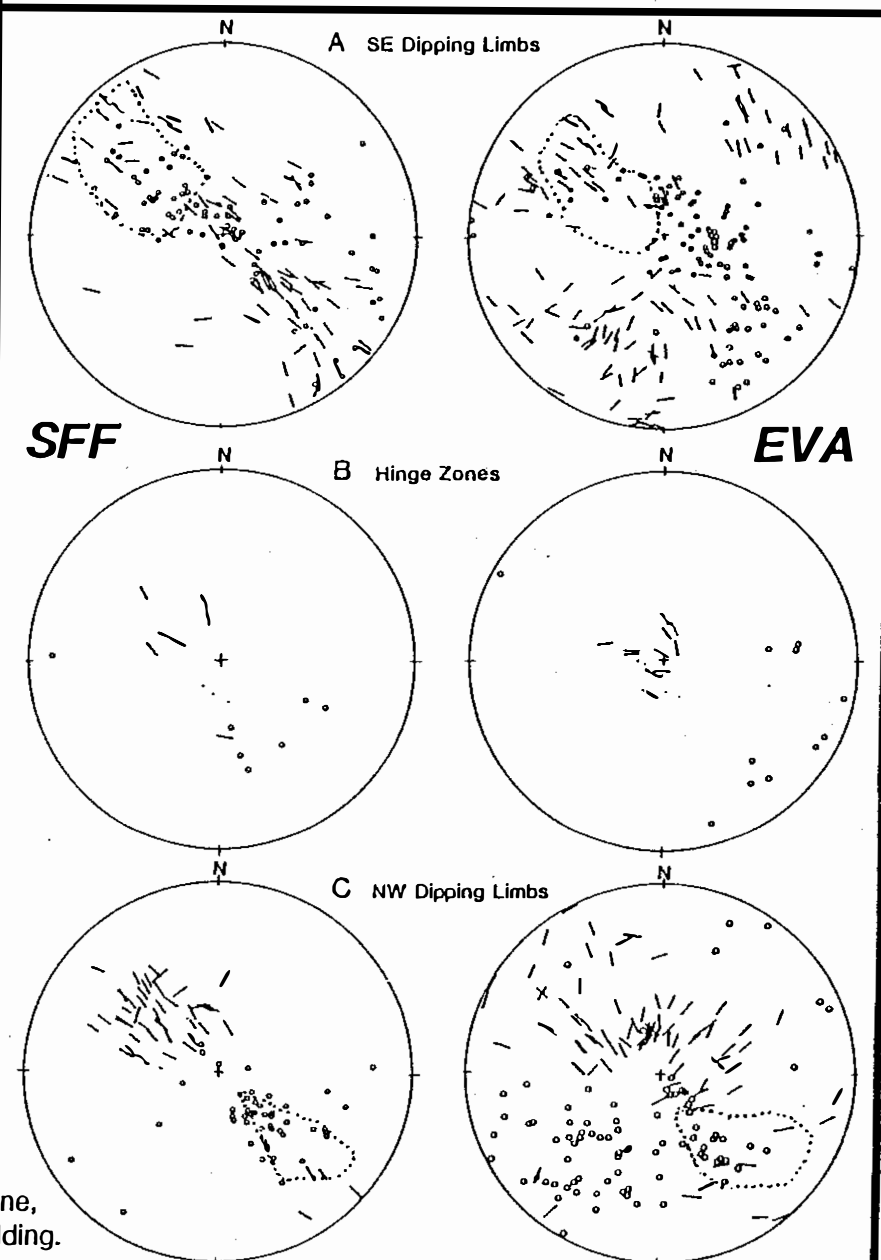

Figure 4b. Plots of max. and min. stress determinations for EVA and SFF. Short lines are max. axes oriented in max.-min. plane, and circles are min. axes. Dotted lines define fields of poles to bedding.

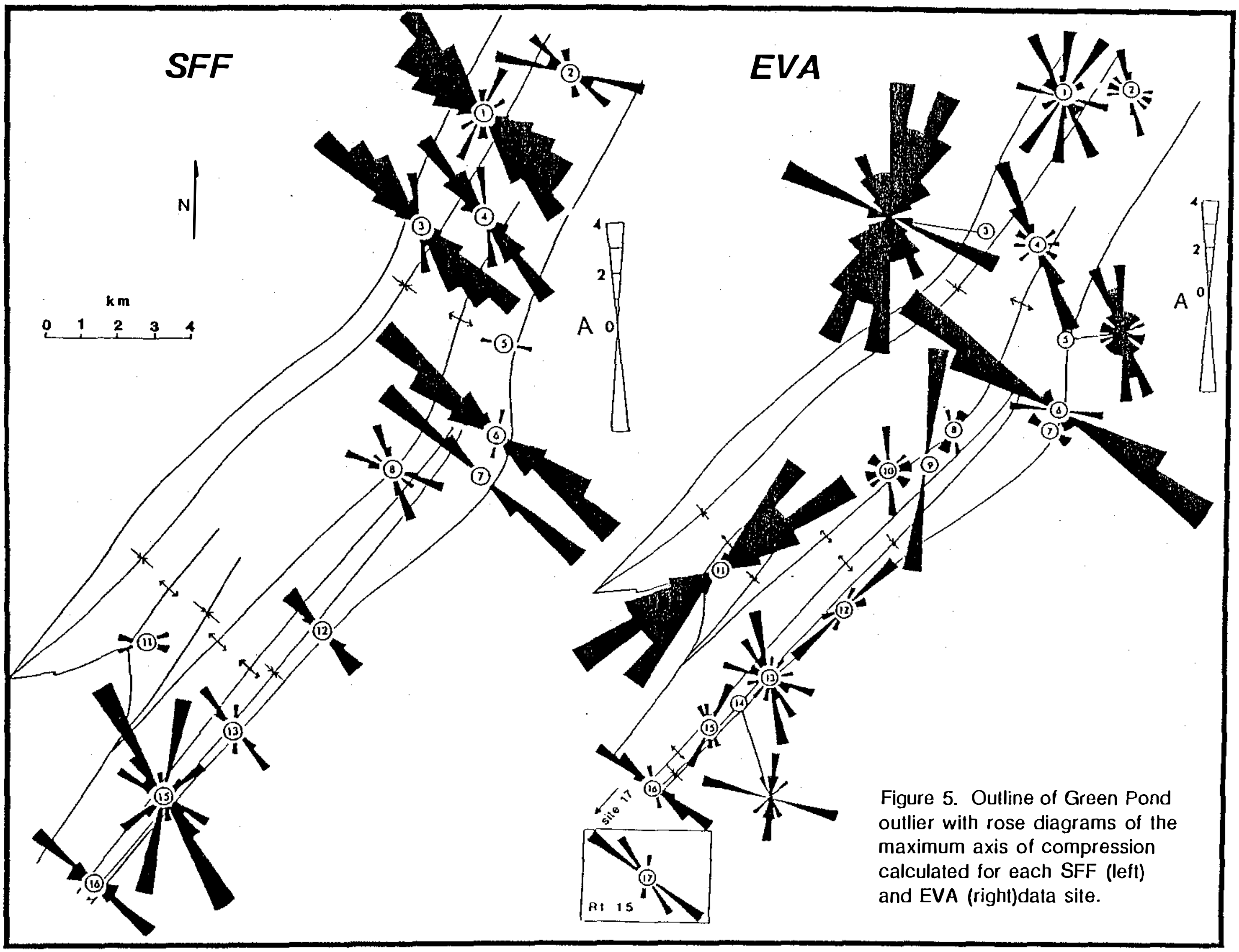

Figure 5. Outline of Green Pond outlier with rose diagrams of the maximum axis of compression calculated for each SFF (left) and EVA (right)data site.

S_3: principal stresses). Second, cumulative rose diagrams for EVA'S and SFF'S indicate that the majority of S_1 axes trend to the NW or SE, essentially at right angles to the fold hinges. Two other modes of S_1 orientations are also present. One trends N-S, the other NE-SW. Third, the SFF data primarily record the dominant NW-SE directed S_1 whereas EVA data have all three modes. Trends for SFF's are much more regionally consistent than those for EVA's, and in addition they show little variation in S_1 trends from different structural positions on the fold structure. However plunges of paleostress axes determined from SFF as well as EVA, do show systematic variations with respect to positions on fold structures. Overall the paleostress data indicated that SFF and EVA have recorded a complex pattern of changing stress conditions in both space and time.

The paleostress data is highly variable. After grouping the data by structural position in the various folds, as well as by general azimuth (of the axis of maximum compression), it is found that much, if not all, of this variability could be accounted for by a combination of two superimposed structural phenomena. The first phenomenon (stage I of Mitchell and Forsythe, 1988) is postulated to be the main phase of folding in the Green Pond outlier under a sub-horizontal NW-SE directed regional maximum compressive stress regime. The second, more illusive, phenomenon (stage II) is thought to have resulted from the rotation of the regional stress field from a NW-SE to a N-S orientation. Under this new field, a phase of noncoaxial shortening took place by the generation of a cross cutting cleavage and numerous additional EVA's that indicate N-S to hinge-parallel maximum compressive stresses.

Stage I (Progressive Buckling). The preponderance of EVA's and SFF's display changes in character and orientation (plunge) that compare favorably with the theoretical computer models of the stress history of folding by Dieterich, 1970 (Figure 6). These models have also compared well to both empirical and other natural folds (Dieterich and Carter, 1969) where there exists significant contrast in the viscosities of the interbedded layers. In the Green Pond outlier variable development of cleavage in the different lithologies, the strong refraction of cleavage from one layer to another, and the general curved parallel fold forms argue for strong inter- and intraformational variations in competency or viscosity during folding. Basically, the argillite and siltstone layers have, through cleavage development, accommodated the buckling

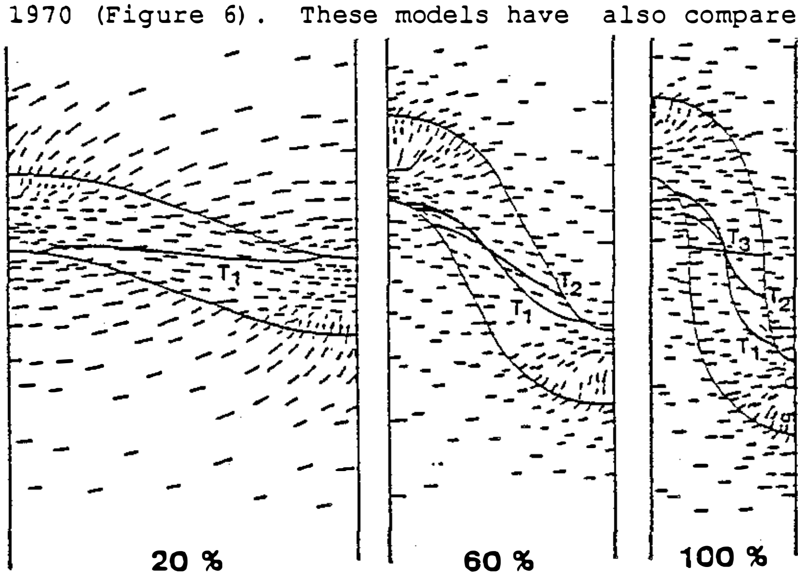

Figure 6. Tracings of maximum principal stress trajectories at three progressive stages of folding based on the computer models of Dieterich and Carter, 1969.

of the more competent sandstone and conglomerate layers. In the theoretical and empirical folding of layers of variable viscosities, it is commonly found that some layer-parallel shortening occurs before appreciable buckling of the higher viscosity layers. As buckling begins there is first a strong guiding of the stresses that keeps the maximum principal axis of compression along the layering. However, as the fold tightens up, the steeply inclined limbs can no longer effectively guide the stresses and the direction of the maximum principal axis returns to take on a position more similar to the external field. The models would argue for three general stages of mesostructural development: an early layer parallel shortening phase, a main buckling phase with the stresses strongly rotated along with bedding, and a late phase with the stresses relaxed in the inclined limbs back to a subhorizontal orientation. These phases are believed to be reflected in the mesostructures of the Green Pond outlier.

Early layer-parallel shortening in the outlier is to be expected to a limited extent because even the resistant conglomerate beds are likely to have had some degree of non-elastic or viscous response to the applied stresses. In the Green Pond outlier most of the slickenfiber faults, and a large percentage of the en echelon vein arrays, indicated paleo-S_1 axes lying at very low angles to bedding (Forsythe and Mitchell, 1988). In addition, several of these EVA's are found bent or folded around small scale folds (Stop 1). The general impression from these relations is that many of the paleostress markers were developed at an early stage in the deformation. Furtherhmore, since buckling reduces the elastic resistance to the applied stress field, it would intuitively follow that nucleation of these brittle ruptures initiated during the threshold stresses that existed in the pre- to early buckling moments. Thus, the first stage of EVA and SFF development in the Green Pond Outlier is postulated to have been represented by an early phase of layer parallel shortening. We suggest as illustrated in Fig. 7, that the preponderance of SFF's that nucleated on the low angle cross beds (mostly west dipping), and EVA's that nucleated in the conjugate orientation, likley accomodated layer parallel shortening in these competent units during the early stages of deformation.

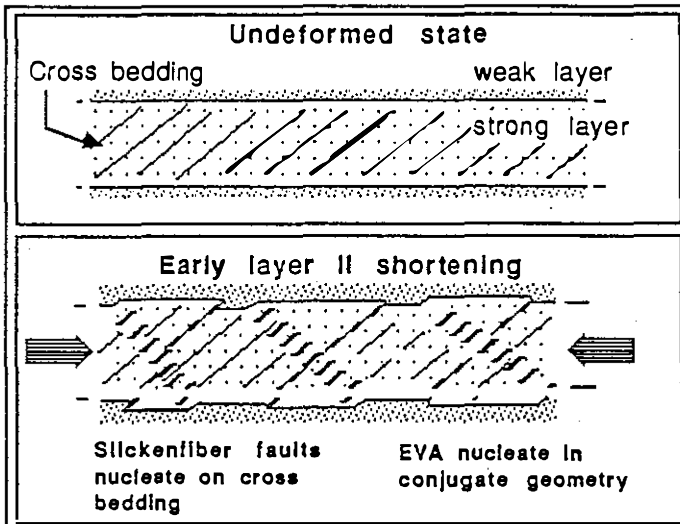

Figure 7. SFF (on cross bedding) and EVA form early conjugate arrays that may have accommodated layer parallel shortening.

As buckling took place, theoretical models of folding would argue for a number of temporal and spatial relationships among mesostructures. The changes to be expected as a function of time (or degree of folding) would best be observed in the limbs of the folds where the rotation of bedding has progressively changed the orientation of the bed with respect to the regional stress field. During progressive folding the competent or higher viscosity layers, acting as stress guides, rotates the local or intrabed stress axis of maximum compression along with the fold limb away from the subhorizontal regional field. With respect to an internal reference frame (fixed to bedding) the rotation of the stress axis is slightly in the opposite direction. This is because the beds are rotated more than the stresses. With progressive tightening of the fold, the ability of the high viscosity beds to guide the stresses in the limbs of the folds are lost and the local axis of maximum compression 'relaxes' back to a direction closer to the regional or 'far' field orientation. This complex rotational history explains two attributes seen in the Green Pond stage I paleostress data. First, as discussed by Mitchell and Forsythe (1988), it explains the systematic asymmetry in the variations of plunge observed for the maximum principal stress determined from EVA data of east and west dipping limbs. For each limb a gap in plunges was observed in the direction opposite to the dip of the beds (see Figure 12 of Mitchell and Forsythe, 1988). The dispersion and gap in the data is precisely what would be predicted by the combined external and internal rotations described above. Secondly, the rotations can help to explain why conjugate sets of EVA's on a given fold limb appear to have dramatically different forms. Once an EVA is nucleated it represents a plane of weakness that will have a protracted history of development during which limb rotation and folding is occuring. As the stresses shift with respect to bedding, the resolved normal and shear stress components on the conjugate arrays are constantly changing. As illustrated in Figure 8 dextral and sinistral EVA shear zones have internal rotations which will either be compatible (syn-rotational) or incompatible (anti-rotational) with the rotation of the stress axes (using an internal reference frame). Thus, on one limb we predict the early dextral EVA to be the syn-rotational array, while on the other limb the early sinistral EVA is the syn-rotational array. Synrotational EVA's have the plane of vein propagation rotating in a sense similar to the stress field, and the anti-rotational veins in the opposite sense. The anti-rotational EVA's should show pronounced sigmoidal shapes, with multiple or superimposed

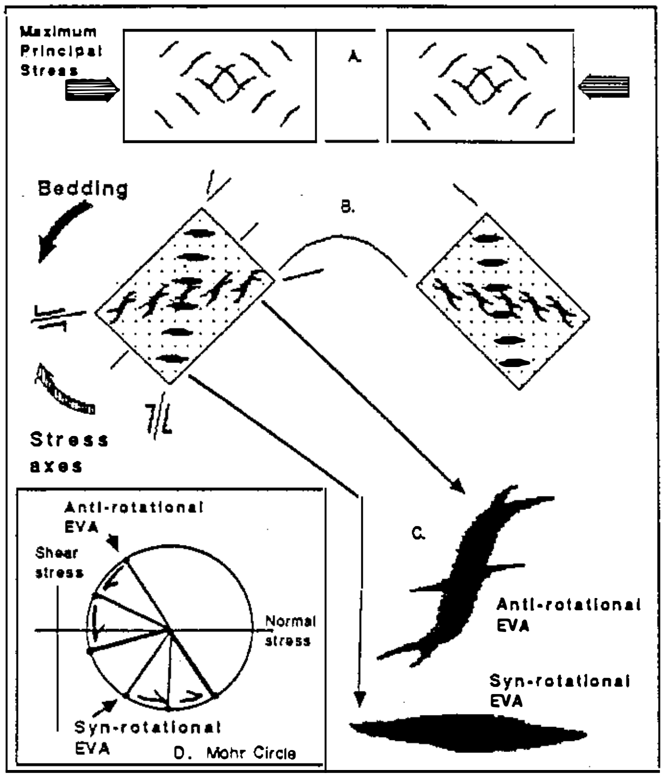

Figure 8. Progressive development of EVA during buckling. A. Early formed conjugate EVA's with thin and planar veins. B. Highly evolved EVA's in advanced stages of buckling that show asymmetry of character as a function of fold limb rotation (with respect to stress axes). C. closeup sketches of the two end-member EVA forms for evolved EVA on fold limbs. D. Mohr Circle illustrating the complex and variable stress histories seen by conjugate pairs of EVA's on fold limbs.

veins, while the syn-rotational EVA's would have more planar forms, and small aspect ratios.

Stage II (Noncoaxial Shortening). This phase is represented by numerous EVA's that indicate N-S to hinge parallel compression, as well as a second cleavage with a rough E-W orientation that trends obliquely across the fold structures. In a few localities (e.g. by Terrace Pond) SFF's have also been found that have multiple orientations of fibers. The superposition of the fibers indicates a similar late stage of N-S oriented regional maximum compression. Thus EVA, cleavage, and limited SFF data, all support a second phase of regional deformation in the outlier under a new N-S regional compressive stress field. Neither the main, stage I, folds nor the axially planar cleavage would have been appropriately oriented to accomodate shortening along this N-S axis. The folds, perhaps like the corrugations in a sheet of metal, would likely have acted as stress guides within this new N-S field. This effect, may in fact explain why there is a significant scatter of Stage II orientations from a N-S to a hinge parallel orientation.

SUMMARY

The abundant development of EVA's and SFF's in the more resistent layers of the Green Pond outlier have provided an opportunity to see how these layers have acted as stress guides throughout the deformation in the outlier, and how the stress fields have changed during the buckling history. The development of an axially planar cleavage in the 'softer' units permits a combined strain and paleostress visualization of the deformation.

Cleavages and folds, while representing long term 'finite' deformation, may not always accomodate the entire deformation. In the Green Pond outlier, as the regional stresses rotated to a new orientation, a second, new cleavage had to form. Therefore, neither the early nor the late cleavage could strictly speaking be said to represent the complete flattening strain in the rock. Each cleavage sees only the strain of its defomation phase. Similarly, the folding has principally accomodated one of two directions of regional shortening. Internally within the folds brittle and ductile structures were both needed to accomodate the deformation. The combination of these structures have given us the unique opportunity to see a combined strain and paleostress record of buckling as well as for the regional superposition of two non-coaxial phases of deformation.

ACKNOWLEDGMENTS

We would like to thank Anna Mitchell and Lisa Chisholm for their help and encouragement. We also thank Gregory Herman, Alexander Gates, and Dennis Weiss for their reviews and comments.

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ROAD LOG FOR THE GREEN POND OUTLIER OF NEW JERSEY AND NEW YORK: A MESOSTRUCTURAL LABORATORY

CUMMULATIVE MILEAGE	MILES FROM LAST POINT	ROUTE DESCRIPTION
0.0	0.0	Begin at Warwick, NY, Route 94 at Burger King on the south end of town. Proceed south on Route 94.
1.5	1.5	Left on Warwick 'Turnpike' at Lloyds Supermarket.
7.2	5.7	Right on Clinton Road at southern tip of Upper Greenwood Lake, NJ.
16.5	9.3	Left on Route 23 East.
18.1	1.6	Pass roadcut at top of hill.
18.6	0.5	Pass Charlotteburg Reservoir on the right and exit right at traffic light for a U-turn. Left on Route 23 westbound.
18.8	0.2	Exit right into rest area and park at far end. This is Stop 1 (two hours).

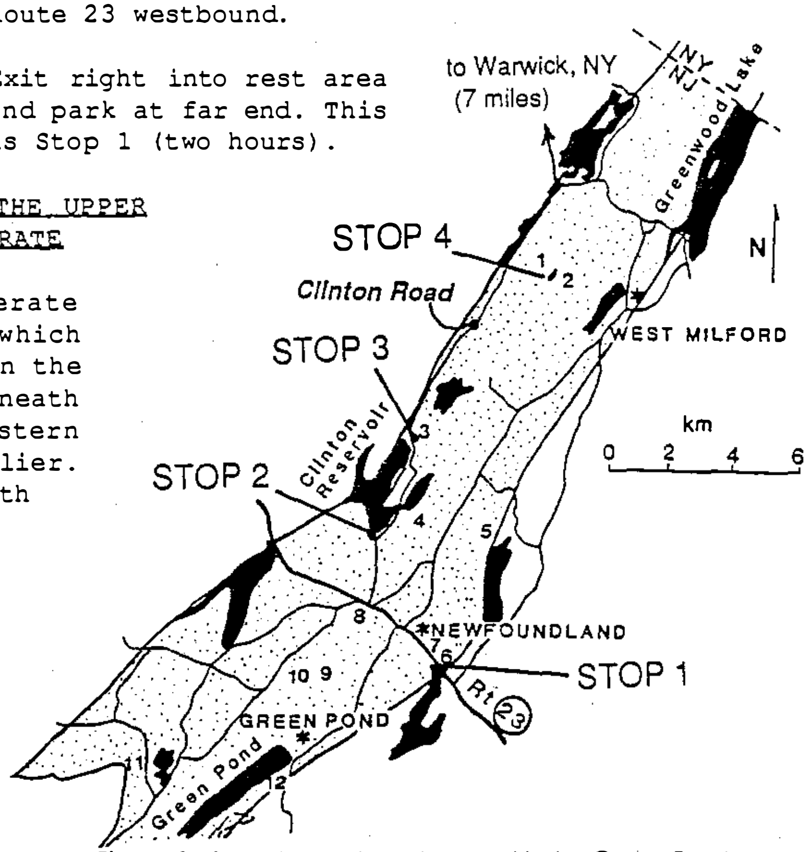


Figure 9. Location and road map with the Green Pond outlier shaded. Small number are data localities, and stop numbers are for this road log.

STOP 1. MESOSTRUCTURES IN THE UPPER GREEN POND CONGLOMERATE

The Green Pond Conglomerate forms the distant cliffs which probably lie unconformably on the Middle Proterozoic rocks beneath the lake. This is the eastern margin of the Green Pond outlier. Bedding dips northwest beneath the mountain. You are also located at the transition between the northern portion of the outlier defined by one large syncline and the central portion defined by three smaller wavelength anticline /syncline pairs. Superimposed on this NW dipping limb are two smaller anticline/syncline pairs that can be seen in this roadcut.

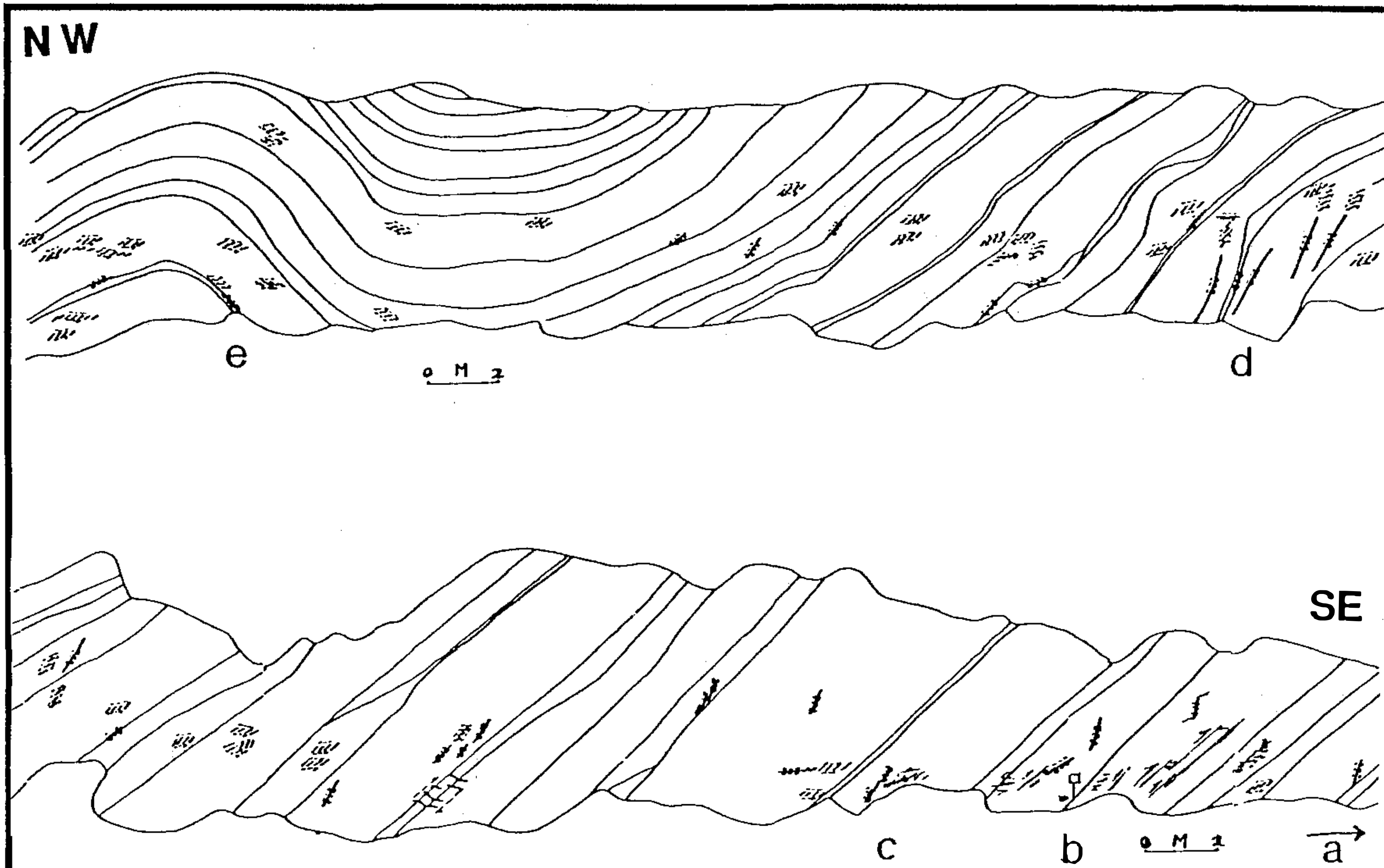


Figure 10. Sketch of the north side of the Rt. 23 outcrop (stop 1) with distribution of the EVA and SFF shown diagramatically. The bench mark and letters are reference point of interest in road log.

The purpose of this stop is to examine associations of the en echelon vein arrays, slickenfiber faults, and cleavage and to relate them to paleostress directions and progressive folding.

Walk along the west bound lane about 100 m to the north face of the exposure. Beginning at the east end of the outcrop, walk about 10 m until you see quartz veins at eye level.

a. En Echelon Vein Arrays. Several subhorizontal arrays of quartz veins occur in 'en echelon' arrangement (Fig. 11). Each array forms within a tabular shear zone and contains veins with a geometry that clearly indicates the sense of shear. In this case the top block moved westward (to the left). These features called en echelon vein arrays (EVA) have also been called tension gash arrays.

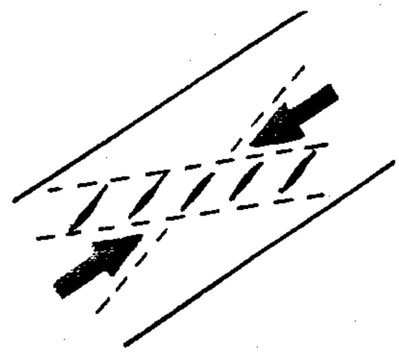


Figure 11. Illustration of how to determine the direction of maximum principal stress from a single en echelon vein array.

EVA Paleostress: The array and vein attitudes indicate (assuming type 1 geometry, see Figure 3b) that bedding parallel compression caused these particular EVA's to form. The direction of maximum compression that forms an EVA parallels the direction of greatest shortening which lies somewhere in the acute angle between the shear zone (array attitude) and a vein member (Fig.11).

Walk 30-40 m to the small Geodetic Survey sign located above a Bench Mark. See Figure 10 for location.

b. Slickenfiber Faults. Examine the many minor faults (Fig. 12) with lineations defined by mineral fibers, called slickenfiber faults (SFF's ; terminology of Wise, et al., 1985). Profiles of these features show overlapping fibers and give a sense of displacement characteristic of SFF's . Note that the quartz filled pull apart structures along some of of the SFF's give a reverse sense of displacement. The "rough-smooth" method (Durney and Ramsay, 1973) also indicates reverse displacement. The

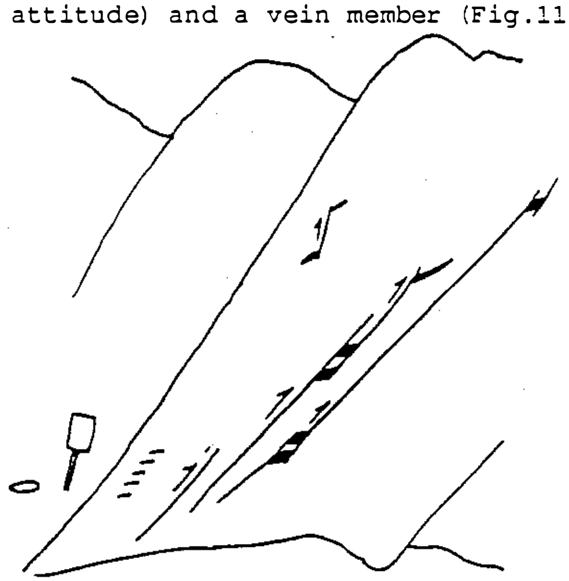


Figure 12. Field drawing showing SFF that nucleated on cross bedding. Pull aparts confirm motion inferred from the SFF fibers.

exposed footwall surface feels smooth to the hand when rubbed upwards along the surface, therefore the hanging wall moved up defining a reverse motion. The rough-smooth technique is reliable for SFF's because fibers always lie in plates that are overlapped in one direction like the shingles on a roof.

SSF Paleostress: Most SFF surfaces here developed on crossbedding surfaces dipping more steeply to the northwest than the bedding. Thus bedding parallel compression indicated by the EVA's, may also represent a conjugate shear system with SFF's following steep planes of weakness in the rocks (crossbeds oriented suitably for slip in the given stress field) and EVA's forming subhorizontal shears across rock anisotropies and in more massive rocks.

Walk 7 m west of the Bench Mark. See Figure 10 for location.

c. Conjugate EVA's and SFF's. As sketched in Figure 13, conjugate SFF's and EVA's occur in concert. A subhorizontal SFF overprints an EVA with the same sense of displacement (top block westward). This is conjugate to a steeply dipping bedding plane-SFF with veins whose displacement is top block upward to the east. These displacements are opposite but complementary such that the principal axes of shortening and compression bisects the two shears and is at a low angle to bedding. Lying in the intersection of the two shears is a quartz vein which is subparallel to the compression direction. This vein is common in conjugate EVA's and is a reliable indicator of paleostress directions.

Progressive Strain: Note the subhorizontal EVA veins in Figure 13 are sigmoidally shaped indicating progressive shear whereas, the steep EVA veins are straight, dilated, and incompletely filled with large quartz crystals (comb structure). This conjugate association can be seen in other places and may represent a slight rotation of the principal axis of compression towards the sigmoidally deformed EVA's (Fig.14). Such a rotation of the stress field, following the initial development of conjugate EVA, would substantially dilate one set of veins. This set will have straight tips which propagate at high angles to their array (synrotational EVA's of Fig. 8). The other set of veins will have curved tips that propagate at low angles to the array (antirotational EVA's of Fig. 8).

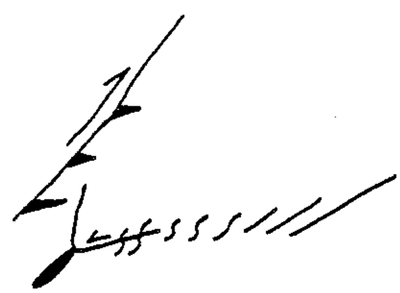


Fig. 13. Sketch of conjugate shears (Stop 1c).

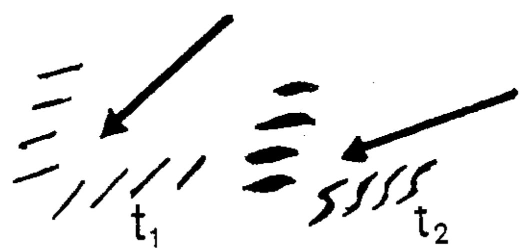


Fig. 14. 2 stage rotation (T1 and T2) for forming asymmetric EVA's.

Walk about 34 m to a position where purple rocks project slightly closer to the curb. See Figure 10 for location.

d. EVA and SFF Links. Again, SFF's occur on steep crossbeds, but this time several SFF's at eye level link up with EVA's higher in the outcrop to form a continuous shear zone with a mutual reverse sense of displacement (Figure 15). Furthermore, the same EVA's are paired with subhorizontal EVA's of the conjugate orientation. This is an outstanding association that clearly indicates the complimentary association of SFF's and EVA's in this area, and the conclusion that many of these features formed in response to maximum compression directed subparallel to bedding.

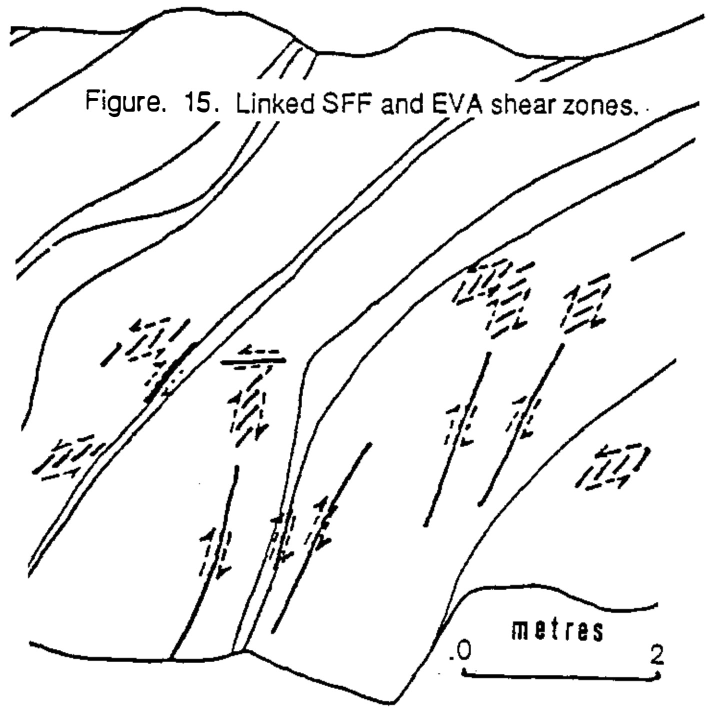


Figure. 15. Linked SFF and EVA shear zones.

Importance of microcracking: What controls whether a shear zone will develop into a SSF or EVA? Certainly planes of weakness in the rock favor SFF development, but sometimes SFF's cut across rock anisotropies. We postulate that the microcracking distribution prior to rock failure determines the thickness of the shear zone and therefore whether SFF's or EVA's will develop (Fig. 16). For example, the linked SFF and EVA shear zones at this site may reflect an incipient zone of microcracking that was narrow at the base where the SFF end developed and wider at the EVA end.

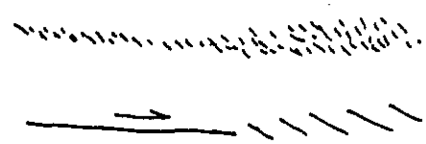


Figure 16. Sketch illustrating a possible cloud of early microcracks which may control eventual SFF or EVA geometries.

Walk about 30 m to a southeast dipping limb and the core of an anticline. See Figure 10 for location.

e. Mesostructures Record Progressive Folding History. Beginning here two anticlines and synclines can be viewed in four sections (the median exposes two sections). Veining here is profuse and complicated. Some EVA's in the hinge zone are bent or curved with pronounced sigmoidal shapes. These veins are interpreted to have been formed early and then incrimentally deformed during folding. In the median exposure (facing the west bound lane), there is a complex set of crosscutting veins that reflects rotation of the veins with respect to stresses during buckling. A late vein found here is suggestive of a late subhorizontal compression subnormal to a steep cleavage in an adjacent bed. Also in the cleaved bed (purple bed at the end of the median), is a conjugate EVA with compression bisecting the obtuse angle. Dissolution of veins at cleavage selvages confirms that this conjugate EVA was later shortened during cleavage development. Also, the veins of the subhorizontal EVA are sigmoidally deformed while the veins of the steep EVA are comparatively straight and dilated (as at site c.). This suggests that the principal stress field rotated away from bedding parallel compression as folding progressed (model shown in Fig. 8). This deformed zone may be very near the fault tip of a buried northwest-dipping reverse fault. It also parallels a steep northwest dipping reverse fault exposed about 100 m to the northwest in the median. In addition to the fault, two anticline-syncline pairs can be seen from both traffic lanes when walking northwestward along the median.

		Return to vehicle and proceed west on Route 23.
20.6	1.8	Right onto Clinton Road.
21.8	1.2	Left on dirt road just before stream bridge and opposite old stone furnace. Proceed around the curve and up the hill for a short distance.
21.9	0.1	Park at the hilltop overlooking Clinton Reservoir and Dam. This is Stop 2.

STOP 2. MESOSTRUCTURES IN THE BELLVALE SANDSTONE

The Bellvale Sandstone floors the road shoulder. Bedding dips NNW towards the axis of the major syncline that extends northeastward into New York State. You are at a position on the fold near the hinge where bedding begins to wrap around the nose of the plunging syncline. The purposes of this stop are to examine EVA's, cleavage and their geometric relations.

Cleavages: A pronounced spaced cleavage strikes about N 80 E subparallel to bedding and a faint cleavage strikes about N 60 E subparallel to the fold axis (time permitting, cleavages and bedding attitudes are more clear at the spillway bridge on Clinton Road, northern stream bank).

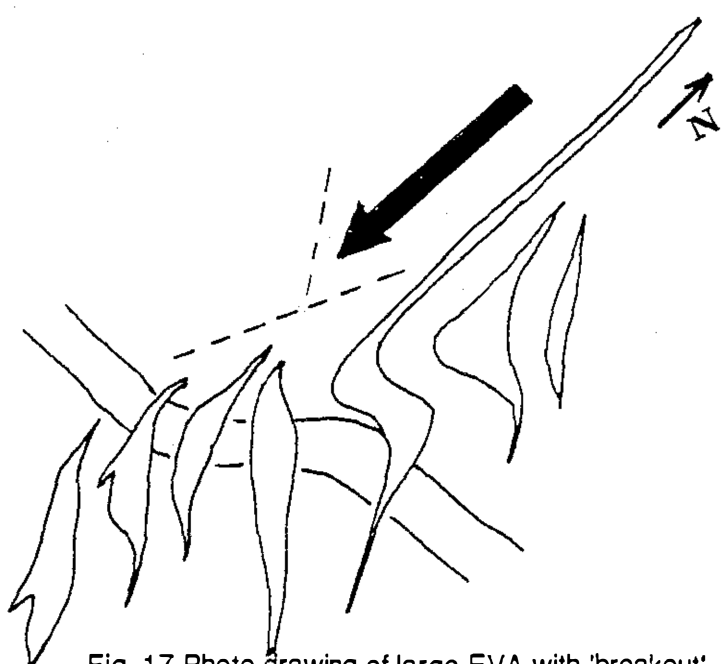


Fig. 17 Photo drawing of large EVA with 'breakout' veins paralleling the acute bisectrix. Pre-existing horizon (transverse band) is displaced sinistrally.

En Echelon Vein Arrays: A large EVA seen here indicates N-S shortening and compression (using the vein/array bisectrix rule for type 1 geometry). This stress field is roughly normal to the spaced cleavages and is postulated to represent a second discrete non-coaxial phase of deformation in the Green Pond outlier. Also note the occasional occurrance of unusually long vein tips extending beyond the EVA shear zone (Fig. 17). These veins when propagating out of the shear zone commonly curve into parallelism with the bisectrix of the vein and its array. This is the direction of the inferred maximum compression. If microcracks in the shear zone in part control vein attitudes then veins that propagate faster than others breakout of the array and may be free of the internal constraints that control intra-EVA crack tips. Thus the breakout veins are probably reasonably good indicators of the paleostress field.

Slickenfiber Faults: SSF offset some veins, most with apparent left lateral displacements. However, some of these have fibers that actually indicate dip slip motions on NW dipping surfaces.

		Return to vehicles, and drive back down the hill
22.0	0.1	Left onto Clinton Road, and over the spillway bridge, where a water fall and stream cut exposure is visible on the right.
24.2	2.2	Left into parking area for a boat landing to the Clinton Reservoir (just after the road curves to the left). Walk from the northern entrance of the parking area across Clinton Road to small cliff exposures near the road. This is Stop 3.

STOP 3. MESOSTRUCTURES IN THE LOWER SKUNNEMUNK CONGLOMERATE

You are in the same syncline as that present at Stop 2 but on the opposite limb. Beds of the lower Skunnemunk dip steeply SE towards the syncline axis. Rocks vary in color (gray, red, purple) and lithology (conglomerate, sandstone, muddy siltstone), typical of the Bellvale-Skunnemunk gradational contact. The main purposes of this stop are to: 1) see two cleavages, and 2) see unusally large EVA's that record hinge parallel compression.

Cleavage: Two spaced cleavages are visible in some of these rocks. One strikes northeastward subparallel to bedding and the fold axis, the other strikes N 85 E. They have the same general strikes as the two cleavages at Stop 2 suggesting two distinct periods of shortening.

En Echelon Vein Arrays: EVA's record a variety of paleostress directions at this site. Figure 10 shows EVA data compiled from Stops 3 and 4 (SE dipping limb only). Some EVA's on this limb of the fold record NW trending (steeply plunging) layer-parallel compression, but other indicate compression at high angles to bedding. These likely record early and late stages, respectively, of a progressive folding history.

Some EVA's in this area record more northerly directed compression. These directions are approximately normal to the N 85 E cleavage.

Unusually large EVA's record NE-SW trending compression directed subparallel to the fold axis. They contain large gently dipping veins often dilated and incompletely filled. Veins are typically one meter long, one meter wide, and spaced up to two meters apart. They likely formed late in the folding history once the fold axis was well defined. They may be an anisotropic response to the N-S directed shortening by other EVA and compatible with the N 85 E cleavage.

Slickenfiber Faults: Most SFF's are bedding plane and cross bedding slip surfaces that record NW trending displacements that are compatible with major folding and NE trending cleavage development.

		Return to vehicles, Proceed north on Clinton Road.
28.5	4.3	Pull off along the road where the forest has been cleared for the gas pipeline. This is Stop 4 (strenuous hike to, and from, Terrace Pond of 1.4 miles round trip)

STOP 4. MESOSTRUCTURES IN THE UPPER SKUNNEMUNK CONGLOMERATE

You are located on the same SE dipping limb as at Stop 3, but in this case we are hiking up the hill to the east (up section) into the axis of the syncline. The purposes are to observe numerous SFF's, EVA's, and a distinct N 85E cross cleavage. Lock the vehicles, and hike eastward up the mountain along the right side of the pipe line clearing. Although the hike is perhaps strenuous to some, the geology and scenery (at Terrace Pond) makes this a worthwhile endeavor.

As you work your way up the ridge keep an eye out for the trail head of the Blue Dot Trail that will take us to Terrace Pond (0.3 miles up the gas line clearing). Before heading in, however, there are exposures under foot in the pipe line clearing near the trail head worth a bit of investigation.

Slickenfiber faults: The outcrops under foot in the pipe line clearing contain exposures of conjugate SFF's. Also, along the forest edge opposite the entrance to the trail is a slip surface worthy of close examination. Here, the SFF surface has two directions of fibers. Upon inspection you should be able to discern their relative timing. A N-S oriented set of fibers has been superimposed over a NW-SE oriented set, and suggests a clockwise shift in the axis of maximum compression.

En echelon vein arrays: The EVA's here largely record NW-SE trending shortening compatible with folding and most SFF's.

Now proceed into the woods on the Blue Dot Trail to Terrace Pond (0.4 miles). Along the way are a discontinous series of exposures within which various mesostructures can be examined.

Terrace Pond.

Cleavages: A distinctive spaced cleavage occurs throughout this area that trends obliquely to the fold axis at about N 80 E. It crosses both limbs and can be seen on both sides of the pond (see Figure 18 for cleavage and bedding) The far cliff descending into the pond is the opposite limb of the syncline (bedding: N 35 E, 75 NW). You are standing on the western limb that forms the west margin of the pond (bedding: N 22 E, 30 SE). The fold axis is submerged. Many mesostructures can be seen along the White Dot Trail that circles the pond. For example, on the west margin an excellent pair of intersecting conjugate EVA's record a near N-S trending compression compatible with cleavage formation in the same rocks (see Fig. 19).

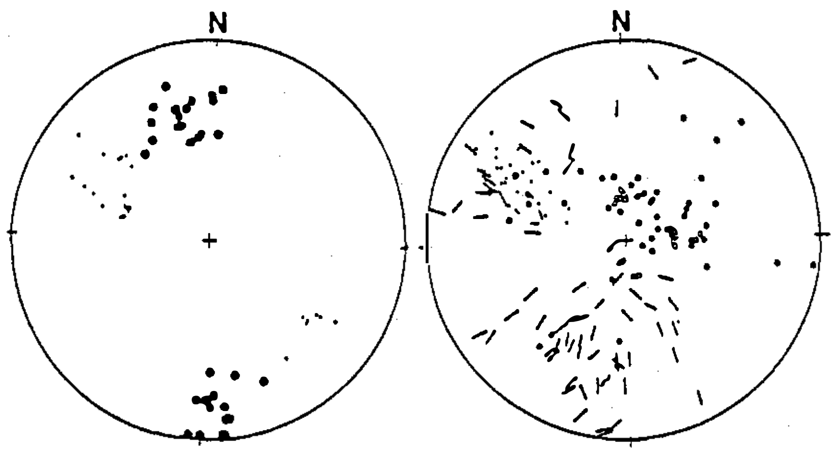


Figure 18. On the left is a stereoplot of poles to cleavage (big dots) & bedding (small dots) from Terrace Pond, and on right is a stereoplot of paleostress data from SE-dipping limbs of Stops 3 & 4. Symbols are as in Fig. 4b.

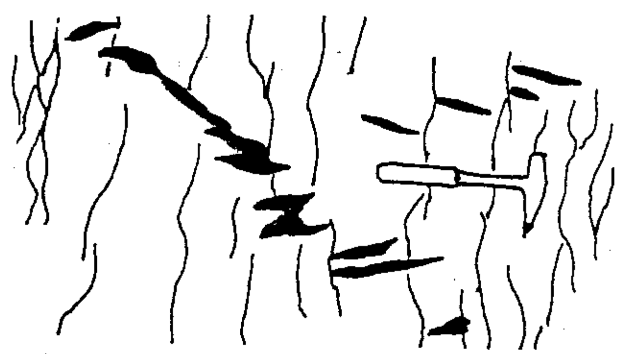


Figure 19. Sketch of field site with conjugate EVA's and spaced cleavage. Both record N-S shortening. Hammer for scale.

Field Excercise for Students. This location is an excellent choice for conducting a field-base lab in mesostructural analysis. Structure students can be taught how to recognize, measure, and interpret SFF's and EVA's along the trail on the way in to Terrace Pond. Then, working around the pond on the White Dot Trail they can collect their own data, first on the west limb, and then on the east limb. The data can be plotted later on stereonets and used for making an interpretation of the paleostress history in the area. Sample data spreadsheets, and a crib sheet for measuring EVA's are shown in Figure 20. The scenery and wilderness setting certainly adds a sense of spirit and adventure to what might otherwise be thought of as a rather tedious excercise.

Hike back to the vehicles, retracing our route along the same trails.

		Proceed northward on Clinton Road
30.1	1.6	Turn left onto Warwick Turnpike, at Upper Greenwood Lake, N.J.
35.8	5.7	Turn right onto Route 94 at Llyods Supermarket.
37.3	1.5	Return to Burger King at Warwick, NY.

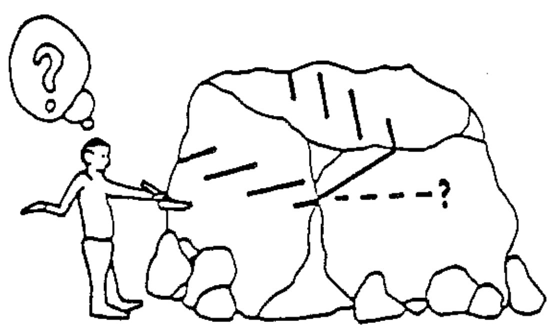
Figure 20. Procedures for the measurement of EVA

- Draw a rough map of trails, power lines, etc., and as you proceed plot your traverse so that you can locate your data collection sites.
- Find an array with at least three en echelon veins.
  Do not measure joints or isolated veins.
  Measure the EVA only if the rock appears to be firmly in place.
- At each locality measure and record onto the data sheet the following information.

1) Note the location number, the distance and direction from the previous locality, & plot the location on your field map.
2) Estimate the average vein length (roughly) for each array to be measured.
3) Take the strike and dip of a representative vein (if they are highly curved use the outer portion of a representative vein, but try not to use any vein tips that wander out of the main array).
4) Take the strike and dip of the array (usually difficult).
5) Assign a rough confidence level for each EVA measurement ; use these symbols:

good	fair	poor
+ $<5°$	Φ $<20°$	− $<60°$

6) Note the rock type, color, & bed thickness.
7) Take the strike and dip of bedding.
8) Take the strike and dip of cleavage (when present; and indicate cleavage type, e.g. slaty, pressure solution, etc.).
9) Sketch the geometry of the EVA & its relationships to bedding (if visible)
10) Make comments such as:
   * possible or definite conjugate EVA's
   * possibly tilted exposure
   * any association with SFF
   * any cross cutting veins, (EVA or SFF, indicate relative ages if apparent).



LOCATION NO.	AVG. VEIN LEN.	STRIKE & DIP OF VEIN	CONFIDENCE	STRIKE & DIP OF ARRAY	CONFIDENCE	ROCK TYPE, COLOR, THICKNESS	STRIKE & DIP OF BEDDING	STRIKE & DIP OF CLEAVAGE	CLEAVAGE TYPE	DRAWINGS

Sample data spread sheet

NOTES

LATE PALEOZOIC DEFORMATION IN THE RESERVOIR FAULT ZONE AND GREEN POND OUTLIER, NEW JERSEY HIGHLANDS

L.D. MALIZZI AND A.E. GATES

Department of Geology
Rutgers University
Newark, New Jersey 07102

INTRODUCTION

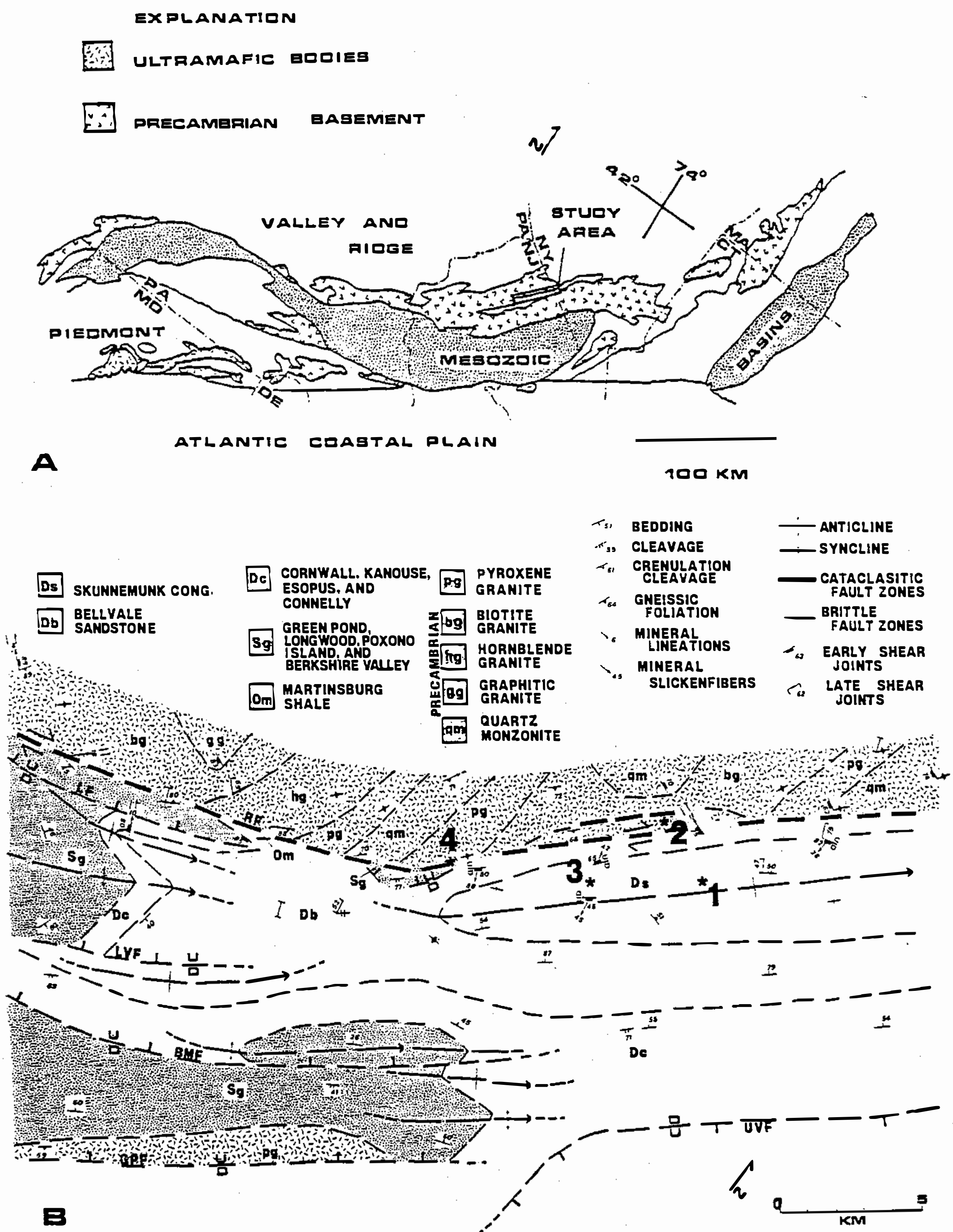
The New Jersey Highlands are part of the Precambrian Reading Prong which extends from Massachusetts to Pennsylvania (Fig. 1a). The field study area is in north-central New Jersey and contains the Reservoir Fault zone and Green Pond outlier.

The Reservoir Fault zone forms the boundary between the western side of the Paleozoic sedimentary rocks of the Green Pond outlier and Grenville gneisses of the New Jersey Highlands (Fig. 1b). The Paleozoic Green Pond outlier locally contains northwest-dipping reverse faults (Herman, 1987; Herman and Mitchell, 1989) and southeast-verging folds. These structures indicate southeast directed transport which is enigmatic with respect to the rest of the Appalachian foreland fold and thrust belt.

The major structure in the area is the northeast-striking Reservoir Fault zone. The fault zone is a high angle 100 m wide cataclasite and semi-brittle mylonite zone, which is composed predominantly of Grenville gneisses. Numerous theories have been presented for the deformational history of the Reservoir Fault zone including Mesozoic normal faulting (Lewis and Kummel, 1912; Ratcliffe, 1980), Proterozoic wrench faulting (Helenek, 1987), Alleghanian sinistral strike-slip faulting (Mitchell and Forsythe, 1988), and Alleghanian reverse faulting (Herman and Mitchell, 1989). Recent work by Malizzi and Gates (1989) indicates that the Reservoir Fault zone and Green Pond outlier form the east side of a positive flower structure that formed through Late Paleozoic dextral transpression with later sinistral strike-slip reactivation.

STRATIGRAPHY

The bedrock stratigraphy in northern New Jersey includes a sequence of Grenville gneisses (locally known as the Byram and Losee gneisses) and Middle Paleozoic clastics and carbonates. The Middle Proterozoic gneisses in the New Jersey Highlands are the oldest units in the study area (Fig. 1b) and have been dated at 913 ma by Rb/Sr whole rock analysis of the Canada Hill granite in New York (correlative of the Byram gneiss of New Jersey) (Helenek and Mose, 1984). The Precambrian units consist of well foliated, medium-grained hornblende, pyroxene, and biotite quartz monzonite gneisses and well foliated, medium-grained hornblende, pyroxene, graphite, and biotite granite to alkali-feldspar granite gneisses. The gneisses contain accessory apatite, magnetite, garnet, zircon, sphene, and ilmenite with secondary and joint-filling epidote, chlorite, quartz, and hematite. The granite gneisses contain zones of locally abundant scapolite, apatite, diopside, potassium feldspar, plagioclase, titanite, and tremolite indicating a possible calc-silicate protolith.


EXPLANATION
ULTRAMAFIC BODIES
PRECAMBRIAN BASEMENT
N
42°
74°
STUDY AREA
VALLEY AND RIDGE
NY
PA NJ
PA
MD
PIEDMONT
MESOZOIC
BASINS
DE
ATLANTIC COASTAL PLAIN
A
100 KM
Ds SKUNNEMUNK CONG.
Db BELLVALE SANDSTONE
Dc CORNWALL, KANOUSE, ESOPUS, AND CONNELLY
Sg GREEN POND, LONGWOOD, POXONO ISLAND, AND BERKSHIRE VALLEY
Om MARTINSBURG SHALE
PRECAMBRIAN
pg PYROXENE GRANITE
bg BIOTITE GRANITE
hg HORNBLENDE GRANITE
gg GRAPHITIC GRANITE
qm QUARTZ MONZONITE
BEDDING
CLEAVAGE
CRENULATION CLEAVAGE
GNEISSIC FOLIATION
MINERAL LINEATIONS
MINERAL SLICKENFIBERS
ANTICLINE
SYNCLINE
CATACLASITIC FAULT ZONES
BRITTLE FAULT ZONES
EARLY SHEAR JOINTS
LATE SHEAR JOINTS
LF
RF
LVF
BMF
UVF
GPF
1
2
3
4
B
0
5
KM

Localized cataclasite zones occur in the quartz monzonite gneiss, biotite granite gneiss, hornblende granite gneiss, and pyroxene granite gneiss. Age relations between the gneisses are unknown but all are cut by zircon, biotite, and monazite bearing granitic pegmatites.

The Paleozoic sedimentary rocks of the Green Pond outlier (Table 1) overlie the Middle Proterozoic rocks. The Green Pond outlier consists of Ordovician-Devonian sedimentary units, typical of the Valley and Ridge Province to the west (Kummel and Weller, 1902). The Ordovician Martinsburg Formation is the oldest unit in the study area. The Martinsburg Formation is overlain by the Middle Silurian Green Pond conglomerate (correlative of the Shawangunk Formation) and Longwood shale (correlative of the Bloomsburg Formation) (Wolfe, 1977). Upper Silurian units include the Poxono Island Formation and Berkshire Valley Formation. The lower Devonian units include the Connelly conglomerate (correlative of the Oriskany Formation), the Esopus Formation, and the Kanouse Formation which are overlain by the Cornwall shale. The Bellvale sandstone and Skunnemunk conglomerate (correlative of the Catskill formation) are the Middle Devonian units. All of the sedimentary units contain en-echelon quartz vein arrays.

The Precambrian gneisses of the New Jersey Highlands were formed from unknown protoliths by a Grenvillian tectonothermal event (Helenek, 1987). The gneisses are overlain by the Ordovician shale which represents sediment that filled a west-facing starved basin (Pollack, 1975). Fluvial molasse deposits shed from a highlands to the southeast, subsequent to the Taconic orogeny, produced the Middle Silurian clastic units. A Late Silurian and Early Devonian marine transgression occurred forming a shallow sea (Wolfe, 1977). Marine carbonates and sandstones were deposited at the margin of the sea. Uplift caused by the Acadian orogeny produced a Middle Devonian marine regression. During the marine regression deltaic clastic units were deposited, and are the youngest units in the Green Pond outlier.

PRE-EXISTING STRUCTURE

A pervasive gneissic foliation and isoclinal folds are the earliest recognizeable structures in the Grenville gneisses. The foliation is nearly vertical and north- to northeast-striking. The foliation is defined by aligned pyroxene, biotite, hornblende, and graphite along with quartz and feldspar ribbons depending upon lithology. It has been proposed that the gneissic foliation represents a medium to high-grade Proterozoic mylonite zone in the granitic units (Hull et al., 1986) but a thorough investigation of Grenville deformation is beyond the scope of this study. The gneisses also contain isoclinal folds in the foliation with northeast-striking, vertical axial planes. The Grenville gneisses are cut by northeast-striking, nearly vertical granitic pegmatites along the Reservoir Fault.

Fig. 1. (A) Regional map of the Central Appalachians indicating the study area. (B) Geologic map of the Reservoir Fault and Green Pond outlier (modified after Herman and Mitchell, 1989). RF-Reservoir Fault, LF-Longwood Fault, LVF-Long Valley Fault, BMF-Brown Mountain Fault, GPF-Green Pond Fault, and UVF-Union Valley Fault. Fault movement sense indicators as indicated.

TABLE 1. Stratigraphy (Barnett, 1976; Herman and Mitchell, 1989; this study).

DEVONIAN	Skunnemunk Conglomerate- Thin to very thick bedded, medium-grained quartz pebble conglomerate with a red-purple medium-grained sandstone matrix with medium-grained red sandstone interbeds. Conglomerate is locally crossbedded. 915 m thick.
	Bellvale Sandstone- Thin to very thick interbedded gray, medium-grained sandstone and gray shale. Locally fossiliferous and crossbedded. 600 m thick.
	Cornwall Shale- Thin to thick bedded, fine-grained fissile black shale interlayered with laminated gray siltstone. Moderately fossiliferous. 300 m thick.
	Kanouse Sandstone- Medium to thick bedded, gray to tan conglomerate and coarse to fine-grained graded sandstone. 15 m thick.
	Esopus Formation- Thin interlayers of gray mudstone and medium-grained sandstone. Fossiliferous. 60 to 100 m thick.
	Connelly Conglomerate- White, medium-grained quartz pebble conglomerate with a tan, medium-grained sandstone matrix. 12 m thick.
SILURIAN	Berkshire Valley Formation- Thin-bedded limestone with interlayers of gray, intraformational dolomitic breccia. Thickness unknown.
	Poxono Island Formation- Medium-bedded gray dolomite interlayered with thin bedded medium-grained calcareous sandstone. 80 to 130 m thick.
	Longwood Shale- Medium-bedded purple shale with interlayers of red crossbedded medium-grained sandstone. 100 m thick.
	Green Pond Conglomerate- Medium-grained quartz pebble conglomerate with a medium-grained sandy matrix and silica cement. The conglomerate contains thin bedded interlayers of crossbedded sandstone. 300 m thick.
ORD.	Martinsburg Shale- Black, slaty fine-grained shale with thin beds of medium-grained sandstone. Moderately fossiliferous and crossbedded. Thickness unknown.
PRECAMBRIAN	Granite Pegmatites- Very coarse-grained quartz, plagioclase, and microcline pegmatites dikes with minor biotite. Accessory zircon and apatite. Secondary epidote, hematite, and monazite. 1 to 10 m thick.
	Quartz Monzonite Gneisses- Medium-grained hornblende, pyroxene, and biotite, quartz monzonite gneisses with quartz, plagioclase (An-37%), and microcline. Accessory magnetite, apatite, garnet, sphene, and zircon with secondary chlorite.
	Granite Gneisses- medium-grained hornblende, pyroxene, graphite, and biotite granite to alkali-feldspar granite gneisses with microcline and plagioclase (An-38%). Accessory sphene, apatite, zircon, and magnetite with secondary epidote, chlorite, monazite, and hematite. Locally contains abundant scapolite, tremolite, and titanite.

LATE PALEOZOIC STRUCTURES

Late Paleozoic structures in both the Green Pond outlier and Reservoir Fault zone consist of dextral strike-slip shear zones and coeval folds, cleavage, and reverse faults. Northeast-striking dextral strike-slip kinematic indicators are present in both the Paleozoic and Precambrian units within the Reservoir Fault zone. The Silurian Green Pond conglomerate contains semi-ductile Type II S-C mylonites (Lister and Snoke, 1984) within the Reservoir fault zone. Exposures exhibit a well defined C-surface and a poorly defined S-surface typical of a Type II S-C mylonites (Fig. 2a). The Green Pond conglomerate also exhibits northwest-striking dextral shear bands that cut the C-planes (Fig. 2a). S-planes are defined by the quartz grains of the rock matrix and C-planes are defined by quartz ribbons.

The sheared Green Pond conglomerate also contains microstructural northeast-striking dextral strike-slip kinematic indicators. Recrystallized σ-Type porphyroclasts (Simpson and Schimd, 1983; Simpson, 1986; Passchier and Simpson, 1986) in quartz ribbons with rotated tails and dragged deformation bands in quartz ribbons (Fig. 2b) indicate dextral movement. The fabric in the quartzite contains a foliation and shear bands aligned subparallel to bedding that indicate movement sense. Similar structures were described in the Moine thrust zone of northwest Scotland by Bowler (1989). Northwest- and northeast-striking conjugate deformation bands and deformation lamellae in quartz grains (Heard and Carter, 1968) occur in the Type II S-C mylonites and indicate a dextral shear sense. Deformation bands and lamellae develop when the rate of dynamic recrystallization of the mineral is low with respect to the shear strain rate (Passchier and Simpson, 1986; Simpson, 1986). The maximum compression direction is indicated by the acute angle between the deformation bands. A west-northwest to east-southeast maximum compression direction is indicated by the conjugate deformation bands.

Sheared Bellvale sandstone within the Reservoir Fault zone consists of a 5 m wide zone of semi-ductile cataclasite and thin minor faults filled with hematite and chlorite. Northeast-striking dextral movement is indicated by offset quartz veins and sandy laminations. Fine-grained kinked chlorite indicates an east-southeast to west-northwest maximum compression. Drill core samples from along the Reservoir Fault in the Cornwall shale contain northeast-striking dextral strike-slip and reverse shear indicators. The shear indicators from the drill core samples include dragged and offset quartz veins and slickenfibers.

The Grenville gneisses within the Reservoir Fault zone also exhibit northeast-striking dextral strike-slip indicators. The Reservoir Fault zone consists of cataclasite with epidote, hematite, and chlorite in the shear planes. The cataclasite contains microcline, quartz, and plagioclase grains with pull-apart textures (Fig. 3a). The cataclasite also contains thin minor faults filled with calcite, epidote, and chlorite that offset quartz and plagioclase grains. The gneisses also contain coarse-grained chlorite fish and fine-grained recrystallized muscovite fish.

The Grenville gneisses contain horizontal mineral lineations indicating northeast-striking strike-slip movement. Dominant northeast-striking and minor northwest-striking conjugate shear joints (Fig. 4a) with horizontal slickenfibers indicate dextral and sinistral movement, respectively (Fig. 3b). The shear joints were analyzed using a technique developed by Hardcastle

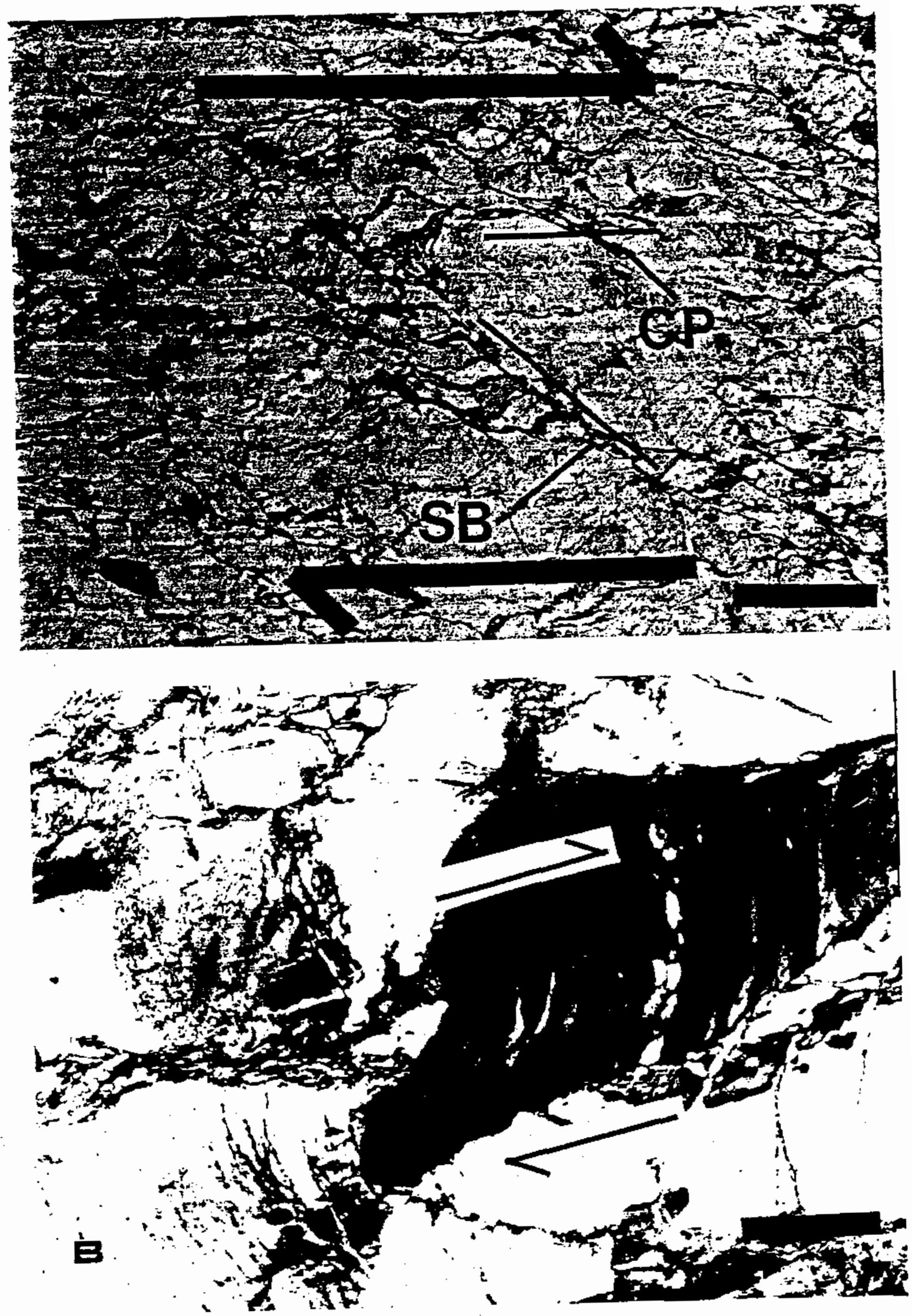


Figure 2. (A) Northeast-striking C-plane (CP) and northwest-striking shear band (SB) in a Type II S-C mylonite from the Silurian Green Pond conglomerate from within the Reservoir fault zone, both indicating dextral drag. Bar scale 10 mm. (B) Deformation bands in a Type II S-C mylonite from the Silurian Green Pond conglomerate in the Reservoir fault zone indicating northeast-striking dextral shearing. Bar scale 1 mm.

Figure 3. (A) Dextral chlorite fish in Grenville granite gneiss from within the Reservoir fault zone. Bar scale 1 mm. (B) Offset quartz ribbon indicating northwest-striking sinistral faulting. Bar scale 5 mm.

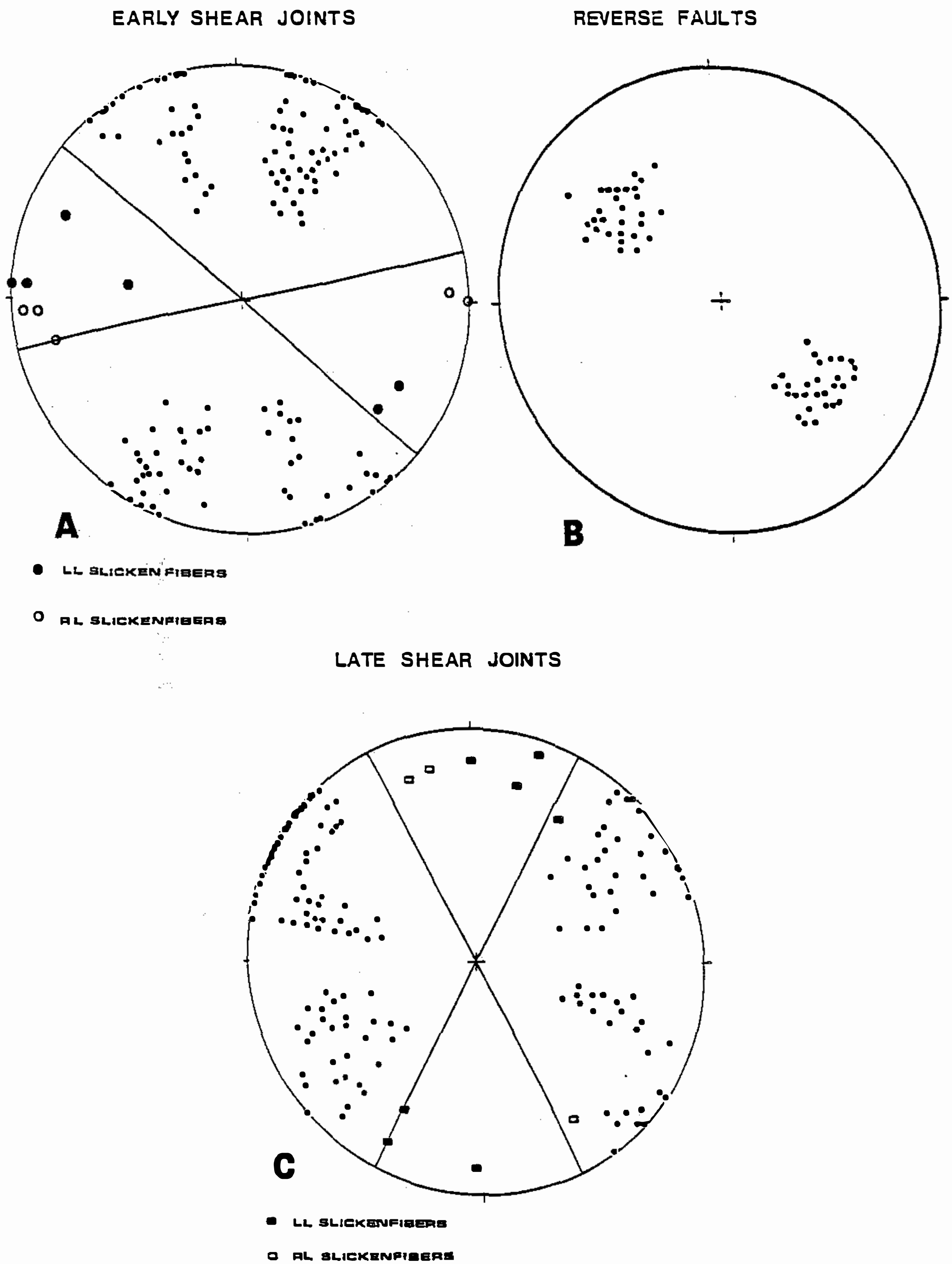
(1989) from methods of Reches (1987), where tensor configurations for fault populations are derived using a least squares regression solution. Only well exposed faults containing slickenfibers with well developed steps were used. A total of 19 minor faults were analyzed using this method and yielded N80W as the average maximum stress direction. In addition, plagioclase kink geometry (Gay and Weiss, 1974) indicates an almost east-west maximum compressive stress. The microstructures and tensor analysis yield very similar stress directions.

The adjacent Green Pond outlier locally contains northwest-dipping reverse faults (Herman, 1987; Herman and Mitchell, 1987; Mitchell and Forsythe, 1988) and southeast-verging folds (Fig. 1b). The reverse faults exhibit top to the southeast movement indicated by fibrous, stepped quartz and chlorite slickenfibers. Detailed mapping indicates apparent steepening of reverse faults with depth. West-side up near vertical reverse movement also occurred on the Reservoir Fault. Southeast-verging folds are indicated by axial plane orientations and bedding-cleavage relations. The mesoscopic folds are assymmetric with northwest-dipping axial planes (Bizub and Hull, 1986; Mitchell and Forsythe, 1988). The sandstones and conglomerates in the outlier contain pervasive solution cleavage sub-parallel to bedding, but the shales exhibit slaty cleavage.

Northeast-striking and northwest-dipping reverse movement indicators are evident in drill core samples of Devonian Cornwall shale from along the Reservoir Fault zone. Mesoscopic conjugate reverse faults, referred to here as small scale keystone structures, consist of northeast-striking, northeast- and northwest-dipping reverse faults (Fig. 4b). The small scale keystone structures range from 0.2 to 10 m wide but apparently reflect large scale structures as well. The sedimentary rocks of the Green Pond outlier contain the majority of the keystones in the study area. Slickenfibers indicate reverse movement sense. In addition, offset quartz veins and quartz pebbles indicate reverse movement in the Skunnemunk conglomerate. The Grenville gneisses also contain small scale keystone structures. Reverse movement sense is indicated by offset gneissic foliation and stepped slickenfibers.

LATEST STRUCTURES

This latest deformation overprints the earlier structures. The Reservoir Fault was reactivated as a sinistral strike-slip brittle fault. Horizontal slickenfibers on northeast-striking faults indicate this latest strike-slip movement. Kinked brittle plagioclase grains and northeast-striking sinistral offset grains are evident in thin sections of the gneisses. En-echelon vein arrays in the Green Pond outlier also indicate northeast-striking sinistral strike-slip movement (Mitchell and Forsythe, 1988). Minor northeast- and northwest-striking conjugate shear joints with horizontal slickenfibers indicate sinistral and dextral strike-slip movement, respectively (Fig. 4c). The Late Paleozoic shear joints are offset by the latest shear joints on the meso- and microscopic scale. The latest shear joints were also analyzed using the Hardcastle (1989) method. A maximum compression of N00E was derived from a total of 30 minor faults. Kinked plagioclase geometry (Gay and Weiss, 1974) and east-west crenulation cleavage also support the late north-south maximum compression.



EARLY SHEAR JOINTS
REVERSE FAULTS
A
B
LL SLICKEN FIBERS
RL SLICKENFIBERS
LATE SHEAR JOINTS
C
LL SLICKENFIBERS
RL SLICKENFIBERS

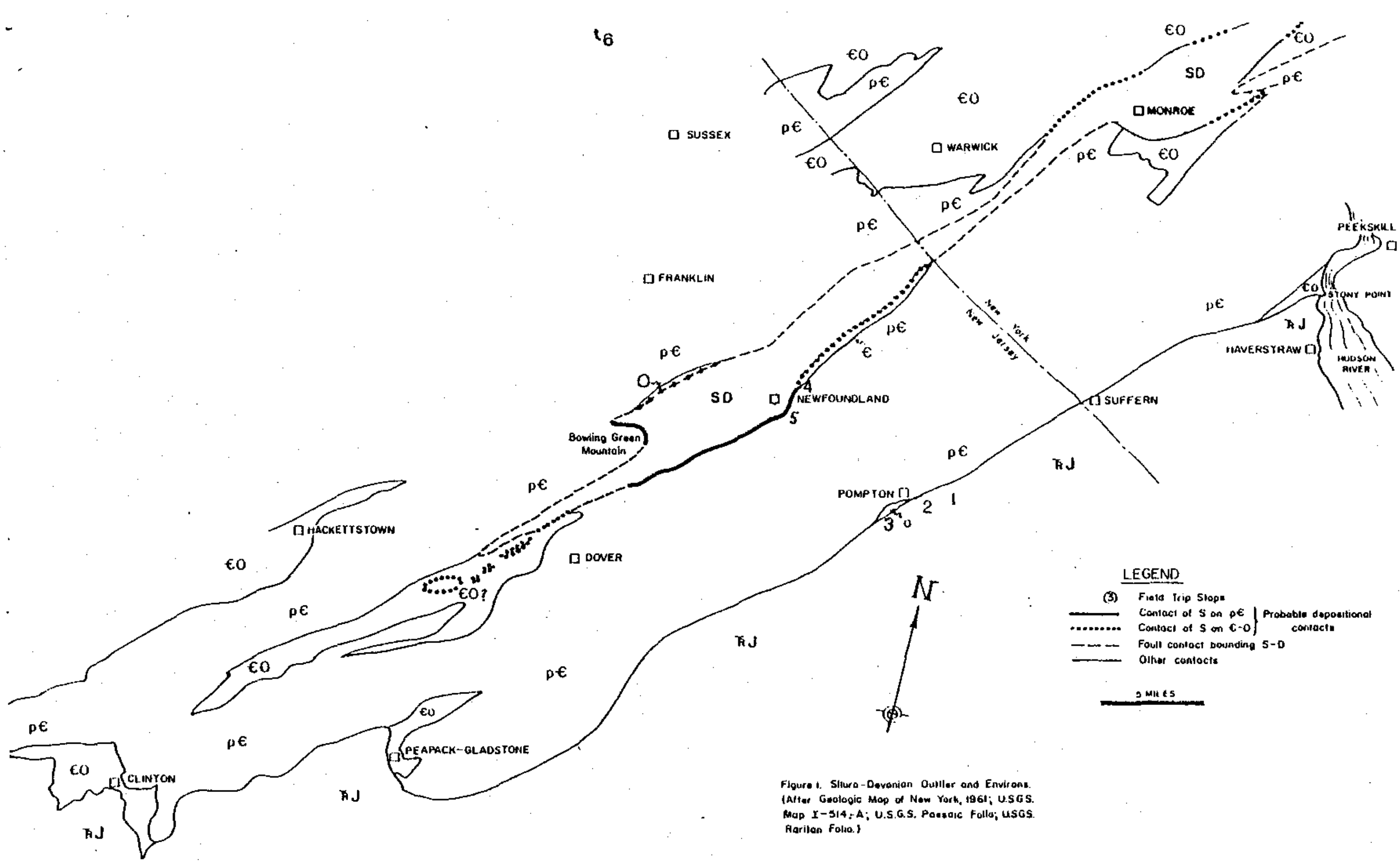


Figure 1. Siluro-Devonian Outlier and Environs. (After Geologic Map of New York, 1961; U.S.G.S. Map I-514-A; U.S.G.S. Passaic Folio; U.S.G.S. Raritan Folio.)

vesicular basalt, presumably Jurassic, and the remainder are sedimentary rocks. A large number of the latter are siliciclastic and most probably come from the upper part of the Paleozoic cover (Siluro-Devonian). The rest are Cambro-Ordovician dolomites from the lower part of the cover. You may be the judge of their relative proportions. A few clasts of Precambrian crystallines have also been found here.

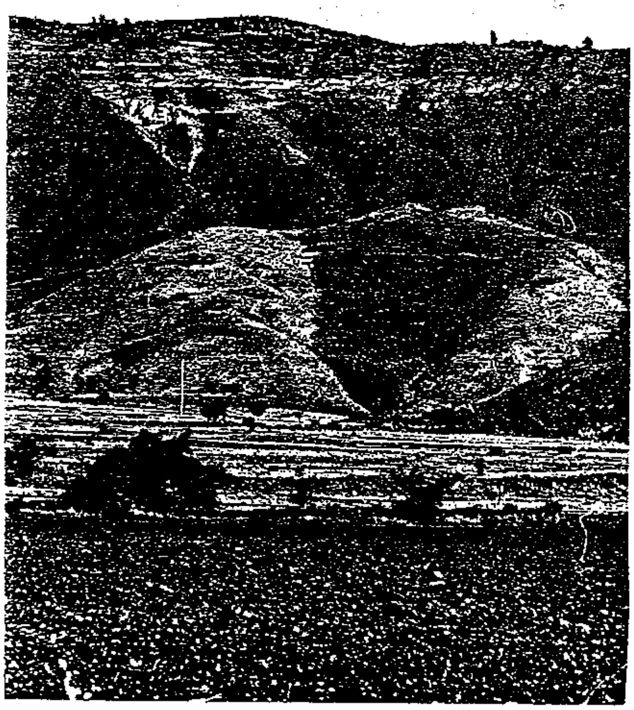
The fault-scarp immediately to the west must have had a cover of Jurassic basalt, beneath which was the Cambrian to Devonian sedimentary sequence, with only a bit of the Precambrian exposed, perhaps at the bottom of the deep valley carved by the river that built the fan.

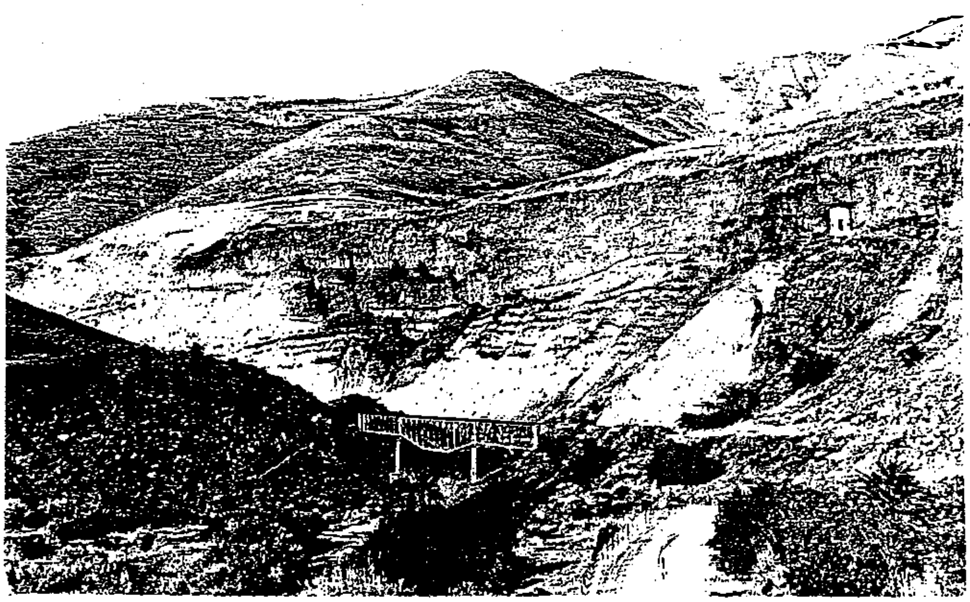
The second conglomerate lies just beneath the Hook Mountain Basalt, and is thus younger than the first. One would expect the stream that formed it to have tapped deeper levels in the adjacent highlands. The fault scarp was continually renewed by uplift, as evidenced by the repeated development of alluvial fans. This, accompanied by erosion, progressively stripped-down the sedimentary cover. Indeed, the basalts are missing in this younger conglomerate, and the Cambro-Ordovician dolomites seem more numerous than the siliciclastics. There are, however, no Precambrian crystallines. Perhaps the stream valley was less deep than the one which produced the earlier fan. It should be noted that the two fans are separated by only 2.5 miles. Thus lithologic differences in the source area are more likely to be related to vertical than to horizontal changes. (A late fanglomerate at Montville, New Jersey, above the Hook Mountain Basalt, is rich in Precambrian cobbles.)

A MODERN STREAM AND THE PRE-JURASSIC BASEMENT

At Riverdale, New Jersey, not far from the last conglomerate, and adjacent to the border fault, is a unique exposure of the pre-Mesozoic basement in the bed and banks of the Pequannock River. This site also affords us the opportunity of comparing the bed load of a modern stream with those of the Jurassic streams that drained the same area.

The pre-Mesozoic rocks here are on the down-dropped side of the border fault. Because the Mesozoic basin-fill is some eight miles thick, the exposure of this rock implies its presence on a less down-dropped sliver of the graben floor. It is a black phyllite identical in lithology and structural position to the Ordovician Annsville Phyllite some 40 miles to the northeast along the border fault at Peekskill, New York (see Finks, 1968, p. 139 for original identification; the river, however, was misidentified as the Wanaque). It is appropriately intermediate in metamorphic grade between the correlative Manhattan Schist to the east and the Martinsburg Slate to the west. If the Green Pond Conglomerate immediately above the Ordovician extended this far east (and the large boulders of it in the fanglomerate only a mile and a half away imply that it was nearby) then this sliver must have been downdropped after the Green Pond Conglomerate had been removed by erosion. It seems less likely that such a resistant unit as the Green Pond would have been removed after downdropping. If this line of reasoning is correct, the time of downdropping would be well after the deposition of the second fanglomerate. Inasmuch as this fanglomerate is close to the top of the Jurassic basin-fill, the faulting could be late Jurassic or even later.





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The Pequannock River drains the upthrown block even as the Jurassic streams did. It does not form an alluvial fan because the scarp is much lower and the climate more humid. Its bedload is fairly coarse, but not so coarse as the Jurassic fanglomerates. Part of the coarseness of the Pequannock sediments is due to reworking of quite nearby glacial tills over which it flows. Most of the cobbles are of Precambrian gneisses and granites, in keeping with the wide present exposure of these rocks on the upthrown block. Green Pond Conglomerate cobbles are also common. The Pequannock flows over the outcrop of the Green Pond about ten miles upstream at Stop 5 of this trip, but at least some of them must come from much nearer glacial deposits. It would be an interesting exercise to determine how many of the cobbles show glacial faceting and striae.

The previous two stops sampled the exposed bedrock of the Jurassic indirectly, whereas the present stop does so directly. All three stops lie within a strip only 4.0 miles long and 1.5 miles into the basin from the main border fault (a position comparable to that of the viewer in Figure 2). The apparent absence (or paucity?) of Annsville Phyllite cobbles in the fanglomerate needs to be explained. Unless a major structural discontinuity existed along the main border fault, such as right-lateral fault transport (the nearest presently-exposed equivalent phyllites on the upthrown side of the Ramapo Fault are some 50 miles northeast in Dutchess County, New York) the Annsville should have been present nearby in the upthrown block. A small patch of a shaly equivalent is, indeed, present in the Outlier (Barnett, 1976). The least radical explanation is the ease with which the rock is broken up during transport. In fact, large shaly clasts of any kind are rare in the fanglomerates.

A SILURIAN STREAM VALLEY?

One of the intriguing mysteries of the Green Pond - Schunemunk Outlier is the fact that the Silurian Green Pond Conglomerate rests directly on the Precambrian in an area some 4 miles wide and 10 miles long between Kanouse Mountain, east of Newfoundland, and Bowling Green Mountain, west of Milton (Figure 1). At the latter locality the mapped contact is nearly horizontal and is unlikely to be a fault. (A thrust plane seems unlikely to follow exactly a conglomerate-gneiss contact over a wide area.) The original extent of the area of Precambrian beneath the Silurian is conjectural, but subject to fairly close constraints. Within the Outlier, it is constrained on the north by the contact of the Green Pond Conglomerate on Cambrian dolomite at the north end of Echo Lake (just south of our Stop 4), and by the Ordovician shale

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Figure 2. (Above) The fault-scarp of the Golan Heights on the east side of the still-active lake Kinneret rift-valley above Kibbutz Ha-On, Israel. The plateau is capped by Pleistocene basalt. This makes a good model for the New Jersey Mesozoic Basin and shows the possible kind of source of the basalt cobbles at Stop 1. Note the Recent fluvial cobbles in the foreground. (Photo courtesy of Zvi Erez, Open University, Tel Aviv, Israel.)

(Below) The Golan Heights looking toward the north bank of the Yarmuk River near Hamat Gader. Note the former tributary stream valley, at mid-height, filled with Pleistocene basalt. Again a model for the source of the basalt boulders in the Jurassic fanglomerate at Stop 1. (Photo courtesy of Zvi Erez.)

valley-filling. The asymmetry of the outcrop bend, with the sharper curvature on the right (north), can be explained by the valley hypothesis, because the valley would have a WSW-ENE trend along the axis of the Precambrian subcrop. The north side of the bend would be nearly perpendicular to the valley trend while the south side would be more nearly parallel to it. If the bend in the unconformity is solely due to a cross-syncline, those curious features which have been enumerated would have no obvious explanation other than coincidence. One should also note in this connection that the WSW trend of the proposed valley, when projected westward, intersects the area of greatest thickness of the Shawangunk Formation south of Delaware Water Gap (Epstein and Lyttle, 1987). This could represent the lower reaches of the valley itself, or a delta at its mouth.

Although it is not necessary for the stream valley hypothesis that the Green Pond Conglomerate be fluvial in origin itself (the stream could have preceded it), the lower part of the formation at Stop 5 does, in fact, show features compatible with fluvial deposition. This Lower Conglomerate Member (Finks, 1968) is characterized by fining-upward cycles of a few feet in thickness each, with coarse pebble conglomerate at the base of a cycle and cross-bedded sand at the top. The cross-laminations indicate a westward flow direction toward the Silurian epeiric sea. There is considerable argillaceous material and hematite in the sediment, more so than in the Shawangunk/ Tuscarora Formations to the west,(Tada and Siever, 1989, fig. 1, p. 95) The conglomerate pebbles are mostly milky quartz, but black, red, and green chert is common (often angular) and some chert pebbles show weathering rinds. Quartzite and shale pebbles are also present.

The higher beds at this locality, and elsewhere, are planar cross-laminated quartzites without pebbles (mostly), characterized by Liesegang rings of hematite (Upper Quartzite Member of Finks, 1968). This unit may or may not be fluvial.

The river valley hypothesis may also account for the discontinuity of the Green Pond Conglomerate along strike. It is missing between the south end of Greenwood Lake in New Jersey and the Monroe/Highland Mills area in New York. Although mapped as a fault (Figure 1), the gap may result from deposition in two separate valleys. This would also account for the lithologic differences between the two areas. The New York Green Pond contains a coarse arkose conglomerate with large feldspar clasts (Middle Arkose Member of Finks, 1968) which is missing in the New Jersey Green Pond.

Although the Green Pond/Schunemunk Outlier is sometimes considered to be a downdropped Mesozoic rift-block (without preserved Mesozoic sediments) it is possible that the original structural frame of the Outlier was a Silurian rift-valley, perhaps formed in a back-arc extensional setting. This would be compatible with the fluvial nature of the basal fill, its strong hematite content, and its locally arkosic nature. The first undoubted marine sediments above the Taconian unconformity are red near-shore muds, bearing *Lingula,* in the Upper Silurian (Upper Shale Member of the Longwood Formation of Finks, 1968). Although the highest Silurian and Devonian sediments of marine origin are sometimes extensions of facies in the main outcrop belt, such as the hematitic crinoidal limestone unit of the Decker Ferry Formation (Skyline Member of Finks, 1968), many highly siliciclastic units are confined to the Outlier, such as the Kanouse Sandstone and Skunnemunk Conglomerate. The postulated stream valley on the Taconian unconformity is also a characteristic of rift-basins. A similar incised valley beneath the Mt. Toby Conglomerate at Roaring Brook in the Mesozoic basin of Massachusetts was described by Bain and Meyerhoff, (1976, fig. 8, and pp. 24 and 119).

Some 12,000 feet of Cambro-Ordovician marine sediments were removed from the rather small, Newfoundland to Bowling Green Mountain, area in order to expose the Precambrian gneiss. The sediments are present in this thickness ten or twenty miles to the north, west and south of this area, and their metamorphic equivalents are a similar distance to the east. The area of removal is broader than the area of Precambrian subcrop, for the Silurian rests on Lower Cambrian for several miles north of Stop 4. One could imagine a persistent major river accomplishing this, but either a eustatic fall in sea level, or an accompanying local upwarp, is necessary, if only to bring these marine sediments up out of the sea. The upwarp (or upthrust) idea is the source of the senior author's 1968 suggestion of a "Taconian island" rising out of the Late Ordovician sea. A sliver of Martinsburg Shale north of Bowling Green Mountain (Barnett, 1976) is so closely adjacent to the place where the Green Pond Conglomerate rests on Precambrian that a pre-Silurian juxtaposition there of Precambrian and Middle Ordovician is implied. (See Figure 1.) This has to be a fault. However, if the fault is invoked to explain the uplift and exposure by erosion of the Precambrian subcrop, it has to be part of a broader upwarp, to account for the extensive Lower Cambrian subcrop to the NE of the Precambrian, in the area of Stop 4.

COEYMANS LIMESTONE REEF

This Lower Devonian reef is exposed as a hill north of Deckertown Turnpike about 1 mile east of Clove Road in Montague Township, New Jersey. The exposed reef outline is about 520 x 200 feet (160 x 70 meters). On the hill south of Deckertown Turnpike similar reef-rock is exposed along strike and may either be a part of the same reef or else an adjacent patch-reef. We will refer to the first hill as "the reef" even though it may be part of a much larger reef-tract.

The beds dip about 20° to the NW. A vertical section through the reef and underlying pre-reef calcarenites is exposed on the northwest side of the reef. The top surface is at present covered with a second-growth mixed hardwood and hemlock forest, with numerous outcrops of bedrock on which observations can be made. Many of the corals are silicified, and much of the information presented here was derived from study of field-oriented blocks etched with hydrochloric acid in the laboratory.

The reef-building fauna consists of *Thamnopora* sp., both branching and massive, slender branching *Cladopora* sp., solitary rugose corals (mainly *Briantelasma americana* Oliver, 1960, and *Tryplasma* sp.), stromatoporoids (probably *Parallelostroma* sp.), and *Aulopora* sp. Small bryozoan colonies, high-spired gastropods, and several species of small brachiopods are also present. The *Aulopora* usually encrusts stromatoporoids, and the *Tryplasma* is often attached to branching *Thamnopora*. The coral fauna of this Lower Devonian (Gedinnian) reef is different from that of the next youngest reefs in this area, namely, those of the Middle Devonian (Eifelian) Edgecliff member of the Onondaga Formation. The Onondaga reefs are dominated by very large colonial rugosa which are absent from the Coeymans reefs. (The nearest Onondaga reefs are in the Hudson valley, south of Albany, New York.)

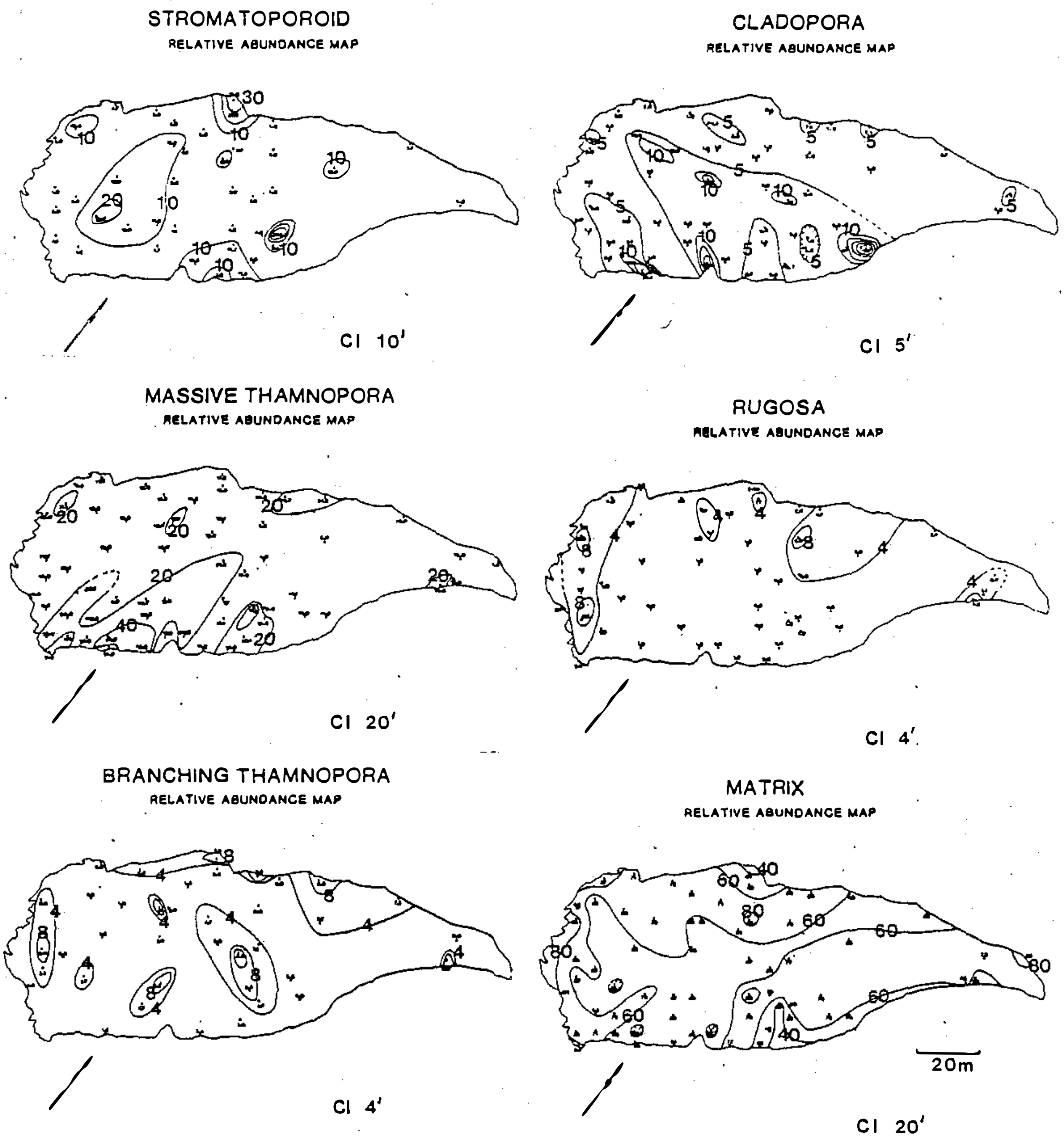

Figure 4. Maps of the Coeymans Limestone reef at Montague, NJ (Stop 6) showing percentages of the major reef-forming species, and matrix, over the top surface of the reef. Relative abundances were determined by point-counts with a one-inch grid at each of the data points shown. See text for discussion.

There are several similar-sized patch reefs, with the same coral fauna, along the belt of the Coeymans limestone outcrop in the Delaware River valley between here and the Tocks Island/Delaware Water Gap area to the south (described in Epstein, Epstein, Spink and Jennings, 1967). Reefs of the same age have also been described from the Knoxboro to Manlius area, south of Oneida, New York (Oliver, 1960) Reefs are apparently missing, however, along the Coeymans outcrop belt between the present reef and those of the Oneida area. Inasmuch as the New York reefs are north-northwest of the Delaware Valley reefs, there is a possibility that a reef-tract in the Coeymans Limestone runs approximately NNW-SSE, and that similar reefs exist in the subsurface beneath the Catskill/Pocono Plateau. The present reef was studied from a petrologic and paleoecologic point of view by Precht (Precht, 1982, 1984, 1987a, 1987b). Our present study is an expansion of Precht's study in a biological direction with special emphasis on species distribution, growth-form distribution and growth orientation.

Coral Distribution

The reef-building species show different patterns of abundance over the exposed surface of the reef (Figure 4) as determined by point-counts made with a one-inch grid over one or more square feet at each of 68 stations. Massive Thamnopora (Figure 7) is most abundant at the southern end of the reef along the southeast edge of the exposed area. The more delicate, branching Cladopora (Figure 9) is also most abundant in a similar, but wider, area, extending farther north. Much of the Cladopora is broken, rather than in growth position.

Branching Thamnopora (Figures 5, 6, and 10) is most abundant more northerly, in the center and along the northwest edge of the preserved reef. Most colonies seem to be in growth position. Solitary rugosa (Figures 5, 6, and 7) are most abundant along the northwest edge of the preserved reef, where they are also more frequently in growth position than they are further south. Stromatoporoids (Figure 8) are most abundant in the northern and central parts of the exposed area, rather similar in their distribution to that of branching Thamnopora. They are especially well displayed in the vertical cliff along the northern edge of the reef and on the top surface just behind this cliff.

The relative abundance of matrix is an inverse measure of total fossil abundance. As can be seen from Figure 4, the matrix is least abundant (less than 60%) in the central and southern parts of the preserved reef, and also in a small patch in the middle of the NW side. Most of the matrix is micrite, rather than the calcarenite of the pre-reef and interreef facies, and presumably results from the baffling effect of the coral colonies. This is strong support for the in-place nature of the corals. There are, however, channels filled with calcarenite, best seen on the vertical NW face.

Coral Orientation

One of the most interesting feature of this reef is the consistent orientation of both in-place corals and loose fragments. The strongest orientations are shown along the entire northwest margin of the reef by the branches of Thamnopora and Cladopora colonies and by attached solitary rugosa.

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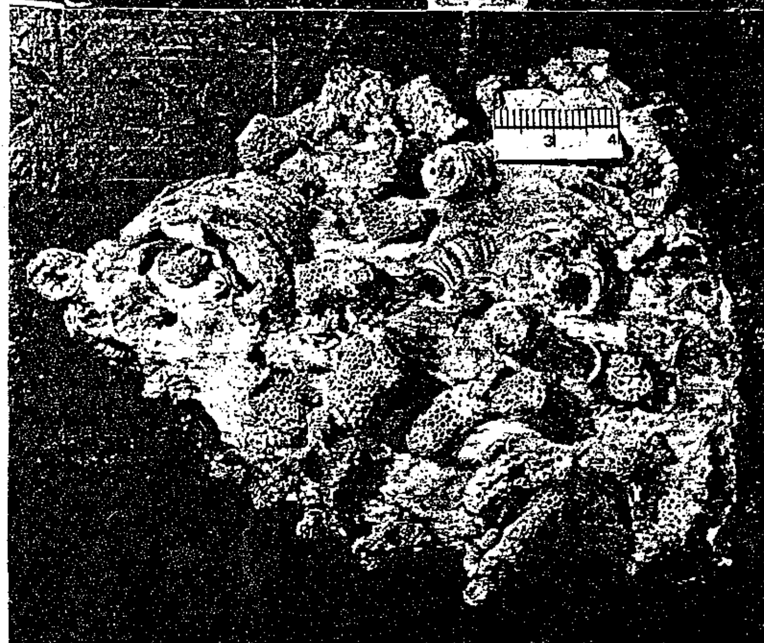

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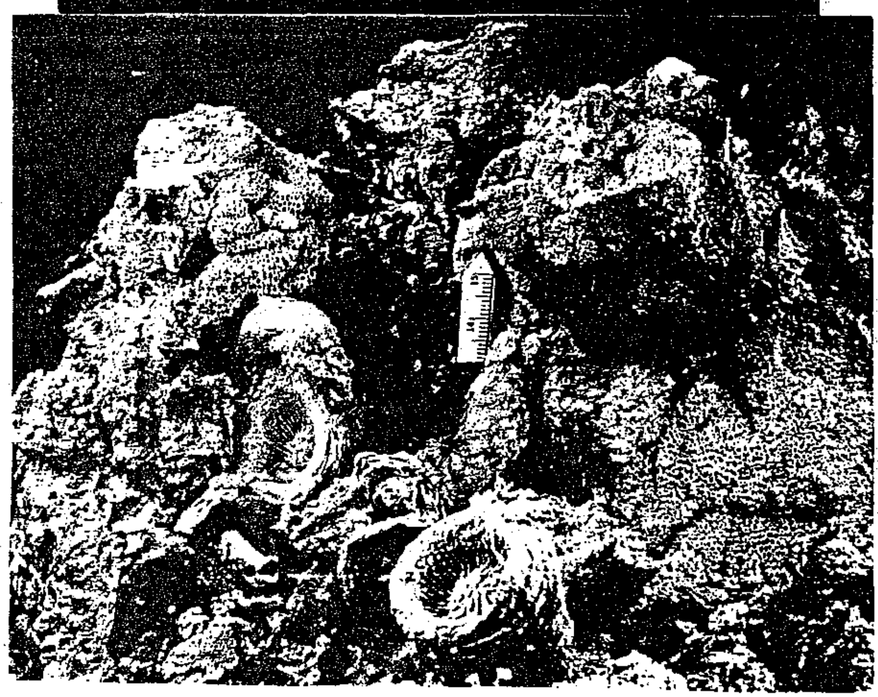

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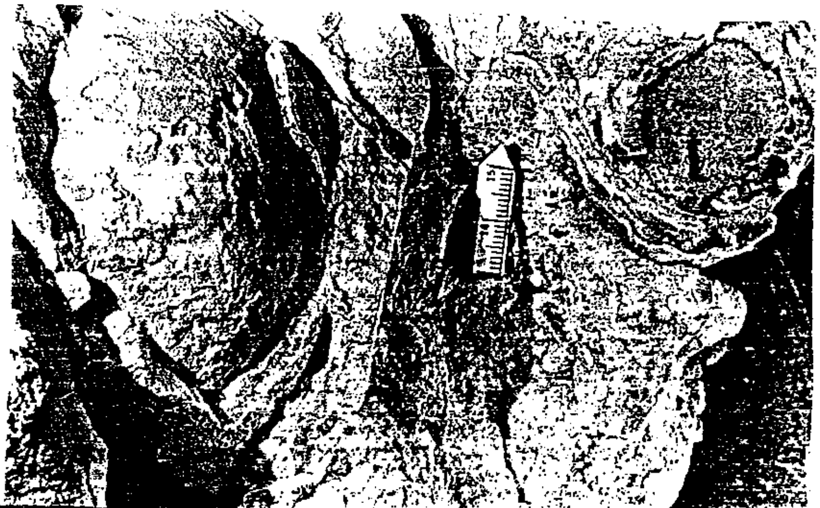
The branches predominantly grow to the north at a low angle, and attached rugosa open to the north. It is clear that whatever determines the growth direction of the Thamnopora branches also determines the direction in which rugose corals open, for Tryplasma attached to Thamnopora uniformly open in the direction of the branch tips (Figure 6). As one crosses the reef to the southeast side, orientation gradually becomes less pronounced (see Figures 12, 13, 14, and compare Figures 6, 7 and 10). Loose, cylindrical fragments of Cladopora and Thamnopora branches are oriented roughly E-W over most of the reef (open bars on Figures 12,13). Unattached solitary rugosa have their apexes pointing north (open bars on Figure 14) as do high-spired gastropod shells on one etched laboratory sample. One can easily demonstrate with paper models of cones and cylinders that in a unidirectional current, cylinders will roll to orient *across* the current and cones will orient with apexes pointing *into* the current (see Figure 11, upper). Thus the data from loose fragments are consistent with a prevailing current from the north. (This would have been paleo-northeast because the Laurentian Plate has rotated counterclockwise by about 45° since the Devonian; see Seyfert and Sirkin, 1979, page 314).

Evidence from modern reefs concerning growth directions of living branching corals has been conflicting. Graus, Chamberlain and Boker (1977) presented theoretical grounds for branching colonies adopting an *up-current* growth in order to reduce mechanical stress on the skeleton of the branch. They also documented such an upcurrent growth in living Acropora palmata on a Caribbean fringing reef (*ibid.*, p. 145) in the zone of strongest unidirectional currents, just lagoonward of the reef-crest. (However, in the "Strong Wave Zone" in front of the reef-crest, the orientation was bimodal, with the greater frequency *down-current*.) Wallace and Schafersman (1977, p. 45) documented the same species oriented *into* the current on the front of a Carribbean patch-reef. On the other hand, Shinn (1963) observed that Acropora palmata grew dominantly *down*-current in front of the reef-crest at Key Largo Dry Rocks on the Florida Reef Tract. Likewise, Bottjer (1980) found that A. cervicornis showed the same *down*-current branch growth in front of the reef-crest on a small barrier-type reef in the Bahamas. It is not clear how these conflicting orientations are to be reconciled. One may note that the *down*

Figure 5. Branching Thamnopora sp. Briantelasma americana Oliver, from the southwest end of the reef. The corals are in life position and more or less vertical. They are silicified and were etched free with hydrochloric acid.

Figure 6. Branching Thamnopora sp. with attached Tryplasma sp., cf. T. fascicularium Oliver, and Briantelasma americana Oliver at the left. Note that all are facing in the same direction. The silicified specimen is float from the top of the reef and was partly etched with hydrochloric acid.

Figure 7. Massive Thamnopora sp., with three specimens of Briantelasma americana Oliver in life-position, oriented toward the north. Note the cylindrical fragments of Tryplasma sp. (?) oriented E-W by the same current. Elsewhere on this block are two high-spired gastropods with apexes pointing north (not visible in this photograph). The silicified specimens were etched out with hydrochloric acid. The sample is from the center of the northwest side of the reef.



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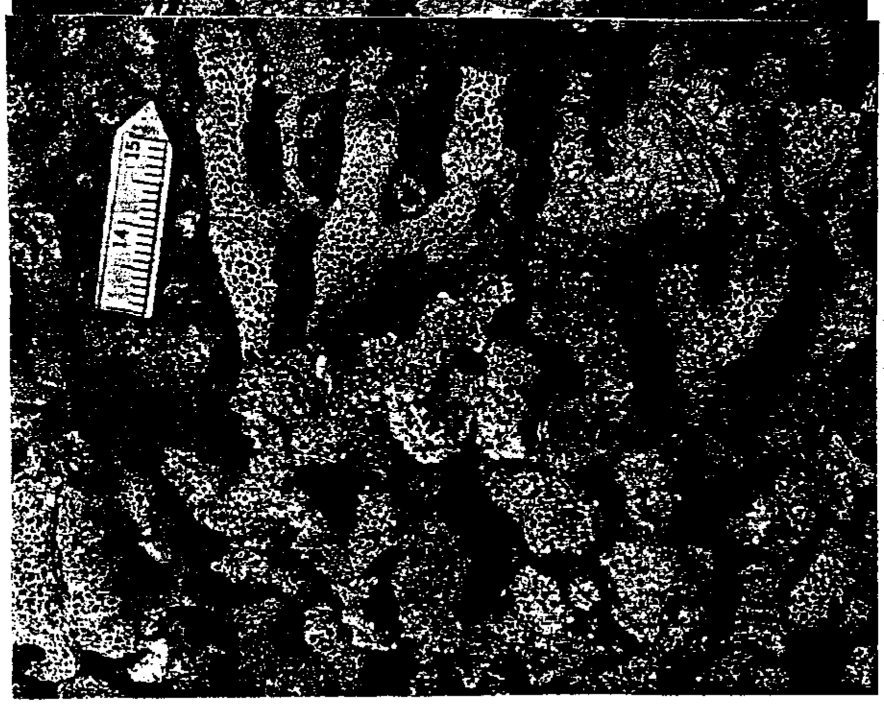

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current orientations are on the wave-dominated front side of elongate reefs, and the up-current orientations on the current-dominated lagoonward side. The cited patch-reef of Wallace and Schafersman (1977) was in an atoll lagoon near a channel through the atoll-reef. This suggests that where there are uni-directional currents, rather than back-and-forth wave-surge (even where one component dominates), the growth of branches will be dominantly into the current. If so, this has implications for the reconstruction of our reef.

A Reef Model

The zone of strongly oriented, nearly-horizontal branches along the northwest side of our reef suggest that this is on the lagoonward side of a reef-crest, close to the crest itself. This would correspond to the "Current Zone" of Graus, Chamberlain and Boker (1977). The direction of wave-approach (i.e., the open sea) would be from the north, and would generate a strong flow to the south over the crest and into the lagoon. Several other lines of evidence favor the reef-crest being on the northward side. One is the presence of shallow calcarenite-filled channels on this side which may represent grooves of a spur-and groove system or, more likely, the less-deep channels through the reef-crest ridge itself (compare Rützler and MacIntyre, 1982, p. 24 figure 13c and p. 26 figure 15). The abundance of domal stromatoporoids here also suggests a reef-crest ridge, for they are noted by many authors to be characteristic of the roughest-water environments on Devonian reefs (Hoffman and Narkiewicz, 1977; Copper, (1974). Likewise domal to massive shape is assigned by James (1983, p. 374, figure 59) to moderate to high wave-energy and low sedimentation environments.

Further south the current would gradually diminish, as described by Graus, Chamberlain and Boker (1977, p. 145). It was still sufficiently strong to orient *Cladopora* branch fragments across it (Figure 13), but by the time one reaches the southern end of the preserved reef the orientations become random. That this area is a back-reef lagoon is also supported by the dominance of the delicate-branching *Cladopora* and the relatively high proportion of matrix (Figure 4). There are patches of stromatoporoid and branching *Thamnopora* dominance (Figure 4) which may represent micro-patch-reefs, so to speak, within the lagoon.

Figure 8. Domal stromatoporoids (*Parallelostroma* sp. ?) from the center of the northwest side of the reef.

Figure 9. A silicified colony of *Cladopora* sp. etched free with hydrochloric acid. The colony is upside-down. The specimen is float from the central area of the reef and its original orientation is not known.

Figure 10. A silicified colony of branching *Thamnopora* sp. freed with hydrochloric acid. The branches are horizontal and strongly oriented toward the north (shown by the arrow on the scale). The oriented specimen is from an outcrop at the northeast end of the reef.

At the southernmost edge of the reef there is an area dominated by massive Thamnopora. This growth-form suggests higher wave-energy (James, 1983, p. 374) and we may be at the back-slope edge of our reef. The general pattern corresponds to that shown by Copper (1974, figure 9, top diagram) as the climax stage of a Devonian reef. We have modeled our reconstruction (Figure 11, bottom) on Copper's diagram. Our reef may well be in a mature successional stage, for there is 3-4 m of reef-rock exposed in the NW cliff-face (at the SW end) above the bedded pre-reef crinoidal calcarenites. A similar recent model, almost identical to the scale of our reef, is the Montastrea-assemblage patch-reef illustrated by Wallace and Schafersman (1977, p. 44, figure 9). This is one of a number of small patch-reefs within the lagoon of an atoll, seaward of the barrier-reef off the coast of Belize. Our lagoon corresponds to their "patch backreef region dominated by Thalassia and low coral growths" (Wallace and Schafersman, 1977, p. 45 figure 11). A small patch-reef model supports the idea of Precht (1987b) that the hill is more or less coterminous with the original outline of the reef.

ACKNOWLEDGEMENTS

We thank Philip LaPorta for first showing us the Coeymans reef, William Precht for sharing his knowledge of the reef with us in the field, Peter Alessi for indispensable assistance in making the road-log, and both Mitchell Algus and Pamela Brock for critical reading of the manuscript. The senior author also thanks the participants in the 1968 trip, as well as many students taken to these locations, for stimulating discussion.

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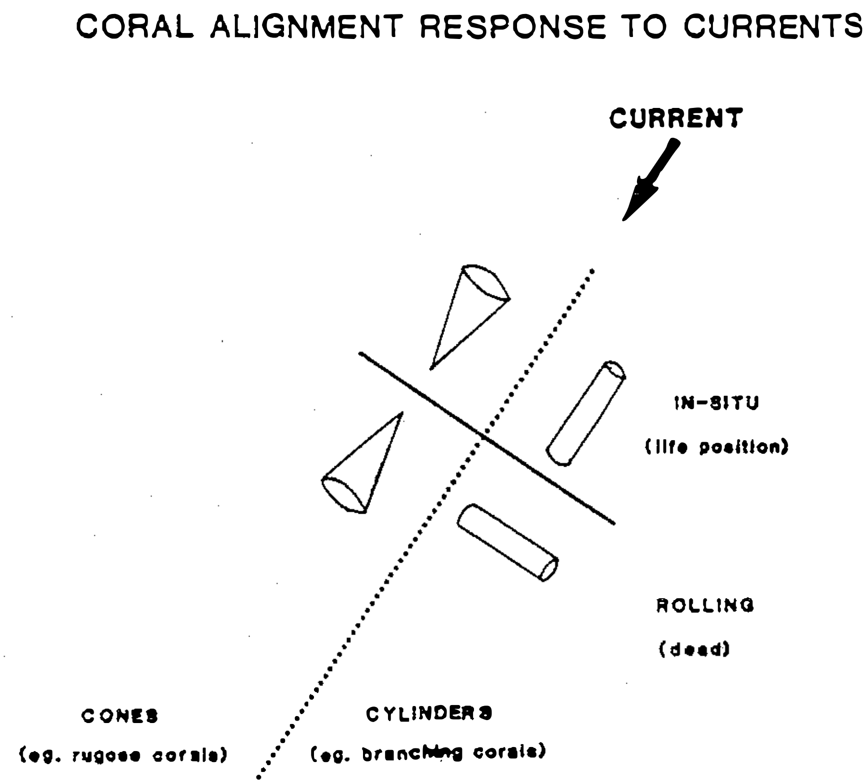


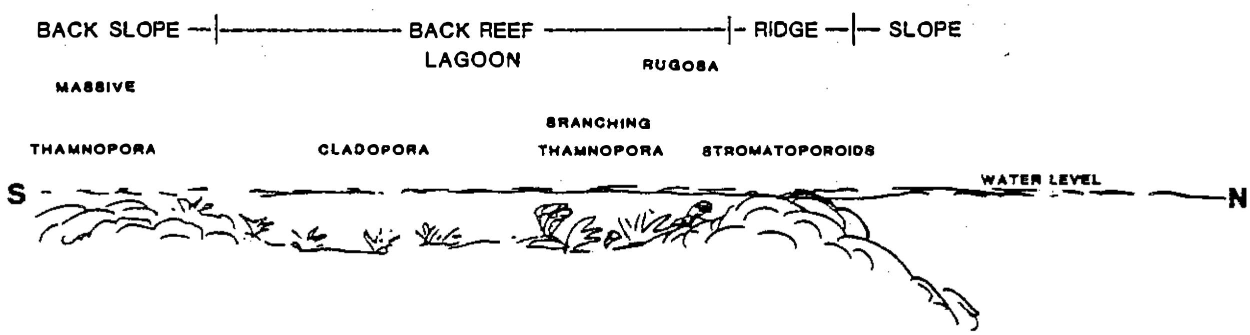
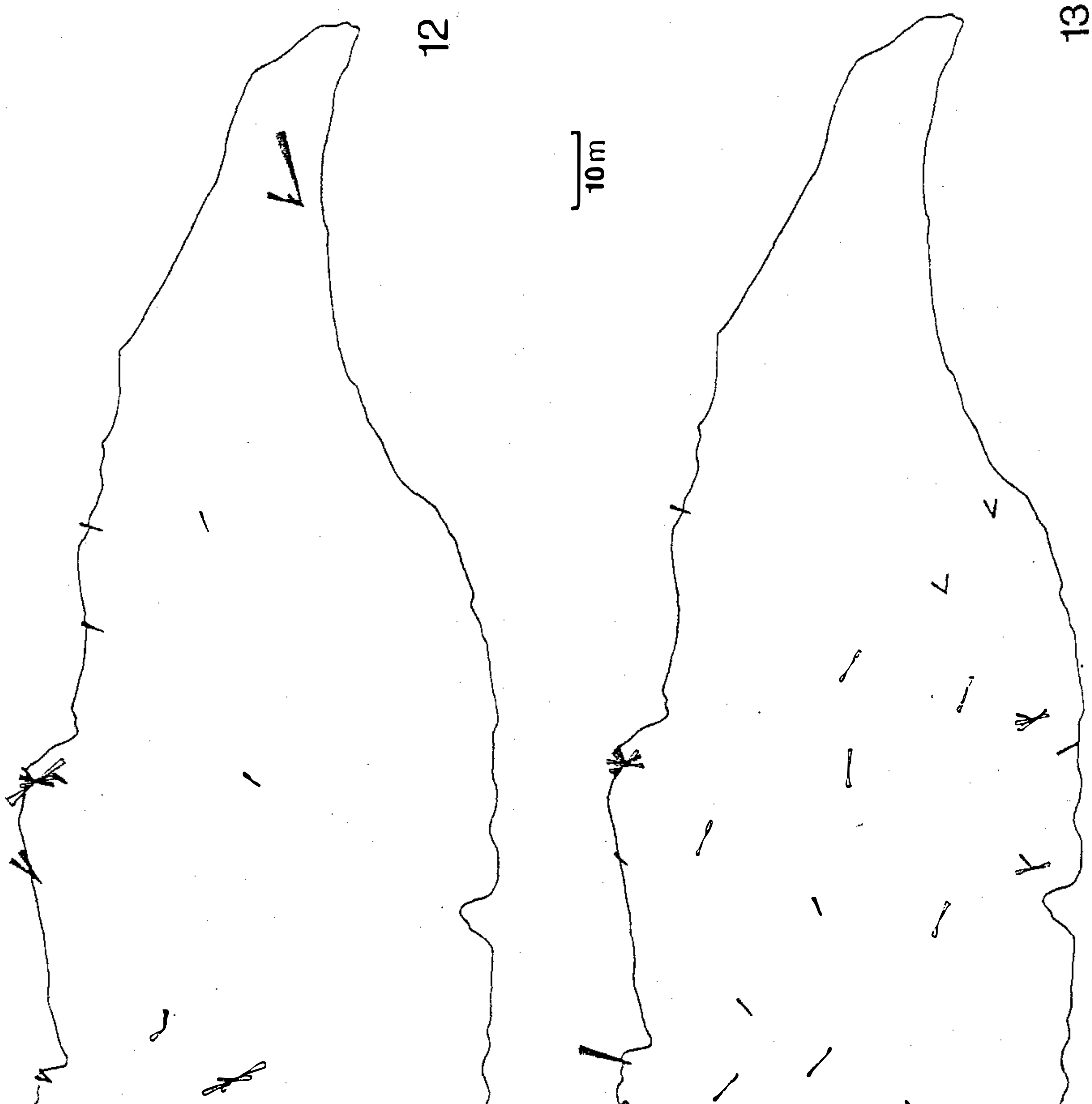


Figure 11. (Above) Diagram showing alignment of different coral shapes in response to the same current. Such alignments were used to construct the orientation maps (Figures 12, 13, 14).

(Below) A proposed model of the reef along a N-S transect, north at right.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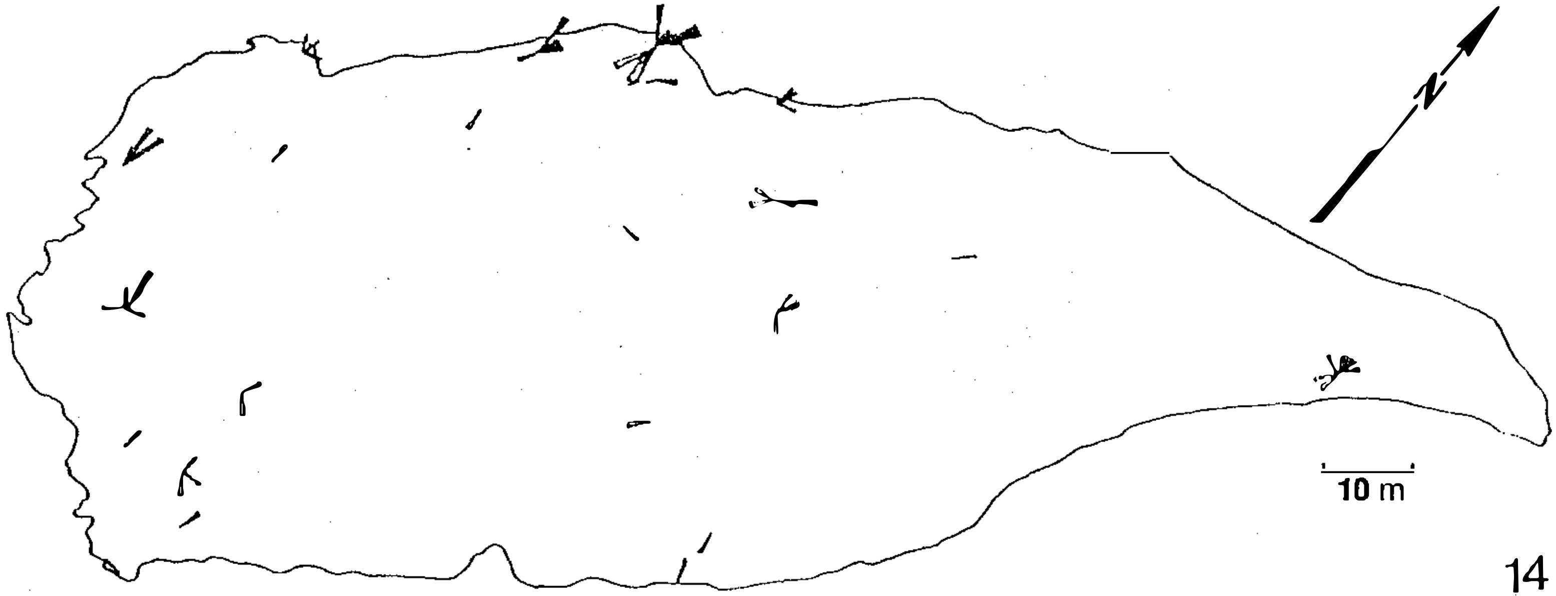

Figure 12. Orientations of branching Thamnopora shown by rose-diagrams at specific localities. Solid bars are in-situ colonies, open bars are fragments of branches. The solid bars expand toward the facing of the branch tips. The open bars are parallel to the branch-fragment axis in both directions. Length of bars proportional to the number of observations within a 10° bin.

Figure 13. Orientations of branching Cladopora. (For explanation see Figure 12.)

Figure 14. Orientations of solitary rugose corals. Bars, both solid and open, expand toward the open end of the corallite. Their appearance in fact mimics the position and shape of the fossils. Solid bars are attached in-situ corals, open bars are loose ones. Length of bar proportional to number of observations within a 10° bin.

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ROAD LOG FOR ANCIENT LAND SURFACES IN AND AROUND THE GREEN POND OUTLIER, A DEVONIAN CORAL REEF, AND "TACONIAN ISLANDS" REVISITED.

CULMULATIVE MILEAGE	MILES FROM LAST POINT	ROUTE DESCRIPTION
0.0	0.0	Juncture of ramp from I-84 (North)East with NY 17 (South)East, Near Middletown, NY. START OF TRIP. Proceed east on NY 17.
1.6	1.6	Anticline in Martinsburg Formation (Ordovician) on left. Zone 13 (possibly Zone 14) graptolites were collected here by Alejandro Vargas and identified by W.B.N. Berry (Vargas, 1976, p. 32 and fig. 5A). Distal turbidites can be seen here.
6.4	4.8	Martinsburg Slate on both sides of the road.
9.2	2.8	Martinsburg Slate on both sides of the road again.
10.1	0.9	Klippe of Precambrian gneiss on right, forming hill.
12.5	2.4	Exit 129. Museum Village Road.
12.7	0.2	Klippe of Precambrian gneiss on left, in thrust contact with underlying Cambro-Ordovician dolomite. The thrust plane and accompanying fault breccias can be seen.
BONUS STOP.		This exposure may be visited by exiting here, turning left (north) on Museum Village Road to cross over NY 17, then taking the first left (Old Mansion Road) for 0.2 mile and parking just before first house on right. Walk into woods on left (south) side of road downhill toward NY 17, crossing the old, low stone wall left of the barbed wire. At highway level the thrust plane is exposed, dipping about 30° SE, with garnet-gneiss above and white-weathering dolomite below. Immediately above the fault

		plane you may collect samples of gneiss mylonite-breccia, and immediately below it, tectonic breccia of dolomite. To resume trip, return to Museum Village Road, turn right over NY 17, then left onto service road, making another sharp left just before bus stop area, to enter NY 17 east at mileage 13.0 of log.
13.0	0.3	Entrance ramp from Museum Village Road. The western border fault of the Schunemunk Outlier crosses approximately here, shown on the 1971 NY State Map as a small offset of the main fault at this point. (Fisher, Isachsen and Rickard, 1971.)
13.3	0.3	Bellvale Sandstone on left, dipping SE. (Mid-Devonian)
13.7	0.4	Axis of Schunemunk Syncline plunging NE. Horizontal pebble beds of Bellvale Sandstone on left. Between 13.7 and 14.1 is a 2,000-foot complete section of the Bellvale Sandstone (measured section in Finks, 1968) from submarine prodelta silts at 14.1 to subaerial piedmont conglomerates at 13.7 It is a part of the Catskill Delta. The beds dip NW on the east limb of the Schunemunk Syncline.
14.1	0.4	Basal beds of Bellvale Sandstone, dipping steeply NW.
17.4	3.3	Exit south on 1-87. Main mass of Ramapo Mountains (Hudson Highlands) are ahead (Precambrian of Reading Prong) and to your left as you swing south. You are in a valley on Cambro-Ordovician dolomitic limestones. This is also part of the Green Pond - Schunemunk Outlier, although the NY State Map has a structural discontinuity (normal fault) between these rocks and the Siluro-Devonian just passed through (Fisher, Isachsen and Rickard, 1971).
19.0	1.6	Outcrop of Cambro-Ordovician limestone on left.
19.8	0.8	Outcrop of Cambro-Ordovician limestone also on left.
20.5	0.7	Greenwood Furnace on left. This smelter served the small magnetite mines in this vicinity during the 19th Century. The ore is from the Precambrian rocks. About three miles further on is another historic locality, the town of Tuxedo Park, where the Tuxedo jacket was invented.
30.6	10.1	Precambrian gneiss outcrop on left.
31.8	1.2	Exit at Suffern.
32.4	0.6	Stay left, and continue south into New Jersey on Route 17.
32.8	0.4	Crossing Ramapo Fault into Mesozoic Basin, a rift-valley formed during the opening of the Atlantic Ocean. The Green Pond-Schunemunk Outlier has a similar structure and may be a Mesozoic basin in which no Mesozoic rocks were preserved. (Or a Siluro-Devonian extensional basin?)

33.9	1.1	Turn right at exit for US 202.
34.0	0.1	Turn left (south) on US 202.
34.8	0.8	Old stone house on right. Historic marker.
35.1	0.3	Good view of Ramapo fault-line scarp on right.
36.5	1.4	Crossing Orange Mountain Basalt (Lower Jurassic).
37.0	0.5	Outcrops of Orange Mountain Basalt on left.
40.1	3.1	Another basalt outcrop on left.
40.6	0.5	Junction with US 208. Continue on US 202.
41.8	1.2	Turn right to continue on US 202.
42.2	0.4	Turn left into parking lot of Hillside Diner.

STOP 1. OAKLAND. JURASSIC FANGLOMERATE BENEATH THE PREAKNESS BASALT.

The most accessible outcrops are on the south side of the diner parking lot on the east side of US 202. Look for the boulders of vesicular basalt. Search for gneiss boulders. Some Paleozoic formations to look for are: white-to-buff-weathering Cambro-Ordovician dolomites, the purple with white quartz Green Pond Conglomerate, the red with white quartz and red quartzite Skunnemunk Conglomerate, the subgreywacke Bellvale Sandstone. Note their proportions.

42.4	0.2	Outcrop of Preakness Basalt on left.
44.0	2.4	This is Stop 2. Do not stop here but proceed to Moyias Road.
44.1	0.1	Moyias Road. Turn right and right again. Park off road next to pumping station. Walk back to outcrop on east side of US 202. LOOK BOTH WAYS WHEN CROSSING ROUTE 202. YOU ARE ON A CURVE AND A HILL. VISIBILITY IS LIMITED. WALK FACING TRAFFIC AND STAY OFF THE ROADWAY.

STOP 2. POMPTON LAKES. JURASSIC FANGLOMERATE BENEATH THE HOOK MOUNTAIN BASALT.

Note the large boulder of Green Pond Conglomerate near the south end of the outcrop, and the many boulders of whitish-weathering Cambro-Ordovician dolomite. Note also the interbeds of finer sand and shale. How do the proportions of lithologies among the clasts compare with those of the preceding stop?

44.4	0.3	Outcrop of Hook Mountain Basalt on left. The outcrop crosses the bed of the Ramapo River on the right, on top of which is a dam and waterfall.
44.5	0.1	Turn right onto Hamburg Turnpike. Cross Ramapo River.
45.2	0.7	Bear left at fork.
45.4	0.2	Crossing Wanaque River.
45.6	0.2	Park in the professional building parking lot on the right.

STOP 3. RIVERDALE. ANNSVILLE PHYLLITE (ORDOVICIAN) AND BED-LOAD OF THE PEQUANNOCK RIVER.

Descend carefully under the bridge and walk to the south side of the east abutment. There is also a path to this point from the south side of Hamburg Turnpike. CAREFUL CROSSING ROAD. The outcrop of the Annsville Phyllite is exposed here. PLEASE DO NOT HAMMER ON THIS RARE OUTCROP! THERE ARE PLENTY OF LOOSE PIECES LYING ABOUT. Another small outcrop lies upstream of the bridge in the channel and on the west bank.

Unless the river is in flood, a large sample of the Pequannock bed-load is exposed beneath and downstream of the bridge. Look for imbricated cobbles. See how many different Precambrian lithologies you can find. Green Pond Conglomerate cobbles are common. Can you find any Cambro-Ordovician dolomites? This river crosses the Green Pond outcrop about ten miles upstream, but much closer glacial moraines may be the source of most of the cobbles.

46.0	0.4	Turn left at light onto County 511. (Before you turned there was a quarry in Precambrian granite visible ahead.)
46.1	0.1	Washington's Headquarters on left.
47.0	0.9	Turn right onto NJ 23 Northwest.
47.3	0.3	Crossing Ramapo Fault. Road climbs hill on Pre-cambrian with extensive cuts exposed in ramps for I-287. A gravel pit was formerly present to the left (south) in a delta built into Glacial Lake Passaic. Moraines on left about a half mile ahead.
54.1	6.8	Outcrops of Precambrian on right, Pequannock River valley on left.
54.6	0.5	Kanouse Mountain ahead, made of Green Pond Con-glomerate on Precambrian. You are on the area of the radar mosaic Figure 3.
54.7	0.1	Turn right (north) onto Echo Lake Road (clearly visible on Figure 3). You are paralleling Kanouse Mountain.
57.0	2.3	Turn left on Macopin Road.
57.8	0.8	Turn left on Gould Road.
58.4	0.6	Bend in road. Hill of Precambrian ahead.
58.8	0.4	This is Stop 4. Do not stop here but proceed to wider shoulder.
58.9	0.1	Park off road on right and walk back to 58.8. WALK TO THE LEFT FACING TRAFFIC AND STAY OFF THE ROADWAY. SIGHT DISTANCE FOR CARS IS LIMITED. Enter woods road running uphill to west (on your present right).

STOP 4. GOULD QUARRY. PRECAMBRIAN GNEISS, HARDYSTON SANDSTONE AND LEITHSVILLE DOLOMITE (LOWER CAMBRIAN), BASAL GREEN POND CONGLOMERATE (SILURIAN) IN KARSTIC POCKETS.

Outcrops at entrance to the road are Precambrian gneiss. Continue uphill on this road to the SW. Where the road bends to the left (south) continue straight ahead into the woods. You will soon reach a small swale. The lower Hardyston Sandstone underlies this swale. As you face SW, the hill on your left is Precambrian gneiss and the low ridge on the right exposes ledges of the upper part of the Hardyston Sandstone and the base of the Leithsville (or Stissing) Dolomite dipping NW. Walcott (1893) reported the trilobite

Olenellus from the same beds a mile or so north, proving their Lower Cambrian age (and for us, their position on the Laurentian Plate).Follow this ridge a short ways to its southern end, where there is a small depression and cross-valley in the ridge. In the depression, against the ledges to the left, is a small pocket of conglomerate in the basal Leithsville Dolomite. (It may be covered by leaves.) This is a karstic cavity filled with the Silurian Green Pond Conglomerate which has infiltrated from above. A thin red shale lines this pocket. Walk to the top of the Leithsville ridge at this point through the small notch and descend its west side. Immediately to the right you will see a small patch of conglomerate on top of the steeply northwestward dipping beds of the Leithsville Dolomite. If you examine the dip-slope below it you can see a narrow extension of this patch apparently filling a fissure in the Leithsville. This must have been the entrance for the gravel into the pocket represented by the patch. PLEASE DO NOT HAMMER ON THIS RARE OUTCROP! IT IS THE ONLY RELIC OF THE KARSTIC INFILLING. PLEASE PRESERVE IT FOR POSTERITY. This outcrop was illustrated by Rodgers (1971, p. 1160, fig. 9B). In the valley to the west is a small outcrop of the Green Pond Conglomerate just west of a small brook. The main mass of Kanouse Mountain, composed of the conglomerate, lies beyond. If you have time, note the ruins of the old lime kiln, north of the quarry, with its partly melted walls of stone blocks. The Leithsville was quarried here for lime. This tree-filled hollow was once an active quarry. The geologic relations here are illustrated in Kümmel and Weller (1902, figure 1).

		Return on Gould Road. Make the U-turn at the driveway of the log house on the left if you wish to preserve road-log mileage. THE U-TURN SHOULD BE DONE WITH <u>EXTREME CAUTION</u> BECAUSE OF THE LIMITED SIGHT DISTANCE FOR CARS ON GOULD ROAD. A SAFER, BUT LONGER, WAY OF GETTING TO STOP 5 IS TO PROCEED ON GOULD ROAD (WITHOUT TURNING BACK) AS FAR AS UNION VALLEY ROAD AT POSTVILLE, TURNING LEFT (SOUTH) ON THAT ROAD TO NJ 23, THEN LEFT (SOUTHEAST) ON NJ 23 TO ECHO VALLEY ROAD, THEN RIGHT ON JUG-HANDLE TO ECHO VALLEY ROAD NORTH (LEFT) AND LEFT AGAIN ONTO NJ 23 NORTH AT TRIP MILEAGE 63.0.
60.0	1.1	Turn right on Macopin Road.
60.8	0.8	Turn right on Echo Lake Road.
63.0	2.2	Turn right on NJ 23 northwest.
63.2	0.2	Turn right into rest area.
63.3	0.1	Park.

<u>STOP 5. NEWFOUNDLAND. GREEN POND CONGLOMERATE (SILURIAN) AND PRECAMBRIAN GNEISS.</u>

The parking area is a loop of old Route 23 and is the original width of the two-way road. Note the historical marker with original relics of local iron smelting. To the south of NJ 23 is Charlotteburg Reservoir. Copperas Mountain to its right (west) comprises the extension of the Green Pond outcrop on the south of the Pequannock watergap. Seen in the distance to the west-southwest, are the ridges of Green Pond Mountain and Browns Mountain, also composed of Green Pond Conglomerate on Precambrian, repeated by folding and faulting.

The nearest ledges north of the parking area show well the steep westward dip and the fining-upward cycles in the Lower Conglomerate Member of the Green Pond Formation. Look for colored chert clasts (red, green, black) in the conglomerate layers. Many show weathering rinds. Follow the base of the cliff uphill to the east (right). A short distance up is a good place to see the cross-bedding. Note the consistent westward dip of the cross-laminations, indicating a westward current.

Continue upward along the base of the cliff and pick up a trail which heads uphill into the woods parallel to the base of the cliff but some distance from it. Keep on this trail in a northeastward direction. After several hundred feet you will come upon ledges of Precambrian gneiss. They are about 50 feet from the base of the cliff, the intervening area being strewn with large talus blocks of the conglomerate. The actual unconformity is not exposed here but it is clear that the conglomerate must rest directly on the gneiss.

When you return via the same trail you will see on the right (west) just above the ledges next to the parking area, beds of the higher non-conglomeratic part of the Green Pond Formation ("Upper Quartzite Member" of Finks, 1968). You can climb up to them if you have rubber-soled shoes or hiking boots (STAY AWAY FROM THE CLIFF EDGE TO THE LEFT) and observe the banded tongues (Liesegang rings) of hematite, showing former groundwater flow to the left (southwest) when the quartzite still had porosity.

63.5	0.2	Drag folds in Upper Quartzite Member of Green Pond Conglomerate. We are continuing northwest on NJ 23.
63.6	0.1	Small outcrop of Longwood red shale (Upper Silurian) on right.
64.8	1.2	Type locality of the Kanouse Sandstone (Lower Devonian) in the village of Newfoundland to the right.
67.2	2.4	Outcrops of Precambrian. We have just crossed the normal fault bounding the Green Pond-Schunemunk Outlier on the northwest. Oak Ridge Reservoir on left. The fault-line scarp bounds the reservoir on the west. This is called the Reservoir Fault (see Malizzi and Gates in this guidebook).
74.5	7.3	Turn left on County 631. Town of Franklin.
74.6	0.1	Franklin Lake on left. Source of the Walkill River.
74.7	0.1	Outcrop of Cambro-Ordovician limestone on right.
75.0	0.3	Old Buckwheat Mine dump (zinc mine in Precambrian Franklin Marble; type locality for Franklinite) and Franklin Mineral Museum on right.
75.1	0.1	Bear right on County 631. Crossing Walkill River. This river flows northeast to join the Hudson River at Kingston, following the Great Valley on the Cambro-Ordovician outcrop belt. Water from Franklin Lake flows 75 miles northeast, then another 100 miles south before it reaches the sea.
75.5	0.4	Wildcat Road on left. 0.5 mile down this road is an exposure of the Precambrian-Lower Cambrian unconformity (described in Finks, 1968). The first

		archaeocyathids from New Jersey were found here in the basal Leithsville Dolomite by Frank Markiewicz on the 1968 NYSGA field trip.
75.7	0.2	Turn left following County 631.
77.1	1.4	Turn left (southwest) on NJ 94.
77.9	0.8	Outcrop of Cambro-Ordovician limestone on right.
78.3	0.4	Outcrop of Cambro-Ordovician Limestone on left.
78.7	0.4	Old Monroe School on right.
82.0	3.3	Turn right on NJ 15 north.
82.3	0.3	Continue north on NJ 15.
83.3	1.0	Bear left on NJ 15 north.
83.5	0.2	Outcrops of Cambro-Ordovician limestone on left.
84.3	0.8	Outcrops of Martinsburg Slate on right.
84.9	0.6	Straight ahead on US 206.
86.8	1.9	Continue on US 206.
90.1	2.2	Culver Lake on right. Ahead is Kittatinny Mountain, held up by the Silurian Shawangunk Conglomerate.
92.6	2.5	Bear right on US 206.
94.7	2.1	Old log house on left.
99.1	4.4	Turn right on Clove Road (County 653).
99.7	0.6	Turn right on Deckertown Turnpike (County 650).
99.8	0.1	Outcrop of New Scotland Limestone (Lower Devonian) on right.
100.0	0.2	Park off road on right. Mile marker 1. Enter woods-trail ahead on left, opposite outcrop of pre-reef limestone on right.

STOP 6. COEYMANS LIMESTONE REEF LOWER DEVONIAN).

Approach the reef by the trail which leads toward the high southeast cliff and attain the top by climbing the slope to the right (east) of it. Here you are at the supposed back-slope area dominated by massive *Thamnopora*. A good outcrop to start with, showing most of the coral species, is just west of the large glacial erratic boulder of shale. From here one may walk along the edge of the high cliff to the west and then north. The reef-ridge area of strong orientations, with dominant branching *Thamnopora*, solitary rugosa, and domal stromatoporoids, is best exposed at about halfway along the northwest face. (Compare Figures 5, 7, and 8.) From here one may walk south along a series of outcrops toward the highest part of the hill, seeing the back-reef lagoon facies and finally reaching the back-slope area at the top of the hill. From here one may walk northeast, past the old stone wall on the right (east) to a fine outcrop at the northeast tip of the reef, with silicified oriented branching *Thamnopora* well displayed, again part of the reef-ridge. (Compare Figure 10.) A walk back along the base of the northwest-facing cliff paralleling the reef-ridge will show some calcarenite channels in cross-section (or grooves of a spur-and-groove system) near the northeast end, some fine stromatoporoid colonies in vertical section about half-way along, and the pre-reef calcarenites at the southwest end. Note that the matrix of the main part of the reef is mostly micrite. Use the maps of Figures 4, 12, 13 and 14 as guides.

END OF TRIP. Return to Clove Road by going ahead (east) for 0.2 mile and making a sharp left onto Birch Tree Road, which reaches Clove Road at 0.85 mile. This is less dangerous than making a U-turn on Deckertown Turnpike because of limited sight distance. Those who wish to return to Middletown, NY should turn right (northeast) on Clove Road and proceed for seven miles to NJ 23. At NJ 23 turn left (north) and follow signs to I-84 East which will return you to Middletown, where I-84 intersects NY 17. Those who wish to return to the New York City area should turn left (southwest) on Clove Road, left again at US 206 (a right turn takes you to Milford, PA and points west) and retrace the trip route as far as NJ 15. Continue southeast on NJ 15 to I-80 and take I-80 east. This is the fastest route to New York City from Stop 6. (Two interesting localities not far from Stop 6 are the kimberlite-like diatreme at Libertyville, with inclusions of Precambrian and Cambro-Ordovician, and the nepheline-syenite at Beemerville. Consult the New Jersey State Geologic Map.)

MAPS

Stops 1 and 2 are on the Wanaque, NJ, 7½-minute quadrangle, Stop 3 on the Pompton Plains, NJ, 7½-minute quadrangle, Stops 4 and 5 on the Newfoundland, NJ, 7½- minute quadrangle, and Stop 6 on the Milford, PA-NJ, 7½-minute quadrangle. The Bonus Stop is on the Monroe, NY, 7½-minute quadrangle.

ADDENDUM

The two expansions in the Outlier, near Newfoundland and Monroe respectively, suggest pull-apart basins along a Silurian right-lateral transform fault. Both expansions are of the same dimensions (about 8 miles in a NE-SW direction). If it were true, it could supply an alternative to the stream-valley hypothesis. The removal of the Cambro-Ordovician in the Newfoundland basin could have resulted from the intial anticlinal upwarp which is often a precursor of a pull-apart. In the Monroe basin the Ordovician shales, which constitute most of the stratigraphic thickness of the Cambro-Ordovician sequence, are in fact apparently missing in the southern part of the basin, for the Green Pond is mapped as resting on the Cambro-Ordovician carbonates east of Highland Mills (Fisher, Isachsen and Rickard, 1971). The present folds within the Outlier are the result of post-Devonian compression.

STRATIGRAPHY AND SEDIMENTOLOGY OF MIDDLE AND UPPER SILURIAN ROCKS AND AN ENIGMATIC DIAMICTITE, SOUTHEASTERN NEW YORK.

ANTHONY R. PRAVE and MOSES L. ALCALA
Dept. of Earth and Planetary Sciences
The City College of New York
New York, NY 10031

JACK B. EPSTEIN
U.S. Geological Survey
Reston, VA 22092

INTRODUCTION

During much of the Silurian Period, sedimentation patterns throughout the northern Appalachian Basin record west to northwest spreading of terrigenous and marginal marine clastics into a deeper water mud and carbonate basin (Cotter, 1988; Middleton, 1987). The clastics were derived from highlands uplifted during the Taconic orogeny in eastern Pennsylvania and southeastern New York. By the Late Silurian, active tectonism had ceased so that the (dis)equilibrium between sediment influx, basin subsidence and relative changes in sea level were the probable first-order variables controlling facies distribution.

STRATIGRAPHY

The middle to Upper Silurian rocks to be examined during the field trip crop out along a northeast-trending belt of exposures in southeastern New York (Fig. 1). The stratigraphic successions at the extremities of this belt are firmly established (Figs. 2 & 3). In northeastern areas the stratigraphy consists of the Shawangunk Formation, High Falls Shale, Binnewater Sandstone of Hartnagel (1905) and Rondout Formation whereas in southwestern areas it is the Shawangunk Formation, Bloomsburg Red Beds, Poxono Island Formation and Bossardville Limestone. Recent mapping by Epstein and Lyttle (1987) indicates that the facies mosaic between those two areas is more complicated than previously reported. They established the stratigraphic framework presented on this field trip as well as determined the structural effects of Taconic and Alleghanian deformations. This field trip is concerned with facies interpretation. The localities to be visited are shown on Figure 1 and generalized descriptions of the stratigraphic units are given in Table 1.

As shown on Figure 2, the Silurian succession thins markedly from Pennsylvania toward southeastern New York. A quartz arenite (Tuscarora Sandstone) and sandstone-shale unit (Clinton Formation) developed in central Pennsylvania undergo a facies change eastward (eastern Pennsylvania) to quartzite-conglomerate units (Weiders, Minsi and Tammany Members) and an intervening

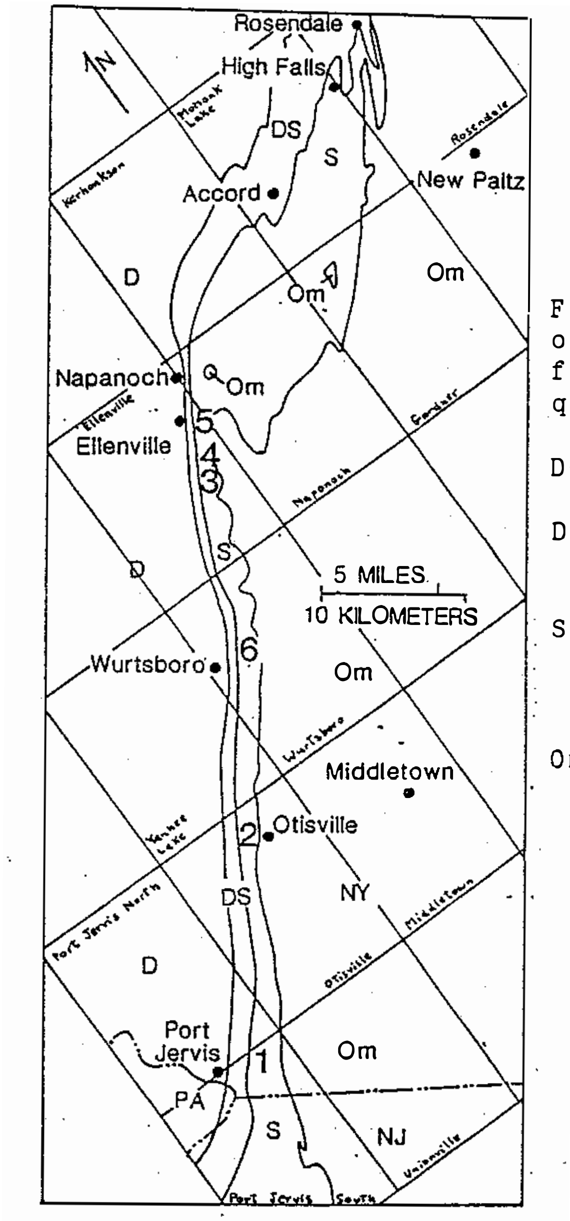

Figure 1. Generalized geologic map of part of southeastern NY showing field trip stop numbers and 7.5' quadrangle coverage.

D-Hamilton Group and younger strata.

DS-Rondout Fm through Onondaga Limestone.

S-High Falls Shale, Bloomsburg Red Beds, Bloomsburg and Shawangunk tongues, Shawangunk Fm.

Om-Martinsburg Formation.

unit containing appreciable shale and some red beds (Lizard Creek Member). Together these members constitute the Shawangunk Formation as described by Epstein and Epstein (1972). The shales of the Lizard Creek Member become less abundant in north-central New Jersey and the member cannot be mapped to the northeast, although scattered intervals of shale. some containing red beds, at various levels within the Shawangunk persist into southeastern New York. The "Otisville Shale" (see Fig. 3), named by Swartz and Swartz (1931) for a unit thought to contain enough shale to be a distinct formation near Otisville, NY (Stop 2), fails the test of mappability and Epstein and Lytle (1987) suggested that the term be abandoned.

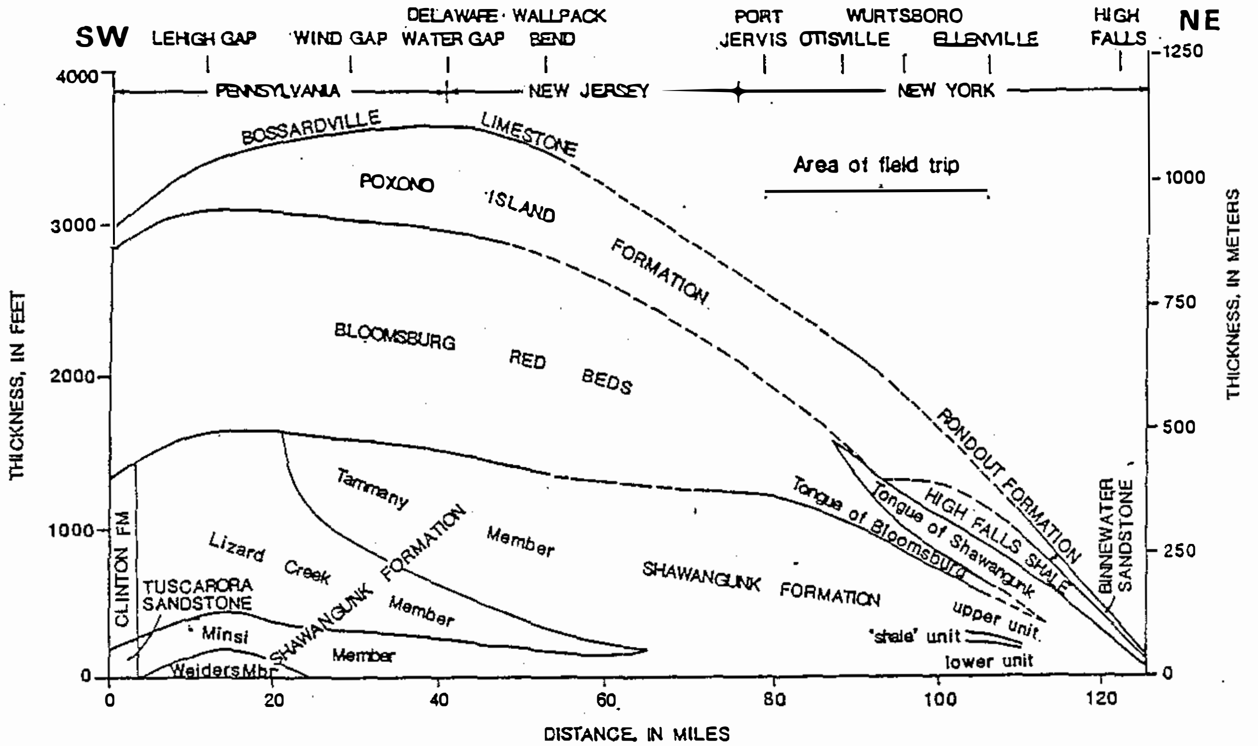

Figure 2. Stratigraphic section of Silurian rocks from Lehigh Gap, PA, to High Falls, NY. Field trip area is shown. Modified from Epstein and Lyttle (1987). Dashes indicate poor exposures of contact.

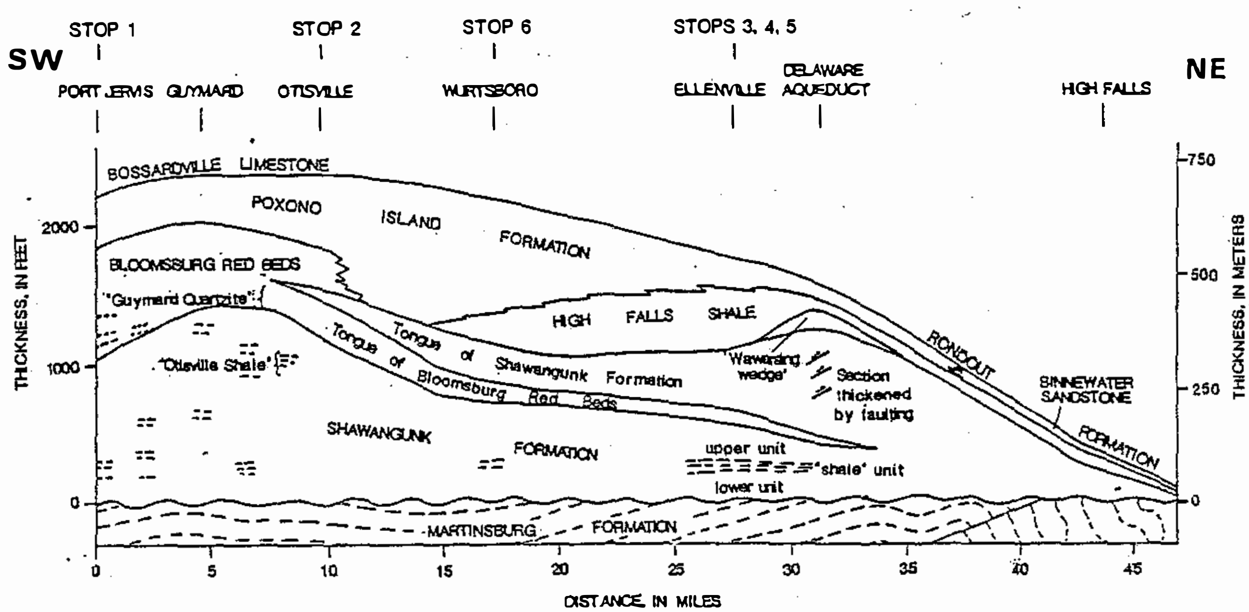

Figure 3. Stratigraphic section of Silurian rocks from Port Jervis to High Falls, NY. Field trip stops are shown. Modified from Epstein and Lyttle (1987).

TABLE 1: GENERALIZED STRATIGRAPHY IN FIELD TRIP AREA.

POXONO ISLAND FM (U. Sil.)-poorly exposed gray and green dolomite and shale, possible red shale near base. 0-150 m thick.

HIGH FALLS SHALE (U. Sil.)-type locality consists of red and green calcareous shale and siltstone; some argillaceous carbonate; 0-25 m thick. In field trip area it consists of fine to medium, trough cross-beded sandstone with thin red and gray shale layers.

BLOOMSBURG REd Beds (U. Sil.)-red and gray shale, siltstone and sandstone in meter-scale fining-upward cycles. 0-335 m thick.

SHAWANGUNK TONGUE (U. Sil.)- crossbedded (light-dark laminae) fine to medium quartzite with some pebbly quartzite; minor green shale and siltstone; rare red shale. 0-105+ m thick.

BLOOMSBURG TONGUE (U. Sil.)-reddish and grayish siltstone, shale and fine to pebbly sandstones in meters-scale fining-upward cycles. 0-50 m thick.

SHAWANGUNK FM (M. Sil.)-crossbedded, tabular to channeled quartzite and conglomerate; minor siltstone and shale, in places red and green; base is an unconformity. 0-425 m thick.

DIAMICTITE-COLLUVIUM (U. Ord.-Lr. Sil.)-"exotic" pebbles in clastic matrix; shale-chip gravel; sheared clay and quartz veins (fault gouge). Locally preserved and <0.3 m thick.

MARTINSBURG FM (Md.-Up. Ord.)-interbedded slate, shale and graywacke. Greater than 3000 m thick.

The contact between the Bloomsburg Red Beds and Shawangunk Formation has been traced without complication from eastern Pennsylvania to Port Jervis, NY (Stops 1A & 1B). Between Wurtsboro and Ellenville, NY, red beds and gray sandstone overlie the Shawangunk and, because of a similar succession of red and gray rocks at High Falls, NY, have been identified as, respectively, the High Falls Shale and Binnewater Sandstone by some workers (Gray, 1961; Smith, 1967). Detailed mapping, however, indicates that these red-gray rocks (Stops 5 & 6B) are facies ("tongues") of the Bloomsburg Red Beds and Shawangunk Formation (Fig. 3), not the High Falls and Binnewater. At Wurtsboro (Stop 6A), the High Falls Shale clearly occurs above the Shawangunk tongue. The Bloomsburg tongue gradually thins northeastward and changes in color from red to gray. The name "Guymard Quartzite" was used by Bryant (1926) for some of the gray Bloomsburg tongue rocks. This unit, like the "Otisville Shale," is unmappable and Epstein and Lyttle (1987) conclude that it also should be abandoned.

The rocks in the Shawangunk tongue are unlike typical quartzites of the Shawangunk Formation in that they are finer, better sorted and rounded, and more distinctly cross-bedded (Stops 5 & 6B). Some of the conglomerate layers in the Shawangunk tongue are similar to those in the Green Pond outlier (located about 40 km southeast of the main outcrop belt). This led Epstein and Lyttle (1987) to suggest that the Shawangunk tongue thickens in that direction to encompass most of the section in the outlier. To the southwest, the Shawangunk tongue pinches out into the Bloomsburg Red Beds (Fig. 3).

The High Falls Shale at High Falls, NY, contains red beds similar to those of the Bloomsburg tongue but also is characterized by a dolostone, fine limestone and shale facies with abundant ripple marks and dessication cracks. The similarity of those rocks to rocks in the Poxono Island Formation suggests that they probably are facies equivalents. Both formations are poorly exposed and the nature of the facies change has not been defined. Similarly, the cross-bedded sandstones of the Binnewater pinchout to the southwest (Rickard, 1962; Waines, 1976) into the Poxono Island Formation (Fig. 3). Thus the Binnewater, High Falls and Poxono Island are facies of a complex mixed carbonate-siliciclastic marginal marine sequence. As a consequence, the name "High Falls Shale" is incorrectly used on the New Jersey State geologic map (Lewis and Kummel, 1912) for rocks that should be referred to as the Bloomsburg Red Beds.

Enigmatic Diamictite

In southeastern New York, the Taconic unconformity (Stops 1B, 2 & 4) between the Martinsburg Formation (late Middle to early Late Ord.) and the overlying Shawangunk Formation (middle Sil.) represents about 10 to 30 million years, nearly the duration of the entire Silurian Period itself. The only apparent records of this period of time are a thin diamictite (Stop 2) and a shale-chip gravel (Stop 4), both patichily preserved beneath the basal unconformity of the Shawangunk Formation. The diamictite, generally less than 30 cm thick, is a mixture of angular to rounded clasts in a sand-silt matrix, dark yellowish orange in color. The clasts consist of fragments of the Martinsburg, quartz pebbles similar to those in the basal Shawangunk and, most interestingly, exotic pebbles that are dissimilar to rock types immediately above or below the unconformity. The source of these pebbles is enigmatic. The shale-chip gravel has been observed only at Stop 4 and consists of semi-consolidated, moderately sorted shale fragments derived from the underlying Martinsburg.

These deposits add an interesting footnote to Late Ordovician paleogeography in that they are evidence for the development of a 'colluvium' on the uplifted and subaerially exposed Martinsburg shales and graywackes. Subsequent pre-Shawangunk erosion resulted in the deposits being preserved only as thin remnants. The source of the exotic pebbles is no longer exposed nearby. Perhaps it was in Taconic nappes and later removed by erosion.

FACIES DESCRIPTIONS AND INTERPRETATIONS

This field trip will deal primarily with the clastic rocks above the Shawangunk Formation and with the puzzling diamictite-colluvium preserved beneath the Taconic unconformity.

The paleoflow data (Fig. 4) and facies characteristics of the Bloomsburg and Shawangunk tongues and High Falls Formation (see stop discussions) indicate deposition in intertidal to subtidal environments. The facies succession indicates an overall transgressive event from the Shawangunk Formation (braided fluvial) through the Bloomsburg (intertidal flats and channels) and Shawangunk (lower intertidal to subtidal sand bodies) tongues. A slight regression marks the lower part of the High Falls Formation (intertidal flats and channels) with resumption of the transgression in the upper part of the High Falls Formation (lower intertidal and subtidal sands) and subsequent development of the Early Devonian carbonate shelf.

Because (1) tectonism had ceased (thereby eliminating tectonically driven rapid uplift or basin subsidence) and (2) climate was for the most part constant---dry and hot given the low-latitude paleoreconstructions of Scotese et al. (1979) and Ziegler et al. (1977)---which in combination with (1) suggests no sudden changes in sedimentation rate, we tentaively infer that sea level changes were the dominant control on the transgression-regression-transgression cycle described above. Given the well-documented hierarchy of sea-level cycles and allostratigraphic correlations for Lower Silurian strata in the Appalachian Basin (Brett and Goodman, 1987; Cotter, 1988; Duke, 1987), it is tempting to suggest that the fining-upward cycles of the Bloomsburg (Fig. 5) are in part also related to high frequency changes in sea level. However, our desire is tempered because (1) the sandstone bodies in the Bloomsburg are laterally discontinuous and (2) accurate correlations to more basinal equivalents have yet to be established.

Nonetheless, we offer two 'end-member' hypotheses as possible explanations of the stratigraphic succession. (1) A state of equilibrium existed among subsidence, sea-level change and sedimentation so that 'static,' vertical aggradation occurred. The fining-upward cycles simply record the lateral migration of tidal channels through intertidal flats (the Clinton Group in eastern Pennsylvania was interpreted thusly by Smith, 1968). (2) The mostly 1-3 m thick cycles (Fig. 5) are due to high-frequency sea level changes and record the displacement during rapid sea level rise of mid-intertidal flat areas over finer, upper intertidal flats. In both scenarios, the Shawangunk tongue represents a transgressive apex (subtidal sands over intertidal flats and channels) prior to the slight sea level drop during lower High Falls time (return of intertidal flats and sands) followed by a continuous rise of sea level. We welcome any suggestions or insights you may have.

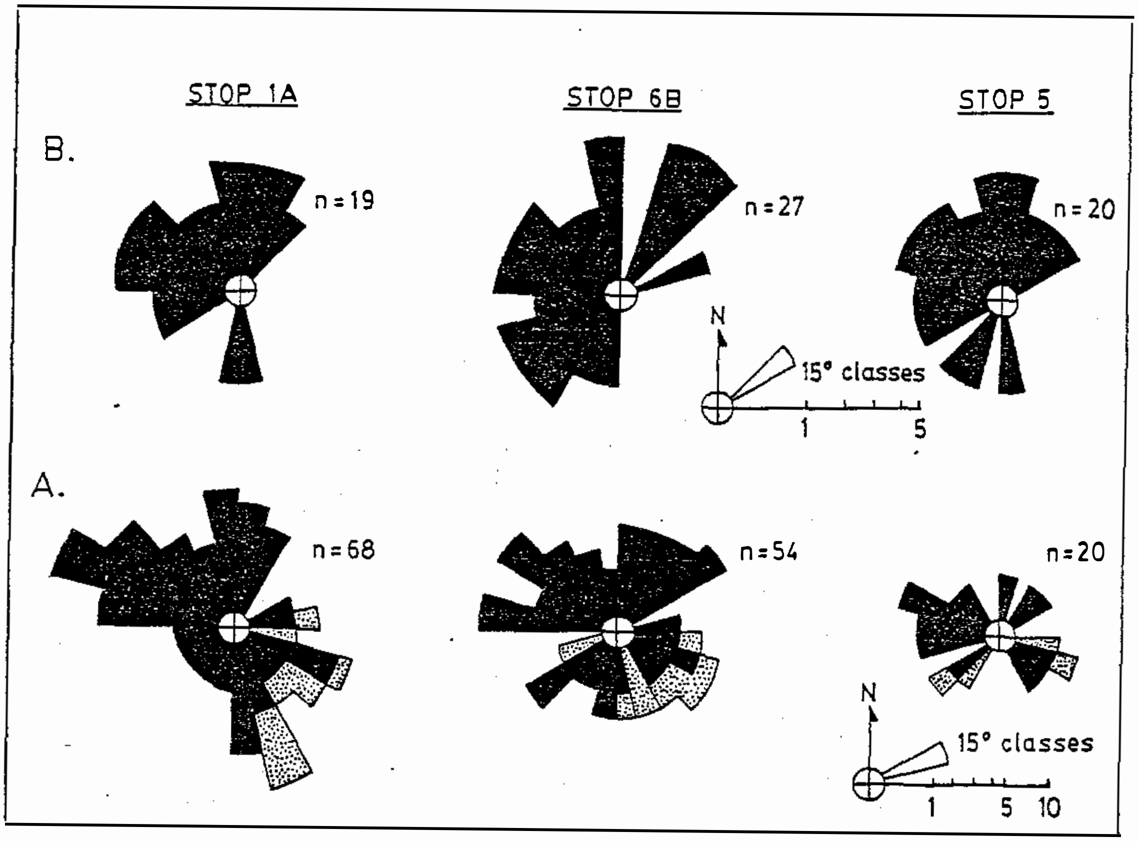

Figure 4. Equal area rosettes. A. Bloomsburg tongue data. Solid pattern-trough axes and edges and lateral accretion surfaces; stippled pattern-small-scale trough cross-strata. B. Shawangunk tongue data. Solid pattern-trough axes and erosive edges.

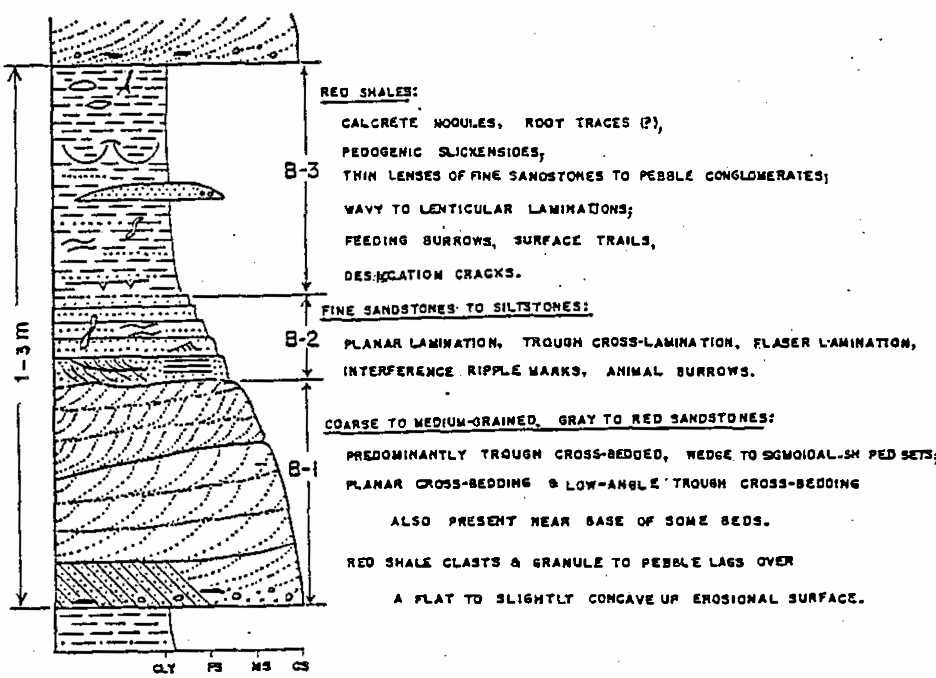

Figure 5. Idealized fining-upward cycle in the Bloomsburg Red Beds. B-1-tidal channel. B-2-3-intertidal flats and paleosols.

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ROAD LOG FOR STRATIGRAPHY AND SEDIMENTOLOGY OF MIDDLE AND UPPER SILURIAN ROCKS AND AN ENIGMATIC DIAMICTITE, SOUTHEASTERN NEW YORK

Cumul. Milage	Miles from Last Point	ROUTE DESCRIPTION
0	0	Proceed SE along Bedford Ave. from South St. intersection at O.C.C.C. Campus, Middleton, NY.
0.4	0.4	Traffic light, turn right onto NY 17M.
1.6	1.2	Turn right onto I84 West. For the next 12 miles there are scattered outcrops of Martinsburg Fm.
13.2	11.6	Martinsburg-Shawangunk contact to left.
14.0	0.8	Parking area to right affords an excellent view of the Pocono Plateau; intermediate ridges are in the Md. Dev. Hamilton Gp; Neversink Valley is floored by Up. Sil-Low. Dev. rocks.
16.3	2.3	Take Exit 1, US6 West. Turn left onto US6 West; at traffic light turn left towards I84 East; Mid. Dev. Esopus Fm on right. Go under overpass and turn left on I84 East. Ascent is on Wisconsinian till covering Up. Sil.-Low. Dev. rocks.
19.3	3.0	Depart vehicles at beginning of outcrop on right and have drivers continue on (0.6 miles) to rest area on right (or drive your vehicle to rest area and walk back down the exposure).

STOP 1A (PORT JERVIS AREA). BLOOMSBURG RED BEDS.

Epstein and Lyttle (1987) estimate that the Bloomsburg Red Beds is about 335 m thick near Port Jervis, NY. These strata can be traced continuously across New Jersey from Pennsylvania with little facies change. The Shawangunk tongue, which separates the main body of the Bloomsburg from the Bloomsburg tongue to the northeast, thins to a feather edge at this locality and is unmappable at a scale of 1:24000. For the sake of brevity, the description and interpretation of the Bloomsburg tongue will be given at this stop and at Stop 5 for the Shawangunk tongue.

Alcala measured the section from the top of the Shawangunk Formation to the top of the Shawangunk tongue (Fig. 6). The Shawangunk tongue is only 5 m thick. It is described in detail at Stops 5 and 6B. The Bloomsburg between those two units is 65 m thick and consists of 45 fining upward cycles. Discontinuous pebbly lags occur above a flat to scoured erosive base which is overlain by fine to coarse quartzose sandstones and lithic wackes. These in turn grade upward into reddish or grayish fine sandstones and siltstones generally overlain by blocky red mudstones. Thin, discontinuous stringers of coarser sandstones are common in these finer clastics.

The fine to coarse sandstones above the erosive base consist of several centimeters- to decimeters-thick cosets of trough cross-beds arranged in sigmoidal- and wedge-shaped bundles bounded by lateral accretion surfaces. The cosets generally thin upward. Planar cross-bedding and reactivation surfaces are present locally. Interference ripples are common on bed tops. The laterally continuous appearance of these sandstone beds is deceiving. They actually are broad, shallow channels that thin and pinchout into finer clastics (this will be seen from the top of the outcrop from a near normal-to-paleoflow view across I-84).

The overlying red and gray fine sandstones and siltstones contain abundant small-scale trough cross-bedding and flat lamination. Flaser and lenticular bedding, reactivation surfaces, interference ripples and burrows are common locally. The red mudstones generally contain desiccation cracks, horizontal and sub-vertical burrows and possible root traces. Calcrete zones and pedogenic slickensides are present in places.

These facies characteristics coupled with the paleocurrent data (Fig. 4) lead us to conclude that the Bloomsburg Red Beds record deposition in an intertidal environment. The broadly-channeled, coarse sandstones were tidal channels that migrated through a finer detrital intertidal flat. Paleosols developed locally in a dry climate as indicated by the calcrete zones and dessication cracks. The thin, coarse sandstone stringers in the shales probably record exceptional tidal or storm events.

From the paleoflow data (Fig. 4), it is evident that ebb flows were dominant. The larger-scale trough cross-beds and lateral accretion surfaces make up the WNW-mode whereas the ESE-mode is more dependent on the presence of smaller bed forms. The lateral accretion surfaces and sigmoidal-shaped bundles indicate that tidal currents continuously mobilized the sandy substrate in a general WNW-ESE trend. These observations corresponds nicely with the pattern of tidal forcing one would expect given the Silurian paleogeographic reconstructions of Scotese et al. (1979) and Ziegler et al (1977) showing the Taconic highlands as a NE-trending belt that had an epeiric sea to the west (Fig. 7).

We interpret the strong NNE mode on the rosette as representing a dominant lateral migration direction of the tidal channels due to the influence of a NE-flowing longshore current. Again this fits well with the paleoclimatic inferences of Ziegler et al. (1977) which indicate a southern, low-latitude high pressure system to the west of the Taconic highlands (Fig. 7); the counterclockwise circulation of surface winds associated with this high would force a NE-flowing longshore current along the western edge of the highlands. Thus, the paleoflow data indicate a dominant WNW ebb tide and a weaker ESE flood tide with a moderately influential NNE-flowing longshore current.

Unfortunately, the exposures are not quite good enough to determine spring-neap cycles. However, many of the coarser sandstones appear to have bundles of well-developed cosets and lesser-developed cosets which we have tentatively interpreted as full-vortex and slackening structures, respectively.

19.9	0.6	Reboard vehicles in rest area; continue on I84.
20.7	0.8	Park vehicles on shoulder to right.

STOP 1B (PORT JERVIS AREA). TACONIC UNCONFORMITY AND BASAL SHAWANGUNK FORMATION.

The Shawangunk Formation is interpreted as recording a NW-flowing braided fluvial system draining the Taconic highlands (Yeakel, 1962; Smith, 1967; Epstein and Epstein, 1967, 1972). The facies and paleoflow data from this exposure (which is representative of the basal Shawangunk throughout the area) support that interpretation (at least for the basal part of the formation); thick-bedded, massive to large-scale trough cross-bedded, quartz-pebble conglomerate and coarse quartzite yielding uniquadrant paleocurrent rosettes (Fig. 8).

Here, the Shawangunk rests unconformably on shale and graywacke of the Martinsburg Formation. The angular discordance is 5^{o}. Along the contact is a 5 to 15 cm thick fault gouge containing angular Martinsburg fragments in a dark-yellowish orange clay matrix. Thus, this contact also represents a zone of tectonic displacement, the amount of which is not known. For a complete discussion of the relative effects of the Taconic and Alleghanian orogenies, see Epstein and Lyttle (1987).

		Reboard vehicles and continue on I84.
21.4	0.7	Take Mountain Rd (Exit 2); at stop sign turn right; you will travel along south slope of Shawangunk Mtn; note occassional Martinsburg outcrops.
26.4	5.0	Bear left on Mountain Road (Rte. 73).
29.6	3.2	Otisville; turn left onto Field Road.
30.1	0.5	Turn left onto Walker Street.
30.5	0.4	Stop sign; turn left onto State Street (NY 211).
30.7	0.2	Park vehicles in broad open area on left.

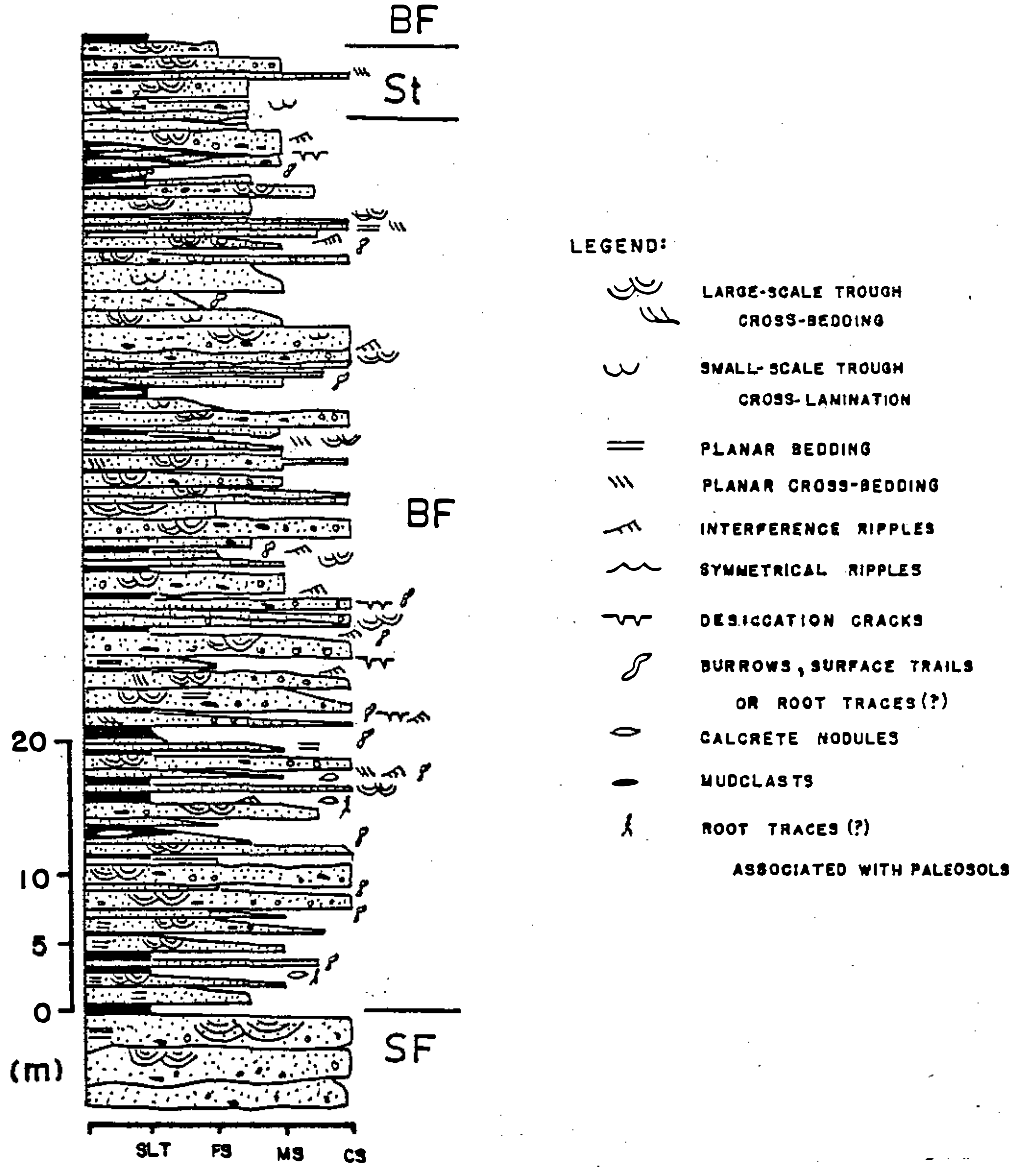

Figure 6. Measured section along I-84 east of Port Jervis, NY. BF-Bloomsburg Red Beds; SF-Shawangunk Fm; St-Shawangunk tongue.

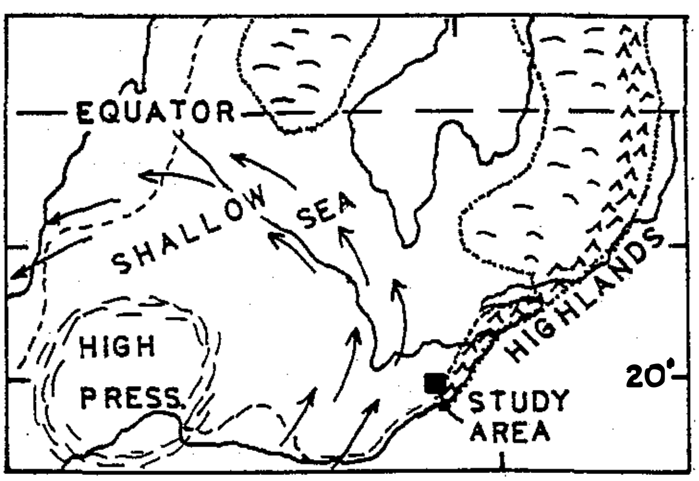

Figure 7. Middle Silurian paleogeography after Zieglar et al. (1977). Arrows indicate circulation pattern of surface winds.

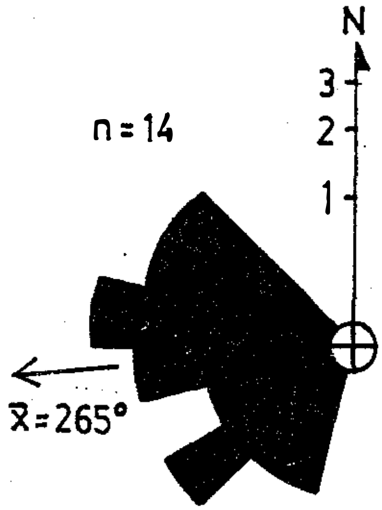

Figure 8. Equal area rosette constructed from trough axes and erosive edges measured in the basal part of the Shawangunk Formation at Stop 1B.

STOP 2 (OTISVILLE). DIAMICTITE AT THE TACONIC UNCONFORMITY.

This classic exposure of the Taconic unconformity along the railroad cut was visited on the 59th NYSGA Meeting (Epstein and Lyttle, 1987). This discussion is a summary of that visit.

The quarry across the road is in the Shawangunk Formation; it is about 2 km along strike from the type locality of the "Otisville Shale" of Swartz and Swartz (1931). As is evident, shale is meager. The "Otisville" is poorly-defined and should be discarded.

The shales and siltstones of the Martinsburg dip moderately to the northwest, whereas, above the unconformity, the quartz-pebble conglomerate of the Shawangunk dips 16^{o} less. The basal surface of the Shawangunk is irregular and displays downward projecting mullions having a few centimeters of relief and spacings up to a few decimeters. Bed-parallel shearing is developed in places above the unconformity.

Along the contact is a diamictite as much as 30 cm thick with sharp, unconformable upper and lower contacts. Within the mud matrix are angular clasts of the Martinsburg (in places large parts of the Martinsburg appear to have been bodily incorporated into this zone) as well as angular to rounded "exotic" pebbles not present in the underlying bedrock: clean and pyritic quartz arenite, feldspathic and chloritic sandstone, cross-laminated feldspathic conglomeratic quartzite, siliceous and micaceous siltstone and red siltstone. Weathering rinds and pits are present on some of the pebbles and, along with the rounding, are evidence for subaerial transport and exposure prior to deposition. The above indicates that the diamictite is a product of mass wasting, possibly a colluvial gravel.

Also in this zone is a grayish gouge with slickensided quartz veins. This gouge and the mullions are typical of many of the Martinsburg-Shawangunk contacts in southeastern New York and indicate that the unconformity is a plane of movement, the displacement along which is not known.

		Reboard vehicles and turn left onto NY 211.
30.8	0.1	Almost immediately turn right onto Orange County 61.; Shawangunk Fm to right.
32.9	2.1	Stop sign; turn right onto US 209 North.
39.7	6.8	Wurtsboro; proceed past traffic light.
42.0	2.3	Wurtsboro airport; Shawangunk mine (Pb-Zn) on top of Shawangunk Mtn to right (Gray, 1961).
46.9	4.9	Wisconsinian morraine; it is over 3 miles wide.
50.8	7.2	Mount Marion Fm to left; deposits of glacial Lake Warwarsing to right.
51.8	1.0	Traffic light in Ellenville; turn right on NY 52 East (Center Street).
52.7	0.9	Stop sign; bear right on NY 52. You will cross North Gulley (Bloomsburg & Shawangunk tongues to left) and proceed up Shawangunk Mtn; exposures of Shawangunk Fm will be on your left.
55.7	3.0	"Shale unit" within Shawangunk Fm on left.
56.8	1.1	Turn left towards Cragsmoor and Ice Caves Mountain National Landmark.
58.2	1.4	Cragsmoor; turn left just before Post Office onto Meadow Lane.
58.5	0.3	Turn right at "T."
58.7	0.2	Turn left at "T."
58.9	0.2	Park vehicles in parking area and follow path to Bear Hill. After lunch return to vehicles.

STOP 3 (BEAR HILL). LUNCH AND SHAWANGUNK FORMATION CROSS-BEDS.

These nearly flat-lying beds of the lower part of the Shawangunk Formation are near the crest of the Ellenville arch, a broad, open fold of Alleghanian age (Epstein and Lyttle, 1987). The basal 30 m of the formation are exposed here and consist primarily of quartz-pebble conglomerate and medium to pebbly quartzite composed of decimeter-scale cosets of trough cross-bedding. A well-developed joint pattern and numerous exposures of bedding surfaces afford an excellent opportunity to examine the relationship between quasi-3-D views and plan views of trough cross-bedding. These exposures can be used to discuss the 'proper' methods of obtaining paleoflow data from troughs.

		Reboard vehicles and continue past parking area.
59.1	0.2	Bear right at intersection.
59.2	0.1	Turn right onto Meadow Lane
59.3	0.1	Stop sign; turn right (downhill).
60.7	1.4	Stop sign; turn right onto NY 52.
64.5	3.8	Turn right onto South Gulley road.
65.3	0.8	Turn left onto Mt. Meenhaga Road; Martinsburg Fm exposure to left.

65.7	0.4	Park vehicles in open area at bend in road and walk 100 m back down road to Stop 4.

STOP 4 (ELLENVILLE). SHALE-CHIP GRAVEL AT TACONIC UNCONFORMITY.

This exposure also was visited on the 59th NYSGA Meeting (Epstein and Lyttle, 1987) but it is a problematic exposure worthy of further puzzlement. Your opinions are encouraged.

The basal Shawangunk forms a slight overhang which affords a view of an in-place, shale-chip gravel developed beneath the unconformity. The shale-chip gravel is at least 60 cm thick and consists of very well sorted shale chips about 3 cm in length derived from the underlying Martinsburg (exposed about 6 m below this outcrop). It is crudely foliated parallel to bedding in the Shawangunk. This shale-chip gravel is similar to those of Pleistocene age derived from Paleozoic shales in many areas in Pennsylvania and New Jersey.

The excellent sorting suggests winnowing of the gravel during transport. This gravel and the diamictite observed at Stop 2, indicate that the Martinsburg surface, uplifted during the Taconic orogeny, was subaerially exposed and covered by colluvium and gravels. Much of this material was subsequently removed by erosion and the remnants were buried by the Shawangunk conglomerates. There are no known nearby sources for the exotic pebbles in the diamictite. As stated earlier, it is possible that they were derived from now-eroded Taconic nappes.

		Reboard vehicles and go back down Mt. Meenhaga Rd
66.0	0.4	Turn right onto South Gulley Road.
66.8	0.8	Stop sign; turn right onto NY 52.
67.2	0.3	Turn left and immediately right into parking area.

STOP 5 (ELLENVILLE). BLOMMSBURG AND SHAWANGUNK TONGUES.

This section was measured by Alcala. The Bloomsburg tongue is 48 m thick and consists of 37 fining upward cycles (Fig. 9). Seven meters of the Shawangunk tongue are exposed. The contact between the Shawangunk Fm and Bloomsburg tongue is gradational as is the contact between the two tongues.

The fining-upward cycles in the Bloomsburg tongue are similar to but contain more shale and mudstone than the Bloomsburg strata seen at Stop 1A. The sandstones forming the basal parts of cycles are the same. Quartz and lithic wackes that consist of sets and cosets of trough cross-strata are arranged in sigmoidal-shaped bundles commonly bounded by lateral accretion surfaces. Given the same facies characteristics and paleoflow trends (Fig. 4), we interpret this section similarly. The dominant WNW-ESE bipolar rosettes indicate tidal flow. The data obtained from the larger bed forms again suggest a stonger ebb tide relative to the flood tide. The NNE mode indicates the continuing influence of northeast flowing longshore currents in forcing a preferred NNE migration direction of tidal channels. The finer clastics represent the intertidal flats. We interpret their greater proportion here as indicating deposition primarily in upper intertidal flat areas.

The first tabular-appearing quartz arenite bed marks the base of the Shawangunk tongue which consists predominantly of thick-bedded, fine to medium quartz arenites with minor greenish shale and rare red shale. Shales occur as thin partings or layers up to 20 cm in thickness. Pebbly quartzite occurs in places (such a bed occurs about 4 m above the base here and at the section near Wurtsboro). Bases of beds are erosive and flat to shallowly scoured and pebbly lags occur locally. Trough cross-bedding is common and occurs in sets and cosets 0.1 to 1 m thick. Planar cross-bedding and flat lamination occur locally; some of the finer sandstones contain small-scale cross-stratification. In all cases, laminations consist of alternating

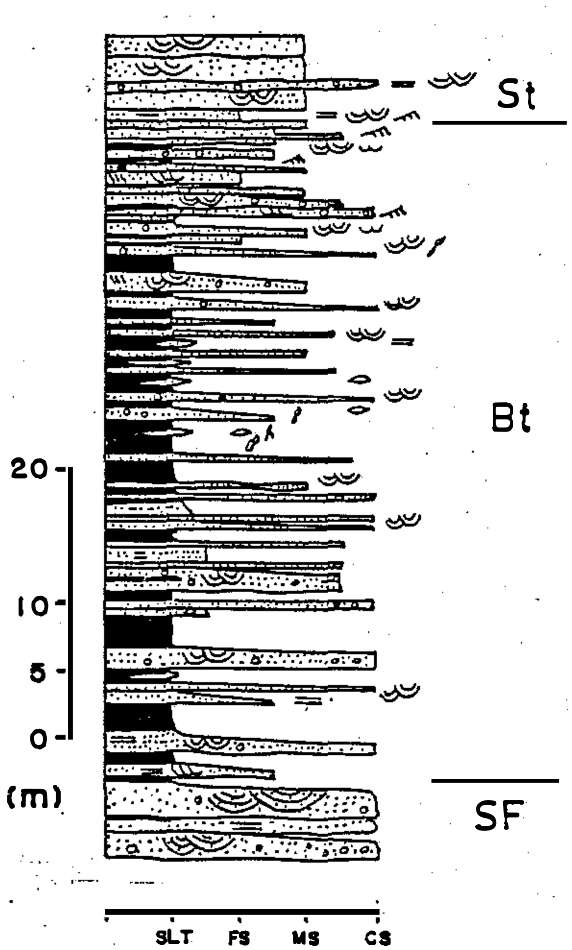

Figure 9. Measured section of Bloomsburg (Bt) and Shawangunk (St) tongues in North Gulley near Ellenville, NY. SF-Shawangunk Fm. See Figure 6 for explanation of symbols.

light-dark couplets. Some of the foresets are defined by flat mud clasts. Interference ripples and, rarely, trails and burrows can be observed on some bed surfaces.

On the basis of the facies characteristics and paleocurrent data (Fig. 4), we interpret the Shawangunk tongue as a subtidal (and possible lower intertidal) sand body. The clustering of flow data in the western hemisphere of the rosettes (i.e., from 180^{o} to 360^{o}) indicates primarily offshore-directed transport, probably by strong ebb tides and possibly by rip and storm currents (note, however, that no hummocky or swaley cross-stratification has been observed to support a storm influence). The NNE mode indicates sediment transport by longshore currents. We interpret the 'clean' quartz arenitic character of these sandstones as resulting from selective sorting by marine currents (the coarser, 'dirtier' sands were 'locked-up' in the tidal channels of the Bloomsburg). The presence of interference ripples and rare trace fossils further supports this interpretation.

Modern tidal flats display a progressive facies change. Upper intertidal areas are finer-grained, less and smaller channeled relative to the well-channeled, mid intertidal areas whereas lower intertidal areas are mostly sandy with less fines; adjacent subtidal zones are sandy (Klein, 1985). Examination of the measured sections (Figs. 6, 9 & 11) indicates such a trend. The lower part of the Bloomsburg tongue contains more fines than does the uppermost part, i.e., deposition initially occurred in middle and upper areas of an intertidal environment with subsequent deposition in lower intertidal environments. As the transgressive event progressed, those deposits were covered by sandy subtidal deposits of the Shawangunk tongue.

		Reboard vehicles and turn left onto NY 52.
67.3	0.1	Turn left onto Center Street.
68.2	0.9	Traffic light; turn left onto US 209 South.
80.3	12.1	Wurtsboro; continue past traffic light unless you would like a cold beer and a pastrami sandwich at Danny's.
81.7	1.4	Turn left onto NY 17 South towards New York City.
83.0	1.3	Park vehicles on shoulder on right.

STOP 6A (WURTSBORO). HIGH FALLS SHALE.

The High Falls Shale and correlative Poxono Island Formation are poorly exposed in this area. The nearest exposures to this locality are 34 km to the northeast near Accord, NY and about a similar distance to the southwest in New Jersey. The sections at Stop 6 (A & B) were measured by Alcala.

The section of the High Falls Shale is 10 m thick (Fig. 10), but it is incomplete (both contacts are covered). The High Falls consists of fine to medium quartzose sandstones with minor, laterally discontinuous greenish and reddish shale layers. Beds are composed of large-scale trough cross-bedded sets and cosets as much as 80 cm thick. Lateral accretion surfaces and, less commonly, reactivation surfaces bound many of the sets. Planar cross-bedding, flat lamination, desiccation cracks, symmetrical ripples, and burrows and trails are moderately common. Bases of beds are flat but erosive and have local lags. Bed tops generally are mantled with interference ripples.

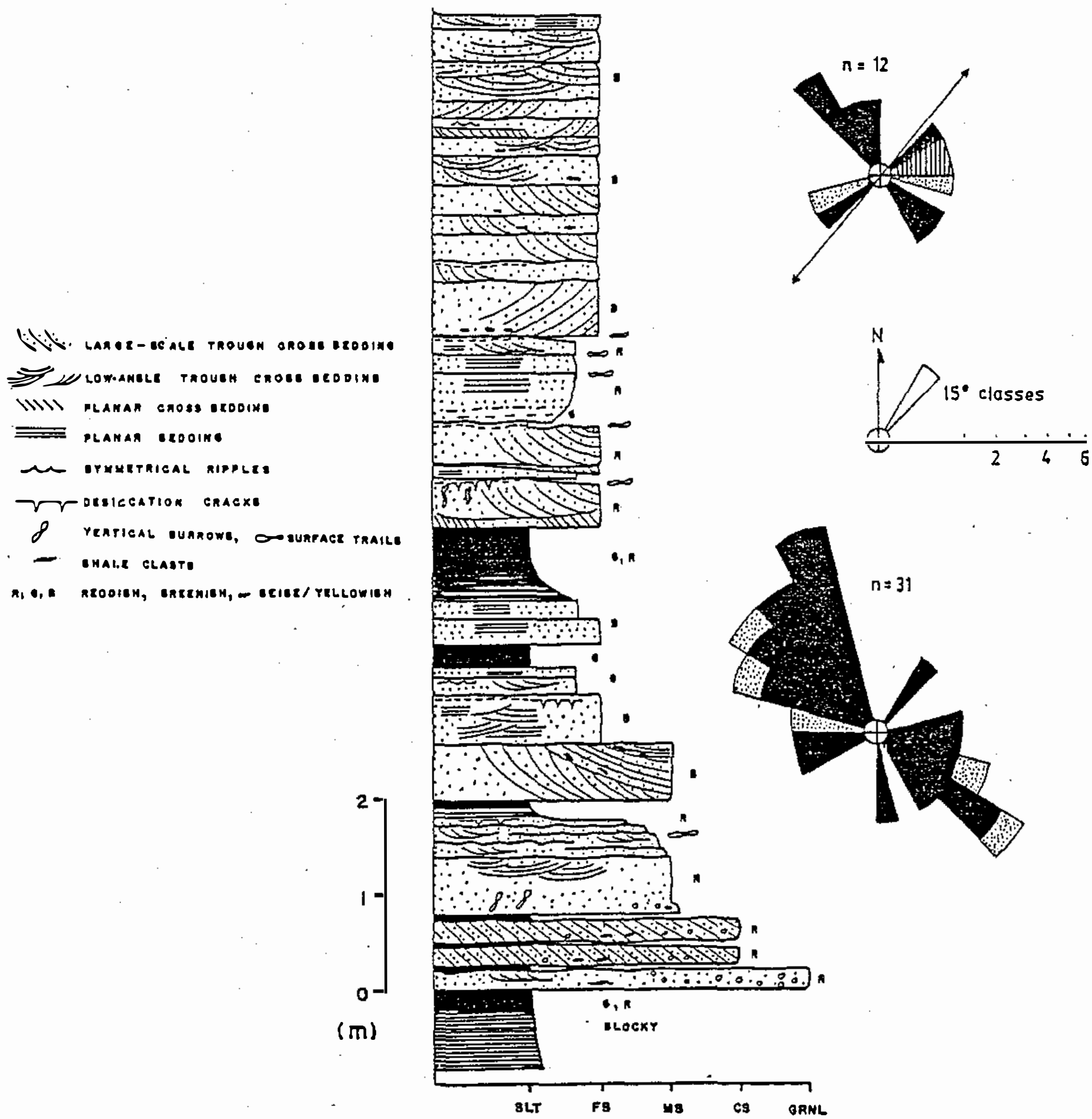

Figure 10. Measured section of the High Falls Formation along Rte. 17 near Wurtsboro, NY. Equal area paleocurrent rosettes: solid pattern-trough axes and edges; stippled pattern-small-scale trough cross-strata; vertical lines-planar cross-bedding. Double arrow represents average trend of three wave ripple crests.

Examination of the paleocurrent rosettes (Fig. 10) clearly indicates that these are tidal deposits. The sedimentary structures and composition suggest lower intertidal to subtidal settings. The WNW-directed ebb tide was dominant. A less-developed NE mode is once again interpreted as recording the influence of longshore currents. Curiously, these flows are rotated slightly clockwise relative to those of the Bloomsburg and Shawangunk tongues. We tentatively conclude that this probably indicates a slight reorientation of the paleostrandline to a more northeasterly direction. The NE-trend of wave ripple crests (Fig. 10) adds some credence to this speculation.

84.0	1.0	Reboard vehicles and continue along NY 17 South. Proceed past curve in road and park along shoulder on right.

STOP 6B (WURTSBORO). BLOOMSBURG AND SHAWANGUNK TONGUES.

The Bloomsburg and Shawangunk tongues are best exposed at this locality which is a normal-to-paleoflow view. Contacts between the "tongues" and the underlying Shawangunk Fm are gradational. The Bloomsburg tongue is 58 m thick and consists of 41 fining-upward cycles (Fig. 11). The Shawangunk tongue is nearly complete (top contact is covered) and is 34 m in thickness. The facies here are similar to those observed at Stops 1A and 5 except that they are thicker and have more paleosol and caliche layers. Thus, given the same facies and paleoflow trends (Fig. 4), the same interpretation is proposed: the Bloomsburg tongue is an intertidal deposit recording WNW-ESE ebb-flood tides and the Shawangunk tongue is a subtidal deposit. Both "tongues" were influenced by a NNE-flowing longshore current.

84.2	0.2	Reboard vehicles and continue along NY 17 South. We will pass the Taconic unconformity and a long exposure of Martinsburg Fm described by Epstein and Lyttle (1987, p. C-71-73); under overpass is a slump structure to the right.
94.9	10.7	Bear right onto I84 West and return home or to the O.C.C.Campus.

END OF TRIP. HAVE A SAFE JOURNEY HOME.

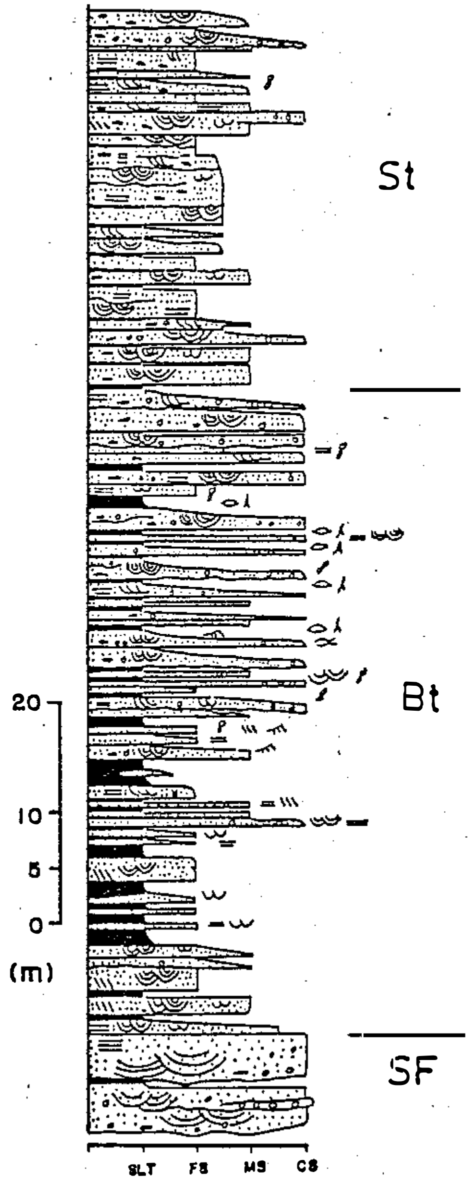

Figure 11. Measured section of the Bloomsburg (Bt) and Shawangunk (St) tongues along Rte. 17 near Wurtsboro, NY. SF-Shawangunk Formation. See Figure 6 for explanation of symbols.

A GEOLOGICAL CLIMB OF SCHUNEMUNK MOUNTAIN

WILLIAM J. TUCCI
Science Department
Valley Central High School
Montgomery, New York 12550

ROBERT KALIN
E.S.S. Department
Suffolk County Community College
Selden, New York 11784

GENERAL INTRODUCTION

This trip focuses on the area known as the Hudson Lowlands. To the east are found the Hudson Highlands, an area of Precambrian terrane of high relief which strikes northeast-southwest. The Catskill Mountains of Devonian age, with Silurian and Lower Devonian outcrops along their eastern front are found to the west. The Hudson Lowlands are an area of relatively low elevation which are underlain by Cambro-Ordovician calcarenites and argillites. It is part of a physiographic trough that extends from the St. Lawrence Valley south into Pennsylvania. This region has a major down-fold referred to as the Schunemunk Outlier. Schunemunk Mountain which is a portion of the central axial region has a height of 507 meters (1664 feet). The synclinal fold that makes up the axial region of the lowlands has flanks upon which older rocks are exposed in topographic highs. The central part is underlain by the youngest rocks which, being resistant sandstones and conglomerates, form ridge-like masses of northeast-southwest trend along the axis. (Kothe, 1960).

The word Schunemunk, pronounced Skun-uh-munk, is the Algonquin Native American name given to the mountain by its early inhabitants. The name means "excellent fire-place" in the Algonquin language and may be the result of the fact that this group at one time occupied a fortified site somewhere near the northern end of the mountain.

Schunemunk is a folded and faulted mountain, unlike the Catskills to the north and west which are erosional remnants of a plateau. Schunemunk is a complexly folded and faulted remnant of the Devonian. Despite the ravages of glaciers, erosion, mass wasting, and the frost action of the Hudson Valley, it stands out as one of the highest peaks in Orange County and one of the highest west of the Hudson River.

HISTORY

Topographically the mountain is almost indistinguishable from the Precambrian Hudson Highlands to the east and Woodcock Hill to the west. However, it is composed of much younger Devonian sediments deposited after the Taconic Orogeny ended the Ordovician Period.

Although several interpretations of the geological evidence may be made, the following ones seem intellectually satisfying. Rodgers (1987) makes the point that the Alleghanian event certainly deformed the entire Valley and Ridge province and transported the Blue Ridge and Inner Piedmont and "probably also those of the Highlands from New York southwest into the Reading Prong."

The equivalence of the Green Pond formation and the Shawangunk conglomerate (Kothe, 1960) and the Shawangunk's occurrence around the flanks of the Schunemunk Outlier relate the Green Pond, Shawangunk and Schunemunk conglomerates stratigraphically. Fraill (1985) further indicates that the folding and faulting of the Green Pond was Alleghanian, with perhaps some further faulting during the Jurassic. It would seem logical to conclude that the folding of the Schunemunk was also associated with this event. The Alleghanian orogeny, approximately 260 m.y.a., marked the last crustal collision of the continents and folded these Devonian beach deposits to form the syncline that was to become Schunemunk Mountain.

Schunemunk may be described as an island of Devonian sediments in a sea of Ordovician and Precambrian metasediments (See Fig. 1. Geologic Map of Schunemunk, after; Jaffe and Jaffe, 1967 and Fisher, 1970) The stratigraphic column (Fig. 2.) shows the relationships among the various rock units which underlie Schunemunk Mountain. Notice that the uppermost layer in the section, the glacial deposits, have been eroded along the ridge of the mountain exposing the Schunemunk Conglomerate.

The cross-section (Fig. 3.) shows the structural relationship between the rock units of the stratigraphic column and their influence on the topography of the region. ("Geology is the mother of topography.") The vertical exaggeration of the cross-section serves to illustrate the resistance to erosion of the Schunemunk conglomerate and the gneisses of the Hudson Highlands.

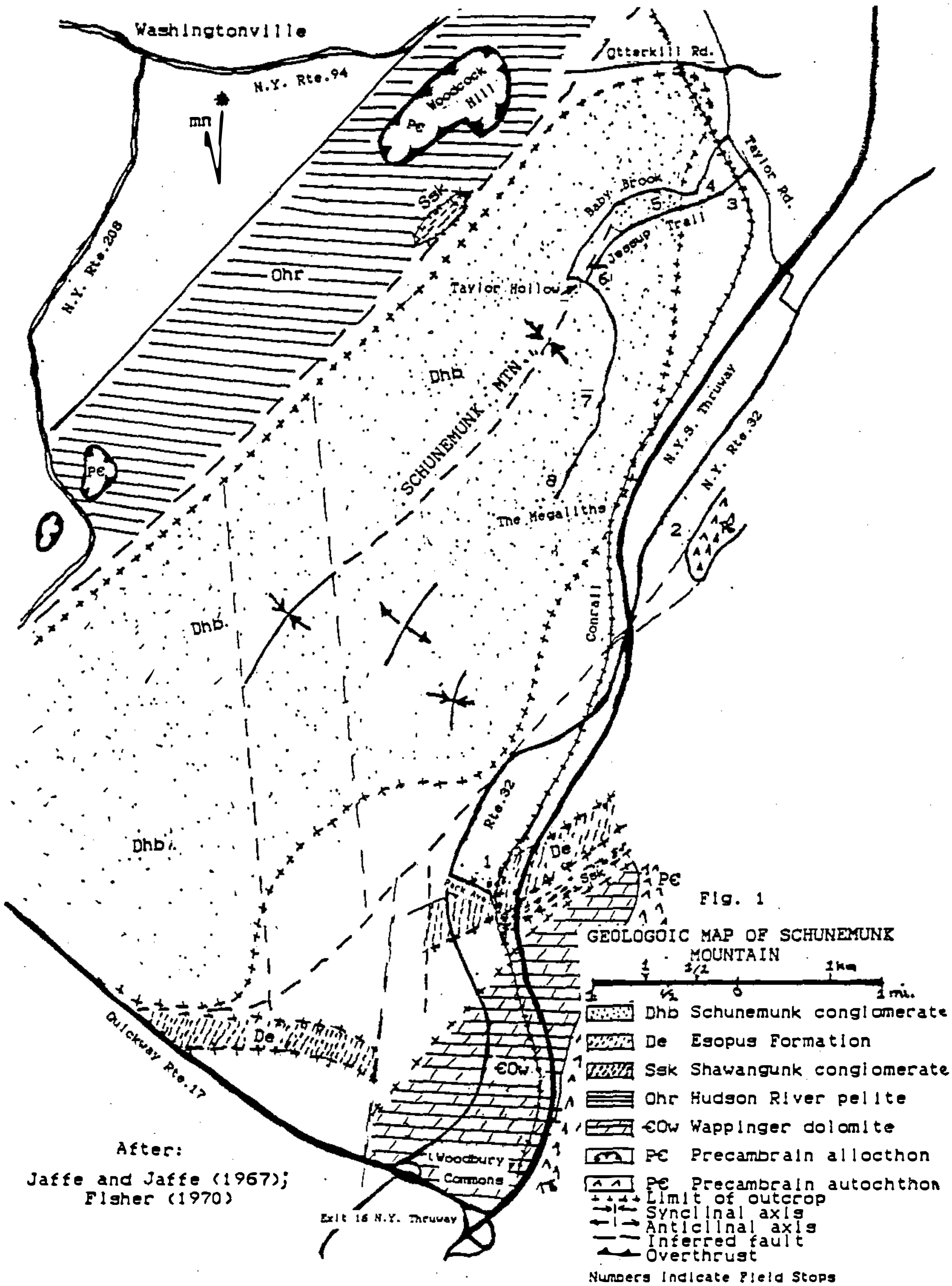

Fig. 1

GEOLOGOIC MAP OF SCHUNEMUNK MOUNTAIN

After:

Jaffe and Jaffe (1967);
Fisher (1970)

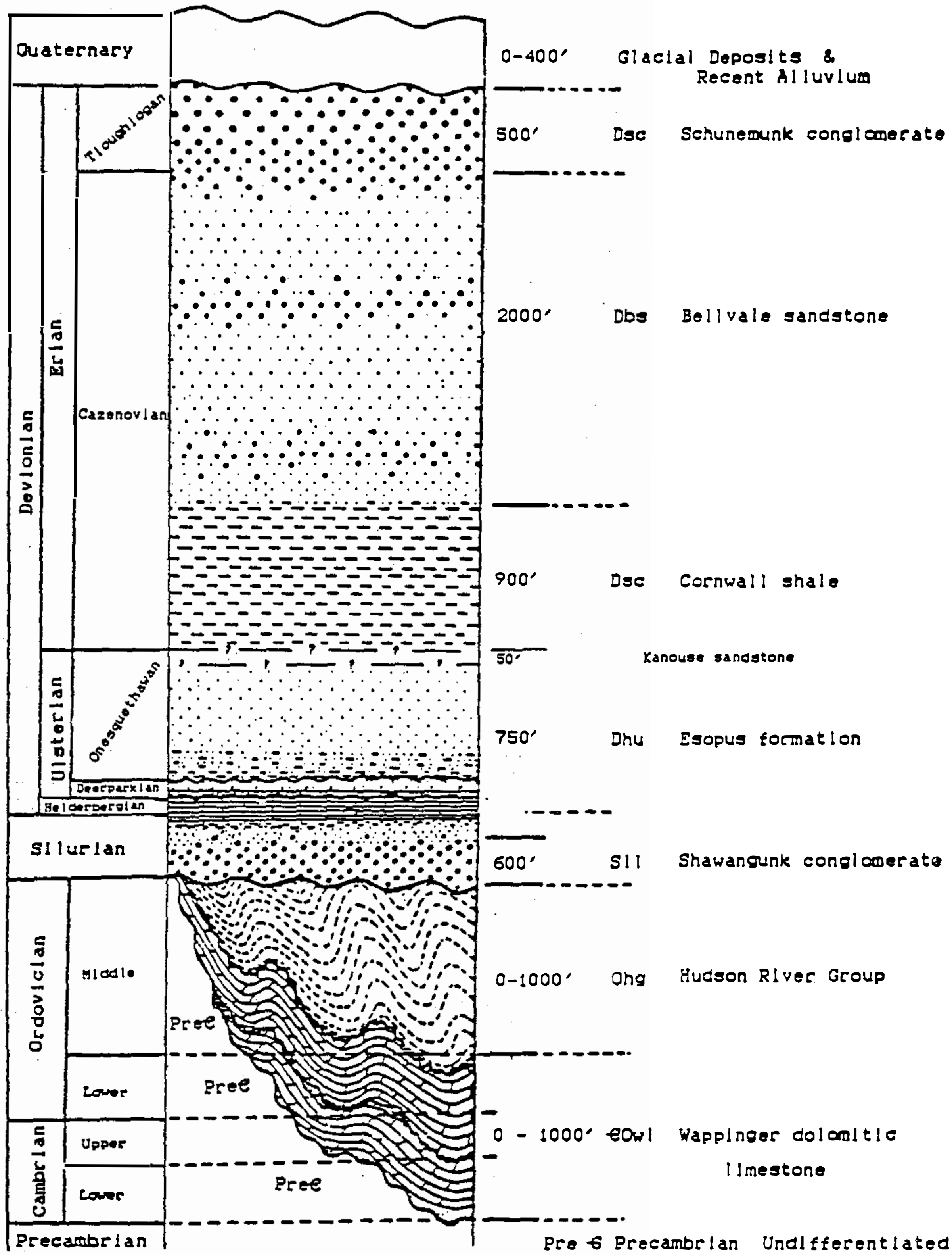

Fig. 2. Generalized Stratigraphic Section in the vicinity of Schunemunk Mountain. (after: Kothe, 1960).

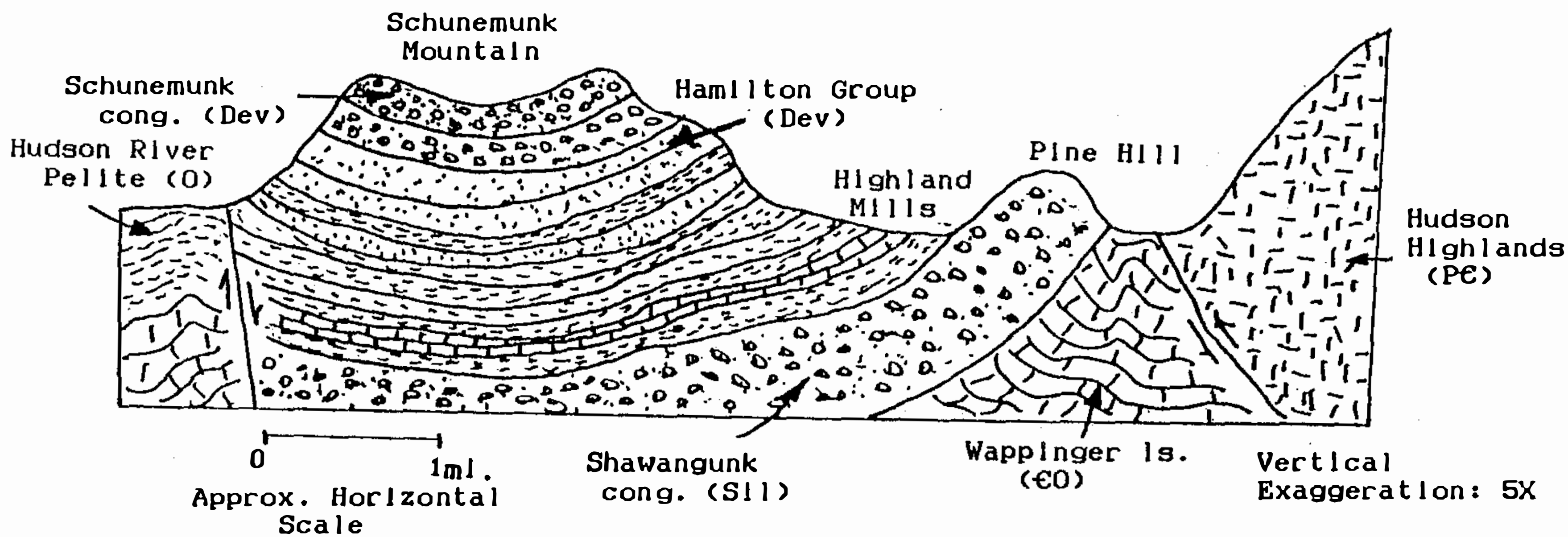

Fig.3. Generalized cross-section of Schunemunk Mountain in the vicinity of Highland Mills. (after: Schuberth, 1968).

Folding and Faulting

The rock units shown in the stratigraphic column were subsequently folded during the Alleghanian event which marked the last closing and opening of the proto-Atlantic Ocean, between 220 and 280 million years ago. The faulting of the rock units on Schunemunk, according to Kothe (1960), occurred as a result of the stresses produced during this folding event. Jaffe and Jaffe (1967) and Faill (1985) further speculate that some of these faults may have been produced or enlarged as a result of the Triassic/Jurassic deformation.

Glaciation

One of the striking features of Schunemunk Mountain is the effect of glacial scouring and polishing along the ridge. The faceted quartz pebbles, the roche-moutonnee, and the glacial striae on the outcrops bear witness to the power of the glacial ice as it moved southward during the ice advance 20,000 plus years ago. The mountain may well have been 90 to 120 meters (300 to 400 feet) higher prior to the glacial erosion (Kothe, 1960).

TOPOGRAPHY

Schunemunk Mountain stretches from the intersection of Otterkill Road and Taylor Road south to New York Route 17. Its total relief is almost 420 meters (1380 feet) from 91m (298 feet) at Taylor Road to its summit 507 meters (1664 feet) above sea level. The actual structure extends beyond Monroe where it is broken by a series of faults and continues into New Jersey.

The Schunemunk Conglomerate occurs in an almost continuous outcrop along Schunemunk Mountain it is a coarse reddish brown conglomerate with lenses of interbedded deep reddish-brown sandstone. The sandstone has the same composition as the matrix of the conglomerate. The conglomerate contains rounded pebbles of milky quartz and quartzite. Some of these elongated pebbles are 15 cm along the A-A axis. In addition, sandstone lenses show crossbedding in many instances. The sandstone and matrix materials are composed of 60% angular quartz grains from 0.3 mm to 0.5 mm in diameter and about 30% finer quartz grains. The remainder is argillaceous material with hematite. These pebbles are remnants of an ancient Taconic beach, deposited 350 m.y.a. The relative age of the Schunemunk Conglomerate has not been determined since no fossil evidence has yet been found. (Kothe, 1960)

The mountain overlaps the following U.S.G.S. 7.5 minute topographic maps: Maybrook, Monroe, Popolopen Lake, and Cornwall. A hike along the Jessup Trail leads to the summit of Schunemunk and one of the mountain's most spectacular features: the Megaliths. These huge blocks have been split from the bedrock by faulting and frost heaving, and have moved down slope. Slickensides can be observed on fault surfaces here. Even in summer some of the deeper crevices will be cool and sometimes continue to retain winter ice. Caution should be exercised at this point in the trip since surfaces tend to be slippery. The views (See Fig.4) of the Hudson Highlands, Catskills, and the Hudson Valley are spectacular from the ridge.

Schunemunk is criss-crossed by a network of seven trails: the Jessup, Western Ridge, Barton Swamp, Sweet Clover, Long Path, Forest, and Dark Hollow Trails -- all marked and maintained by the New York-New Jersey Trail Conference.

The trail used for this field trip is the longest, the Jessup Trail. It is almost 7 miles long and runs the entire length of the ridge from Taylor Road to Seven Springs Road (Zimmerman, 1987). The Jessup Trail was selected since it leads quickly to the summit, provides access to the greatest number of features, and provides clear views of the surrounding topography. A secondary reason for the selection of the Jessup Trail is that it is one of the easiest walks to the summit. This is an important consideration since there is a wide diversity in the physical conditioning of the participants who make the trip. People whose medical condition would preclude a strenuous walk should not take this trip.

CAUTIONS

There are no serious physical threats on the mountain from vegetation (no observations of poison ivy have been made). However, there is the threat of venomous snakes. Copperheads and Rattle Snakes are known to inhabit rocky ledges and ridges. It is best to remain on well used trails. The Deer Tick is also known to be found in the region. Precautions followed should be to: a) wear long pants, b) tuck pants legs into socks, c) wear a long sleeved shirt and keep the collar turned down.

The trip is best made during early spring before the nights become warm and the snakes emerge or late in the fall after the trees have lost their leaves and the reptiles have hibernated for the winter.

Acknowledgements

We wish to thank the following people for their contributions: Mortin Strassberg and Sheldon Penn of Suffolk County Community College for reading the manuscript, Dennis Weiss of City College his intrest and helpful suggestions, and Martin Rutstein of S.U.N.Y. New Paltz for the inspiration to begin the project.

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ROAD LOG FOR A GEOLOGICAL CLIMB OF SCHUNEMUNK MOUNTAIN

CUMULATIVE MILAGE	MILES FROM LAST POINT	ROUTE DESCRIPTION
0.0	0.0	Start from Woodbury Commons Shopping Mall, Central Valley, N.Y. at the junction of N.Y. Rte. 32, Rte. 17 and the N.Y. Thruway at Exit 16. Proceed north on N.Y. Rte. 32.
0.8	0.8	Bright Star Diner on the right. This is a good pit stop.
2.2	1.4	Right turn on Park Ave., Highland Mills, N. Y. Notice the use of the Schunemunk Conglomerate as a building material.
2.6	0.4	Stop 1. Pull off on the left into the old Railroad Station.

STOP 1: DEVONIAN FOSSILS

USE CAUTION HERE. THIS IS AN ACTIVE RAIL LINE. Park vehicles west of the tracks and proceed along the tracks about 110 meters (363 feet) to a worked outcrop of the Lower Devonian Esopus Formation (Lowe, 1958). This outcrop contains a wide variety of brachiopod and pelecypod fossils and (rare) trilobite fossils, as well as ripple marks in the tilted bedding planes. Note that Schunemunk Mountain is visible at N 15 E from this point.

2.9	0.3	Return to N.Y. Rte. 32 along Park Ave. turn right and continue north.
3.2	0.3	Note glacial erratics and stratified sand and gravel deposits of glacial origin on the left of N.Y. Rte. 32.
3.6	0.4	Cemetery of the Highlands. Note the use of Schunemunk Conglomerate to build cemetery walls.

5.3	1.7	N.Y.S. Thruway underpass.
5.5	0.2	STOP 2. Pull off on the left to a small unsurfaced road and walk north to the abandoned sand and gravel quarry.

STOP 2: KAME TERRACE

Proceed north along the unsurfaced quarry road to a large outcrop of Precambrian schist and gneiss exposed by a quarrying operation. Glacial striae, grooves and chatter marks are common on the glacially polished bedrock. A fine view of Schunemunk Mountain can be observed to the southeast from the top of the exposure. This site is an abandoned sand and gravel quarry. The glacial deposits appear to have been formed as melt waters flowed from the edge of the ice and deposited their sedimentary load at the base of the Precambrian outcrop (Sorrentino, 1979). Much of the original formation has been removed by the quarrying operation. A variety of cobbles, pebbles and sand has been deposited against the bedrock of the Hudson Highlands. Return to the vehicles and continue north on N.Y. Rte. 32.

6.0	0.5	Notice the view of Schunemunk on the left.
7.7	1.7	Star Expansion Co. on the left.
8.0	0.3	Turn left at Black Rock Fish and Game Club.
8.1	0.1	Left turn past Club House.
8.3	0.2	N.Y.S. Thruway overpass.
8.4	0.1	Parking lot on the right of Taylor Road.
8.8	0.4	Walk north on Taylor Road to the beginning of Jessup Trail.

From this point: Follow the well-marked (yellow markers) Jessup Trail to the Megaliths.

STOP 3: GROUND WATER

Follow the faint foot path through the cornfield. Observe the hillside to the south (left) of the dirt road for traces of surface water. Proceed about 160 to 170 meters (approx. 500 ft.) along the trail to a large oak tree bordering the woodland on the north. At this point the land to the south slopes upward and a small spring, which flows throughout the year will be observed emerging from the plowed field. About 100 meters (300 feet) further along the trail note a concentration of erratic boulders along the north side of the path. The boulders are part of a ground moraine, a deposit left by glaciers melting in place. Take note of the size and shape of the boulders.

STOP 4: RAPIDS OF BABY BROOK

Baby Brook is a youthful stream and is actively downcutting its bed. Several important observations and definitions can be made at this stop: earth flows, gravity falls, soilification, "V" shaped valleys, mass wasting, and hillside creep. Observe the large boulders in the stream bed and the curved boles of the trees along the sides of the stream.

STOP 5: WATERFALL OF BABY BROOK

Cross the railroad track and continue along the trail following the yellow markers. Walking time expended about 40 to 45 minutes. You will hear Baby Brook far below in a steep valley it has cut into the softer sandstone (Bellvale) below the resistant Schunemunk Conglomerate capstone.

Note the nature of the bedrock that is visible on the trail surface. This is part of the Hamilton Group which consists of Devonian sandstones. At about 50 minutes walking time a small side trail to the west will appear. This trail leads to the Baby Brook Falls. Here one can observe the Schunemunk Conglomerate as bedrock for the first time. The conglomerate acts as a resistant cap-rock which produced Baby Brook Falls. Beyond this point note the preponderance of Schunemunk Conglomerate (with its dark reddish brown matrix and milky quartz pebbles). Continue along the Jessup Trail (yellow markers).

STOP 6: APPROACHING THE SUMMIT OF SCHUNEMUNK

This site is said to be the site of an Algonquin Indian encampment. The connection between geology, topography, ecology, and the Algonquin's adaptation to these natural features can be explored. The participants should be encouraged to find the connection between these factors by placing themselves in their position and exploring their needs for food, shelter, wood, game, water, and protection from their enemies. Return to the Jessup Trail (yellow markers) and continue toward the summit. Along the way the many glacial features of the mountain should be noted: glacial polish, faceted pebbles of the Schunemunk conglomerate, chatter marks, plucking, and "white" carbonate erratics which have been transported from the north.

STOP 7: THE SUMMIT

From this elevation much of the topography of the region can be observed: the Catskills, the Shawangunks, and the Hudson Highlands. A question poses itself at this point: Why have Schunemunk and Woodcock Mountain, part of a nappe, (a remnant of a sheet like, allochtonous rock unit that has moved in a horizontal surface) remained? Continue along the trail to the Megaliths.

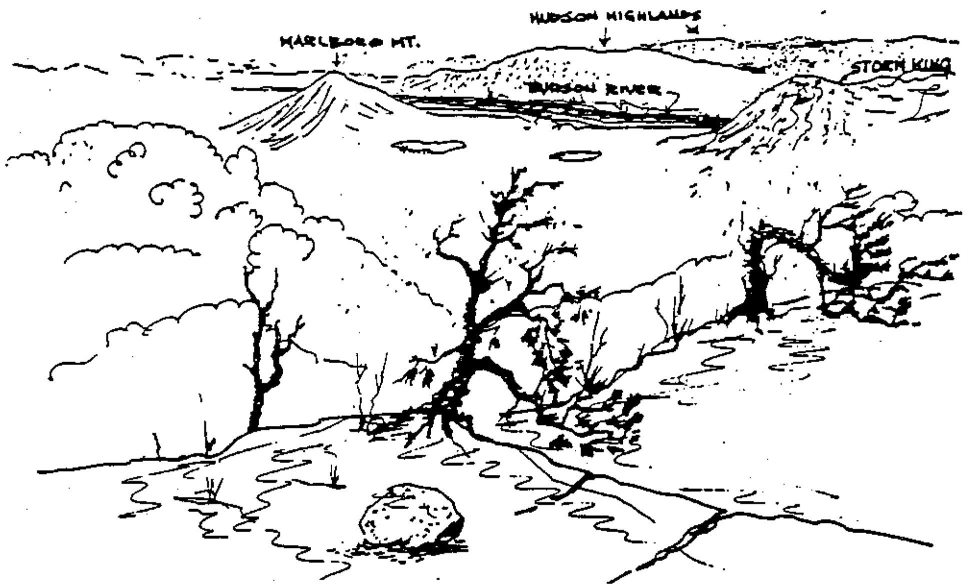

Fig. 4. View to the north from the summit of Schunemunk.

STOP 8: THE MEGALITHS

The Megaliths are the result of faulting. These large blocks, originally fractured by faulting, have tumbled together to form "caves" as a result of glacial plucking and/or mass wasting. Snow can sometimes be found well into spring in these "caves". At this location many features of faulting and jointing can be found. Joints, joint pairs, faults, slickensides (show direction of movement), frost heaving, and weathering can be identified.

The return to the cars can be made by several different routes. The most direct return is along the Jessup Trail. However, if arrangements have been made in advance, a return can be made to Route 32 via Long Path (turquoise markers), Dark Hollow Trail (white on black markers), or Sweet Clover Trail (white markers). The return trip can be dangereous because the temptation to travel quickly (gravity is on our side) can cause falls, etc.

THE WATCHUNG BASALTS OF NORTHERN NEW JERSEY

JOHN H. PUFFER
Geology Department
Rutgers University
Newark, New Jersey 07102

Introduction

This field trip through the northern end of the Watchung Mountains of New Jersey will include six good exposures of quartz tholeiitic basalt. The Watchung basalt flows of the Mesozoic Newark Basin are exposed as three northeast/southwest trending ridges. The three basalt units dip to the west at about 15 degrees and are known as the Orange Mountain, Preakness, and Hook Mountain Basalts (Fig. 1).

The lower contact of the first flow unit of the Orange Mountain Basalt with the underlying Passaic Formation will be seen at Stop 2. The three flow units of the Orange Mountain Basalt are locally separated from each other by thin layers of sediment and comprise an aggregate thickness averaging 183 m (Faust, 1975). The upper flow is characteristically pillowed and amygdaloidal, whereas the lower two flows typically display well-developed columnar joints (lower colonnade and entablature).

Each of the three flow units of the Preakness Basalt are geochemically distinct, and each will be visited during the fieldtrip. The massive, very coarse grained lower flow of the Preakness will be seen at Stop 3, the second flow together with a thin layer of sediment separating it from the first flow will be seen at Stop 4, and the second and third flows separated by another thin layer of sediment will be seen at Stop 5. The aggregate thickness of the three Preakness flows averages 215 m (Olsen, 1980).

The lower contact of the Hook Mountain Basalt with the underlying Towaco Formation will be seen at Stop 6. The Hook Mountain Basalt extruded about 550,000 years after the extrusion of the Orange Mountain Basalt (Olsen and Fedosh, 1988) and consists of at least two flows with an aggregate thickness of 91 m (Faust, 1975).

The hydrous mineral assemblages that accumulated in the vesicles and vugs between pillows of Watchung Basalt, presumably mixtures of carbonates, clays and alteration products, have responded to low temperature burial metamorphic effects. The resulting zeolite facies assemblage includes some very well developed crystal aggregates of stilbite, heulandite, chabazite, and datolite typically precipitated with calcite, quartz, prehnite, and sulfides (principally chalcopyrite and chalcocite).

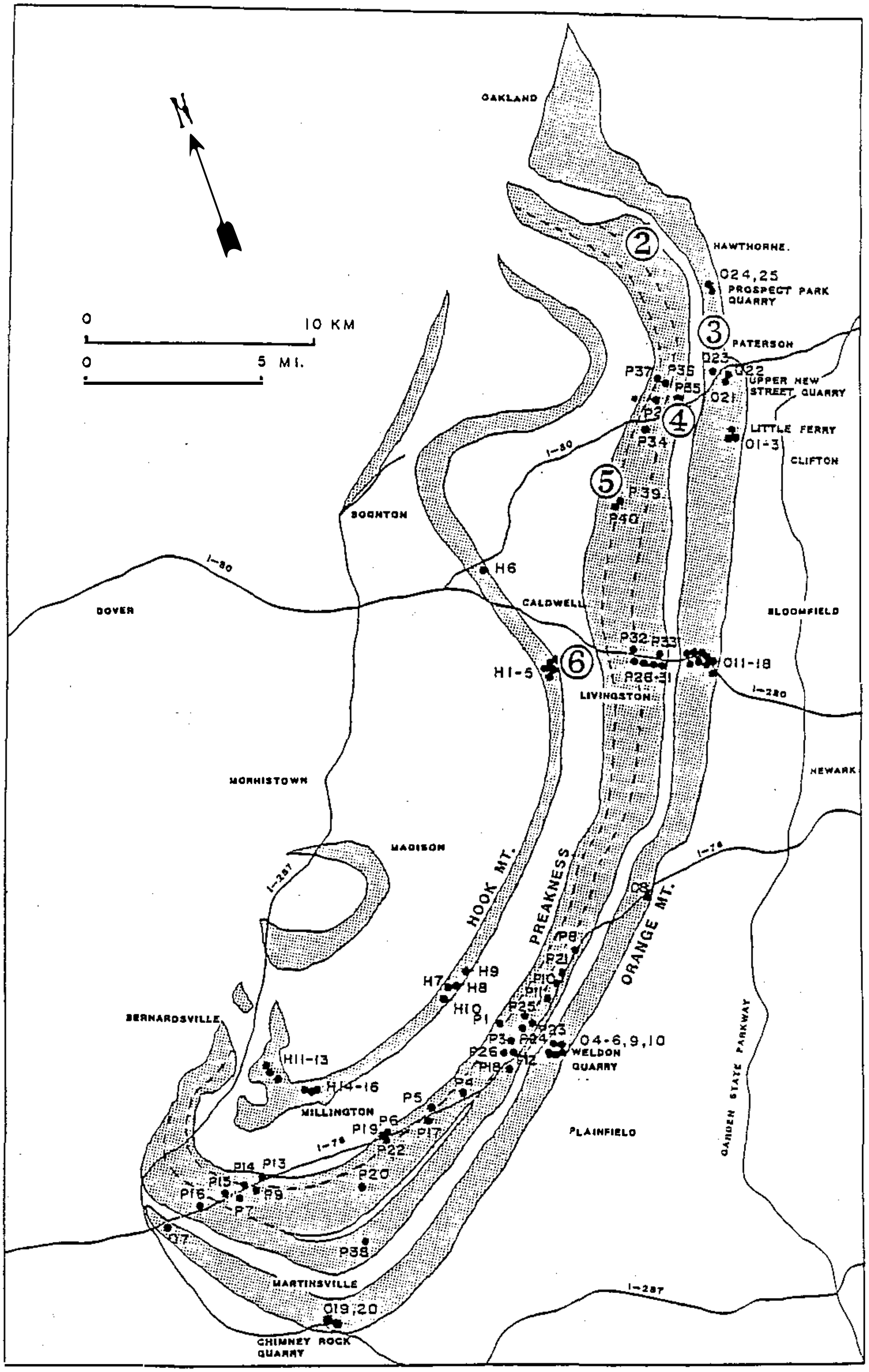

Figure 1. Map of Watchung Basalts with locations of field trip stops, samples chosen for chemical analysis (see Table 1 and Fig. 2), and approximate contacts (dashed lines) of the three Preakness Basalt flows.

The distribution of these secondary minerals has been described by Laskowich and Puffer (1986) but unfortunately the best zeolite and copper collecting localities are found in the active trap-rock quarries of New Jersey, and entry permission for NYSGA was denied at each quarry. Copper sulfides are particularly abundant in the amygdules and sediments at the base of and between the flow units of the Orange Mountain Basalt.

Discussions held during the field trip will be directed toward reinterpretations of the petrogenesis of the Watchung Basalts that are currently being developed as new data is made available.

Petrography

Orange Mountain Basalt. The Orange Mountain Basalt is a quartz normative tholeiite composed of plagioclase and augite with minor orthopyroxene and altered olivine in a glassy mesostasis containing quench dendrites of Fe-Ti oxides. Augite phenocrysts, glomeroporphyritic aggregates of augite, orthopyroxene, altered olivine, and a few plagioclase phenocrysts are characteristic of the basalt. Typical modes average 35 percent plagioclase (An65), 35 percent pyroxene (augite (Wo34En55Fs10), pigeonite, and minor hypersthene), 28 percent glassy mesostasis, and 3 percent opaque Fe-Ti oxides. Accessory and trace minerals include apatite, biotite, alkali feldspar, and pyrite.

Preakness Basalt. The very coarse-grained appearance of the interior of the first or lowermost of the Preakness, resembling a diabase, may be related to the unusual flow thickness that may have included intrusive pulses similar to those proposed by Philpotts and Burkett (1988). Typical samples from the base of the first flow consist of about 50 percent pyroxene and 43 percent plagioclase as an intergranular mixture with about 3 percent plagioclase phenocrysts and 5 percent dark, fine-grained glassy mesostasis enriched in quench oxides. Typical unaltered coarse grained samples from the interior consist of an intergranular mixture of about 45 percent pyroxene, 50 percent plagioclase, 3 percent opaque oxides, with only about 2 percent brown glass. The plagioclase composition of the least altered samples averages about An57 on the basis of 40 microprobe analyses, with a range from An50 to An60, excluding secondary albite determinations. On the basis of 13 microprobe analyses the augite averages (Wo35En45Fs20). On the basis of 20 microprobe analyses, the magnetite composition averages Usp60, and on the basis of 12 microprobe analyses, the ilmenite is Hem8.

Most samples from the middle flow are medium-grained, intergranular mixtures of 35 percent pyroxene (both pigeonite and augite), 59 percent plagioclase, 4 percent opaque oxides, and 2 percent brown glass. Some samples are aphyric, but large plagioclase phenocrysts make up as much as 10 percent of other middle flow samples.

Table I
Chemical Composition of Watchung Basalts

(typical samples)

	O12 Orange Mt. first flow	O6 Orange Mt. second flow	O9 Orange Mt. third flow	P5b Preakness first flow	P18 Preakness second flow	P6 Preakness third flow	H1 Hook Mt. first flow	H9 Hook Mt. second flow
SiO_2	51.44	51.88	51.31	51.65	51.20	50.83	49.89	49.51
TiO_2	1.19	1.09	1.12	1.05	1.21	0.83	1.44	1.40
Al_2O_3	14.75	14.70	14.31	14.19	15.30	15.65	13.26	13.94
FeO	10.54	10.40	10.60	11.41	12.45	10.15	14.89	14.59
MnO	0.16	0.15	0.17	0.19	0.21	0.17	0.19	0.21
MgO	7.77	7.32	7.61	6.48	5.41	7.60	5.65	5.58
CaO	10.61	9.75	10.15	9.48	9.55	10.31	9.88	10.23
Na_2O	2.35	2.40	2.98	3.05	3.21	2.49	2.53	2.23
K_2O	0.32	0.47	0.52	0.63	0.59	0.36	0.32	0.44
P_2O_5	0.11	0.12	0.12	0.12	0.12	0.08	0.14	0.17
H_2O^+	0.42	0.69	0.91	0.76	0.48	0.61	0.69	1.03
Total	99.66	98.97	99.60	99.01	99.73	99.08	98.88	99.33
Ba	162	165	153	142	135	88	110	119
Co	37	39	39	38	39	38	43	49
Cr	374	330	435	76	53	251	122	72
Cu	120	205	137	78	80	70	167	183
Ni	84	77	69	38	25	55	50	55
Sr	206	191	194	136	151	141	118	128
V	250	250	257	315	340	248	324	351
Zn	72	82	82	84	95	72	115	129
Zr	78	100	97	88	91	56	98	77
La	10.1	10.0	-	9	11	7.4	8.1	-
Ce	22.5	22.0	-	19	26	15	19	-
Nd	13.5	13.0	-	14	16	9	12	-
Sm	3.8	4.0	-	2.8	3.0	2.5	2.9	-
Eu	1.11	1.00	-	1.02	1.25	0.90	1.00	-
Gd	3.6	3.5	-	3.25	3.8	4.0	4.0	-
Dy	3.8	3.9	-	5.0	5.0	4.2	5.5	-
Er	2.3	2.3	-	3.5	3.8	2.5	4.0	-
Yb	2.3	2.35	-	3.1	3.2	2.2	3.1	-

The uppermost flow of the Preakness is exposed best along Interstate 78 (Fig. 1); elsewhere exposures typically are altered. The somewhat vuggy nature of the flow seems to have accelerated alteration. The flow is medium grained and slightly porphyritic, consisting of glomerorphyritic clots of pyroxene (largely augite) and plagioclase in a dark mesostasis enriched in quench oxides. The pyroxene (Wo24En56Fs20) and plagioclase (An73) content of the rock is approximately equal. Large phenocrysts of plagioclase and pyroxene are not uncommon.

Hook Mountain Basalt. The Hook Mountain Basalt consists of at least two amygdaloidal and deeply altered flows. The basalt is composed of plagioclase, clinopyroxene, and Fe-Ti oxides in a fine-grained to glassy and typically vesicular mesostasis. Phenocrysts of plagioclase and pyroxene are common. The plagioclase composition of samples taken at the base of the lower Hook Mountain flow averages An68 and the augite composition from the same samples averages Wo31En51Fs18.

Geochemistry

Orange Mountain Basalt.The Orange Mountain Basalt fits into the HTQ type of ENA tholeiites as proposed by Weigand and Ragland (1970). The chemistry of the basalt is rather uniform throughout (Table 1) and virtually is equivalent in all respects to samples of Palisades chill analysed by Walker (1969), Shirley (1987), and Husch (1988). The REE content of the Orange Mountain Basalt plots close to and parallel with the REE distribution pattern of the lower chill margin of the Palisades sill.

The chemistry, mineralogy, and texture of the Orange Mountain Basalt is equivalent to the first Early Jurassic basalts in the other basins of the Newark Supergroup, such as the Talcott Basalt of the Hartford Basin, Connecticut, and the Mount Zion Church Basalt of the Culpeper Basin, Virginia (Puffer and others, 1981; Puffer, 1984).

Preakness Basalt. The chemical composition of the first flow of the Preakness Basalt (Table 1) is distinctly enriched in most incompatable elements compaired to the Orange Mountain Basalt and is virtually identical to the Holyoke Basalt of Connecticut. The middle flow is even more highly enriched with a TiO_2 content averaging 1.21 (Table 1). Accumulation of plagioclase phenocrysts in some samples of the middle flow is responsible for a slightly positive Eu anomaly. These plagioclase-rich samples plot at the most chemically evolved end of the Preakness field on a TiO_2 versus MgO diagram (Fig. 2).

The chemical composition of the uppermost of the three Preakness flows (Table 1) is characterized by a consistently low TiO_2 content ranging from 0.7 to 0.9 and averages 0.8 percent. The basalt qualifies in all respects as a typical ENA-LTQ basalt as defined by Ragland and Whittington (1983). The occurrence of

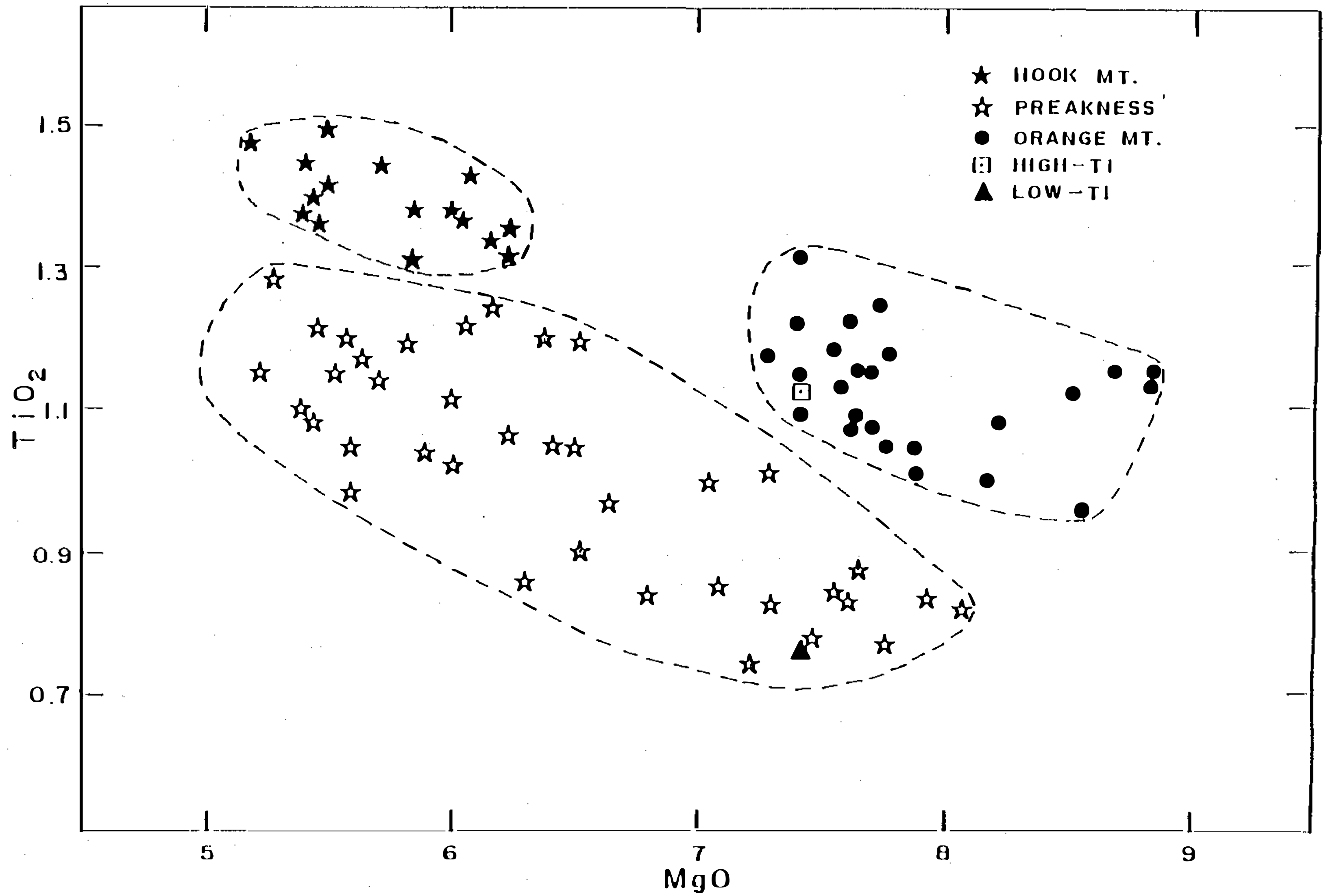

Figure 2. TiO_2 versus MgO diagram showing distribution of Watchung Basalt samples located on Figure 1. High-Ti (HTQ) and Low-Ti (LTQ) values after Weigand and Ragland (1970).

LTQ basalt within the Newark Basin is not surprising in light of its occurrence in the Gettysburg Basin to the south (Smith and others, 1975) and in the Hartford Basin to the north (Philpotts and Martello, 1986).

The overlapping compositions of the three Preakness Basalt flows in Fig. 2 supports interpretations that all three Preakness flows are genetically related to each other

The REE content of the third flow of the Preakness Basalt is within the "low-TI" or LTQ range of Ragland and others (1971), but the first and second flows contain distinctly higher REE concentrations. The distribution patterns (Fig.3), however, are close to each other and reasonably parallel furthur supporting a genetic relationship perhaps controlled by fractionation. The slightly positive Eu anomaly displayed by the second flow (Fig. 3) is consistent with plagioclase enrichment that approximately is balanced by the relatively low plagioclase content of the first Preakness flow.

Hook Mountain Basalt. The SiO_2, Na_2O, Cr, Ni Rb, and Sr contents of the Hook Mountain Basalt (Table 1) are intermediate between those of the Orange Mountain and Preakness Basalts. The REE distribution pattern of the Hook Mountain Basalt (Fig. 3) plots close to that of the Orange Mountain Basalt and is within the "high-Ti" or HTQ range of Ragland and others (1971). The Hook Mountain Basalt, however, contains less light REEs than the Orange Mountain Basalt despite its more highly evolved major element concentrations (including iron and titanium). There is also a distinct cross over in the distribution pattern resulting in a higher heavy REE content for the Hook Mountain Basalt than the Orange Mountain samples.

The composition of the Hook Mountain Basalt particularly the REE content resembles that of the first flow of the Preakness more closely than any of the other Watchung Basalts although a genetic relationship is not clear.

Petrogenesis

Orange Mountain Basalt. Despite the fact that the Orange Mountain Basalt is a quartz normative tholeiite, an interesting although somewhat radical case can be made in support of its assignment as a primary magma. The chief obstacle to a primary magma assignment is the low Mg' value of the Orange Mountain Basalt (0.51) which is considerably lower than the 0.68 to 0.75 range for primary magmas proposed by O'Hara and others (1975) and Frey and others (1978). As they noted, if a magma is primary the forsterite content of its liquidus olivine should be the same as that of olivine in the magma's residual source. Only melts with an Mg' range of 0.68 to 0.75 would be in equilibrium with the Fo90 olivine typically found in harzburgite and lherzolite inclusions interpreted as mantle xenoliths. The currently popular view that only picritic basalts or komatiites qualify as

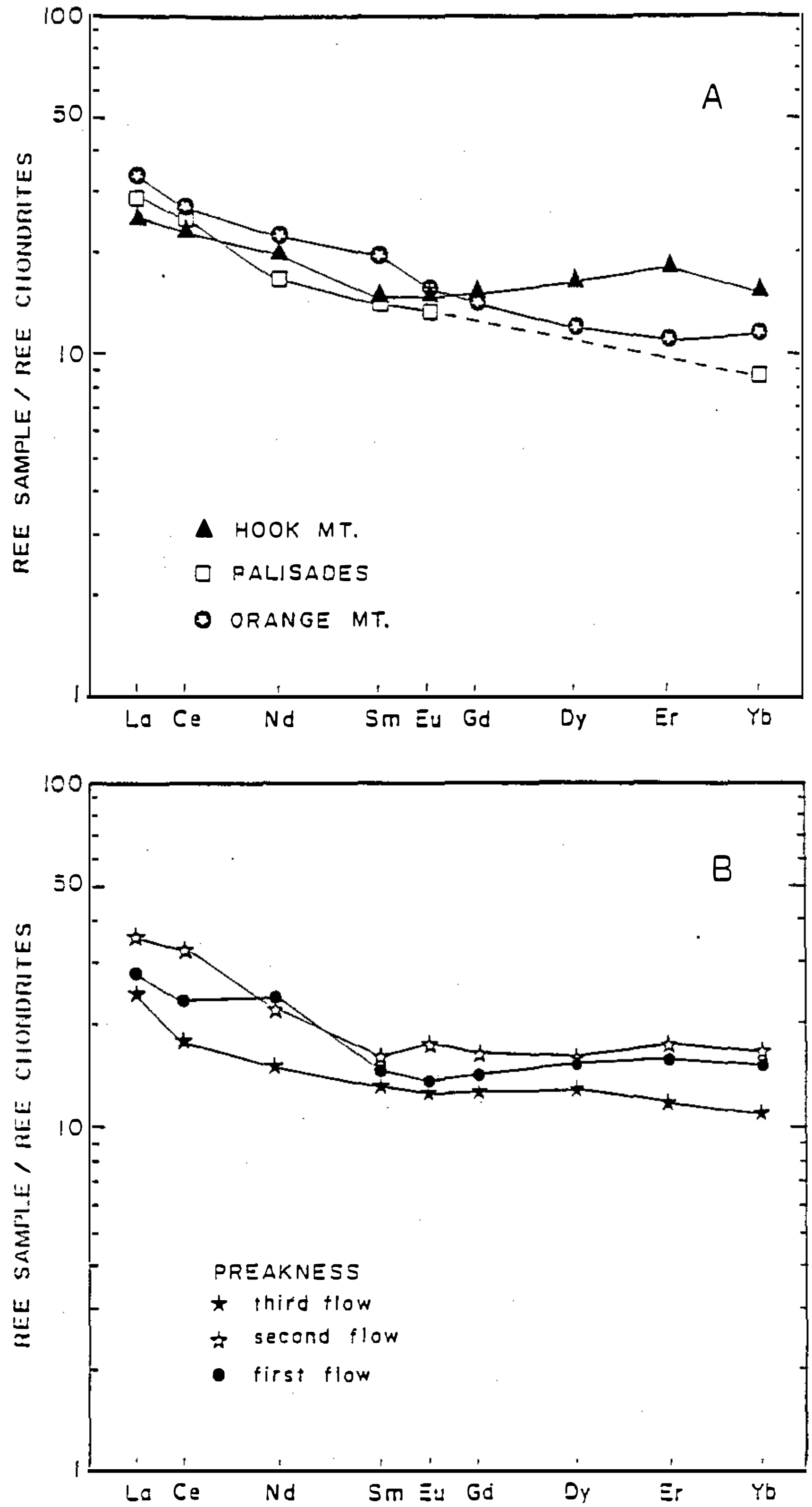

Figure 3. Chondrite normalized (after Masuda and others, 1973) REE distributions of five typical Watchung Basalt samples and the Palisades sill lower chill (sample 80B1 of Shirley (1987)).

primary basaltic magmas supports the interpretation that the ENA quartz tholeiites are derived from the more primative ENA olivine normative magmas (perhaps through fractionation of HLO magma; Fig. 4 after Whittington, 1988).

Mounting evidence, however, suggests that fractionation mechanisms are not useful in genetically relating the various ENA magma types to each other. It now appears that on geochemical grounds none of the Watchung Basalts, for example, can be derived from any other Watchung Basalt or Newark Basin intrusive rocks through fractionation. There is typically some geochemical difference beyond the range of plausable fractionation mechanisms. It might be argued, therefore, that if fractionation has failed as a mechanism capable of genetically relating some closely spaced quartz tholeiites it is even less likely that it will be successful in genetically relating highly diverse magma types such as quartz tholeiite and picrite.

Although a primary magma proposal is clearly highly speculative it is safe to say that not enough is known about all the factors that effect upper mantle partial melting processes to totally reject it. Particularly little is known about the volatile content of upper mantle, subcontinental rocks that may have been the source of the Watchung Basalts. There is also the possibility that the iron content of the source may have been much higher than typically suspected.

Evidence supporting the interpretation of the Orange Mountain Basalt as a primary magma include:

1) <u>Great Magma Volume</u>. Carmichael and others (1974) suggest that great magma volume is one of two principal characteristics of primary magmas. They have observed that "Continental tholeiitic flood basalts and related diabases more than any other class of volcanic rocks satisfy the two criteria postulated for magmas generated directly by fusion - great volume of and compositional homogeneity within each magma province... To derive the magma from picritic basalt of deep-seated origin by low pressure fractionation... requires that again and again each successive draught of magma must rid itself cleanly, while still largely liquid, of the same fraction of crystalline olivine along some identical source of ascent. This seems highly improbable." The same logic also argues against derivatiuon through mantle or crustal assimilation or through filter pressing.

The volume of magma represented by the Orange Mountain Basalt is very greatly extended if each of the ENA high-Ti tholeiites are included together. It was shown that the chemical composition of the Talcott Basalt of the Hartford Basin of Connecticut, and the Mount Zion Church Basalt of the Culpeper Basin, Virginia (Puffer, 1984) are both chemically equilivent to the Orange Mountain Basalt and in each case represent the first of several Mesozoic ENA extrusive events.

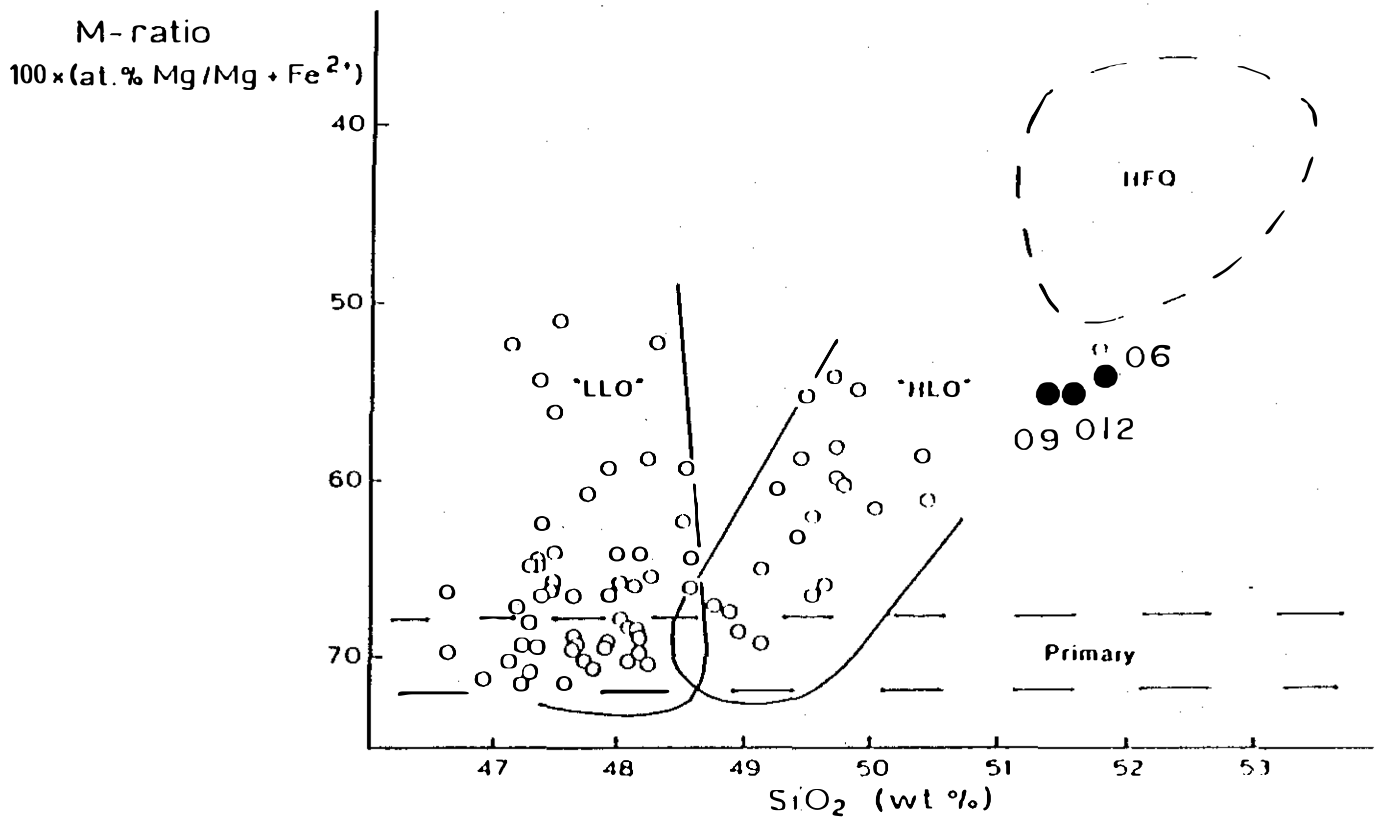

Figure 4. Weight percent SiO_2 vs M-ratio (100 x atomic % Mg/Mg + Fe) for aphyric North Carolina diabase shown with fields for quartz diabase (HFQ) from North Carolina, after Whittington (1988). The vertical trend of no silica enrichment and the inclined trend of silica enrichment correspond to Ragland and Whittington's (1983) LLO and HLO groups, respectively. Orange Mt. samples 06, 09, and 012 (Table 1) are also plotted.

2) Chemical Homogeneity. The second characteristic of primary magmas suggested by Carmichel and others (1974), "compositional homogeneity" is also a characteristic of Orange Mountain Basalt and HTQ basaltic rock in general. Despite the huge volume of magma represented the the entire HTQ population, occurrences of ENA basaltic rock of a composition intermediate between the Orange Mountain and picritic rocks are rare if they exist at all. The entire HTQ population is instead tightly clustered around the original average determined by Weigand and Ragland (1970).

3) Close Resemblance to Other Early Jurassic Basalts of Possible Primary Nature. When the chemistry of the Orange Mountain Basalt is compared with other basalts on a world-wide basis a remarkable coincidence becomes apparent. Of all the known world-wide basalts located beyond the Newark Supergroup the basalts that most closely resemble the Orange Mountain are also Early Jurassic. The close resemblance with the High-Atlas Basalts of Morocco was first reported by Manspeizer and Puffer (1974) and an equally close resemblance with the Lesotho Basalt of South Africa has been recognized (Fig. 5). If the early Jurassic basin containing the High Atlas Basalt was contiguous with the Newark Basin the "great volume and compositional homogeneity" arguments are strengthened and together constitute a kind of igneous super province implying that magma was generated by an early Jurassic tectonic event of major proportions. Melting of the Lesotho may have been triggered by the same event or perhaps by a highly similar event. It has been proposed by Marsh (1987) that the Lesotho magma was a primary type derived from old and enriched subcontinental lithosphere. Sr and Nd isotopic data have been interpreted by Bristow and others (1981) as indicating subcontinental enrichment events beneath southern Africa between approximately 1 and 2 b.y.

4) Extended Insulation of an Old Enriched Subcontinental Lithospheric Source. The isotopic data of Pegram (1983) indicate that the ENA quartz tholeiites are enriched to a degree approximately equivalent to that of the Lesotho. His initial 87Sr/86Sr value of the ENA quartz tholeiites range from 0.7054 to 0.7072 and overlap the 0.7059 average of Lesotho data reported by Compston and others (1968). Pegram (1983) suggests that the ENA quartz tholeiites were derived from a source isotopically distinct from the MORB and reflect a subcontinental mantle with a complex history involving the long term (approximately 1 b.y.) enrichment in Rb/Sr, Nd/Sm, and U/Pb. Pegram (1983) also concludes that his data are inconsistent with crustal contamination.

Elevated temperatures within the upper mantle caused by the prolonged insulation effects of a thick continental cover may have resulted in melting in a shallow, low pressure regime. Elthon and Scarfe (1984) have shown that advanced melting of the mantle at pressures less than 10 kbar would result in primary melts that become increasingly enriched in silica as the pressure decreases. Widespread melting under an attenuated Pangea

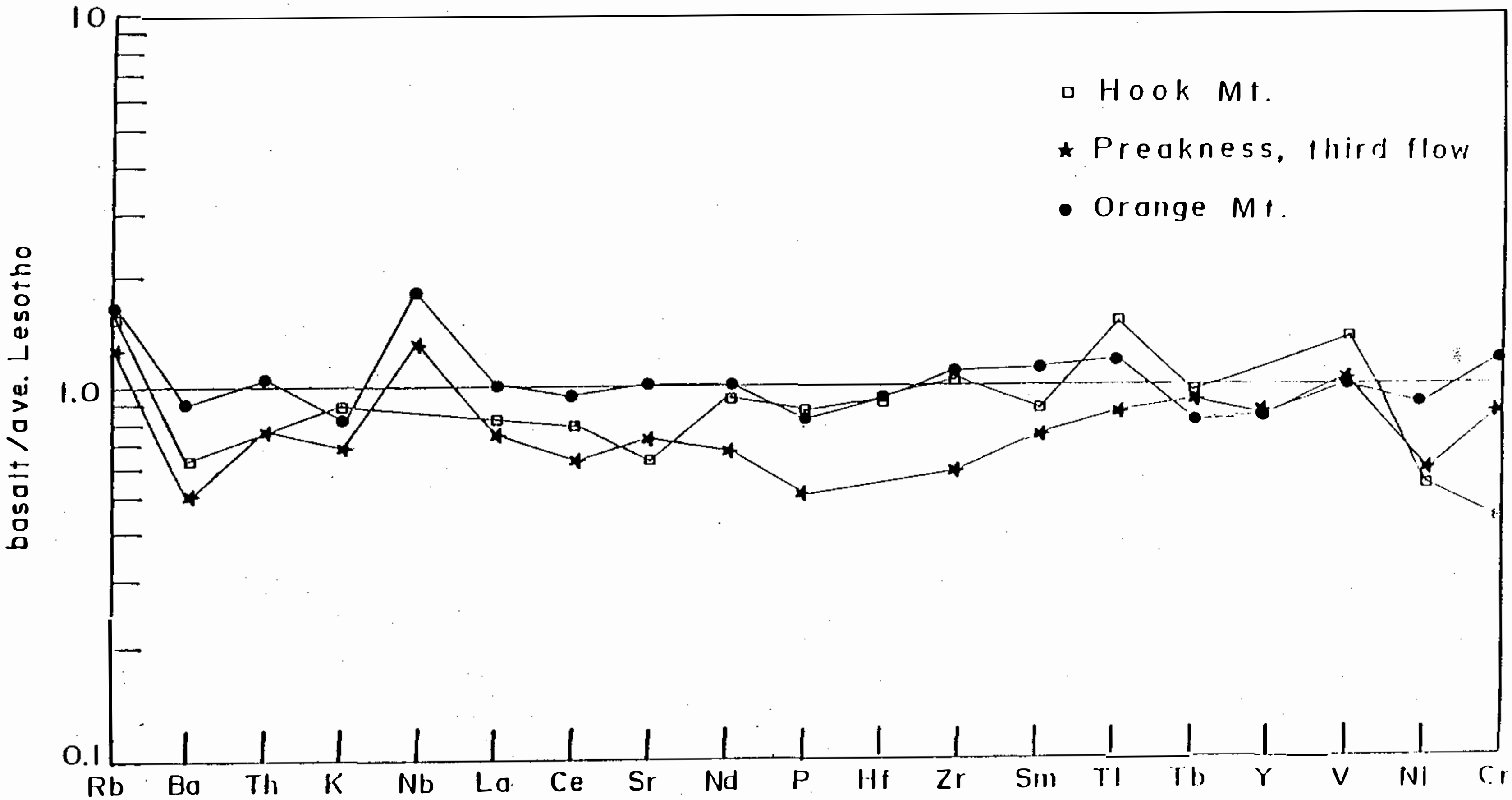

Figure 5. Lesotho normalized diagram comparing average (Karoo) Lesotho basalt (data of Duncan, 1987) with Watchung Basalts, samples 012, P6, and H1 (Table 1)

triggered by Mesozoic rifting, therefore, may have occurred in an enriched, low-pressure environment unlike the environment of oceanic or other basaltic sources.

5) Monovalent Cation Enrichment. Kushiro (1975) and Mysen (1977) have shown that if lherzolite or harzburgite is enriched in monovalent cations such as H_2O, Na_2O, and K_2O, the liquidus boundary shifts towards silica, but if the source rocks are rich in 4 or 5 valent cations such as TiO_2, CO_2, or P_2O_5, the liquidus boundary shifts away from silica (Fig. 6). Compaired with most basalts on a world-wide basis the Orange Mountain contains a distinct high ratio of $K_2O + Na_2O/TiO_2 + P_2O_5$, clearly much higher than typical oceanic basalt although it is difficult to speculate about water contents. Kushiro (1975) suggests that partial melting of peridotite enriched in water and other monovalent elements would form silica rich magmas such as quartz tholeiite and suggests that such elements may be contained in minerals such as phlogopite with stability fields that extend into the high pressure conditions of the upper mantle. Kaersutitic amphibole (Basu and Murthy, 1977) and beta-Mg_2SiO_4 (Smyth, 1987) have also been proposed as potential sources of water and monovalent cations in the mantle.

Preakness Basalt. The Preakness Basalt is a somewhat less viable candidate for a primary magma designation although it contains even less TiO_2 and P_2O_5 than the Orange Mountain Basalt. Preakness Basalt compositions, unlike Orange Mountain, are spread out over a wide range (Fig. 2) generating a trend that may be derived from a more primative source (Fig. 4). The compositional range, however, overlaps or closely resembles the Sander Basalt of the Culpeper Basin, Virginia (Puffer, 1984), and the Holyoke Basalt of the Hartford Basin (Puffer and others 1981). The composition of the first and second of the three Preakness flows is generally more chemically evolved than the Orange Mountain Basalt to an extent approximately equilivent to the degree the interior of the Palisades Sill differs from the chill-margin of the Palisades Sill. These relationahips were interpreted by Puffer and Lechler (1980) and Walker (1969) as due to fractionation processes. The uppermost of the three Preakness flows, however, has been determined to be a Low-Ti ENA type (Puffer, 1989) unrelated to the underlying high-Ti Orange Mountain Basalt. Analyses of samples from each of the three Preakness flows combine to generate a chemically diverse range that when plotted on MgO variation diagrams (such as Fig. 2) is distinctly depleted in several incompatable elements (Ti, Zr, and light REE) compared to the Orange Mountain Basalt (Table 1).

Although the three Preakness flows are apparently not related to the Orange Mountain Basalt they may be related to each other through fractionation processes that took place in a shallow sill. Philpotts and Asher (1989) present convincing evidence that decompression melting during ascent through the lithosphere has prevented fractionation from effecting some of the igneous rocks of the Hartford Basin, Connecticut. The effects of

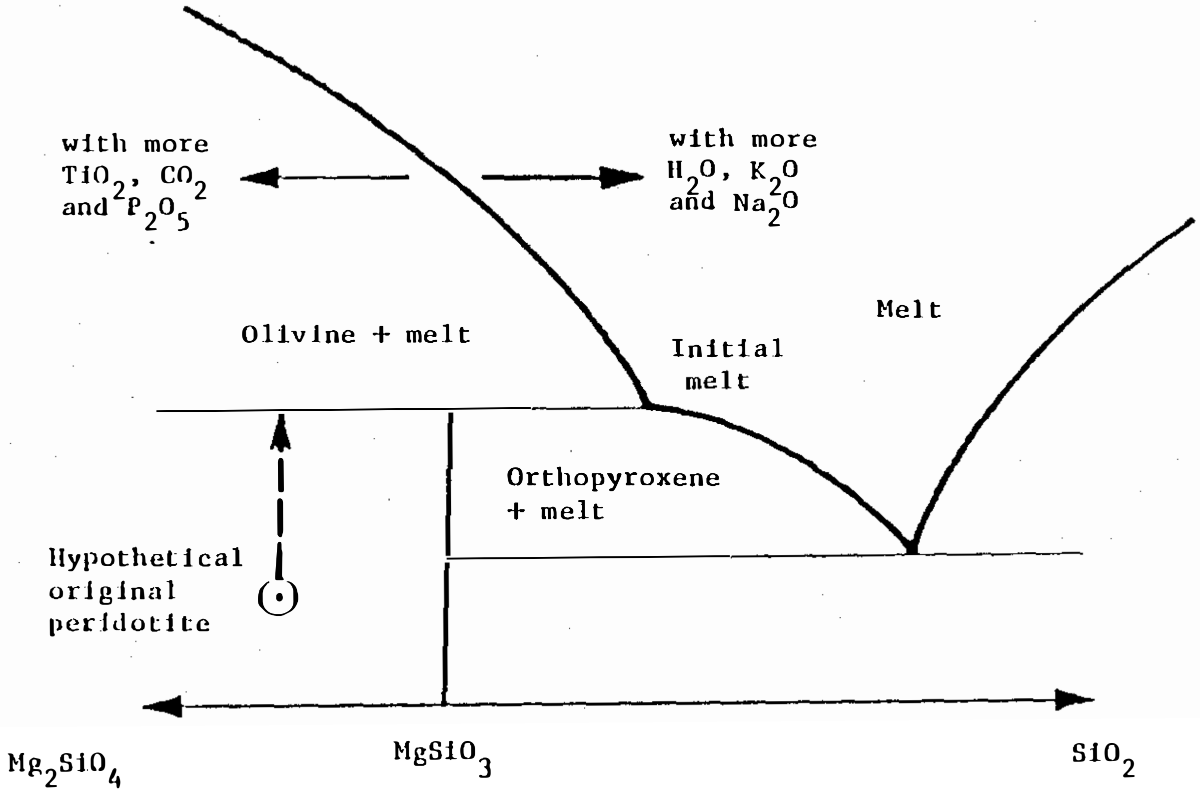

Figure 6. Shift in liquidus boundary with monovalent cations versus 4- or 5-valent cations, after Hyndman (1985). A high ratio of K_2O + Na_2O/TrO_2 + P_2O_5 is a characteristic of Orange Mt. Basalt.

decompression melting, however, are presumably limited to assent. Once shallow sills such as the Palisades are reached, shallow, in-situ pyroxene controlled fractionation may have occurred.

Hook Mountain Basalt. Fractionation processes are also incapable of relating the Hook Mountain Basalt to any of the underlying Watchung Basalts. Puffer and Lechler (1980) have shown that the Hook Mountain Basalt could not have fractionated out of Preakness magma largely because of higher Cr/Mg and Ni/Mg ratios than the Preakness. More recently it has been shown by Gottfried and Tollo (1989) that the Hook Mountain Basalt could not have fractionated out of Orange Mountain Basalt largely because of a lack of expected enrichment in Zr and Nb.

The Hook Mountain Basalt is a relatively minor magma type compaired to the HTQ or LTQ magmas. The Hook Mountain is chemically equivalent to the Hampden Basalt of Connecticut (Puffer and others, 1981), but in both cases the relatively minor magma volume and the upper stratigraphic position makes it more vulernable to a wide range of processes, particularly assimilation, that are capable of affecting magma compositions.

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ROAD LOG FOR EXPOSURES OF WATCHUNG BASALT, NOTHERN NEW JERSEY

CUMULATIVE MILEAGE	MILES FROM LAST POINT	ROUTE DESCRIPTION
0	0	From the intersection of Interstate 84 and Rt. 17 near the Orange County College Campus proceed southeast on Rt. 17 which becomes Rt. 6 near Goshen. Continue

		on Rt. 6 to Exit 131.
18.0	18.0	Proceed south on Rt. 17/ 32 one traffic light. Make left turn onto Rt 6 and continue toward Bear Mountain.
25.0	7.0	Make second right on circle onto the Palisades Interstate Parkway (south). Continue on parkway to Exit 13.
34.0	9.0	Turn right (west) onto US 202.
36.5	2.5	Park along US 202 opposite the basalt outcrop on the left.

STOP 1. LADENTOWN BASALT

The closely-spaced curved cooling columns displayed by the Ladentown basalt along US 202 contrast with the thick massive columns typical of the Palisades sill that underlied the basalt and was probably its source of magma. The basalt exposed at this locality is fine grained, contains large plagioclase phenocrysts, and is slightly vesicular. Chemical analysis of the Ladentown Basalt (Puffer and others, 1982) compare closely with the fractionated interior (Walker's (1969) second magma pulse of the Palisades sill. The Ladentown Basalt, appears to have extruded onto sediments of the Passaic Formation, perhaps forced to the surface by the injection of a second magma pulse within the Palisades sill. This occurred before the Feltville Formation was deposited, and before the Preakness Basalt was extruded. A physical connection between the Ladentown flow and the western end of the Palisades sill at Mount Ivy is indicated by magnetic and gravity data (Kostsomitis, 1980; Kodama, 1983). However, this connection is not seen at the surface.

Good exposures of coarse boulder conglomerate are located another 0.4 mi. south along US 202. The coarse clast size of the sediment coincides with their close proximity to the western border fault in the valley running parallel to US 202. Note the distinct change in topography on the opposite side of the border fault where Precambrian gneisses are the dominant lithology.

36.8	0.3	Continue south on US 202, observe the coarse-rained fanglomerate exposed on the left and the Precambrian Highlands of the Reading Prong exposed west of the Ramapo fault valley

		on the right.
51.0	14.2	Turn left (southeast onto Rt. 208.
53.8	2.8	Turn right (south) onto Rt. 502 (Ewing Ave.).
55.8	2.0	Turn left onto High Mountain Rd. Bear right at the fork onto Belmont Ave.
57.0	0.2	Park at the base of the jeep trail to the summit of High Mountain.

STOP 2. PREAKNESS BASALT AT HIGH MOUNTAIN, NORTH HALEDON

High Mountain is the highest point (970 ft; 296 m.) in the Preakeness Mountain chain. The upper portion of the hiking or jeep trail to the top of High Mountaom cuts through closely-spaced columnar-jointed basalt typical of the entablature of the Preakness Basalt. At the top of the mountain is a beautiful alpine meadow that on a clear day affords a spectacular view of most of northern New Jersey. The basalt exposed in the meadow is very coarse-grained; it is typical of the interior of the first flow of the Preakness Basalt and resembles diabase.

The view to the southeast includes the New York City skyline in the background behind the ridge formed by the Palisades sill. (A copy of 1981 Paterson 7 1/2-minute Quadrangle is recommended). The city of Paterson is in the foreground where the Passaic River cuts through the First Watchung Mountain ridge formed by the westward dipping Orange Mountain Basalt. The New Street trap-rock quarries and Garret Mountain are clearly visible just north of Paterson. The view directly to the south includes the city of Newark in the distant background on the far side of the First Watchung Mountain, Montclair State College southeast of Paterson, and Paterson State College in the foreground. The view to the southwest includes the Precambrian New Jersey Highlands province west of the Ramapo Fault scarp in the background and the curved "inverted S" shaped Hook Mountain ridge formed by the Hook Mountain Basalt in the foreground.

57.2	0.2	Return to Belmont Ave., turn right.
60.3	3.1	Turn left onto West Broadway.
60.4	0.1	Turn right (west) onto Totowa Ave.
60.7	0.3	Turn left at stadium and park at Passaic Falls.

STOP 3. ORANGE MOUNTAIN BASALT AT PASSAIC FALLS

An excellent exposure of Orange Mountain Basalt is located here in the park above Passaic Falls. Observe the columnar jointing in the basalt and some large convex upward, or half-moon vesicles that typically occur near flow-tops. Carefully climb into the narrow notch in the basalt eroded along the strike-slip fault for a closer inspection of the basalt. Most of the slickensides have been eroded away but a few still remain.

Cross the footbridge, or if blocked, walk down Spruce Street to the statue of Alexander Hamilton. At the exposure along the north edge of the parking lot near the statue of Alexander Hamilton, the lower contact of the Orange Mountain Basalt with the underlying Passaic Formation is seen. Southwest plunging pipe-amygdules and vesicles are exposed in the basalt near the contact. These vesicles occur entirely within the basalt above the basal contact.

From the statue of Alexander Hamilton observe the vertical strike-slip fault planes through the lower flow unit of the Orange Mountain Basalt and the contact between the lower colonnade and the overlying entablature. The "S.U.M." over the door of the historic building near the river at the base of the view area stands for "Society of Useful Manufactures," an organization founded in 1789 to promote local trade.

From the falls, walk south along Spruce Street, cross McBride Avenue and continue one-half block to the Paterson Museum. Some of the best examples of the secondary minerals found in the Paterson area trap-rock quarries are on display at the museum.

63.0	2.3	Continue west on Totowa Ave., turn right onto Green Ave.
63.0	0.0	Turn left onto Claremont Ave.
63.2	0.2	Proceed northwest to a sharp right curve in the road, and park along outcrop.

STOP 4. PREAKNESS PILLOW BASALT, TOTOWA

The lower portion of a Preakness Mountain Basalt flow unit, the second of three, is exposed along the west side of Claremont Avenue. The base of the flow is a subaqueous flow lobe containing ellipsoidal pillows and pahoehoe toes. Secondary mineralization includes calcite, quartz, and minor heulandite in small stretched amygdules. The pillowed base of the flow rests on a thin layer of red silstone exposed (depending on the amount of refuse present) at road level. The flow grades upward into massive columnar basalt. Further north on Claremont Avenue the upper part of the flow, in contrast to the bottom is not pillowed and contains large spherical amygdules mineralized with prehnite and pectolite.

63.4	0.2	Return southeast on Claremont Ave. to Green Ave., and turn right.
63.5	0.1	Turn left on Totowa Road.
63.7	0.2	Turn right (south) onto Union Blvd.
65.5	1.8	Turn right into Walnut St.
65.6	0.1	Turn left onto Montclair Rd.
65.9	0.4	Turn left onto Rt. 23 (Pompton Tpk.).
66.0	0.1	Turn right onto Rt. 527.
67.2	1.2	Turn right onto Greenbrook Rd.
67.7	0.5	Park at intersection of Greenbrook Road with Central Ave. just north of the bridge over Green Brook.

STOP 5. UPPER (LOW-TI) FLOW OF PREAKNESS BASALT, NORTH CALDWELL

The middle and upper flows of the Preakness Basalt are exposed here together with a layer of red siltstone that was deposited between the two flows. The middle flow is well exposed along the banks of Green Pond (sample site P39, Fig. 1) and can be examined by carefully walking down the slope near the bridge. Be careful of the poison ivy. The middle flow is greatly enriched in incompatable elements compaired to the upper high-Ti or HTO flow of the Preakness (Table 1).

The upper flow can be examined along the road-cut just south of the bridge (Sample site P40, Fig 1). It is exposed above the thin red stilstone layer along the road. Good dinosaur footprints were found in this siltstone by Chris Laskovich. Columnar jointing is reasonably well displayed at this stop but both flows are other-wise quite massive with little evidence of secondary mineralization.

69.7	2.0	Proceed south on Central Ave. turn left on Bloomfield Ave.
70.2	0.5	Turn left onto Rt. 527.
72.7	2.5	Turn right (west) onto Interstate 280.
73.9	1.2	Turn left (south) onto Eisenhower Parkway at exit 4A.
74.8	0.9	Turn right into parking lot.

STOP 6. HOOK MOUNTAIN BASALT, ROSELAND

This Hook Mountain exposure displays columnar joints, but a well-defined Tomkeieff (1940) sequence is not apparent. An extensively altered volcano-clastic layer is exposed near the base of the Hook Mountain Basalt. In addition, some bleaching and evidence of low-grade thermal metamorphism is seen at the lower contact.

In contrast to the Orange Mountain (Stop 2) and Preakness (Stop 3) Basalts, the Hook Mountain Basalt is relatively enriched in amygdules and vesicles. Prehnite is abundant at this locality and is easily collected, particularly near the north end of the exposure.

75.7	0.9	Proceed north on Eisenhower Parkway and turn right onto I-280 east.
78.7	3.0	Type section of Preakness Basalt on both sides of highway.
80.2	1.5	Type section of Orange Mountain Basalt on both sides of highway.
83.6	3.4	Exit onto Garden State Parkway north or south.

NOTES

EARTHQUAKE ACTIVITY IN THE GREATER NEW YORK CITY AREA: A FAULTFINDER'S GUIDE

ALAN L. KAFKA
Weston Observatory
Department of Geology and Geophysics
Boston College
Weston, MA 02193

MARGARET A. WINSLOW
Department of Earth and Planetary Sciences
City College of CUNY
New York, NY 10031

NOEL L. BARSTOW
Rondout Associates, Inc.
Stone Ridge, NY 12484

INTRODUCTION: EARTHQUAKES IN THE NORTHEASTERN UNITED STATES

Although the area within a 100 km radius of New York City is clearly not as seismically active as most areas near plate boundaries, this area has had its share of moderate earthquake activity. For most earthquakes in the interior of the North American plate--unlike the situation along plate boundaries--it is not clear whether geologic faults mapped at the surface are the faults along which the earthquakes are occurring. The purpose of this field trip is to examine several geologic features in the greater New York City (NYC) area that might be related to earthquake activity. As we proceed, the following question will be addressed:

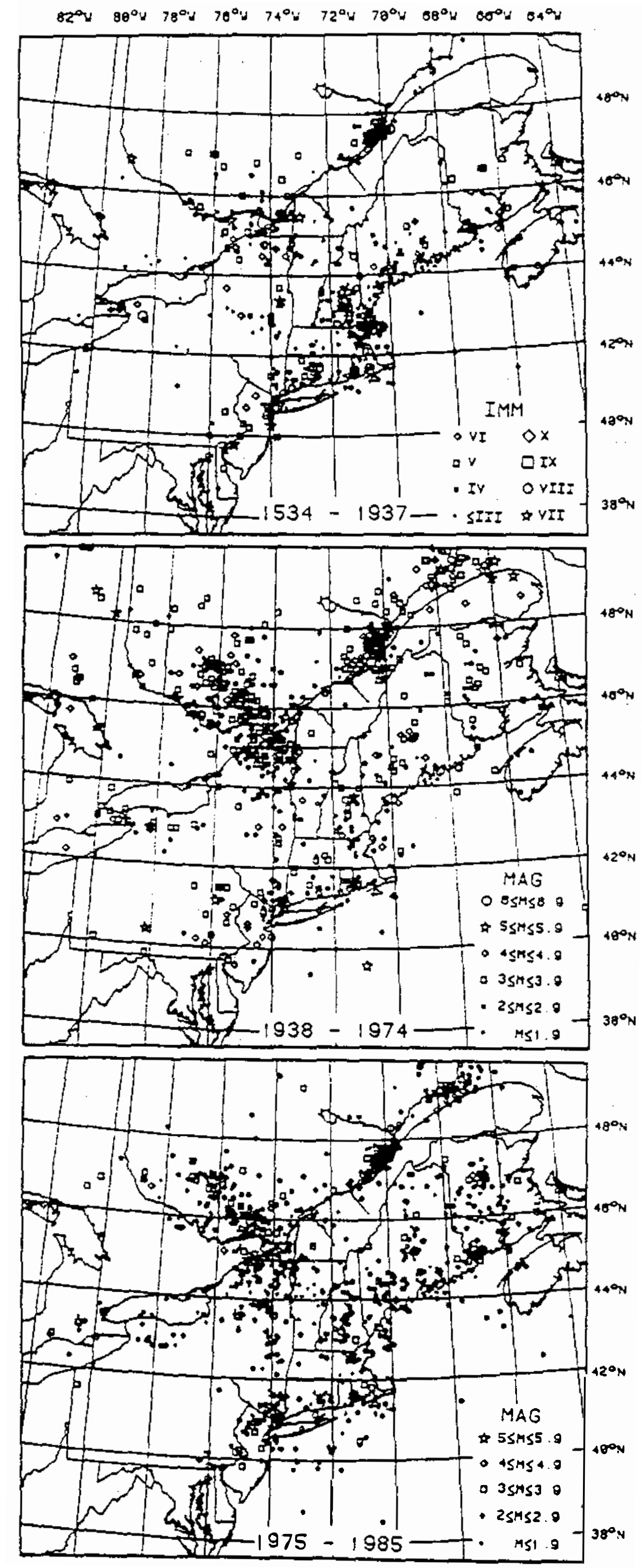
Are seismically active faults in the NYC area delineated by geological features that are observable at the earth's surface?

To understand earthquake phenomena in the NYC area, it is helpful to picture this area in the context of earthquake activity throughout the northeastern United States (NEUS). The NEUS has one of the longest records of reported earthquake activity in North America (Fig. 1). Earthquake activity was noticed in this region by early European settlers, and as the population density grew, an increasing number of earthquakes were reported. These earthquakes were usually minor, but sometimes not-so-minor, and occasionally even caused damage. Instrumental seismic monitoring in the NEUS began in the early 1900's, and routine reporting of instrumentally recorded earthquakes began in 1938 with the initiation of the Northeastern Seismic Association (Linehan and Leet, 1942). The first telemetered regional seismic network was operated in northern New England by Weston Observatory from 1962 to 1968, but it was not until the early 1970's that the present regional networks were established. The number of seismic stations in the NEUS steadily increased between 1970 and 1974. By 1975, a number of institutions operating seismic networks in the region formed a cooperative group known as the Northeastern United States Seismic Network (NEUSSN; Fig. 2). The data recorded by the NEUSSN has enhanced our ability to study the regional seismic activity. Analysis of these data provides insight into the possible causes of the earthquakes.

a

b

c



HISTORICAL

EARLY INSTRUMENTAL

NETWORK

Figure 1.

Seismicity in the northeastern United States and adjacent areas.

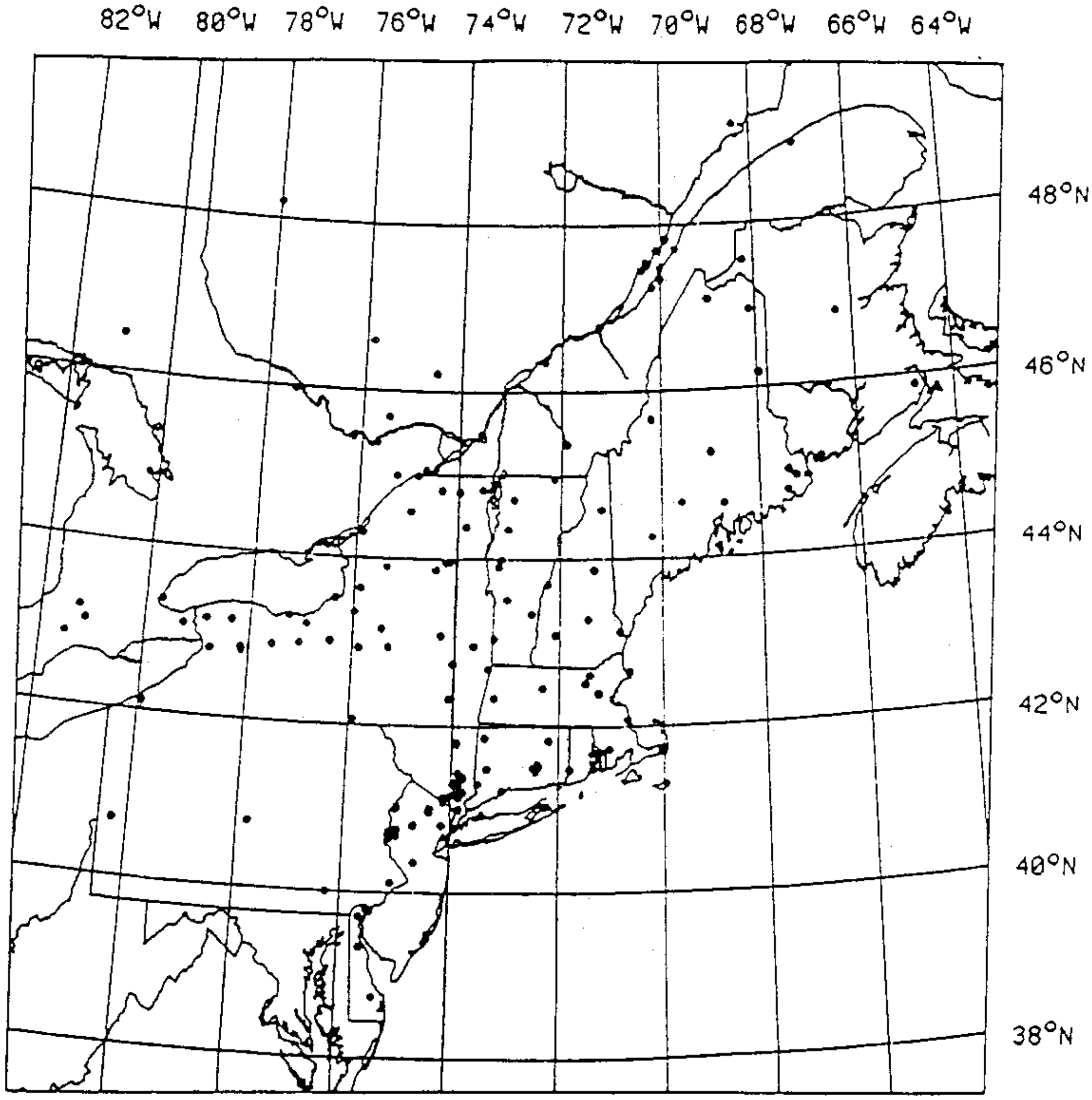


Figure 2. Stations of the Northeastern United States Seismic Network (NEUSSN). These stations are operated by Weston Observatory of Boston College, Lamont-Doherty Geological Observatory of Columbia University, Massachusetts Institute of Technology, Woodward-Clyde Consultants, State University of New York at Stony Brook, Pennsylvania State University, and Delaware Geological Survey.

Since at present the NEUS is not located on or near an active plate boundary, there is no obvious plate tectonic interpretation of why earthquakes occur here. The present-day tectonic setting of the region is a passive continental margin. The nearest plate boundaries are the mid-Atlantic ridge to the east, the subduction/transform system along the western margin of North America to the west, and the northern boundary of the Caribbean plate to the south. All of these plate boundaries are located several thousand kilometers from the NEUS. The last major tectonic activity in this region postdates the Triassic separation of North America from Africa and occurred in the latter part of the Mesozoic. This activity is delineated by igneous rocks found in the Connecticut and Newark Mesozoic basins, in the White Mountain magma series of New England, in the Montregian Hills of southern Quebec, and along the New England seamount chain. In the more distant geologic past, the area experienced at least two continental collision and rifting episodes, of which the Appalachian mountain system is a remnant expression (Bird and Dewey, 1970; Skehan, 1988; Skehan and Rast, 1983). While the present-day seismicity is quite low compared to that of most plate boundaries, its persistence and its potential for producing damaging earthquakes make it the subject of both scientific inquiry and public concern.

The following three fundamental, and as yet unanswered, questions provide the underlying framework for our analysis of NEUS earthquakes:

- What are the forces that cause earthquakes in this area?
- Which geologic or tectonic features are seismically active?
- What is the potential for future large earthquakes?

With these questions in mind, we will summarize the current understanding of earthquake phenomena in the NEUS, with particular emphasis on the NYC area.

THE PROBLEM WITH MAGNITUDES

Before we summarize the seismic activity in the NEUS, it is important to mention that there are some unresolved issues regarding magnitudes of earthquakes in the NEUS. Magnitudes reported in the Bulletins of the NEUSSN are reported as m_N, where "N" refers to the Nuttli (1973) m_{bLg} magnitude scale. The Nuttli magnitude formulas were developed for estimating body wave magnitude (m_b) from 1 Hz Lg waves (a superposition of Love waves and higher-mode Rayleigh waves), which are recorded from more distant earthquakes. However, the seismograms of small earthquakes (m_b < about 4.0) recorded at shorter distances by stations of the NEUSSN are generally dominated by higher-frequency (~5-10 Hz) Lg waves. It is the amplitudes of these higher-frequency waves that are used along with the Nuttli (1973) formulas to estimate magnitudes for reporting earthquakes in the NEUSSN bulletins. The relationship between these reported m_N magnitudes and m_b is not well known.

For earthquakes that are large enough that the recorded seismograms are clipped (i.e. off-scale), but not large enough to be recorded globally, magnitude formulas were developed that use signal duration (sometimes called "coda-length") instead of amplitude as an estimate of the size of the event. The m_C magnitude scale was developed for the NEUS from a study of the relationship between coda-length and m_N (Rosario, 1979; Chaplin et al., 1980). The purpose of using m_C is to provide an estimate of m_N when amplitude measurements are not available, and m_C is reported as the magnitude for some of the events in the NEUSSN bulletins.

The magnitudes shown for events in Fig. 1(c) are those listed in the NEUSSN computer files, and most are listed as m_N or m_C. To compare the seismicity shown in Fig. 1(c) with that in other parts of the world, it is necessary to estimate the relationship between "m_N or m_C" and some more generally used magnitude scale. Fig. 3(a) shows the relationship between "m_N or m_C" (as reported in the NEUSSN Bulletins) and m_{bLg} for 19 events in the NEUS and adjacent Canada. The slope of the regression line in Fig. 3(a) is 1.14, and the average difference between "m_N or m_C" and m_{bLg} is 0.4 magnitude units (with "m_N or m_C" overestimating m_{bLg}). Thus, although the problem with NEUS magnitudes is not yet resolved, the magnitudes shown in Fig. 1(c) should be considered to be approximately correct in a relative sense, but they appear to overestimate m_b by about 0.4 magnitude units.

SPATIAL DISTRIBUTION OF EARTHQUAKE ACTIVITY IN THE NORTHEASTERN UNITED STATES

In the next three sections, we summarize the current understanding of earthquake activity in the NEUS to provide a framework for our discussion of earthquake activity in the NYC area. Locations of the older epicenters in Fig. 1(a) are based upon felt reports, while the locations of earthquakes since the 1930's have been determined primarily from instrumental data. The earthquakes shown in Fig. 1 indicate that much of the area is seismically active but that the distribution of earthquake activity is not spatially uniform. All of the zones that have been seismically active throughout the historical record (including the NYC area) are also seen to be active in the instrumental records. The largest earthquakes in the NEUS have all occurred in one or another of these active areas. There is a general correlation between the spatial distribution of epicenters determined from the network data and that of the historical seismicity. The general features of the pattern of seismicity have been fairly stable since about the mid-1500's. Thus, it appears that whatever the process is that causes earthquakes in this region, that process has been spatially stable for the past several hundred years.

The largest earthquakes recorded in the NEUS are: the 1727 and 1755 earthquakes located near Cape Ann, Massachusetts; the 1904 earthquake in eastern Maine; the 1929 earthquake in western New York; the 1940 earthquakes in central New Hampshire; the 1944 earthquake at the northern New York/Canadian border; the 1983 earthquake in north central New York, and (most relevant to this field trip) the 1884 earthquake near NYC. All of these earthquakes caused at least minor damage, and all of them (except perhaps the 1884 NYC event) probably exceeded m_b 5.0 (Street and Turcotte, 1977; Street and Lacroix, 1979). The largest earthquake known to have occurred in the NEUS is the 1755 Cape Ann earthquake. The maximum intensity for that event has been estimated to be VIII on the Modified Mercalli (MM) scale (Weston Geophysical, Inc., 1977). The Cape Ann earthquake caused damage in Massachusetts, New Hampshire and Maine and was felt as far away as Washington, DC. Street and Lacroix (1979) estimated the magnitude of the Cape Ann earthquake to be approximately m_b 6.0.

Rockwood (1885) investigated the 1884 NYC earthquake, and he reported fallen bricks and cracked plaster from eastern Pennsylvania to central Connecticut. The a

b

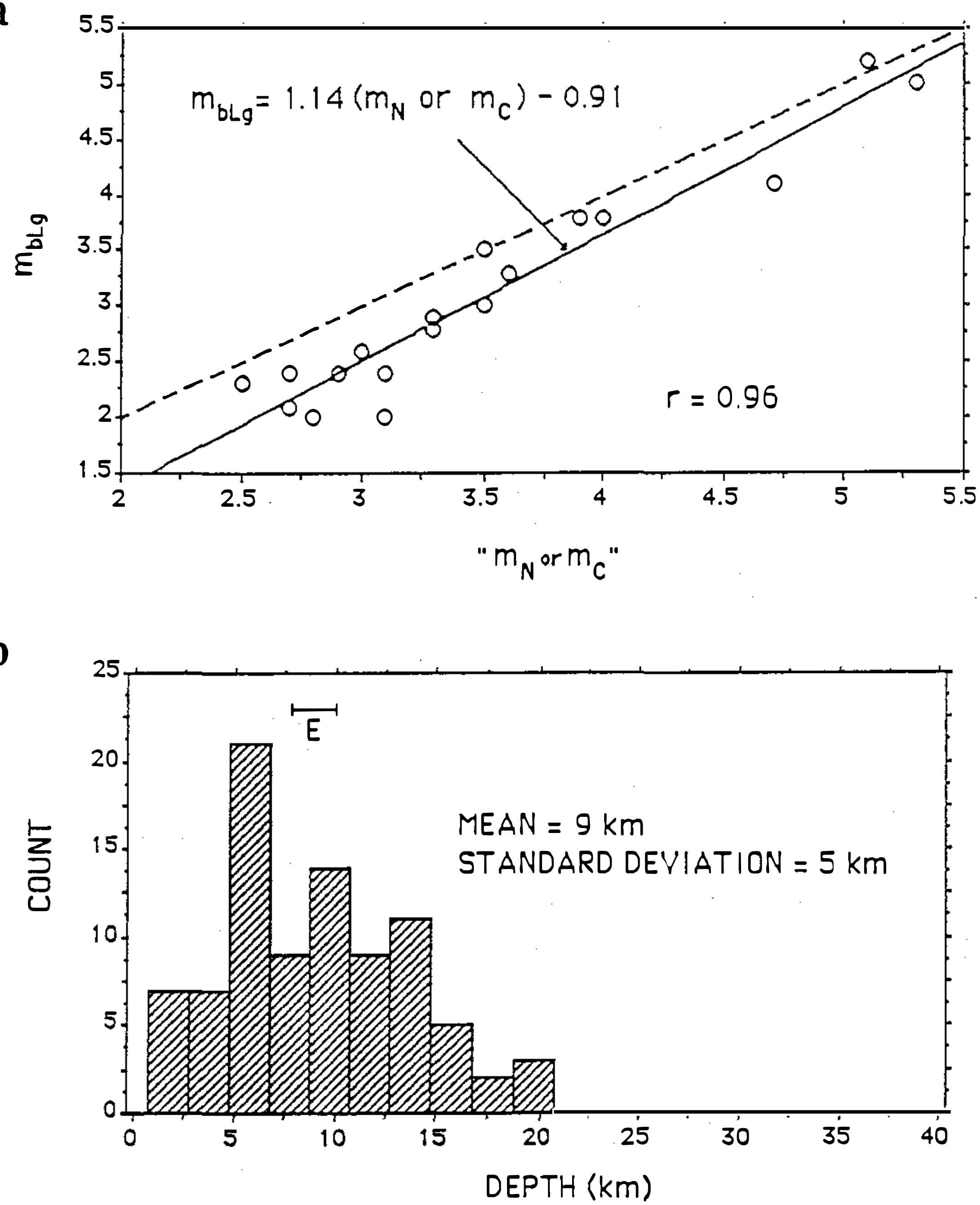


Figure 3. (a) Relationship between "m_N or m_C" and m_{bLg} (from Kafka, 1988). The value of "m_N or m_C" is the magnitude that was sorted from the NEUSSN data (as described in the text). The dashed line is the locus of points for which "m_N or m_C" equals m_{bLg}. (b) Histogram of depths of earthquakes recorded by the NEUSSN in the northeastern United States from1975 to 1983. Bar labelled E indicates the range of depths that Ebel et al. (1986) found from teleseismic waveform modelling of some of the largest earthquakes in North America.

maximum intensity reported by Rockwood was at two sites on western Long Island (Jamaica, New York and Amityville, NY). The felt area of the 1884 earthquake, measured from the map published by Rockwood, was about 270,000 km^2. Smith (1966) reported a maximum MM intensity of VII for the 1884 NYC earthquake, which suggests a magnitude on the order of m_b 5.0 for that event.

The occurrence of the 1755 Cape Ann earthquake and the 1884 NYC earthquake provides minimum values for the size of earthquakes that can be expected to occur in the NEUS in general (m_b of about 6.0) and in the NYC area in particular (m_b of about 5.0). In order to make more specific predictions of the maximum magnitude earthquakes that can occur at a particular location in the NEUS, it is necessary to know which faults (or other features) are active, as well as the record of seismic activity for those features. Unfortunately, however, there are few (if any) surface-mapped faults in the NEUS that have been confirmed to be active. Thus, at the present time, we must be satisfied with only general conclusions regarding where and when future large earthquakes will occur in the NEUS.

DEPTH OF EARTHQUAKE ACTIVITY

Although the relatively low density of regional seismic stations limits the resolution of focal depth for many of the events, the station density is sufficient to conclude that earthquakes in the area generally occur in the upper half of the crust. The deepest hypocenters located by the NEUSSN stations generally occur at about 20 km, and the shallowest earthquakes are about 1 km deep (Pomeroy et al., 1976; Ebel et al., 1982; Mrotek et al., 1988). Events deeper than 20 km tend to occur more frequently in adjacent Canada rather in the NEUS. Earthquakes as deep as 33 km have been reported in the Charlevoix seismic zone--a cluster of activity centered at about 47.5° N, 70.3° W (Fig. 1).

To summarize the depth distribution of earthquakes recorded in the NEUS, Ebel and Kafka (1989) analyzed the depths reported for a sample of earthquakes from the NEUSSN bulletins. This sample consisted of all events recorded between 1975 and 1983. From those events they extracted all earthquakes for which there was (a) at least one station located within a distance of twice the depth; and (b) a total of at least four stations recording the event. A histogram of those data is shown in Fig. 3(b). For that set of data, the mean depth is 9 km and the standard deviation is 5 km. Because of the requirement of at least one nearby station, the depths shown in the histogram are probably biased toward deeper events (which would be more likely to be recorded by a station within twice the depth). From Fig. 3(b) it appears that the earthquake activity tends to cluster in a zone between about 5 and 15 km beneath the earth's surface (although events shallower than 5 km could be more numerous than Fig. 3(b) implies since they were systematically excluded from these statistics).

An independent measure of the depth range where the larger earthquakes tend to occur in this region can be obtained from the results of waveform modelling of larger earthquakes that were recorded teleseismically. Also shown in Fig. 3(b) is the range of depths that Ebel et al. (1986) found from teleseismic waveform modelling of the largest earthquakes of this century in northeastern North America. Those events were found to occur at depths ranging from 8 to 10 km. In addition, Nabelek (1984) and Basham and Kind (1986) modelled teleseismic waveforms from the January 9, 1982 earthquake in New Brunswick, Canada (m_b=5.7). They concluded that the depth of that event was 6 km. Similar depth ranges were observed from aftershock surveys of larger earthquakes. For example, Seeber et al. (1984) found that aftershocks of the 1983 Goodnow, NY earthquake (m_b=5.1) were confined to a depth range of 7 to 8.5 km.

All of these depth estimates suggest that NEUS earthquakes occur in the upper 20 km of the crust. This depth range for NEUS earthquakes is consistent with the idea that only the upper half of the crust behaves in a brittle fashion (Kirby, 1980).

STRESS FIELD AND MAPPED FAULTS

The most commonly accepted explanation for the cause of earthquakes in the NEUS is that ancient zones of weakness are being reactivated in the present-day stress field (Sykes, 1978). In this model, preexisting faults and zones of weakness from earlier orogenic episodes persist in the intraplate crust, and, by way of analogy with plate boundary seismicity, earthquakes occur when the present-day stress is released along these zones of weakness. Much of the recent research on the cause of NEUS earthquakes has, therefore, involved attempts to identify preexisting faults and determine whether they are favorably oriented so as to be reactivated by the present-day stress field. While this concept of reactivation of old zones of weakness is commonly assumed, the identification of individual active geologic features has proven to be quite difficult. It is not at all clear whether faults mapped at the earth's surface in the NEUS are the same faults along which the earthquakes are occurring.

State of Stress

The intraplate stress field in the NEUS is generally assumed to be the result of a combination of forces generated by plate tectonic processes. Two large-scale sources of stress that have been considered in a number of studies are asthenospheric drag and ridge push (Richardson et al., 1979). Another source of stress that could be significant in plate interiors is asthenospheric counterflow (Chase, 1979; Hager and O'Connell, 1979). Within the context of the ancient zones of weakness hypothesis, it is important to characterize accurately the observed modern stress field to determine whether preexisting faults are favorably oriented to be reactivated.

In an effort to summarize the present state of knowledge regarding the stress field in the NEUS, Ebel and Kafka (1989) reviewed data from focal mechanism studies in the NEUS and adjacent parts of Canada (Figs. 4 and 5). Focal mechanisms can be described by specifying the directions of the so-called *pressure axis* (or P-axis) located in the center of the dilatational quadrant and *tension axis* (or T-axis) located in the center of the compressional quadrant. Fig. 4 shows the azimuths of the horizontal component of P-axes from numerous studies superimposed on a map of the region. Fig. 5 shows equal area stereographic plots of the P-axes and the T-axes from the various studies analyzed by Ebel and Kafka (1989). The average P-axis and T-axis for this data set was determined by vectorially averaging all of the axes which intercept a given focal hemisphere. The P-axes tend to cluster toward the east and west directions. The average P-axis was found to have an azimuth of 266° and a plunge of 1°. The average T-axis was found to have an azimuth of 105° and a plunge of 88°.

These focal mechanism results are consistent with a regional stress field for the entire NEUS and adjacent Canada in which the maximum compressive stress is essentially horizontal and trends approximately east. The minimium stress, as

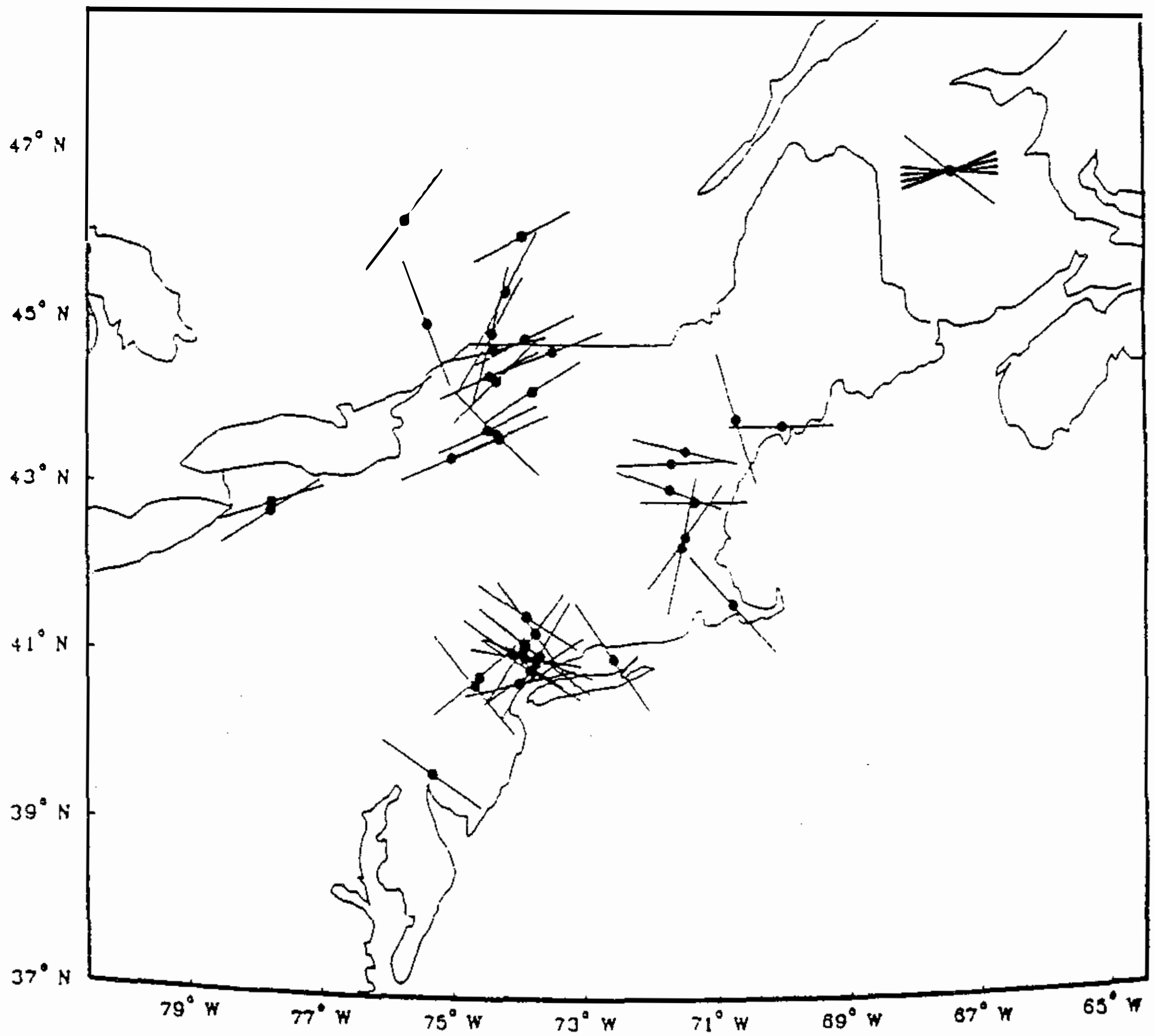
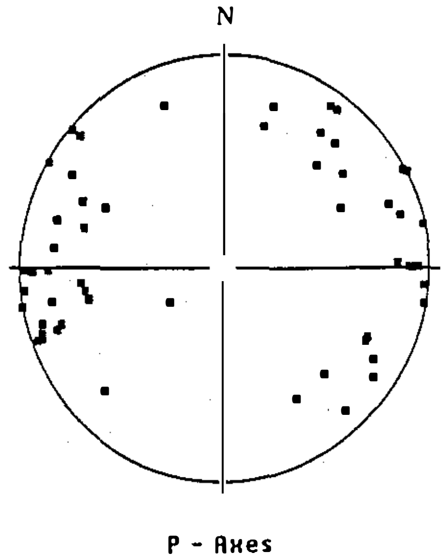


Figure 4. Map showing azimuths of P-axes from focal mechanisms of earthquakes in the northeastern United States. The data shown here were taken from references listed in caption of Figure 5.



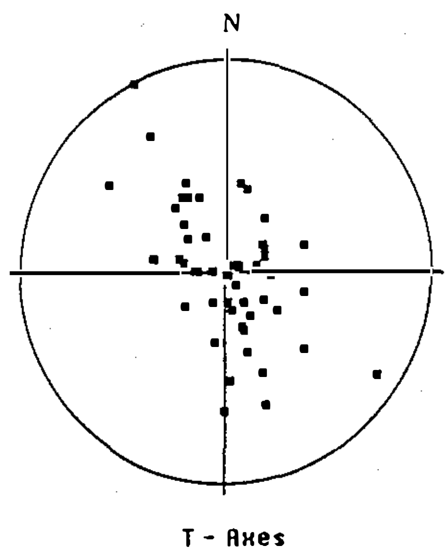


Figure 5. Lower hemisphere stereographic plots of P-axes and T-axes of earthquakes in the northeastern United States. The data shown here were taken from the following studies: Sbar et al. (1970, 1972, 1975), Herrmann (1979), Horner et al. (1978), Graham and Chiburis (1980), Yang and Aggarwal (1981), Pulli and Toksoz (1981), Kafka (1982), Kafka et al. (1982), Schlesinger-Miller et al. (1983), Wetmiller et al. (1984), Wahlstrom (1985), Quittmeyer et al. (1985), and Filipkowski (1986). The P-axes tend to cluster in the east and west directions, and the T-axes tend to cluster in the vertical direction. The average P-axis for this data set has an azimuth of 266° and a plunge of 1°. The average T-axis for this data set has an azimuth of 105° and a plunge of 88°.

inferred from the average T-axis, is nearly vertical. There is, however, no reason to assume that the stress field doesn't vary (temporally and/or spatially) across the region studied.

Seismicity and Mapped Faults

If the stress field in the upper crust can be accurately characterized, and if reactivation of ancient faults is the cause of NEUS earthquakes, then seismically active features could be delineated by identifying the faults and determining their orientation relative to the stress field. The situation is apparently more complex, however, since for most areas in the NEUS, the seismicity does not reveal any obvious alignments of epicenters along mapped faults or other known structural boundaries. It is thus quite difficult to prove the existence of any seismically active faults on a regional basis. Detailed studies of seismicity on a more local scale, especially results of aftershock monitoring of larger events, also yield ambiguous results with regard to the identification of active features. In this section, we discuss several examples of such detailed studies, and we conclude that an unequivocal correlation between mapped faults and earthquake activity is, at best, the exception rather than the rule.

Two relatively large earthquakes that occurred since the installation of the NEUSSN stations were the m_b 4.4 Gaza, NH earthquake of 1982 and the m_b 4.1 Dixfield, ME earthquake of 1983. In the case of these two earthquakes, as well as other relatively large earthquakes discussed in this section, m_b is given here rather than m_N because the events were large enough that m_b magnitudes were reported in the Regional Catalogue of Earthquakes (published by the International Seismological Centre). Both the Gaza, NH earthquake and the Dixfield, ME earthquake occurred in locations where field mapping has failed to show surface faults in the vicinity of the epicenters (Ebel and McCaffrey, 1984; Brown and Ebel, 1985).

Microearthquake swarms near Moodus, CT have occurred near a locality where a fault has been inferred but never confirmed (Ebel, 1985). Aftershocks of the m_b 3.8 earthquake near Bath, ME in 1979 occurred on or very near the Cape Elizabeth fault (Ebel, 1983). This may be evidence that, at least in some cases, earthquakes in the NEUS are associated with mapped faults. The mere spatial association of earthquakes with faults that are mapped near the surface, however, does not necessarily imply that the earthquakes are occurring on those faults. The faults could, for example, be the nucleation points for the earthquakes, yet the earthquake movements themselves may not have actually occurred on the preexisting fault surfaces. Furthermore, the larger earthquakes appear to be occurring at depths exceeding 5 km, and the orientation of faults mapped at the surface may be quite different from the orienation of faults at depth.

Another event of interest is the m_b 5.1 Goodnow, NY earthquake that occurred in the central Adirondack Mountains in 1983. That event was large enough to be felt throughout the NEUS and in southeastern Canada. Aftershock surveys provided a well constrained location for the source region. Seeber et al. (1984) argued that the aftershocks of the Goodnow earthquake occurred on a steeply-dipping fault striking N $15\pm10°$W beneath a NNW-striking surface lineament. The focal mechanism for this earthquake supports the trend seen in the aftershocks, but no surface fault breakage was found from the event. The aftershocks were confined to a depth range of 7 to 8.5 km, and there is no detailed information regarding the orientation of faults at that depth beneath the epicenter.

In the case of an earthquake near Lancaster, PA (m_{bLg} 4.1, April 23, 1984), the locations of the main event, aftershocks, and previous events in that region correlate well with mapped geologic lineaments. The mapped geologic lineaments also correlate with one fault plane of the focal mechanisms found for the 1984 events (Stockar and Alexander, 1986; Armbruster and Seeber, 1987). While such studies have indicated the possible orientations and positions of active faults, there are few, if any, cases in the NEUS where surface-mapped faults have been confirmed to be active.

EARTHQUAKE ACTIVITY IN THE NEW YORK CITY AREA

One of the proposed candidates for present-day seismic activity resulting from reactivation of ancient zones of weakness is found in the NYC area: the Ramapo Fault system in northern New Jersey and southeastern New York [Fig. 6(b)]. The Ramapo Fault system forms the northwestern margin of the Triassic-Jurassic Newark Basin. This fault system is about 120 km long, and it appears to have been active at various times throughout geologic history including the Precambrian, Paleozoic, Triassic, and Jurassic (Ratcliffe, 1971). Thus, the Ramapo Fault system could possibly be an example of a major throughgoing fault system that is being reactivated by the present-day stress field in the NEUS.

Aggarwal and Sykes (1978) concluded from analyzing locations, depths and focal mechanisms of earthquakes in the greater NYC area that seismic activity in that region is concentrated along several NE trending faults of which the Ramapo Fault appears to be the most active. More recent studies of earthquakes in the NYC area, however, suggest a more complicated relationship between earthquakes and geological features in this region. For example, Seborowski et al. (1982) argued that focal mechanisms of three earthquakes that occurred on or very near the Ramapo Fault have fault planes that are transverse to the mapped trace of the fault. Moreover, they argued that the microearthquake seismicity near the northern end of the Ramapo Fault trends NW, transverse to the trend of major geologic structures mapped at the surface. Also, Kafka et al. (1985) argued that earthquakes at least as large as those recorded near the Ramapo Fault have occurred as far as 50 km from that fault in a variety of geologic structures that surround the Newark Basin.

Based on the distribution of earthquakes recorded by the NEUSSN between 1975 and 1986, the Ramapo Fault does not appear to be any more active than numerous other locations in the NYC area [Fig. 6(b)]. There appear to be a number of locations in the NYC area that are at least as active as the area surrounding the Ramapo Fault. Indeed, the largest earthquake recorded by the NEUSSN stations in the NYC area occurred about 25 km from the mapped trace of the Ramapo Fault, near Ardsley, NY (October 19, 1985; m_b 3.6).

Fault plane solutions have been determined for the main shock, a foreshock and an aftershock of the Ardsley, NY earthquake (Filipkowski, 1986; Ebel and Kafka, 1989). These fault plane solutions suggest strike-slip faulting, with one possible fault plane trending NE (i.e. parallel to the trend of the Ramapo Fault) and the other plane trending NW (i.e. transverse to the Ramapo fault). Seeber and Dawers (1987) showed that there is a spatial correlation between a NW-striking mapped fault (the Dobbs Ferry Fault) and the earthquake activity near Ardsley, NY. The Ardsley earthquake

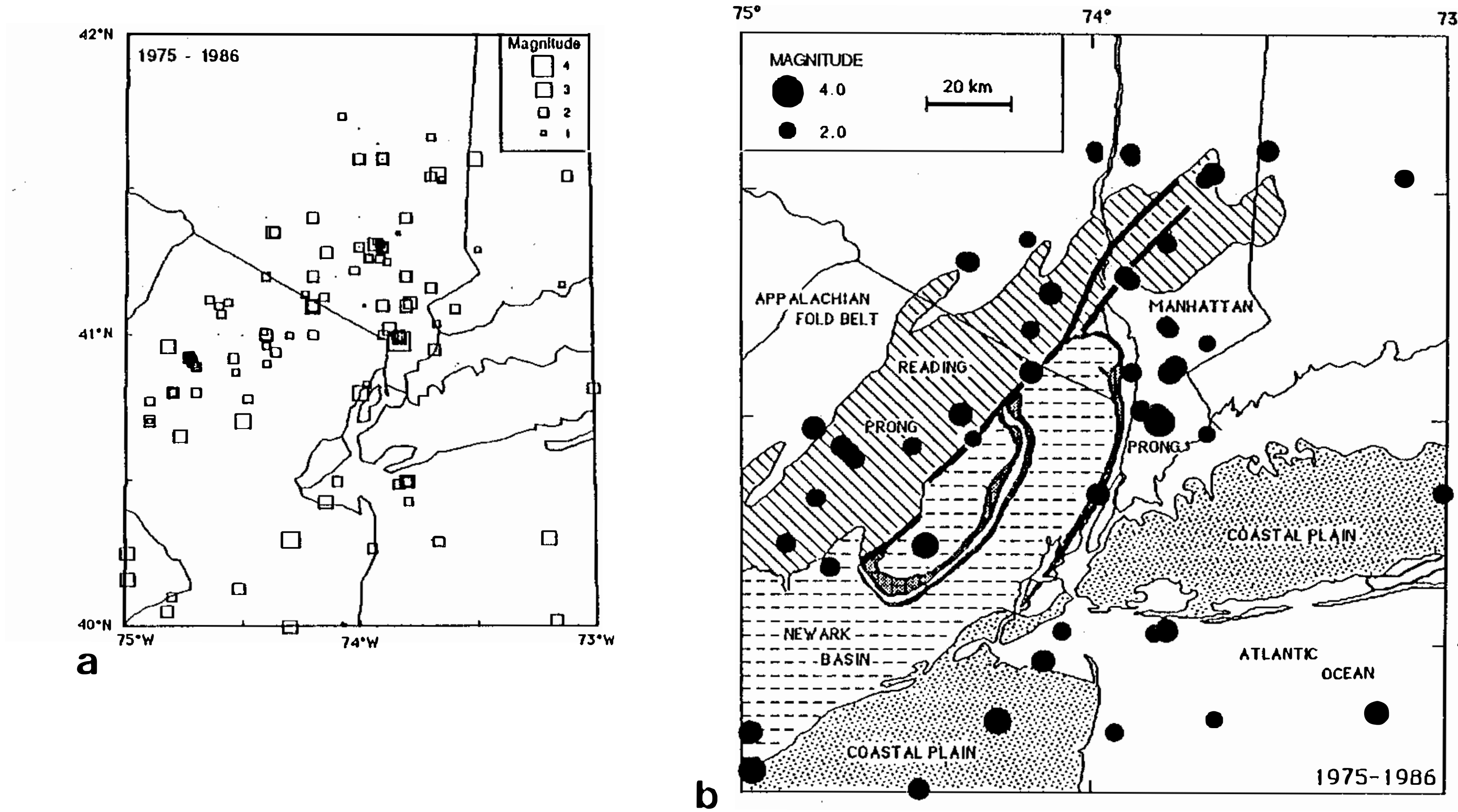


Figure 6. (a) Earthquakes recorded by stations of the NEUSSN in the greater NYC area. Magnitudes are m_N or m_C as published in the bulletins of the NEUSSN. (b) Earthquakes with m_N or $m_C \geq 2$ in the NYC area (1975-1986) and major geological features. Thick solid lines represent the surface trace of the Ramapo fault.

occurred at a depth of about 5 km, however, so that geological features mapped at the surface may be quite different from those that are in the vicinity of the hypocenter.

In the NYC area, it appears that the earthquake activity is broadly distributed throughout the geological features that surround the Newark Basin [Fig. 6(b)]. As in other parts of the NEUS, there are few, if any, cases in the NYC area where surface-mapped faults have been confirmed to be seismically active.

In this field trip we examine evidence for faulting (throughout geologic time) in the vicinity of the Ramapo Fault. As we have discussed above, the area surrounding the Ramapo Fault does not appear to be any more active than numerous other parts of the NYC area (Fig. 6). We have chosen this fault as a focus for this trip, however, not because it is has been confirmed to be seismically active, but rather because it is a prominent feature that exhibits many of the characteristics of geologic faults. This fault has been studied extensively, and it appears to have been active at various times throughout geologic history (Ratcliffe, 1971). In addition, this fault zone has been the focus of controversy related to issues of earthquake hazards at the Indian Point nuclear power plant (Aggarwal and Sykes, 1978), making it of some interest politically as well as scientifically.

GEOLOGY OF THE RAMAPO FAULT SYSTEM

The Ramapo Fault system marks the NE-trending boundary between the Newark Basin and the Hudson Highlands (Figs. 6b and 7). The Ramapo Fault proper extends from Peapack, NJ to the Hudson River near Stony Point, NY (Ratcliffe, 1971). The overall trend of the fault seems controlled by Grenville structures, i.e. by planes of weakness developed during late Precambrian. Outcrops along the Ramapo Fault west of the Hudson River place Precambrian metamorphic and igneous rocks of the Hudson Highlands against Triassic and Jurassic sedimentary rocks of the Newark Basin. Conglomerates near the border fault contain abundant clasts of Precambrian gneiss and Paleozoic cover rocks eroded from the uplifted Hudson Highlands (Carlson, 1946; Savage, 1968). Near Stony Point, outcrops of Paleozoic as well as Triassic rocks lie adjacent to the fault (Savage, 1968). Late Precambrian through Mesozoic displacements along the Ramapo Fault system include right-lateral, normal and reverse slip (Ratcliffe, 1981).

The Ramapo Fault can be followed as a clear, northeast-trending topographic lineament until north of Ladentown, NY (Fig. 7), where it splits into a N20°E-trending segment and one at N60°E connecting with the Thiells fault (Ratcliffe, 1980). The Mott Farm Road Fault extends northeastward and rejoins the main fault north of Tompkins Cove, NY (Ratcliffe, 1980). Alternatively, the main fault may be rotated westward and offset by the Willow Grove fault. If so, a NE-trending fault picks up again along Thiells Road. This second alternative would suggest that the Ramapo Fault is offset 4 km right-laterally. The age of this offset, however, is unknown.

Continuation of the Ramapo Fault across the Hudson River is problematic. It appears to follow either or both the Canopus Creek and Peekskill River valleys. In Peekskill Hollow, Paleozoic rocks lie next to Precambrian gneisses, marking the boundary of the Manhattan Prong. It is mainly a Paleozoic fault, although some Triassic and Jurassic reactivation may have occurred, each time resulting in the southeast side moving down (Ratcliffe, 1980). A complex, semi-ductile shear zone and a fracture zone extend along Canopus Creek, cutting across the structural grain of the Precambrian rocks. This fault zone locally places Paleozoic rocks next to Precambrian rocks (Ratcliffe, 1971). There is no direct evidence to support post-

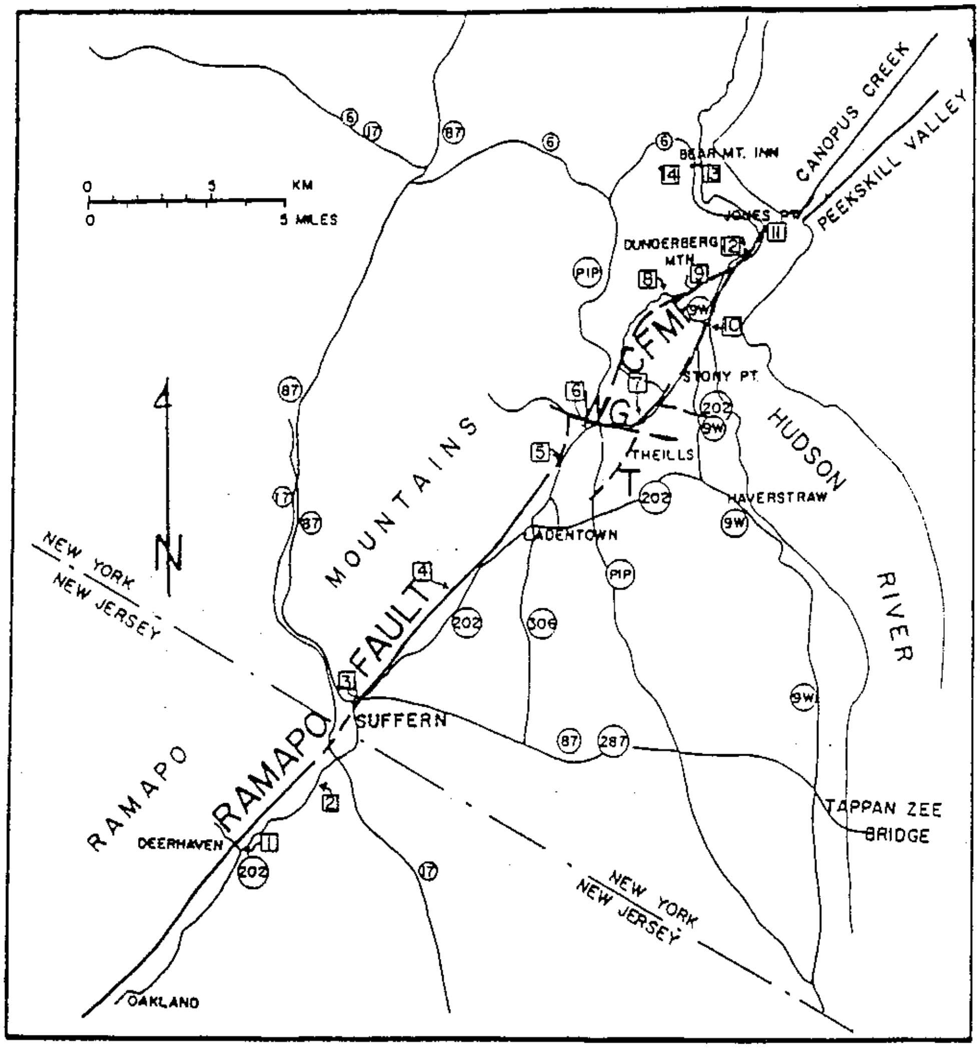


Figure 7. Road map and stops for field trip. Major faults shown as heavy lines may have Quaternary motion. WG = Willow Grove Fault; CFMF = Cedar Flats-Mott Farm Fault. T = Thiells Fault. Stop numbers in squares; Route numbers in circles.

Paleozoic motion east of the Hudson River along these faults (Ratcliffe, 1971; Ratcliffe, 1980). Structural and seismic evidence, however, support both valleys as possibly active faults (Seborowski et al., 1982).

Small-scale brittle structures within 100-200 m of the Ramapo Fault in several localities include conjugate oblique-normal faults (Ratcliffe, 1980). Both sets strike NE. Slickensides measured on the fault surfaces suggest that the SW-dipping set has right-oblique slip, while the NW-dipping set has predominantly left-oblique slip (Ratcliffe, 1980).

The dip of the fault, measured from drilling, is 45-70° SE, steepening toward the NE from Bernardsville, NJ to Stony Point, NY (Ratcliffe, 1980). Unhealed, unconsolidated rock gouge is found at each drilled locality. The fracture patterns, micro-offsets, and slickensides indicate that the northwest side has moved up. The dominant sense is right-oblique faulting. The maximum vertical offset is about 500 m (Ratcliffe, 1980).

GEOLOGIC EVIDENCE OF MOVEMENT ON THE RAMAPO FAULT SYSTEM DURING THE QUATERNARY

The Ramapo Fault stands out as a major structural feature in the NYC area. We examined evidence of Quaternary fault movements at some localities along the Ramapo Fault. The best geomorphic evidence of Quaternary fault movements consists of terrace development on one side of a valley, valley tilting, and systematic tributary offsets. These all indicate moderate slip of the Ramapo fault during the Quaternary.

To date, there is no clear evidence, however, of offset post-glacial material across faults in the NYC area (B. Stone, pers. comm., 1989). Pollen stratigraphy in wetlands along the Ramapo Fault near Ladentown, NY (Nelson, 1980) reveals no sudden changes in the post-glacial record. Unconsolidated cataclasites and gouge along the Ramapo Fault, however, may suggest post-Jurassic (Ratcliffe and Burton, 1984; 1985), or even younger, fault motion. Cores from glacial lake Passaic reveal slump folding, brecciation, and microfaulting (Forsythe and Chisholm, 1989). The features suggest antipodal reverse and normal components. Cumulative vertical displacements may be 1-2 meters.

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RAMAPO FAULT FIELD TRIP

In this field trip, we will examine geologic features that might be related to earthquakes, and we will look for characteristics that might identify active and non-active faults. Of several geomorphological features possibly associated with Quaternary age faulting in the NEUS, a few well-defined examples stand out in the NYC area (Figs. 6b and 7). These features appear to concentrate along the Ramapo Fault zone and the Hudson River valley. This field trip will visit several localities along the distinctive topographic lineament of the Ramapo fault system where geomorphic features suggest Quaternary displacements. These features include the following:

(1) Fault scarps (primary and secondary) along the length of the Ramapo Fault system.
(2) Different heights, numbers, and ages of river terraces across the Ramapo and Mahwah Rivers.
(3) Asymmetric valleys with river running against the faulted side of the valley. This suggests continued tilting of fault blocks (e.g. Ramapo and Mahwah Rivers).
(4) Narrow zones of unconsolidated fault gouge in soil along the Ramapo Fault.
(5) Repeated offset tributaries to the Ramapo river suggesting dextral shear.
(6) Migrated meander loops of the Ramapo River suggesting dextral shear.

Some features, such as offset tributaries, valleys, and asymmetric meanders are better observed on maps and aerial photos than on a field trip. Other tectonically-controlled landforms, such as reversal of river drainage direction, changes in tilt of the Palisades sill, offsets in the sill, and abrupt changes in height of its basal contact, post-date the early Mesozoic, but cannot be proven to be Quaternary in age.

The field trip follows the trace of the Ramapo fault along the Ramapo River in NJ and the Mahwah River in NY. South of Suffern, the Ramapo fault system forms a well-defined linear feature. At Suffern, however, it appears to split into two or more splays beneath a floodplain. The lineament following the Mahwah River marks the boundary between the Hudson Highlands and the Newark Basin. The field trip will follow the Mahwah River to its headwaters, where the Ramapo fault appears to split or be truncated by the Willow Grove Fault. North of Willow Grove, a NE-trending fault can be followed to the Hudson River.

The presence of greater numbers and heights of river terraces on the west side of the Ramapo and Mahwah River valleys along the boundary between the Hudson Highlands and Newark Basin support uplift of the NW side of the valley, the Hudson Highlands, along the Ramapo fault. The regular spacing of the terraces down to the present river edge suggests episodic uplift during the Quaternary. The absence of terraces on the SE side of the river valleys appears to be independent of lithology. In other words, the river terraces on the NW side appear to be tectonic.

Several right-slip offsets of cross-strike (SE-flowing) tributaries occur where they empty into the Ramapo River. Two meanders of the Ramapo also appear to be structurally controlled. They also support Quaternary age right-slip motion.

ROAD LOG FOR FIELD TRIP: EARTHQUAKE ACTIVITY IN THE GREATER NEW YORK CITY AREA

To begin this field trip: from the North, take Routes 6 and 17 east to the NY State Thruway I-87 (Exit 16). Take the Thruway south 16 miles to Exit 15 (Route 17 South to NJ). From the South, follow the Thruway north (I-87 and I-287) from the Tappan Zee Bridge or intersect the Thruway northbound from Exit 9 of the Palisades Interstate Parkway. Take Exit 15 of the Thruway, the second Suffern exit. Follow signs to Route 17 south toward NJ. Mileage begins once exit ramp merges with Route 17 South.

CUMULATIVE MILEAGE	MILES FROM LAST POINT	ROUTE DESCRIPTION
0.0	0.0	Route 17 southbound lane.
1.0	1.0	Exit to Route 202 on right. At the stop sign turn left onto Route 202 South. The Ramapo fault lies to the west between the Ramapo Mountains and the Ramapo River.
3.0	2.0	Pass Ramapo Reservation sign on right.
4.6	1.6	Turn right at small bridge with historical sign and horse crossing marker. Cross the Ramapo River.

STOP 1. DEERHAVEN.

The topography rises toward the west in discrete steps forming several discontinuous terraces on the NW side of the Ramapo River valley. They are not matched on the SE side. The terraces have formed on both unconsolidated sediment and bedrock. Notice how the river flows to the NW side of the valley along most of its length, regardless of the topographic and bedrock changes on the SE side of the river. This suggests westward tilting of the valley toward the fault. Bear Swamp Brook cuts perpendicular to the structure (SE). Upstream its banks are inequal in height. The NE side appears to be higher, as is the case for several other cross-strike streams that cross the Ramapo fault trace.

4.7	0.1	Turn left and cross small creek. The houses are built on a higher level. Follow the road south then west up to a higher level. Take a left at Deerhaven Road which follows a higher level.
5.2	0.5	At T-intersection make a right to go up hill. At fork make right and go up hill to discontinuous outcrops of crystalline rocks.
5.5	0.3	End of road. Return by same route. Notice the asymmetry of the cross-sectional view of the Ramapo River valley.
6.4	0.9	Return to Route 202 and proceed north.
8.0	1.6	Turn left at Ramapo Reservation sign, park at the south end of the parking lot.

STOP 2. RAMAPO RESERVATION.

This stop involves a 30 minute round-trip hike. Follow a trail directly west to a bridge over the Ramapo River. Cross abandoned meander loops and marshy areas. Across the river is a large lake in a wide, flat flood plain. Begin to climb upward. Return to parking area. Exit to south only.

8.2	0.2	Proceed north on Route 202.
9.0	0.8	Pass entrance to Ramapo College on right.
10.3	1.3	Pass under Route 17. Stay on Route 202.
11.3	1.0	Pass under railroad bridge to stop. LOW CLEARANCE! Turn left to remain on Route 202.
12.0	0.7	Take a sharp left immediately after Thruway overpass onto Pavilion Road. Proceed uphill and park halfway up the hill in the Knights of Columbus parking area.

STOP 3. PAVILION ROAD.

Near the bottom of the hill are exposures of slickensided crystalline rock, one of the few outcrops of the Ramapo fault. These well-developed slickensides of unknown age suggest dip-slip movement with a minor right-slip component. Slickensided fault surface averages N 55-65°E, and dips 60-65°SE. Slickenlines pitch N 15°E, 55°SE.

12.7	0.2	Return to the intersection with Route 202 and proceed north. Route 202 runs along one terrace level above the Mahwah River that is not present on the eastern side. Some discontinuous, higher terrace levels can be seen on the west side of the road.
16.3	3.6	Turn left onto Sky Meadow Road.

STOP 4. SKY MEADOW ROAD.

Proceed southwest and west down across the Mahwah River, driving toward the fault. The road climbs four well-developed levels between the river and the fault. Power lines on the mountain follow a natural bench for several kilometers. The river valley is terraced on the NW side, and is hummocky and irregular on the SE side. Drilling data near here suggest that a Triassic age vesicular lava flow, over 450 ft thick, accumulated near the border fault (Ratcliffe, 1980). Local relief at the edge of the basin at the time was approximately 450 feet (Ratcliffe, 1980).

16.6	0.3	Intersection. Turn around. Return to Route 202.
16.9	0.3	Intersection of Sky Meadow Road and Route 202. Turn left onto Route 202.
19.4	2.5	At traffic light turn left. Maps say Ladentown.
19.5	0.1	Turn right at intersection onto Old Route 306.
19.8	0.3	Turn left at sign for Call Hollow Road. This road runs parallel to the Ramapo fault. The river valley narrows toward the north. The

		fault lies on the west side of the river valley. The topography rises on both sides of the road. However, on the east side, the landforms are hummocky and irregular, whereas on the west side of the river, there are distinct levels. The power lines follow one wide terrace.
21.5	0.7	Park on the right side where the power lines cross the road.

STOP 5. CALL HOLLOW ROAD.

North of Ladentown and just south of Willow Grove on Call Hollow Road the ridge defining the eastern edge of the Hudson Highlands begins to turn north as you approach the Willow Grove Valley. The Mahwah River valley pinches out. The terrace along which the power lines have run also ends here. The lines cross the road to join a buried gas line that follows the NW-trending steep ridges and narrow valley. These features also truncate in Willow Grove valley. The gas line is buried in intensely fractured material. Triassic basalts on the W side of the road tilt SE.

21.5	0.0	Proceed north on Call Hollow Road. The Ramapo fault no longer forms a clear topographic lineament toward the north.
22.4	0.9	Park directly under the power lines.

STOP 6. WILLOW GROVE.

The Ramapo fault cannot be easily followed north of Willow Grove valley. The fault appears to trend more toward the north and to be offset about 4 km right laterally by an E-SE-trending fault that follows Minisceongo Creek through Willow Grove valley. The creek experiences several 90° bends near Willow Grove. The terrain to the north and west of Willow Grove consists of massive, rounded hills with short, non-linear valleys as is typical of the topography west of the Ramapo fault. An alternative suggestion is that Willow Grove valley could be a pull-apart developed between the diverging splays of the Ramapo and Theills faults. The saddle to the south consists of highly brecciated rock fragments.

22.6	0.6	At the stop sign at the bottom of the hill, turn right onto Willow Grove Road. Proceed east along Minisceongo Creek.
23.1	0.5	Pass under Palisades Parkway.
24.2	1.1	At Hammond Road (unmarked), bear left.
24.4	0.2	Turn left onto Theills Road; park on the right.

STOP 7. THEILLS ROAD.

Here the best candidate for a continuation of the Ramapo fault is seen to follow the NE-trending section of Cedar Pond Brook. Terraces are seen only on the NW side of the stream valley. Once again it represents hilly Hudson Highlands juxtaposed next to Quaternary sediments. Notice here, too, that there are river terraces only on the NW side of the valley, and that the SE side is lower. Between Call Hollow Road and Theills Road is a possible horst block of Proterozoic gneiss (Ratcliffe, 1980). NE-striking features in the area do not seem to offset lava flows, but discontinuous exposures do not allow confirmation of any post-Mesozoic movements, except for tilting of basaltic layers.

25.2	0.8	Turn left (west) onto Cedar Pond Road (Route 210)
26.5	1.3	Turn right onto Cedar Flats Road. Old schoolhouse on the right. Cedar Flats Road and Mott Farm Road parallel the fault.
28.2	1.7	Past Queensboro Road turn left toward the sign for "Camp Addisone-Boyce". This is Mott Farm Road (unmarked). The road detours around Lake Bullowc. The fault continues on NW side of the lake. The road then resumes its trend parallel to the possible fault.
29.2	1.0	Pass sign for Camp Addisone-Boyce.
29.5	0.3	Park on right after 20 ft high outcrop on left.

STOP 8. MOTT FARM ROAD.

Slickensided surfaces, vertical to steeply dipping. Some slickensided surfaces dip steeply SW with slickenlines plunging 65-75°SW. Other surfaces dip steeply NW with slickenlines plunging 75-80°NE. Several fracture patterns are also evident.

29.8	0.3	Pass outcrop of brecciated rock with complex fracture patterns.
30.1	0.3	Birdhill Road. Left on Fairview.
30.5	0.4	At Tompkins Lake, follow road around toward north on natural terrace 2-3 meters high which is not present on the other side. Steep talus slope rises to west.
30.8	0.3	Stop 9.

STOP 9. TOMPKINS LAKE.

View of terrace and former lake level. Return to main road via Birdhill Road.

31.7	0.9	Birdhill Road intersection with Mott Farm Road. Turn left (east).
32.4	0.7	Pass intersection with Gays Hill Road.
32.65	0.25	Mott Farm Road and Route 9W intersection. Turn right and proceed south.
33.25	0.6	Turn left onto Elm Street. It is a sharp turn. Take the first left. Follow the road down the hill. At the power station, turn right.
33.8	0.55	Park at the large limestone outcrop at the entrance to the Tompkins Cove quarry next to the Lovett power plant.

STOP 10. TOMPKINS COVE QUARRY.

Cambro-Ordovician limestone beds with black chert interbeds and bedding-parallel stylolites trend N 10-20°E and dip 55-65°SE. Calcite-filled fractures cut the outcrop at N 20-30°E and dip 40-42°SW. The brittle features could be related to the Ramapo fault system. See Ratcliffe (1980) for further information. The quarry reveals eight fault zones that disrupt the structural continuity (Ratcliffe, 1980). The faults are identifiable as red or green zones of clay gouge and phyllonite, best seen on the south face of the quarry on all levels. These faults are SE-dipping reverse faults with left- or right-lateral offsets. Despite a youthful appearance, they are associated with folds (Ratcliffe, 1980) although they appear to have been reactivated to produce the clay gouge and slickensides. Ratcliffe suggests that the faults are pre-Mesozoic thrusts that were reactivated during the Triassic as normal and strike-slip faults.

34.2	0.4	At intersection of Elm Street and Route 9W, turn right (north).
34.8	0.6	Pass intersection with Mott Farm Road.
35.8	1.0	Intersection of Gays Hill Road and Route 9W. Turn right (north) onto Route 9W. Slickensided surfaces on a new outcrop exposed on left at mile point 36.0. No stop here. As you proceed north on Route 9W, the Ramapo fault or Mott Farm Road fault runs close to the river. Route 9W is adjacent to or on the fault. Recent construction on Route 9W has revealed rich zones of slickensides, intensely fractured rock, and brecciated zones along this stretch of road.
36.8	1.0	Bear right leaving Route 9W for Jones Point Road. The road follows along a river terrace. There is also a lower level, and the remnant of a higher one. Turn right at the "T", then left to continue north along the lower terrace. These terraces are not present on the east side of the Hudson River valley in this vicinity.
37.5	0.7	Park along the railroad right-of-way at the end of the road.

STOP 11. JONES POINT.

Jones Point is built on a series of low terraces. Walk along the railroad bed toward the north. STAY OFF THE TRACKS! The fault also outcrops immediately behind the house. At the intersection of the road and the railroad tracks is a polished and grooved outcrop that marks the end of the crystalline rocks in this area. The steep, SE-facing fault surface is polished and has sub-horizontal glacial grooves. Small normal faults are seem on the exposure. Note that where the fault crosses the Hudson River, the river abruptly changes direction. The Indian Point power plant is located directly across the Hudson River to the east. The Canopus and Peekskill River valleys are visible from here.

37.8	0.3	Return by the same route, take right fork to view a possible older terrace and till remnant near a former chapel.

38..1 0.3 Cross Route 9W and pull immediately into a wide pull-over and park.

STOP 12. ABANDONDED QUARRY.

The Ramapo fault cuts the south side of the hill. Between this site and the river are several low terraces. Note truncated (over-steepened) spurs here. Walking west over a wide flat area, there is a possible scarp and sag pond in the area not exploited by the former quarry. The small scarp is 1-1.5 meters high. The terrace is developed on stream cobbles of a former river level.

38.1	0.0	Proceed north on Route 9W.
40.8	2.7	Take left fork in road for Bear Mountain State Park and Inn (Route 6).
41.5	0.7	At traffic circle, take first right for Bear Mountain Inn.
41.9	0.4	Turn left into parking lot.

STOP 13. BEAR MOUNTAIN INN.

Refreshments, etc. The only scheduled rest stop.

42.0	0.1	Leave Bear Mountain Inn. Turn right onto Route 6. At 42.3 mile point, Bear Mt. Circle, take first right; continue on Route 6 west.
44.0	2.1	Turn right at sign to Perkins Memorial Drive. Proceed to top of mountain to lookout. No mileage.

STOP 14. PERKINS MEMORIAL DRIVE LOOKOUT.

The Ramapo fault crosses the Hudson River to enter the Peekskill Hollow or Canopus River valleys. Sharp bends in the Hudson River appear to be structurally controlled. Each linear river bend, when projected onshore in Westchester or Rockland County, is associated with an anomalously linear river valley that follows the same strike. Some examples include the Ramapo, Mahwah, Cedar Pond, Minisceongo, Canopus, Leeds Cove, and Peekskill Hollow valleys. The terrace on the NW side of the Peekskill river appears to be higher than on the east side.

On a regional scale, the Hudson River is linear and trends N-S, except where crystalline rocks cross it. At least one of the NE-trending segments occurs where the Ramapo fault crosses the Hudson. Another SE jog occurs where the Willow Grove fault crosses the river. The trend of another sharp bend in the Hudson River toward the SE follows a lineament onshore into Westchester County along the linear Leeds Cove valley. This latter example suggests that the Ramapo fault offsets one kilometer right-laterally a possible fault beneath the Hudson River.

Turn right onto Route 6 and proceed west. After a forced merge with the Palisades Interstate Parkway either proceed south toward New York City or take Exit 18 to remain on Route 6 West. At traffic circle, take first right onto Route 6 West and continue to Routes 6, 17, or the New York Thruway.

NOTES

DEGLACIAL HISTORY AND ENVIRONMENTS OF THE UPPER WALLKILL VALLEY

G. GORDON CONNALLY
12 University Avenue
Buffalo, NY 14214

LES SIRKIN
Department of Earth Sciences
Adelphi University
Garden City, NY 11530

DONALD H. CADWELL
New York State Geological Survey
Room 3140
Albany, NY 12230

INTRODUCTION

The glacial history of the Wallkill Valley of eastern New York first was described by Connally and Sirkin (1967) in the guidebook for the 39th annual meeting of NYSGA. Prior to their interest, the area had received scant attention. It was mentioned peripherally by Woodworth (1905) and discussed in abstract by Connally (1966). The only previous detailed work was a study of proglacial lakes reported by Adams (1934).

Before commencing the joint study, Connally and Sirkin established restrictive criteria for locating the bogs that have become the standard sections: 1) some aspect of the peat bog must be directly linked to some specific geomorphic event, 2) the bog must have a high, unbreached rim with exclusively internal drainage, and 3) a deep section must directly overlie glacial sediments. We refer to this strategy as an integrated approach and contrast it with the grab-sample approach of others.

Following the tenets of the integrated approach, Connally and Sirkin (1970) refined interpretations. They discussed the valley again in the context of the entire Hudson-Champlain Lobe (Connally and Sirkin, 1973) and then Connally (1983) and Connally and Sirkin (1986) discussed regional correlations.

In the summer of 1988, Connally completed a 25-year project, mapping the entire Wallkill Valley at a scale of 1:24,000. The 7½' maps that cover the New York portion are on open file in the New York State Geological Survey in Albany, New York. NYSGS supported the later phases of this project, from 1983 through 1988, under the aegis of Don Cadwell. The maps are part of the data base for the Lower Hudson sheet of the Surficial Geologic Map of New York. During NYSGS sponsorship, new landforms have been recognized, new ideas have been generated in the field of glacial geology, and new interpretations have been applied to the Wallkill Valley

GEOLOGIC SETTING

The Wallkill Valley is a northeast-southwest trending lowland approximately 100 km long and 30 km wide, as shown in Figure 1. It narrows to the south. The western boundary is the scarp of the Shawangunk Mountain cuesta. The eastern boundary, separating the Wallkill Valley from the Hudson River trench, is the Marlboro Mountains. The crystalline rocks of the Hudson Highlands, and appended Schunnemunk Mountain, form the southeastern boundary.

Most of the valley is drained by the northerly flowing Wallkill River and its tributaries. At Sussex, New Jersey the trunk stream bifurcates. Papakating Creek drains the physiographic continuation of the valley for 12 km, from Augusta north to the bifurcation at Sussex. However, the Wallkill River rises almost due south, in a valley that cleaves the crystalline uplands. The southern end of the physiographic Wallkill Valley is approximated by the Culvers Gap Moraine.

The other major drainage system in the Wallkill Valley is Moodna Creek. This stream, and its tributary Otter Creek, flows eastward and empties into the Hudson River between the southern end of the Marlboro Mountains and the Hudson Highlands. Seeley Brook is a small southern tributary to Moodna Creek.

The lowest point in the Wallkill Valley is 140 ft at the confluence with lower Rondout Creek, north of Rosendale, New York. The highest point is 2289 ft at Sam's Point on the crest of the Shawangunks. Though the total relief is more than 2000 ft, most of the valley bottom lies between 200 and 600 ft above sea level, exhibiting low to moderate relief.

Bedrock

The valley is underlain by rocks of the Ordovician Martinsburg Formation. Resistant sandstone and siltstone beds account for the prominent strike ridges and roches moutonees in the valley bottom. On the east, the Marlboro Mountains result from very coarse and very resistent sandstones in the Quassaic Group. On the west, the softer sedimentary rocks are buttressed by the overlying, very competent, Shawangunk Formation, producing the spectacular Shawangunk escarpment (STOP 5). On the south and east, Schunemunk Mountain is upheld by resistant clastic sedimentary rocks of Devonian age.

Drift

There is an almost continuous blanket of supraglacial meltout till throughout the valley. This till is clogged by clasts and channers of Martinsburg lithologies. It usually occurs as clast-supported diamict. Matrix-supported diamict has been observed only in drumlin cores where sedimentary structures suggest that it has been disturbed or resedimented. Unequivocal subglacial meltout till has not been observed in the valley.

Ice-contact stratified drift is common. It is primarily shale-gravel and sand. However, prominent white sandstone and quartzite clasts derived from the Shawangunk Formation are common in the Wallkill Moraine and immediately adjacent to the Shawangunk Mountains south of that moraine. Lacustrine

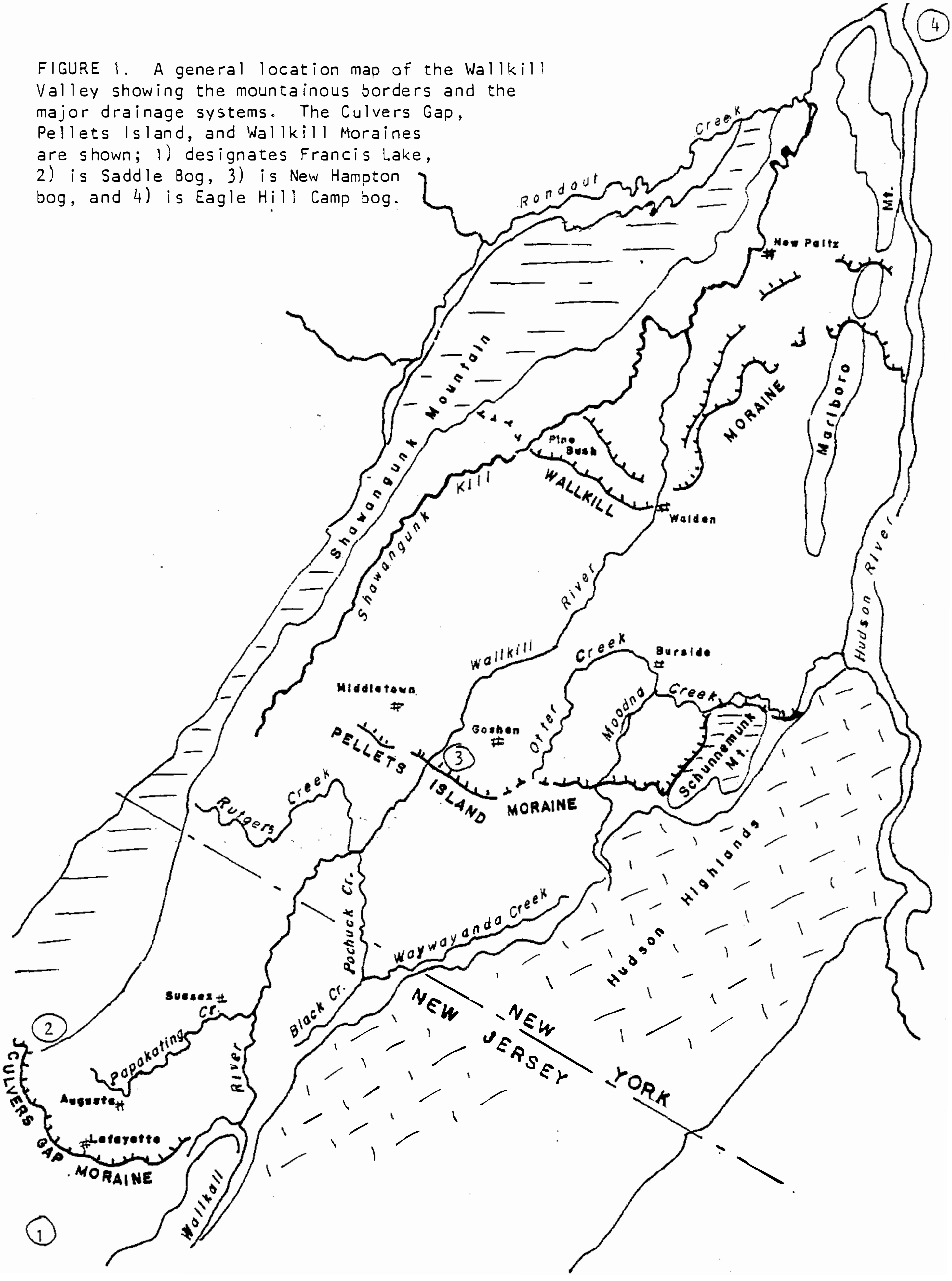

FIGURE 1. A general location map of the Wallkill Valley showing the mountainous borders and the major drainage systems. The Culvers Gap, Pellets Island, and Wallkill Moraines are shown; 1) designates Francis Lake, 2) is Saddle Bog, 3) is New Hampton bog, and 4) is Eagle Hill Camp bog.

GLACIAL GEOLOGY

Connally and Sirkin (1967) described 5 ice margin positions, from oldest to youngest, the Ogdensburg-Culvers Gap, Augusta, Pellets Island, New Hampton, and Wallkill Moraines. In 1970, they added a brief discussion of the Rosendale readvance, described by Connally (1968), recognizing it as the youngest deglacial event in the valley. By 1973, they had realized that the "New Hampton Moraine" was only a minor recessional feature closely related to the Pellets Island Moraine and abandoned it as an independent moraine or even as a distinct ice margin position. Also by 1973, Connally had become suspicous that the western Culvers Gap Moraine described by Salisbury (1902) might not correlate with the feature at Ogdensburg and thus Connally and Sirkin (1973) dropped "Ogdensburg" from the previously hyphenated name. Unfortunately, there was no discussion of the reason for that change. Finally, still in 1973, they added the "Sussex Moraine" that was inferred to be recessional to the "Augusta Moraine" in northern New Jersey.

Because of the new data and new ideas generated during the most recent mapping efforts, we here abandon both the Augusta and Sussex "moraines", no longer recognizing them as ice margin positions, and also modify previous views about the Culvers Gap Moraine. We now recognize the Culvers Gap, Pellets Island, and Wallkill Moraines as the only significant ice margin positions in the Wallkill Valley. However, we continue to recognize the Rosendale readvance, an event lacking a definitive ice margin position, as the youngest event to affect the lower Wallkill Valley. The evolution of terminology, from 1967 to the present, is summarized in Figure 2.

CONNALLY and SIRKIN (1967)	CONNALLY AND SIRKIN (1970)	CONNALLY AND SIRKIN (1973)	CONNALLY-SIRKIN-CADWELL (this Guidebook)
not recognized	ROSENDALE READVANCE	ROSENDALE READVANCE	ROSENDALE READVANCE
WALLKILL MORAINE	WALLKILL MORAINE	WALLKILL MORAINE	WALLKILL MORAINE
NEW HAMPTON MORAINE	NEW HAMPTON MORAINE	*abandoned*	
PELLETS ISLAND MORAINE	PELLETS ISLAND MORAINE	PELLETS ISLAND MORAINE	PELLETS ISLAND MORAINE
not recognized	*not recognized*	SUSSEX MORAINE	*abandoned*
AUGUSTA MORAINE	AUGUSTA MORAINE	AUGUSTA MORAINE	*abandoned*
OGDENSBURG-CULVERS GAP MORAINE	OGDENSBURG-CULVERS GAP MORAINE	CULVERS GAP MORAINE	CULVERS GAP MORAINE

FIGURE 2. Ice margins and events in the Wallkill Valley as they have been interpreted from 1967 to the present.

sands and rhythmically bedded silt and clay are common in bottom lands. Extensive organic deposits are present in the mucklands of southern Orange County. These evidently mark the final phase of sedimentation in once prominent lakes. Organic sediments also are present in many kettle bogs, including New Hampton bog where Sirkin (Connally and Sirkin, 1967) established the pollen stratigraphic standard section for the valley.

Inversion Ridges

Before examining the glacial history of the Wallkill Valley, it is necessary to consider a newly recognized landform that has increased our understanding of events in the valley. This landform is named an inversion ridge. Inversion ridges, and inversion terraces, are quite common flanking the Shawangunk Mountains. Some of these deposits were referred to as inwash and were recognized by Connally and Sirkin in 1967, although they were not fully understood at that time.

The concept of inverted topography certainly is not new to glacial geology (e.g., see Lobeck, 1939, p. 312). It has long been recognized that depressions, or negative areas, on ice sheets collect and concentrate sediment. When the glacier melts, the sediment is let down to form topographic highs, or positive topographic areas. As a corollary, the positive regions on a glacier, comprising mainly unmelted glacier ice, become nagative topographic regions exhibiting a minimum of sediment when the glacier disappears. The definition of the inversion ridge is rooted in this long established concept.

An inversion ridge is a linear accumulation of drift that is located in a depression or valley bottom and is parallel or subparallel to an existing stream channel. It may be prominent or subdued; dissected or intact. It heads at a subareal drainage channel, or at a col or upland valley. It is composed of stratified drift and/or resedimented, clast-supported diamict. Matrix-supported diamict also may occur, but at present it is considered to be a minor constituent. The downslope, or downstream, end of the inversion ridge may merge with another inversion ridge, develop into an inwash or outwash fan (STOP 1), or terminate at what Fleisher (1986) called a dead ice sink. An inversion terrace is similar except that it is banked against a valley wall with the present stream channel on the valley side.

The recognition of inversion ridges as separate landforms restricts the use of the term esker. Inversion ridges represent drainage that was initiated in drainage systems above or beyond stagnant ice masses. Es kers are restricted to former drainage systems that were entirely englacial or subglacial in origin and which therefore may be unrelated to present drainage systems.

Because inversion ridges are so well preserved, and because they often are associated with dead ice sinks, they almost certainly resulted from deposition in or on stagnant ice. They are inferred to document drainage systems that functioned much as they do today after the ice has disappeared. The only difference between modern drainageways and those reconstructed from the inverted topography is that stagnant ice filled the depressions, or valley bottoms, or stream channels and channelized sediments were confined by walls of ice and preserved. Sedimentary structures in rare exposures imply downslope sediment transport, either by water currents or by mass movement.

Culvers Gap Moraine

Salisbury (1902, p. 270) describes

> "... a fairly well defined belt of recessional moraine ... from Ogdensburg, *via* Balesville, to Culver's Gap."

and continues with a more detailed description on pages 350 to 355. In summarizing the glacial history of the Hudson-Champlain Lobe, Connally and Sirkin (1973) inferred that the Culvers Gap Moraine, as illustrated in Figure 3, marked the maximum Woodfordian position. Because of doubts that were expressed in an unpublished manuscript by Connally in the late 1960's, the Ogdensburg portion of Salisbury's Ogdensburg-Culver's Gap Moraine was not included and the name was shortened to "Culvers Gap Moraine".

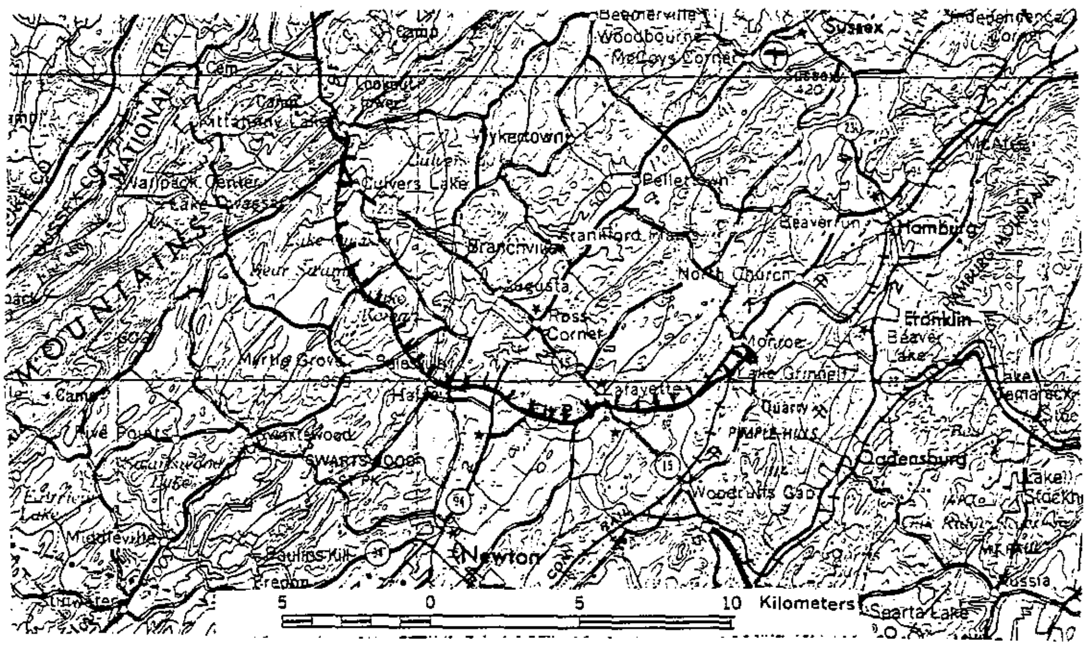

FIGURE 3. The Culvers Gap Moraine from the Kittatiny Mountains on the west to the Pimple Hills on the east. An inversion ridge at Ogdensburg is here considered to be somewhat older than the moraine.

Once Fleisher (1984) introduced the concept of a dead ice sink, later refined by Fleisher (1986), it became obvious that there was a dead ice sink immediately south of the morainal segment southwest of Lafayette, New Jersey. The Lafayette segment exhibits truncated foreset beds that were deposited against, or upon, a stagnant ice mass 3 km long that occupied the Paulins Kill valley from Lafayette south to Newton. If stagnant ice was present as the moraine was being deposited, the moraine hardly could have marked the maximum position of the Hudson-Champlain Lobe. In fact, at least the higher parts of the moraine probably are better interpreted as inversion ridges than as end moraine. Traced westward to the Shawangunk Mountains (Kittatiny Mountain in New Jersey), it appears to be inverted topography formed when drainage from the mountain encountered stagnant ice in the valley.

We still consider the Culvers Gap Moraine to be an end moraine, at least in part. At Lafayette, a large volume of sediment was furnished from a northern "up-ice" source, presumably active ice in the tradition of "the dirt machine" of Koteff (1974). From Lafayette eastward, the ice margin position disappears in the pitted outwash on the west side of Germany Flats (not shown on Fig. 3). The relationship of Germany Flats outwash to the drift at Ogdensburg is problematic.

The Ogdensburg segment probably represents a situation similar to that at Lafayette. The upper Wallkill Valley, southwest of Ogdensburg, probably was a dead ice sink. Drainage from Franklin Creek was channeled southwest onto stagnant ice, ultimately separating a stagnant ice block to the north from another to the south. The detailed mapping by Connally suggests that the outwash at Germany Flats, clearly related to the Culvers Gap Moraine, may be younger than the stagnant ice that remained north of the Ogdensburg inversion ridge. Thus, the drift at Ogdensburg probably is not a moraine segment and may well be older than the Culvers Gap Moraine at Lafayette.

Drift At Augusta

Salisbury (1902, p. 374) first reported the "Augusta Moraine" as follows:

> "A small moraine crosses the valley ... a mile or so north of Augusta. In front of the moraine the surface is strewn with large cobbles, but the gravel decreases notably in size to the southward ... this plain partakes of an overwash plain or a wide valley train.
>
> The southern edge of this plain is continuous with interrupted terraces along Paulinskill,.... The lack of continuity here does not appear to be due altogether to erosion ... the valley was partly occupied with ice when the gravel was deposited. Strongly marked stagnant ice forms are present ..."

The deposits described by Salisbury now are interpreted as an inversion ridge formed when Papakating Creek flowed out onto stagnant ice in the main valley. Initial drainage probably was graded to ±520 ft, though the graded plain referred to by Salisbury clearly is graded to ±500 ft. More stagnant ice topography, also previously mentioned by Salisbury, is present between Augusta and Sussex. Most deposits crest at ±520 ft, suggesting initial presence of a threshold to the south that impounded waters at that elevation.

The "Augusta Moraine" is here abandoned both as a moraine and as an ice margin position.

Drift At Sussex

Salisbury (1902, p. 423) described "The Sussex Delta" as follows:

> "A little east of Sussex is an elevated sand and gravel plain ... Its elevation about 535 feet ... it rises by a steep slope eighty to 100 feet above the low land at its border ... On the top, the delta form is less distinct. Instead of being flat or gently sloping, it is pitted by sinks, twenty-five to forty feet deep, and one knoll rises ten feet above the general level ... no exposures reveal the structure ..."

Connally and Sirkin (1973) reinterpreted this feature, located on the west side of the Wallkill Valley, as an end moraine and renamed it the "Sussex Moraine". It was inferred to be recessional to the "moraine" at Augusta. This feature (STOP 1) now is considered to be an inversion ridge deposited by Clove Brook drainage flowing out onto stagnant ice in the valley. It clearly is not graded to either the 520 ft or 500 ft water levels that existed to the south during deglaciation. This inversion ridge must have preceded the ice margin that impounded the lakes to the south. Younger deposits on the east side of the Wallkill River are graded to ±520 ft, with an erosional terrace at ±500 ft. The eastern deposits (STOP 1) are in an inversion ridge from an upland drainage basin and evidently formed after the western inversion ridge, as the stagnant ice surface lowered to the water planes.

The "Sussex Moraine" is here abandoned, both as a moraine and as an ice margin position.

Though there are no true recessional moraines between the Culvers Gap Moraine and the Pellets Island Moraine 42 km to the northeast, there are three prominent stagnant ice complexes, one west and two east of the mucklands surrounding the Wallkill River. To the west, in the valley of Rutgers Creek, ice-contact stratified drift and inversion ridges south of Westtown document a dead ice sink. As the ice retreated, or melted, water was impounded first at ±540 ft and later at ±500 ft. To the east, adjacent to the valley of Pochuck Creek, there is massive ice-contact drift just north of the New Jersey border. Northeast of Warwick, New York, there is a huge dead ice sink of about 20 km^2 that is almost completely collared with outwash and/or inwash gravel. The valley bottom is above 520 ft, precluding evidence of a 500 ft lake level. However, there is abundant evidence that as the stagnant ice melted it was replaced by a lake with a level of ±530 ft in the south and ±580 ft at the Pellets Island Moraine. The only distinctive deposits in the main Wallkill Valley are in a remnant esker system at Breeze Hill, 4 km south of Pellets Island and adjacent to the mucklands.

Pellets Island Moraine

The Pellets Island Moraine, illustrated in Figure 4, was described originally by Connally and Sirkin (1967) as a massive ridge of till and stratified drift. It was named for exposures of sand and gravel in a gravel pit at Pellets Island. The type locality is badly slumped but several newer exposures are present (STOP 2) about 3 km southwest of Goshen. The moraine is traced continuously eastward at the head of the mucklands to the massive esker-fed moraine at Chester; then to the western flank of Schunnemunk Mountain; and thence northeast to the Hudson Highlands west of the Hudson River. Connally and Sirkin (1986) correlated the Pellets Island Moraine with the Shenandoah Moraine, banked against the Hudson Highlands east of the Hudson River.

The Pellets Island Moraine cannot be traced west of the Wallkill River channel. There are esker-fed morainal remnants about 3 km northwest of Pellets Island, in the valley of Catlin Creek, that are undoubted equivalents. However, there is little ice-contact material present between Catlin Creek and the Shawangunk Mountains. Does this suggest a general absence of meltwater in the vicinity of the Shawangunks during deglaciation? Perhaps the meltwater, largely sediment-free, was channeled east to the Wallkill River eskers via englacial drainage.

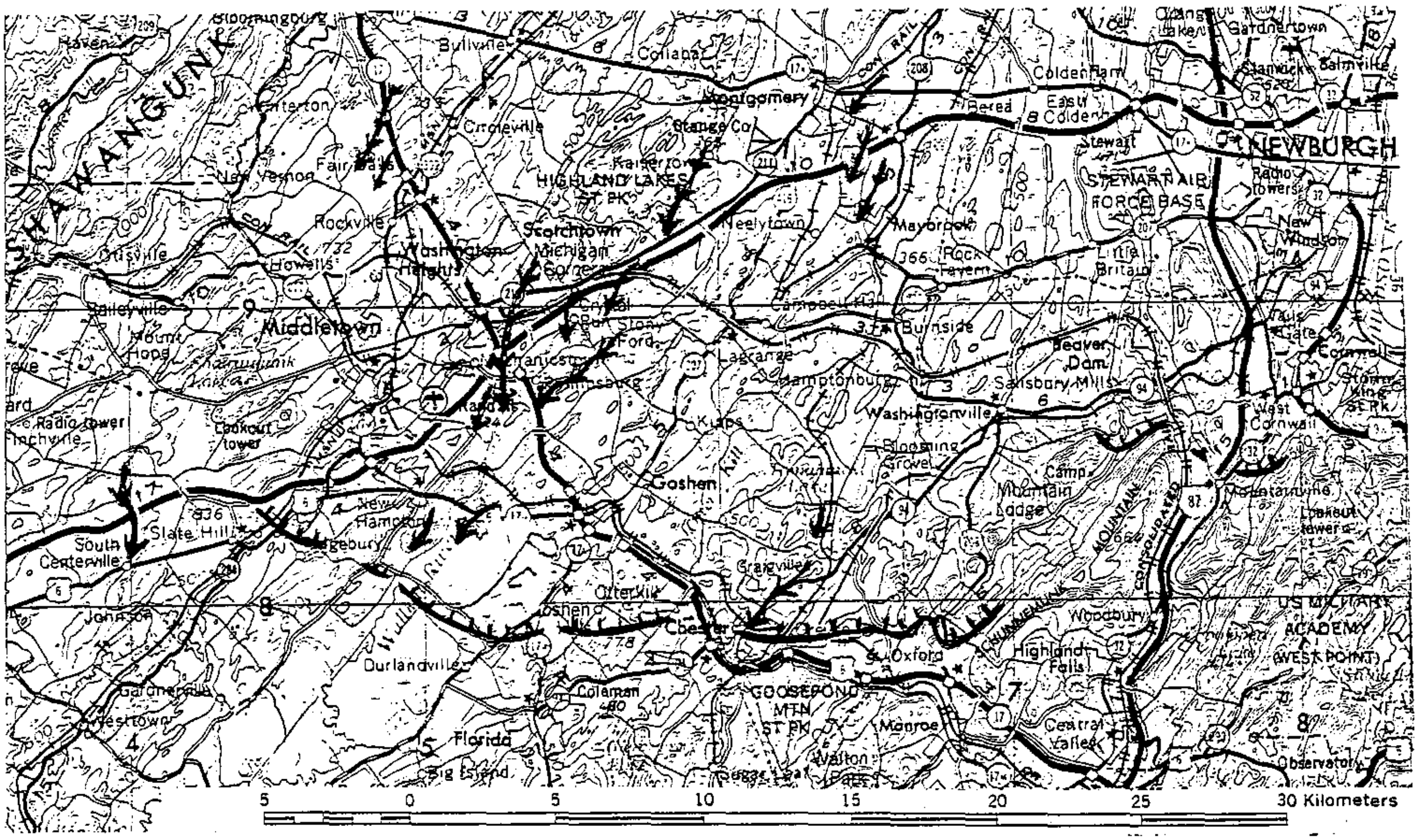

FIGURE 4. The Pellets Island Moraine from Rutgers Creek, southwest of Middletown, to Mountainville in the Hudson Highlands. Arrows indicate the major esker systems that fed sediment to the ice margin.

For 22 km northeast of the Pellets Island Moraine, the Wallkill Valley is congested with esker remnants (STOP 3) and attendant ice-contact deposits. It appears that most sediment was fed to the ice margin by subglacial drainage that funneled into the Wallkill River channels. Outwash and lacustrine deposits flank the ice-contact drift, attesting to a northward-expanding proglacial lake in the lowlands that followed the receding ice margin back to the Wallkill Moraine.

Glacial Lake Fairchild

When the Woodfordian ice margin began to retreat from the Lafayette segment of the Culvers Gap Moraine, meltwater was impounded between the ice margin and the moraine. Evidence between Augusta and Sussex suggests that the initial impoundment was at ±520 ft, but later lowered to ±500 ft. The outlet for these waters was westward to Paulins Kill, either through a channel at Augusta or through an alternate channel 1 km south. Because the southern channel lacks erosional features and is filled with sediment that is smoothly graded to the 500 ft level, we infer that it handled the initial effluent at ±520 ft. Then, much later, the northerly channel at Augusta became the master outlet, the lake level droped to ±500 ft, and the southern channel aggraded.

At first, the impoundment was confined to the lower Papakating Creek valley. Then, when the ice margin retreated north of the bifurcation at Sussex, one of two things happened. Either the waters at ±520 ft expanded southward sown the Wallkill River channel, or a local lake at a superior elevation lowered and merged with the expanding 520 ft lake. Similarly, when the ice margin later retreated north of Pochuck Mountain (Fig. 1), the 520 ft lake expanded east and south, up the Pochuck Creek valley to the vicinity of Sand Hills in the tributary Black Creek valley.

When the Woodfordian glacier stood at the Pellets Island Moraine, the lake filled the entire Wallkill Valley, including lower Papakating Creek and Pochuck Creek, at an elevation of ±520 ft. It filled the valley of Rutgers Creek-Catlin Creek southwest of Middletown at an elevation of ±540 ft. The lake extended eastward along the moraine to Chester, where it was blocked at first by the crystalline highlands. In the center of the Wallkill Valley, foreset beds from the moraine document initial deposition at ±580 ft. To the east and west, along the moraine, the elevation was ±560 ft. The lake expanded south from Chester to occupy the dead ice sink at Warwick at ±540 ft. The relationship of this lake and the Pellets Island Moraine is shown in Figure 5.

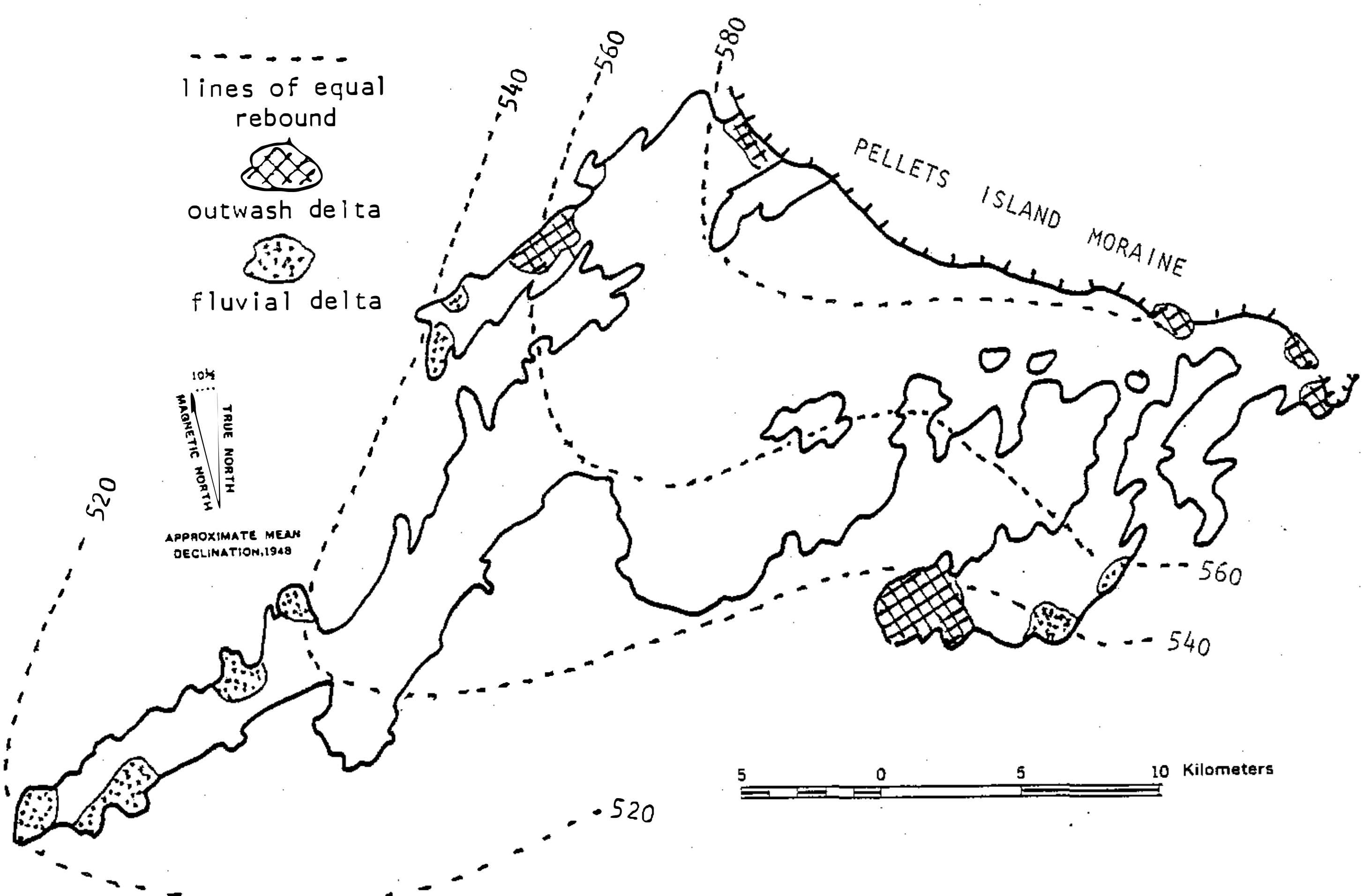

FIGURE 5. Glacial Lake Fairchild when dammed by the ice margin at the Pellets Island Moraine. Outwash deltas and fluvial deltas that were marginal to the lake are illustrated. Isobases suggest lines of equal rebound. An overhead projector transparency of this figure makes an excellent overlay for a 1:250,000 map.

In 1970, Connally submitted a manuscript report to NYSGS and requested that it be considered for publication in expanded form. Since he has had no subsequent word, evidently it still is under consideration. In that manuscript, he proposed the name Lake Fairchild in honor of the late Herman LeRoy Fairchild, pioneering glacial geologist in New York and former chair of the Department of Geology at the University of Rochester. Although the name has been used on an informal basis since that time, we here propose the name Glacial Lake Fairchild for the lake impounded by the Woodfordian glacier as it stood at the Pellets Island Moraine and is illustrated in Figure 5 and described below.

Glacial Lake Fairchild filled parts of four parallel, southwest-trending valleys. The major portion occupied the Wallkill/Papakating valley with a southern outlet to New Jersey's Kittatiny Valley. It occupied the Rutgers Creek valley to the northwest, the upper Waywayanda Creek valley between Chester and Warwick, and the Seeley Brook valley on the east.

The 520 ft Frankford Plains delta near Augusta and the alluvial plain from Armstrong to Pelletstown, on the east side of the Papakating valley, mark the southern end of Lake Fairchild. Another delta 8 km north at McCoys Corners also crests at ±520 ft. West of Sussex, two ice-contact deltas crest at ±540 ft in the tributary valley of Clove Creek, and two more, farther north at Owens and Quarryville, are similar. The only delta north of Sussex is at ±560 ft southwest of Florida, New York, though a ±500 ft delta north of Florida probably represents the lowered, post-rebound, 500 ft phase.

In the Rutgers Creek valley, many lacustrine features are present. Deltas at ±540 ft at Waterloo Mills, ±560 ft and ±500 ft at Westtown; inwash at ±570 ft and a delta at ±500 ft at Millsburg; a spit(?) at ±560 ft west of Slate Hill; and flat-topped outwash at ±580 ft at the Pellets Island Moraine suggest a northerly rebound of 40 ft in 16 km. A similar gradient is evident between Chester and Warwick. South of Warwick, at New Milford, deltaic deposits developed within stagnant ice topography at ±530 ft; north of Warwick several summits are graded to ±560 ft; south of Durland Hill, and at Smith's Clove on the moraine, there are 580 ft deltas. However, the straight connecting the dead ice sink at Warwick with the main lake must have been quite narrow.

As the Woodfordian ice margin retreated north, down the Wallkill Valley, outwash graded to ±580 ft was deposited east and west of New Hampton and farther north at Phillipsburg. Thus, Lake Fairchild continued to expand northward during deglaciation. Following deposition of the Phillipsburg outwash, the new master channel at Augusta appears to have lowered the lake level to a uniform(?) ±500 ft. Because 500 ft features are ubiquitous, local rebound must have followed very closely the removal of ice and the draining of several million litres of proglacial meltwater. Figure 5 shows suggested isolines of equal rebound (isobases) following draining of Lake Fairchild.

Glacial Lake Woodworth

Shawangunk Kill is a mid-elevation tributary of the Wallkill River. It drains the western Wallkill Valley at the base of the Shawangunk Mountains. South of the Wallkill Moraine, the Wallkill River and Shawangunk Kill channels are separated by a high bedrock strike ridge. As the glacier was retreating from the Pellets Island Moraine to the Wallkill Moraine, it impounded water in the valley of Shawangunk Kill at ±600 ft. That lake, confined to the valley of Shawangunk Kill by the glacier and the bedrock ridge, is here named Glacial Lake Woodworth in honor of the late John Brainard Woodworth, the first glacial geologist to report on the Hudson Valley region.

At its maximum extent, Lake Woodworth was just 14 km long, from Winterton to the southwest to near Ulsterville on the northeast. As the ice margin retreated down valley, the level appears to have lowered first to ±520 ft, then to ±440 ft, and finally to the regional level of ±400 ft. The lake probably was penecontemporaneous with Lake Dyson, described on following pages, and is illustrated with that lake in Figure 7.

Wallkill Moraine

The Wallkill Moraine was named and described by Connally and Sirkin (1967). However, detailed mapping has confirmed neither the original description nor the original tracing of the ice margins. Only two ice margins are recognized and illustrated in Figure 6, rather than the three that were proposed in 1967. The outer moraine crosses the Wallkill River at the village of Walden. It was traced westward to Pine Bush as a series of closely spaced ice-contact deposits, outwash deltas, or outwash fans, most fed by eskers. The moraine does not extend south to Bloomingburg as originally suggested. Rather, outwash fans suggest a margin near Walker Valley at the base of the Shawangunk Mountains, 8 km north of Bloomingburg. A recessional position exists 2 to 3 km northeast of the outer margin, tracable from Allards Corners to Red Mills. The outer position is traced east and north from Walden to Modena using outwash aprons. From Modena, the margin turns eastward to Baileys Gap, a col in the Marlboro Mountains, and then southeast to the Milton delta of Lake Albany at the Hudson River trench. The recessional position is traced northward to Clintondale Station, then to Lloyd, and then to a notch at the north end of Illinois Mountain. The recessional margin appears to correlate with the Poughkeepsie Moraine east of the Hudson, as proposed by Connally and Sirkin (1986). However, the outer Wallkill Moraine at Milton may predate the Poughkeepsie ice margin by a short span of time.

The Wallkill Moraine, like the Pellets Island Moraine, seems to have derived its sediment from an esker system along the Wallkill River channel (STOP 5). Swarms of feeder eskers head at or near the base of the spectacular Shawangunk climbing cliffs. It is probable that the fresh rock face of the northern Shawangunks was created by hydraulic plucking in a bergschrund-like situation that developed between the Woodfordian glacier and the mountain, as it stood at the Wallkill Moraine. Many of these feeder eskers were constructed of huge blocks of Shawangunk conglomerate that were frozen in time as they were moving down the mountain side (STOP 6).

Glacial Lake Dyson

As the ice margin retreated from the Pellets Island Moraine, Lake Fairchild expanded northward in contact with the glacier at ±580 ft. After the glacier had retreated north of Phillipsburg in the main valley, the level of Lake Fairchild lowered to ±500 ft. When the ice margin retreated north of the

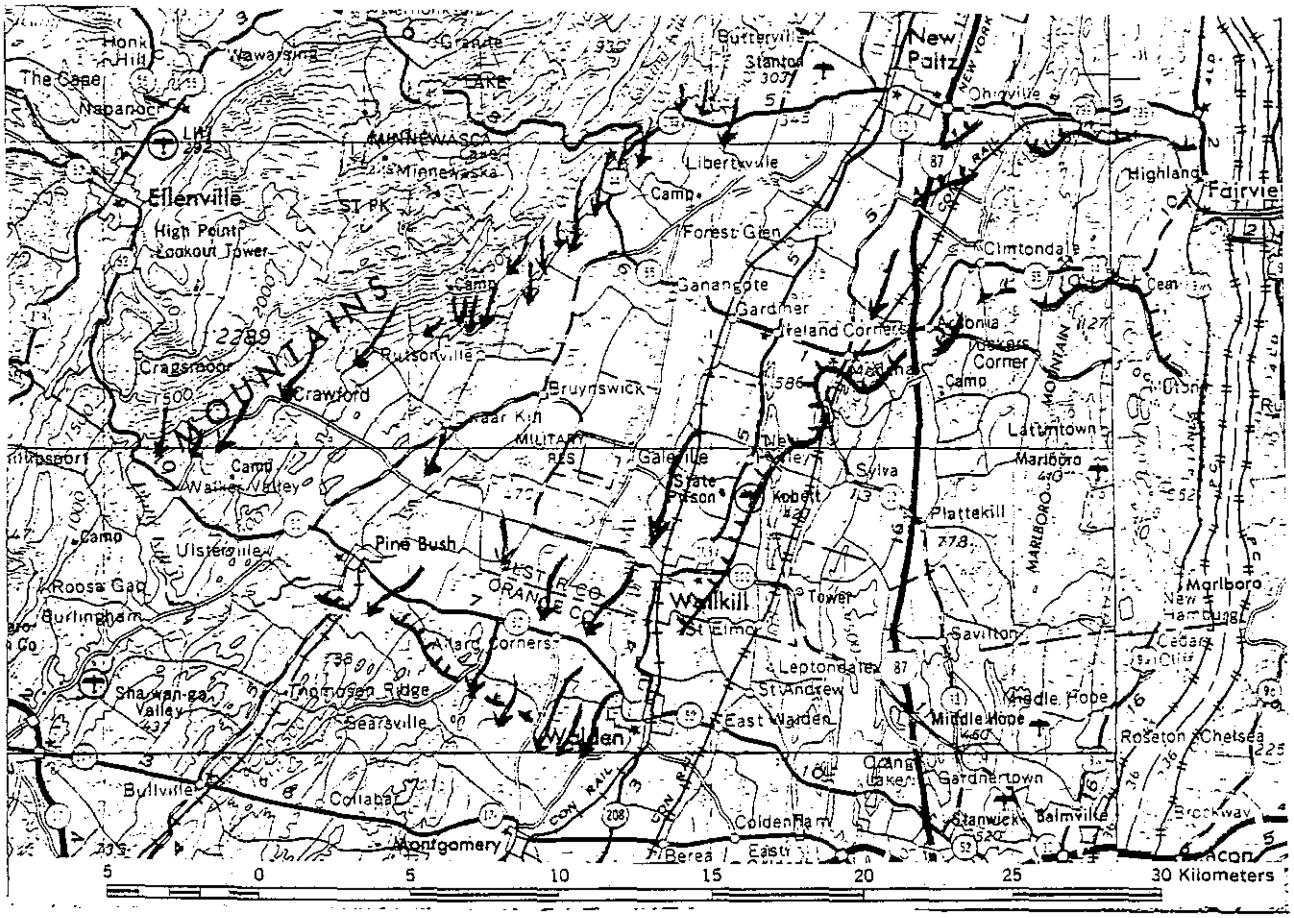

FIGURE 6. The Wallkill Moraine from Pine Bush on the west to the Milton delta of Lake Albany on the east. Arrows indicate eskers that fed sediment to both the outer ice margin and also the recessional position.

Otter Creek channel (Fig. 1), it opened a new eastward drainageway into Lake Albany in the Hudson River trench, via Moodna Creek. A new lake level of ±400 ft was established. The threshold must have been controlled by the drumlinized topography 3 km west of the village of Burnside. Both upstream and downstream of Burnside, there are extensive fluvial remnants aggraded to ±360 ft, thus precluding the obvious bedrock threshold at Burnside as a spillway for the higher 400 ft lake.

Sediments that have been interpreted as deltas and aggraded alluvial plains are present at ±400 ft from Sussex, New Jersey to the Pellets Island Moraine in the main Wallkill Valley. At the Wallkill Moraine, the lake spread westward into Shawangunk Kill and eastward into the headwaters of Otter Creek (Fig. 1). Outwash graded to ±400 ft comprises a large portion of the Wallkill Moraine. Extensive outwash aprons developed between the glacier and the eastern side of the Wallkill River valley (STOP 4) as far north as Clintonville Station and Ohioville. At its maximum extent, the lake was 70 km long, though it might be interpreted as a 40 km long southern lake and a 30 km long northern lake separated by a narrow straight at the Pellets Island Moraine, as shown in Figure 7.

In the unpublished 1970 manuscript report, Connally proposed the name Lake Dyson for this lake. We here propose formally the name Glacial Lake Dyson in honor of the late James Lindsey Dyson, eminent glacial geologist and chair of the Department of Geology at Lafayette College, for the proglacial lake impounded by the Woodfordian glacier as it stood at the Wallkill Moraine, as illustrated in Figure 7.

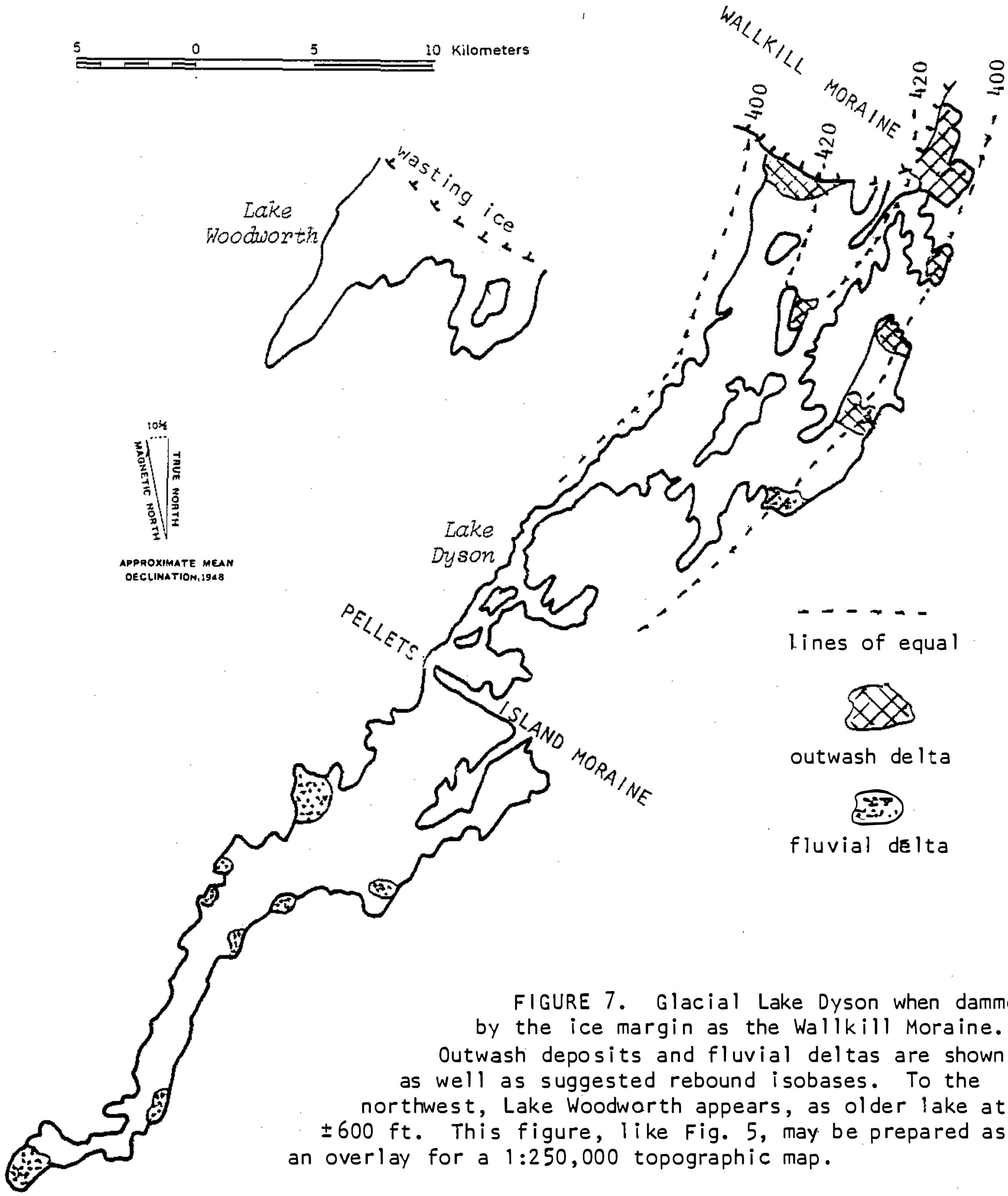

FIGURE 7. Glacial Lake Dyson when dammed by the ice margin as the Wallkill Moraine. Outwash deposits and fluvial deltas are shown, as well as suggested rebound isobases. To the northwest, Lake Woodworth appears, as older lake at ±600 ft. This figure, like Fig. 5, may be prepared as an overlay for a 1:250,000 topographic map.

Except for outwash at Montgomery, Campbell Hall, and Coldenham that is graded to ±420 ft, all peripheral features appear to be graded to elevations between 400 and 407 ft. These central features suggest a slight local rebound in the center of the Wallkill Valley following the demise of Lake Dyson. A south-to-north gradient of less than 20 ft can be inferred for the 70 km between Sussex and Clintonville Station.

There are three possibilities that might account for the consistent 400 ft elevation over a distance of 70 km. Perhaps most rebound, local and regional, had taken place prior to deposition of the Wallkill Moraine and only the very center of the valley remained out of adjustment. In this case, rebound in the Wallkill Valley was independent of that in the Hudson Valley to the east. Perhaps regional rebound had an east-to-west gradient and Lake Dyson developed parallel to the iso-adjustment lines. Both of the first two possibilities are consistent with the model developed for Lake Fairchild. The third possibility is that there was little or no rebound in the Wallkill Valley and that the deposits and elevations assigned to Lake Fairchild are misinterpreted.

Once the glacier retreated from the inner Wallkill Moraine, Lake Dyson evidently drained northward through decaying ice. An esker-fed outwash apron at ±340 ft, 1 km west of Libertyville, might represent a lowering lake level, but it is isolated and may just as well be attributed to subglacial deposition. We suggest that Lake Dyson drained quickly and completely, leaving the landscape clear for the Rosendale readvance and related events in the lower Wallkill Valley.

LATE-GLACIAL AND POSTGLACIAL ENVIRONMENTS

In 1964, Connally and Sirkin commenced a cooperative study that eventually encompassed all of the Hudson and Champlain Valleys, as well as adjacent portions of New Jersey and eastern Pennsylvania. They employed the integrated stratigraphic and geomorphic approach that had proven successful in local areas. The research plan was based on the pollen-stratigraphic work of Sirkin, ultimately published as Sirkin (1967a, 1967b, and 1971). Cadwell began working in New Jersey in 1974 and joined the lower Hudson Valley work in 1983.

Research Strategy

Connally and Sirkin chose to study and report only on closed peat bogs. Using the criteria mentioned earlier, they restricted their studies to produce a set of strictly controlled, radiocarbon dated, stratigraphic standard sections throughout the length of the Hudson and Champlain lowlands. The three restrictive criteria assured them *a priori* of the most complete, reliable stratigraphic records possible as standard sections. In each case, the stratigraphic record commenced very shortly after deglaciation. Two bog sites in the Wallkill Valley (New Hampton, Saddle), one adjacent to the Delaware River valley (Wigwam Run), and two in the Hudson Valley (Pine Log Camp, Eagle Hill Camp) all contribute to our understanding of the upper Wallkill Valley.

After preliminary probing to locate the thickest section of pollen-bearing sediment, each bog was sampled with sequential core segments (STOP 4) until coarse clastic sediment was encountered at the base. Each section consisted of an upper, dominantly organic (organic-rich) portion and a lower, dominantly inorganic (suborganic) portion.

The terms organic-rich and suborganic are applied to bog and lacustrine deposits ranging from pure peat to fine-grained clastics that incorporate varying amounts of organic material. Organic-rich deposits range from 90 percent organic matter in peat, to 10 percent for gradational sediments such as clay gyttja. Suborganic deposits generally contain ±5 percent organic matter as shown by ignition studies.

Dating

After completing sequential coring for each section, we retrieved between 5 and 12 additional core segments from the deepest organic-rich sediments. These samples were pooled and submitted to a laboratory for radiocarbon dating. They yielded absolute dates for specific depths at or near the base of each organic-rich portion. The dates are listed in Figure 8.

Using lithologic relationships in each core, we established correlations with specific geomorphic events near each site. We assigned ages to the herb (T), spruce (A), pine (B), and oak (C) pollen zones based on pollen zone correlations with established, dated sections. The radiocarbon date from the base of each organic-rich portion provided a basis for absolute age assignment.

To estimate the age of the base of each section, it is necessary to have a calculated sedimentation rate for suborganic sediments. Davis and Deevey (1964, p. 1293) calculated an average accumulation rate of .036 cm/yr for the basal 2 m of suborganic sediment from the south basin of Rogers Lake in south-central Connecticut. Connally and Sirkin (1986, p. 68) established an independent rate for Pine Log Camp bog near Glens Falls, New York. In 1968, they obtained a date of 12,400 yrs BP from the spruce zone between -8.00 and -7.85 m. In 1970, they obtained a date of 13,150 yrs BP from the base of the herb zone between -8.07 and -8.10 m. Thus, it took 750 years to accumulate the lowest 25 cm of suborganic herb zone sediment at Pine Log Camp bog. This yields an accumulation rate of .033 cm/yr. In estimating the basal age of each section, we now use .036 cm/yr to estimate the minimum time necessary to accumulate suborganic sediment and .033 cm/yr to estimate the maximum.

BOG SECTION	RADIOCARBON DATE in yrs BP	LAB NO.	DEPTH in m	POLLEN ZONE	AGE AT BASE in yrs BP
Pine Log Camp	12,400 ± 200	I-3199	7.85	A_{1-2}	
	13,150 ± 200	I-4986	8.10	T	13,150
Eagle Hill Camp	13,670 ± 170	SI-4082	10.25	T	16,020
New Hampton	12,850 ± 250	L-1157A	7.00	A_3	17,210
Wigwam Run	11,430 ± 300	W-2893	5.25	A_4	17,675
Saddle Bog	12,300 ± 300	W-2562	5.65	A_4	18,360

FIGURE 8. Dates and calculated ages for pollen-stratigraphic standard sections.

Pollen-Stratigraphic Standard Sections

Following is a brief description of each of the bog sections that has been used to help understand late-glacial and postglacial environments in the Wallkill Valley. The data was summarized in Figure 8.

New Hampton Bog. The standard section for the Wallkill Valley is New Hampton bog No. 1 (Connally and Sirkin, 1970, p. 3300-3303). Accumulation began when the Woodfordian glacier retreated north of Phillipsburg, Lake Fairchild drained, and the Lake Dyson level was established below the rim of the bog. The Wallkill Moraine is slightly younger.

The suborganic portion, from -8.50 to -7.00 m comprises the herb zone and the base of the spruce zone. The organic-rich portion, from -7.00 m to the top, comprises the upper spruce zone through the oak zone. The radiocarbon date at the base of the organic-rich portion is 12,850 yrs BP. We estimate a minimum age of 17,020 and a maximum of 17,400 yrs BP for the base of the suborganic portion. We posit an average age of 17,210 yrs BP for the establishment of the Wallkill Moraine and Glacial Lake Dyson.

Saddle Bog. In 1971, we extended our stratigraphic studies westward to Saddle Bog (Sirkin and Minard, 1972, p. D53). This is an upland bog on the side of Kittatiny Mountain, about 400 ft above the Wallkill Valley. It is bordered by a ridge that represents either a minor readvance or the first signs of upland drainage onto newly stagnating ice. Accumulation began after the glacier had receded from the Culvers Gap Moraine, or had begun downwasting there.

The suborganic portion, from -7.65 to -5.65 m, comprises the undifferentiated herb zone and the base of the spruce zone. The organic-rich portion, from -5.65 m to the top, comprises the upper spruce zone through the oak zone. The radiocarbon date at the base of the organic-rich portion is 12,300 yrs BP. We estimate a minimum age of 17,860 and a maximum of 18,360 yrs BP for the base of the suborganic portion. Because of the high quantity of non-arboreal pollen (NAP) throughout the core, and the corresponding lack of forest pollen (AP), during accumulation of the suborganic portion, we posit an age near the maximum 18,360 yrs BP for recession from the Culvers Gap Moraine.

Wigwam Run Bog. In 1974, the three of us extended our studies farther west to Wigwam Run bog (Sirkin, 1977, p. 210-212). The bog is in a kettle that is isolated by an inversion ridge west of the Delaware River and north of the Pennsylvania equivalent of Kittatiny Mountain. Accumulation began when the ice had receded from the terminal position of the Delaware-Minisink Lobe. Connally and others (1979) suggested that this was a post-Culvers Gap, but pre-Pellets Island, deposit.

The suborganic portion, from -7.40 to -5.25 m, comprises the herb zone and most of the spruce zone. The organic-rich portion, from -5.25 m to the top, comprises the top of the spruce zone through the oak zone. The radiocarbon date at the base of the organic-rich portion is 11,430 yrs BP. We estimate a minimum age of 17,400 and a maximum of 17,950 yrs BP for the base of the suborganic portion. We posit an average age of 17,675 yrs BP for deglaciation.

Eagle Hill Camp Bog. The standard section for the mid-Hudson Valley is Eagle Hill Camp bog (Connally and Sirkin, 1986, p. 64-69). Accumulation began when the Woodfordian glacier receded from the Red Hook Moraine that was emplaced east of the Hudson River during the Rosendale readvance.

The suborganic portion, from -11.15 to -10.40 m, comprises only the herb zone. The organic-rich portion, from -10.40 to -1.00 m comprises the herb through oak zones. The upper 1.00 m was too wet to sample. The radiocarbon date at -10.25 m is 13,670 yrs BP. We estimate an age of 13,890 yrs BP for the base of the organic-rich portion. We estimate a minimum age of 15,670 and a maximum of 16,070 yrs BP for the base of the suborganic portion. We posit an average age of 16,020 yrs BP for recession from the Rosendale readvance.

Pine Log Camp Bog. The standard section for the upper Hudson Valley is Pine Log Camp bog (Connally and Sirkin, 1971, p. 998-1003). Accumulation began when kettle ice in outwash from the Luzerne readvance had melted.

The suborganic portion, from -8.10 to -7.85 m, comprises the undifferentiated herb zone. The organic-rich portion, from -7.85 m to the top, comprises the spruce through oak zones. The radiocarbon date at the base of the organic-rich portion is 12,400 yrs BP. The radiocarbon date at the base of the suborganic portion is 13,150 yrs BP, closely approximating deglaciation in the upper Hudson Valley.

Other Radiocarbon Dates In The region

According to Sirkin (1977), the Woodfordian glacier reached a maximum stand on Long Island about 21,750 yrs BP and had begun receding north, up the Hudson Valley, by 21,200 yrs BP. However, the oldest finite radiocarbon dates for pollen stratigraphy are reported by Cotter (1983) for Francis Lake. The lake, while not conforming to our restrictive criteria, is located in northwestern New Jersey, about 20 km south of the Culvers Gap Moraine. Its relationship with the moraine is problematic because it is located between valley trains in Paulins Kill to the west and the Pequest River to the east, but clearly related to neither. Accumulation began following deposition of outwash that probably predates the Culvers Gap outwash in the other valleys by one "event". Dates of 18,570 and 18,390 yrs BP document the oldest herb pollen-bearing sediments in the northeastern U.S. and confirm the presence of tundra vegetation during Woodfordian deglaciation. The herb pollen zone is dated between 18,570 and 14,250 yrs BP and the spruce pollen zone between 14,250 and 11,250 yrs BP (Cotter and others, 1986, p. 42-47).

Cadwell obtained a date in the 19,000 year range from the base of a peat bog on Jenny Jump Mountain, near The Terminal Moraine, in north-central New Jersey. However, the pollen stratigraphy was not consistent with the 19,000 year age and the deposit remains under investigation. An early late-glacial date of 17,950 yrs BP was reported by Weiss (1971, Table 4). The dated sample was from "varved" silt and clay in the lower Hudson River channel, though the sediments are described as pollen-free.

SUMMARY

During late Wisconsinan time, the Woodfordian glacier reached a maximum stand on Long Island about 21,750 yrs BP. Recession was underway by

21,000 yrs BP. In the Wallkill Valley, the glacier advance farther south than Newton, New Jersey, and perhaps as far south as The Terminal Moraine. By 18,570 yrs BP the glacier had established a position south of Newton, shedding outwash past Francis Lake. Subsequently, the Culvers Gap Moraine was deposited. By 18,360 yrs BP, the ice of Kittatiny Mountain had retreated or downwasted and sediment began to accumulate in Saddle Bog. Thus, the age of the Culvers Gap Moraine probably is close to 18,500 yrs BP. By 18,360 yrs BP, tundra vegetation was established in the valley bottom and on the uplands.

As the Woodfordian glacier retreated northward, down the Wallkill Valley, Glacial Lake Fairchild was impounded between the Culvers Gap Moraine and the retreating ice margin, with an outlet at ±520 ft. By 17,675 yrs BP, the ice margin was well north of the Culvers Gap Moraine and by about 17,500 yrs BP, the Pellets Island Moraine was deposited. By 17,210 yrs BP, the glacier had retreated north of the Pellets Island Moraine, an eastward outlet drained the valley, and Glacial Lake Dyson was established at ±400 ft. Shortly thereafter, perhaps about 17,200 yrs BP, the Wallkill Moraine was deposited. When the ice margin resumed northward recession, Lake Dyson drained north into the wasting ice, leaving a large undrained depression to the south that filled with organic-rich sediment to become the Orange County mucklands.

The final event to affect the Wallkill Valley was the Rosendale readvance. That event was completed prior to 16,020 yrs BP and the entire valley was ice-free. Tundra vegetation occupied the valley throughout this interval. Spruce trees began to encroach in northern New Jersey by 14,250 yrs BP. The spruce forest migrated to the Pellets Island Moraine perhaps 13,700 years ago, to the mid-Hudson Valley about 13,000 years ago, and to the upper Hudson Valley by 12,400 yrs BP. Interestingly, pollen stratigraphy from Saddle Bog and Wigwam Run bog indicate that tundra vegetation still dominated the uplands until about 12,500 years ago, long after the spruce forest had migrated up the valley bottom. These relationships are summarized in Figure 9, below.

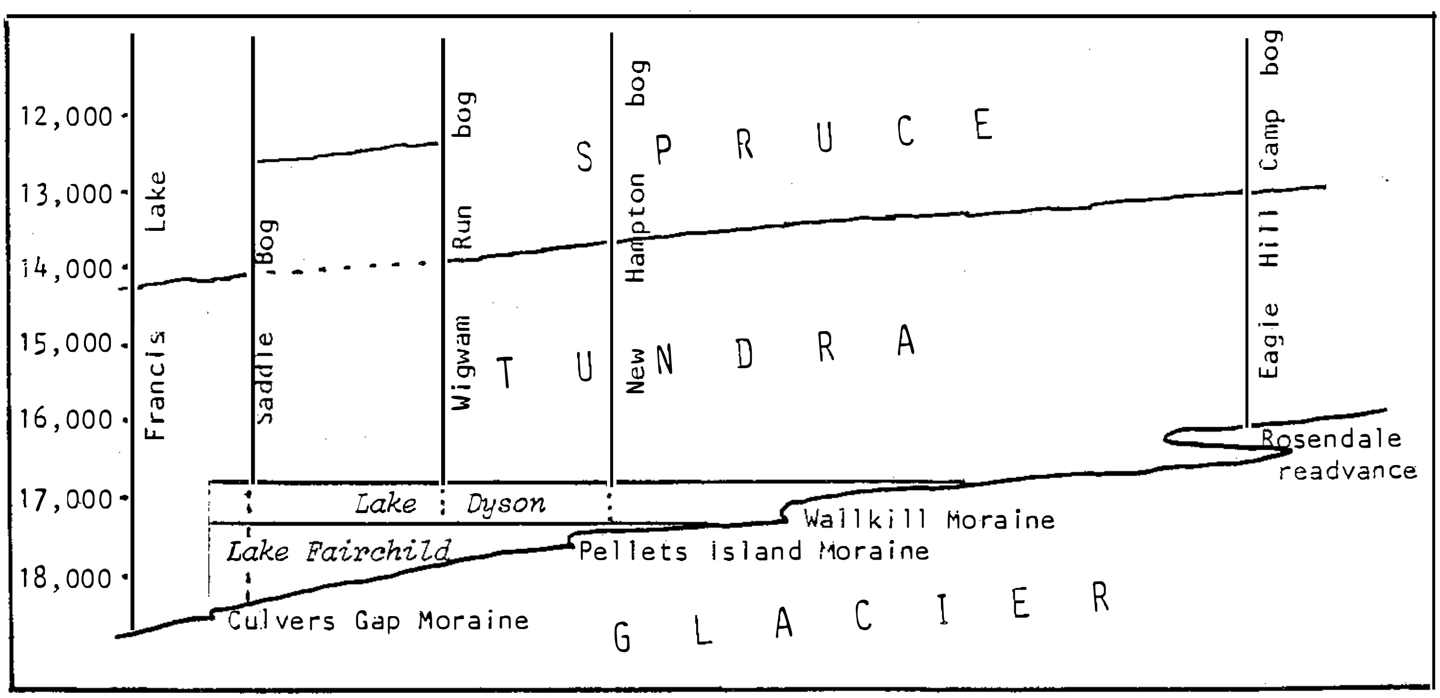

FIGURE 9. A space/time diagram showing the relationship between deglacial events and the migration of vegetation.

ACKNOWLEDGEMENTS

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ROAD LOG FOR DEGLACIAL HISTORY AND ENVIRONMENTS OF THE UPPER WALLKILL VALLEY

CUMULATIVE MILAGE	MILES FROM LAST POINT	ROUTE DESCRIPTION
		Begin trip, OCCC Parking Lot
0.0	0.0	Turn left (west onto Grandview Road)
0.1	0.1	STOP SIGN, Turn left (south onto Waywayanda Avenue)
1.1	1.0	Turn left (east onto Co. Route 78)
1.8	0.7	TRAFFIC LIGHT, Turn right (south onto Route 17M)
2.1	0.3	TRAFFIC LIGHT, Turn right (west onto Route 6)
3.9	1.8	Esker complex on left marking the Pellets Island Moraine in the Rutgers Creek valley
5.7	1.8	Turn left (south onto Route 284)
9.9	4.2	Dead ice sink on right; inwash is mined for sand and gravel
11.1	1.2	Village of Westtown
14.2	3.1	Continue left on Route 284; Village of Unionville
14.8	0.6	New Jersey State line
15.9	1.1	Lake Fairchild/Lake Dyson lake bottom at left, now the Wallkill River floodplain
20.6	4.7	The inversion ridge formerly identified as the "Sussex Moraine" is visible at 11:00 o'clock

22.0	1.4	TRAFFIC LIGHT, Turn left (east onto Route 23)
22.6	0.6	Bridge across Papakating Creek
23.4	0.8	TRAFFIC LIGHT, Turn left (north onto NJ Route 565)
24.1	0.7	Bridge across upper Wallkill River
24.9	0.8	Stay left (north on NJ Route 565)
25.4	0.5	STOP 1 at Landrud Road, RAIA Industries pit

STOP 1. INTERNAL STRUCTURE OF AN INVERSION RIDGE

NOTE! Permission to enter this property has been obtained for this trip ONLY! RAIA Industries, in Hamburg, New Jersey, probably will NOT grant permission to individuals.

Walk down the inversion ridge to the exposures in the active sand and gravel pit. This feature formerly was described as the "Sussex Moraine". What type of depositional environment can you infer from the sedimentary structures? What and where was the source for this sediment? Where was the glacier? In what condition was the ice?

As you travel to Stop 2, you will cross the upland drainage system that fed onto the ice to form this feature. Keep your eye peeled for this at mile 26.1.

	0.0	Continue north on NJ Route 565
26.0	0.6	Bear left (north onto NJ Route 667)
26.1	0.1	The head of the inversion ridge drainage channel; upland drainage system strewn with boulders to the right, dead ice and inversion ridges (then, stream channels in the ice) to the left.
27.7	1.6	Lake Fairchild/Lake Dyson lake bottom ahead, on left.
28.1	0.4	Stay left
29.2	1.1	Stay left
30.2	1.0	Continue straight ahead
30.5	0.3	Continue straight ahead
31.2	0.7	New York State line (Phew!)
33.7	2.5	Pochuck Creek valley to right at 2:00 o'clock
33.9	0.2	INTERSECTION, continue straight
34.2	0.3	Bridge across Pochuck Creek
34.7	0.5	Turn left (north onto C. Route 6); Village of Pine Island
		Orange County mucklands on both sides of the road for next 5 mi. This is the last remnant of Glacial Lakes Fairchild and Dyson.
38.7	4.0	A cluster of 5 drumlins forms "Big Island".
39.6	0.9	The Pellets Island Moraine crosses the Orange County mucklands.
41.2	1.6	Turn left (west onto Co. Route 42, "Cross Road")

41.7	0.5	Drive up the distal slope of the Pellets Island Moraine
42.4	0.7	Turn left (west onto Co. Route 37, "Maple Road")
42.6	0.2	Stop 2 at Cedar Swamp Road, Gurda Gardens pit

STOP 2. INTERNAL STRUCTURE OF THE PELLETS ISLAND MORAINE

NOTE! You must obtain permission from Gurda Gardens, Ltd. to enter this sand and gravel pit.

Walk into the sand pit on the west side of Cedar Swamp Road. This pit is about 5 km east of the type locality for the Pellets Island Moraine at Pellets Island. The old pit, visitied in 1967, is badly slumped today. What type of depositional environment can you infer from the sedimentary structures? What and where was the source for the sediment? Where was the glacier? In what condition was the ice? How does this deposit contrast with that at Stop 1?

	0.0	Return east on Co. Route 37
45.3	2.7	Bear right onto Maple Avenue Extension
46.0	0.7	Turn right (east onto Route 17)
46.5	0.5	EXIT 124B (north onto Route 207)
46.8	0.3	Turn right (continue north on Route 207)
47.6	0.8	TRAFFIC LIGHT (continue north on Route 207);Village of Goshen
50.0	2.4	Turn left (west onto Everett Road
51.3	1.3	STOP SIGN, Turn left (south onto Hill Road)
51.4	0.1	STOP SIGN, Turn right (west onto Scotchtown Road)
52.3	0.9	Bridge across Wallkill River
52.9	0.6	Turn right (north onto Stony Ford Road)
53.4	0.5	Bear right on Stony Ford Road
53.7	0.4	Turn left (west onto Stage Road)
54.3	0.6	STOP 3 at the Smiley Farm, dirt farm lane

STOP 3. SMILEY FARM ESKER(?)

NOTE! DO NOT EVEN ASK! This popular stop was included on the 1967 trip. Since then, the Smilev's have been deluged with visitors - most of whom have not even esked permission. For many reasons, the Smiley's DO NOT WANT PEOPLE VISITING THEIR WORKING FARM. DO NOT EVEN BOTHER TO ASK PERMISSION! We have obtained permission for this trip only -- and only because of our long association with the Smileys and with this stop.

Climb the hill on the east side of the dirt access road. Gather at the top of the hill, that is located about midway between the Pellets Island and Wallkill Moraines. See if you can come up with an explanation better than 1967's when the term "moulin kame" was bandied about.

	0.0	Return on Stage Road
54.9	0.6	Turn left (north onto Stony Ford Road
56.0	1.1	Turn right (following Stony Ford Road east)
56.3	0.3	Bridge across Wallkill River

57.3	1.0	STOP SIGN, Turn left (north onto Route 207)
58.0	0.7	Bear left immediately after underpass (north onto Route 416)
59.4	1.4	LUNCH STOP at Orange County Picnic Area.

LUNCH STOP. THOMAS BULL MEMORIAL PARK

This is the same lunch stop used in 1967. In May of 1967 it was so cold (it snowed in the Catskills) that Orange County thought we would cancel lunch and didn't even have the restrooms open for us. They are up the hill, on the left. If the weather is bad, we have reserved the picnic pavillion.

	0.0	Continue north on Route 416
62.3	2.9	Bear right (north onto Route 211)
63.7	1.4	STOP SIGN, Turn right (east onto Route 17K); Village of Montgomery
65.6	1.9	TRAFFIC LIGHT, Turn left (north onto Route 208)
68.3	2.7	STOP SIGN, (continue north on Route 208); Village of Walden
68.4	0.1	TRAFFIC LIGHT, Turn right (north with Route 208)
71.8	3.4	YIELD SIGN, Bear right (north with Route 208); Village of Waldkill
72.1	0.3	Wallkill Moraine visible on right from 1:00 o'clock to 4:00 o'clock.
72.7	0.6	Turn right (east onto Route 300); continue up the proximal slope of the Wallkill Moraine
73.6	0.9	Wallkill Moraine outwash apron deposited at the edge of Glacial Lake Dyson at ±400 ft.
73.9	0.3	STOP 4 at boggy area on the right side of the road

STOP 4. A ROADSIDE BOG

At this stop we will demonstrate the technique used in retrieving a core from a subsurface organic deposit. Although this site does not meet the restrictive criteria established by Connally and Sirkin, it is conveniently close to the road as a demonstration site. First we will probe the depth of the organic sediment and then retrieve a sequential core using a Davis Piston Corer. We also will discuss surface vegetation and the subsurface pollen stratigraphy of the Wallkill Valley.

	0.0	Continue east on Route 300
74.0	0.1	Turn left (north onto St. Andrews Road)
74.7	0.7	Turn left (west onto Reservoir Road)
75.3	0.6	Climb the distal slope of the Wallkill Moraine
75.5	0.2	STOP SIGN, Turn right (north onto Route 208)
75.9	0.4	Turn left (west onto Birch Road); the road climbs the distal slope of the "inner" Wallkill Moraine
77.0	1.1	Straight ahead onto paved road
77.3	0.3	Stay right at fork
77.7	0.4	STOP 5 at abandoned sand and gravel pit

STOP 5. ESKER FEEDER FOR THE WALLKILL MORAINE

CAUTION! You are on the property of the Wallkill State Prison. If you do not obtain permission before visiting this pit you may expect a visit from one or more armed officers.

There is little exposure left at this site that is about 4 km north of the inner Wallkill Moraine. Look about for clasts that you can identify as white Shawangunk conglomerate or orthoquartzite sandstone. Take in the splendid view of the bare rock cliffs of Shawangunk conglomerate -- these are made fameous by mountain climbers from all over the northeast. Why are these cliffs so fresh? Why is the escarpment to the south covered by a blanket of talus? What is the source of the sediment for this huge esker that flanks the bottom land channel of the Wallkill River?

Look for evidence of Shawangunk conglomerate as you travel to Stop 6.

	0.0	Turn right (north onto Sand Hill Road)
81.6	3.9	STOP SIGN, Village of Gardner; take a FULL left turn (west onto Main Street; Routes 44 & 55)
82.7	1.1	Bridge across Wallkill River
84.4	1.7	Continue straight (west on Routes 44 & 55)
85.4	1.0	The foot of a feeder esker appears on the left; the road parallels this esker up the hill to the base of the Shawangunk cliffs
86.5	1.1	STOP 6 parking lot of restauratnt

STOP 6. HEAD OF FEEDER ESKER AND VIEW OF SHAWANGUNK CLIFFS

This stop enables you to view the huge, block-sized clasts of Shawangunk conglomerate that were frozen in time as they were moving down into or onto the Woodfordian glacier. Evidently the hydraulic situation along the cliff was similar to a bergschrund on a cirque headwall. The plucking of these huge blocks resulted in a cleanly exposed rock face (or free face) once the glacier receded. Why don't you see these large blocks down in the valley? Why are these cliffs not covered with talus as are those to the south?

As you drive down into the valley, you will pass a second massive esker on your left. Look at the change in the size of the Shawangunk conglomerate lasts as you continue down to the valley floor.

	0.0	Turn right (east onto Route 299)
92.4	5.9	Bridge across Wallkill River
92.6	0.2	TRAFFIC LIGHT (contue straight up hill on Route 299; Route 208 joins from the right); Village of New Paltz
93.1	0.5	TRAFFIC LIGHT (continue east of Route 299; Route 32 leaves to right)
93.9	0.8	Overpass over NYS Thruway
94.0	0.1	New York State Thruway, New Paltz Interchange

END OF TRIP

NOTES

PLEISTOCENE GEOLOGY OF THE EASTERN, LOWER HUDSON VALLEY, NEW YORK

Les Sirkin
Department of Earth Sciences
Adelphi University
Garden City, NY 11530

Donald H. Cadwell
New York State Geological Survey
Room 3140
Albany, NY 12230

G. Gordon Connally
12 University Avenue
Buffalo, NY 14214

INTRODUCTION

A preliminary understanding of the style of glacial retreat, in the eastern lower Hudson Valley, has been gained through mapping the surficial deposits between the Long Island-Staten Island end moraines and the Hudson Highlands. The present model of late Pleistocene glaciation for the region includes 1) a lobate ice front, 2) two distinct glaciations separated by a warm interval, and 3) final deglaciation with the sequential development of recessional moraines at some ice margins and extensive proglacial lake and valley fill deposits. This model is used in conjunction with field evidence to reconstruct the glacial history of the region. This field guide provides the basic framework for the glacial history and illustrates typical deposits and morphologic features. This article complements that of Connally, Sirkin, and Cadwell (this volume). For convenience, certain sections and descriptions are not repeated in both articles, but are cross-referenced.

PREVIOUS WORK

This model of late Pleistocene glaciation and deglaciation was developed during field examination of glacial deposits of the Lower Hudson Valley, Long Island, and Block Island, Rhode Island (Sirkin, 1982, 1986). A Late Wisconsinan Hudson Lobe probably existed during advance and retreat through this region (Sirkin, 1982). Two distinct drifts are represented in end moraines of western Long Island, (Sirkin and Mills, 1975; Sirkin, 1982, 1986).

The chronology of late Pleistocene glaciation and rate of deglaciation of the Hudson Lobe is derived from glacial ice-margin positions, radiocarbon ages and pollen stratigraphy (Connally and Sirkin, 1986). Connally and Sirkin, 1986, also trace the northward wasting of the glacier margin in the mid-Hudson Valley using moraines north of the Hudson Highlands in Dutchess and Columbia Counties. The deglacial chronology and pollen stratigraphy for the mid-Hudson Valley, west of the Hudson River, is summarized in Connally and others (1989; this volume).

BEDROCK GEOLOGY

Manhattan, Bronx, and Westchester Counties are underlain by rock formations of the Appalachian Fold Belt. Here the structure is similar to that of the Appalachian Blue Ridge with northeast-southwest trends and folds recumbently overturned to the northwest. Rock units include the Fordham Gneiss (generally associated with Grenville age tectonics and metamorphism), and the Lowerre Quartzite, Inwood Marble and Manhattan Schist of Cambro-Ordovician age. These units, the New York Group, were deformed and metamorphosed during the Taconic Orogeny. In southern Westchester County and adjacent Connecticut an additional sequence of early Paleozoic metamorphics, the Hartland Formation and the Harrison Gneiss, represent an accretionary facies. The Fordham and Hartland units exhibit structures of partial melting which give their outcrops a dramatically swirled appearance. The Fordham Gneiss is the most resistant rock unit in Westchester County and forms the highest ridges. The younger metamorphics of the New York Group are less resistant and support low ridges, or as the Inwood Marble, underlie the valleys. The Inwood is occasionally found folded against the resistant Fordham ridges and may directly underlie glacial deposits along the valley walls. The Lowerre Quartzite, originally a coarse clastic unit of fluvial origin, is discontinuously preserved along the trend of the Fordham. Vein quartz and pegmatite dikes of varying thickness cross-cut the Lowerre Quartzite and Fordham Gneiss.

In northern Westchester County (near Peekskill) the New York Group is intruded by an early Paleozoic basic rock unit known as the Cortlandt Complex.

GLACIAL DEPOSITS

Lower Till. In general the drumlin-shaped hills that form the uplands of southeastern New York are capped by a layer of glacial till up to 10 m thick. This appears to be a basal till deposited during the earlier of two glaciations. This lower till forms the north-south trending drumlins in Westchester County. It is usually medium to dark brown, blocky and oxidized on the blocky surfaces. Uplands formed on the resistant ridges of Fordham Gneiss and Cortlandt Complex have only thin patchy deposits of till. The lower till may be correlative with the "drumlin till" of southern New England.

Upper Till. The upper till appears poorly sorted, with occasional crude stratification and is interpreted to be a recessional deposit, such as ablation or meltout till. This unit is generally 1-3 m thick and may overlie the lower till. It may also occur as a meltout or flow till in ice-contact deposits. The upper till represents wasting of the Woodfordian glacier.

Many glacial erratics in the study area were quarried by the glacier from the rectangularly jointed bedrock units. Some erratics retain a rectangular shape and presumably were not transported far. Striations on bedrock surfaces generally trend north-northwest to south-southeast. Glacier flow from the north-northwest would account for Palisades basalt and Brunswick sandstone erratics of Hudson Lobe deposits in southern Westchester and western Long Island.

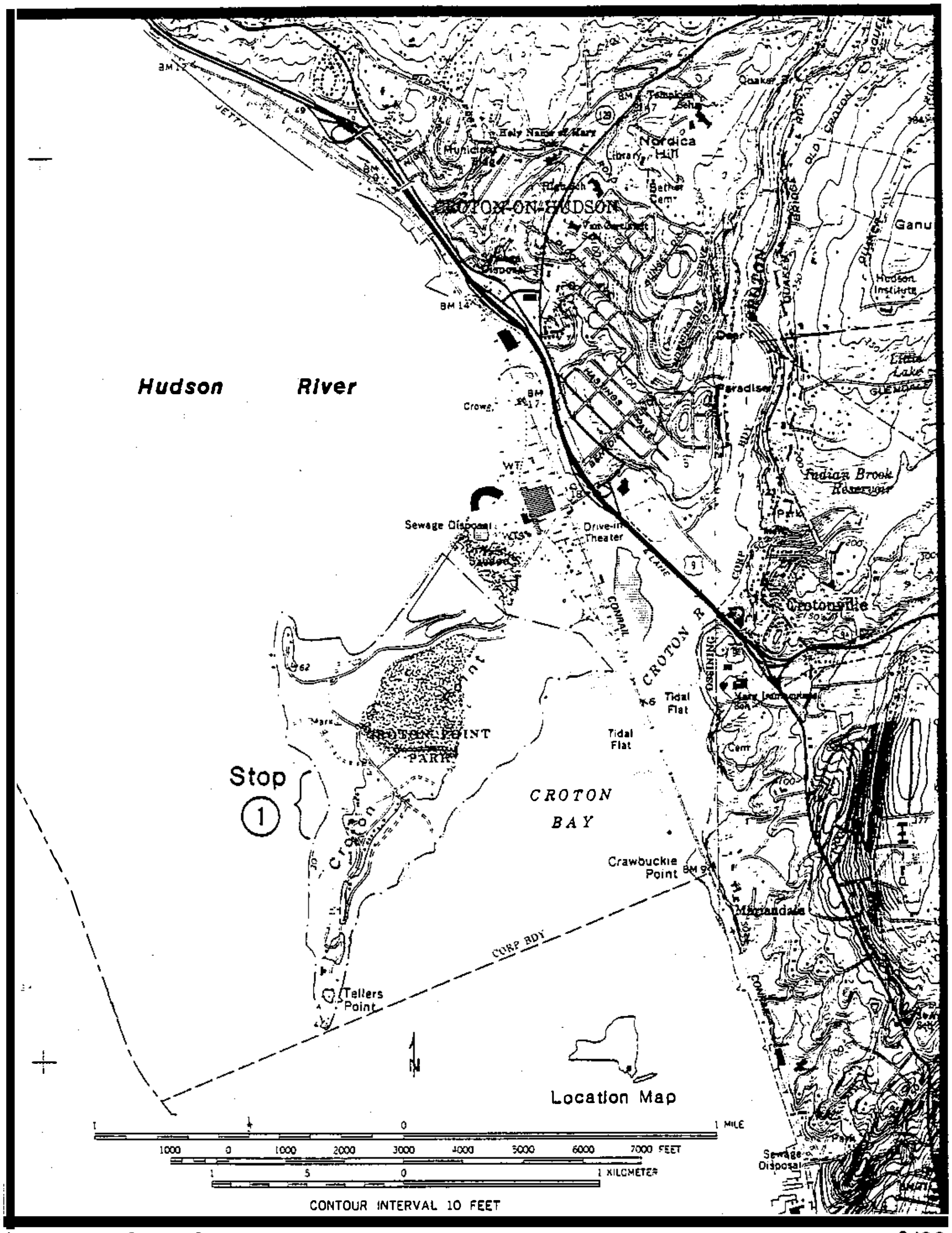
Stratified Drift. Outwash deposits are generally concentrated in main valleys, although occasionally found on the uplands too. Kame terraces occur in the Saw Mill, Grassy Sprain and Bronx River valleys, as well as in south-trending valleys in the Mamaroneck Quadrangle. A small kame is located against bedrock in the Kings Bridge area of the south Bronx.

An east-southeast trending valley with sand and gravel kame deposits extends from Tarrytown to Elmsford and White Plains. The White Plains part of this valley is the confluence of three north-south valleys intersecting the east-west valley and forming three distinct umlaufbergs.

Outwash and kame terraces fill many of the valleys in mid- and northern Westchester County, particularly in the Mianus River and Chappaqua Brook valleys. Kame deltas protrude laterally and southward into the Mount Kisco-Bedford Hills-Katonah valley.

The more prominent man-made lakes in Westchester County belong to the New York City reservoir system. The water bodies in the Amawalk and Titicus Creeks have drowned heads of outwash, kame and delta deposits, while active sand mining for new housing and industrial park sites have eliminated others. A kame delta located in apparent safety at Memorial Park in Bedford Hills vanished to the drag line in preparation for a condominium site, leaving only a smear of lake clays. Some heads of outwash are located east of Bedford, east of Jay High School, southeast of Somers, at Granite Springs, just north of Yorktown Heights, north of Katonah, and north of New Croton Reservoir.

The Croton Reservoir masks a possible stillstand of the ice margin. If real, this stillstand occurred when the retreating glacier formed the Croton Point Moraine (new name), and the Croton Point Delta (Figure 1, STOP 1). Similarly, the east-west valley trends at Yorktown Heights, Crompond, Jefferson Valley, Mohegan Lake, Lake Peekskill and Peekskill may denote ice margin positions. The valley of Peekskill Hollow Creek, has numerous ice contact kame deposits, one is a kame delta capped with coarse



New York State Geological Survey 6/89

Figure 1. Portion of the Haverstraw 7.5 minute U.S.G.S. topographic quadrangle illustrating STOP 1, Croton Point Park and vicinity.

gravel (Figure 2; and Fieldtrip Stop 3).

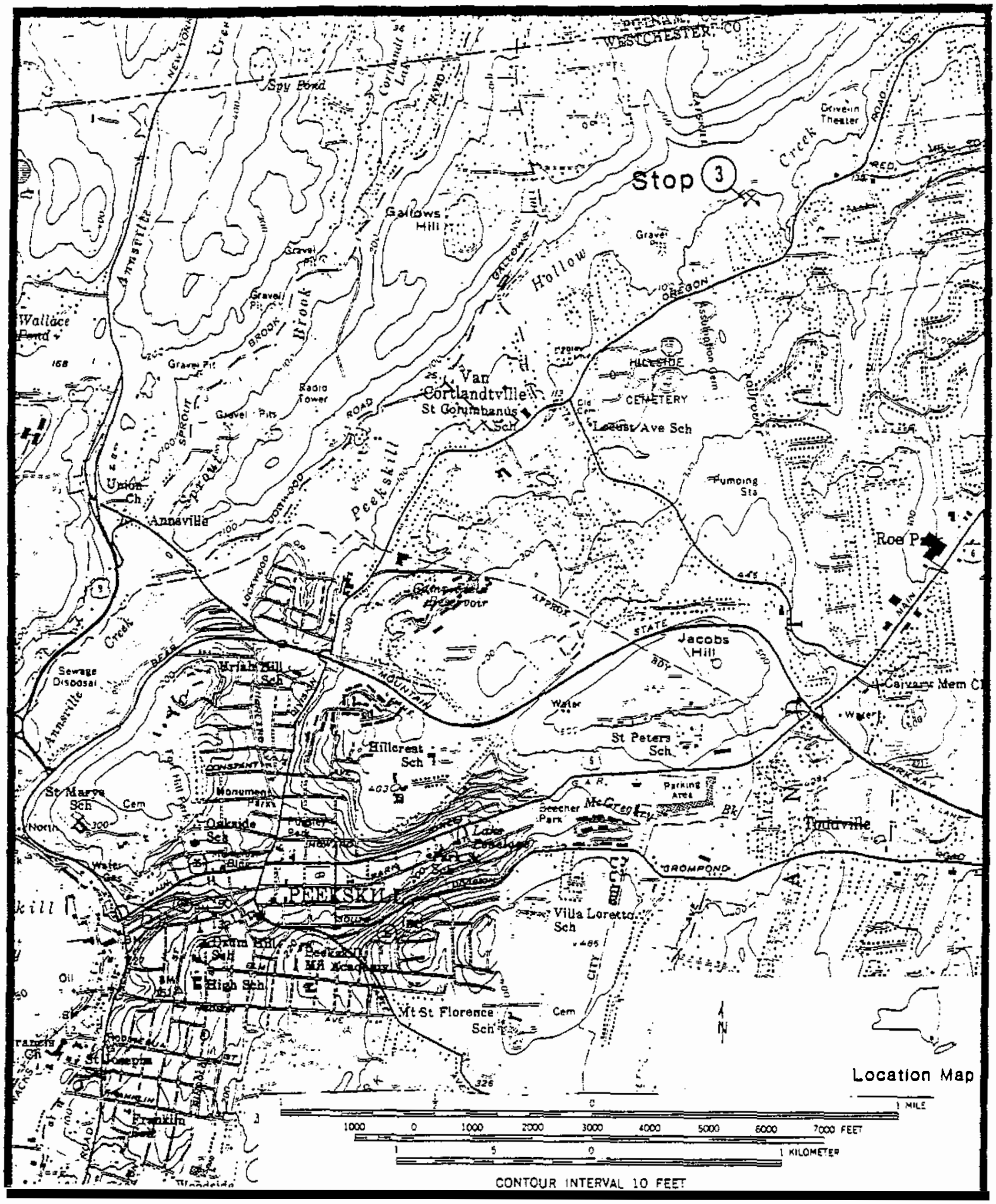
LACUSTRINE DEPOSITS

Lake-sand deposits are located adjacent to the Hudson River between Spuyten Duyvil and Irvington at an elevation of 100 ft. Artificial fill masks these deposits in Manhattan to the south and between Irvington and Ossining to the north. Lake sands are also preserved in Van Cortlandt Park, and the Saw Mill and Bronx River valleys. The Van Cortlandt Park deposits were formed during the development of Lake Hudson, while the Bronx River valley deposits may have formed during a northward extension of Lake Flushing. Glacial Lake Hudson developed between the Terminal Moraine and the retreating Woodfordian glacier margin and Lake Flushing formed in the position of the present Long Island Sound behind the end moraines on Long Island.

The best evidence of Lake Hudson in northern Westchester County is Croton Point, a large delta that juts into Haverstraw Bay (Figure 1.) This bay, just south of the Hudson Highlands, is a widened and overdeepened part of the Hudson estuary. Some of the glacial deposits can still be seen even though much of the delta has been eroded, mined for the brick-making industry, or covered by a landfill. Stratified drift is exposed locally on the hillslopes adjacent to the railroad. Grain size analyses of these deposits show fining toward the delta and the sediments may represent the topsets of the delta. The west end of Croton Point is composed of ablation till overlying stratified drift. Here, erosion of these deposits has left a lag of boulders, dominantly composed of Cortlandt, Palisades, Brunswick, and Highlands lithologies as a beach. Lake clays are also exposed in the cliffs along the west side of the Point a few feet above river level and overlying glacial drift. The clays appear to be rhythmites possibly deformed during initial glacial retreat from the Croton Point ice margin. Exposures of lake clay occur in the valley west of Harmon, near the mouth of Furnace Brook south of Crugers and in the valley just to the northeast of Crugers.

ICE MARGINS

The Sands Point Moraine of northwestern Long Island extends westward from the Sands Point Delta to the College Point Delta in northern Queens between Flushing Bay and the East River. This ice margin may continue westward along the east-west valley south of Steinway and then cross the East River to Manhattan. Lack of exposures of ice contact deposits in Manhattan's concrete massif, however, makes tracing this margin westward a mute effort. Only a small patch of outwash in northwestern Central Park and another at Mount Morris Park, both somewhat north of the projected Manhattan landfall, could mark the ice stand.



New York State Geological Survey 6/89

Figure 2. Portion of the Peekskill 7.5 minute U.S.G.S. topographic quadrangle illustrating STOP 3, Peekskill Hollow Creek and vicinity.

Similarly, the City Island Moraine, including Hart Island to the east, may extend westward to the south Bronx where thick till was exposed in a construction site on Castle Hill Avenue in Unionport. While this till could not be differentiated from lower till, the moraine might extend westward across the south Bronx and be correlated with two small till exposures in cemeteries near Trinity Church in northern Manhattan. It is apparent that the evidence for these ice margins west of the nominative moraines becomes highly speculative in the "high rise" terrane of New York City.

North of the City Island Moraine, there are no continuous features that permit the identification of moraines. Heads of outwash, kame terraces and kame deltas offer, however, tantalizing lures for speculation of ice margins in the Elmsford-White Plains Valley; the Bedford Hills-Mount Kisco Valley; and the Bedford-Mianus River Valley.

In the region of this field trip the most prominent ice margin is the Croton Point Moraine at Croton Point (Figure 1). This moraine denotes a position within the Hudson Valley and may possibly be extended eastward into the Croton Reservoir Valley. The Croton Point and Peekskill deltas substantiate the existence of Lake Hudson with a water plane elevation of 100 ft. The valley at Yorktown Heights forms another tempting location for an margin, but with only an outwash head as control. The ice contact kames in Peekskill Hollow possibly mark an ice margin in northern Westchester.

The Woodfordian glacier continued its retreat and deposited the Shenandoah Moraine on the north flank of the Hudson Highlands. Much of the glacial meltwater flowed directly into the expanding Glacial Lake Hudson, but during the early stages some meltwater flowed across the Highlands where sediments completely filled the upper Clove Creek valley (STOP 4).

ACKNOWLEDGEMENTS

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ROAD LOG FOR THE PLEISTOCENE GEOLOGY OF THE EASTERN, LOWER HUDSON VALLEY, NEW YORK

The Road log for this trip begins in Westchester County at the east side of the Bear Mountain Bridge. Figure 3 is a generalized County street map with the locations of Fieldtrip STOPS 1 and 3.

CUMULATIVE MILEAGE	MILES FROM LAST POINT	ROUTE DESCRIPTION
0.0	0.0	East side of the Bear Mountain Bridge, Route 6 and 202; proceed south.
3.8	3.8	Junction Route 9 south, proceed on Route 9.
12.4	8.6	Exit Rt. 9 to Croton Point Avenue
12.5	0.1	Right on Croton Point Avenue
12.6	0.1	Traffic light, proceed straight onto one-lane bridge
13.8	1.2	Stay right after bridge to Croton Point Park; park at the west end of the large parking lot.

STOP 1. CROTON POINT PARK. Walk westward along the beach about 0.5 mi. to exposures of till and clay beds. Discussion will center on the evidence for the Croton Moraine and the Croton Point Delta (see Figure 1 in text). Note the lithologies of the boulders.

Return to cars.

15.3	1.5	Return to Rt. 9 north bound ramp. Proceed north on Rt.9

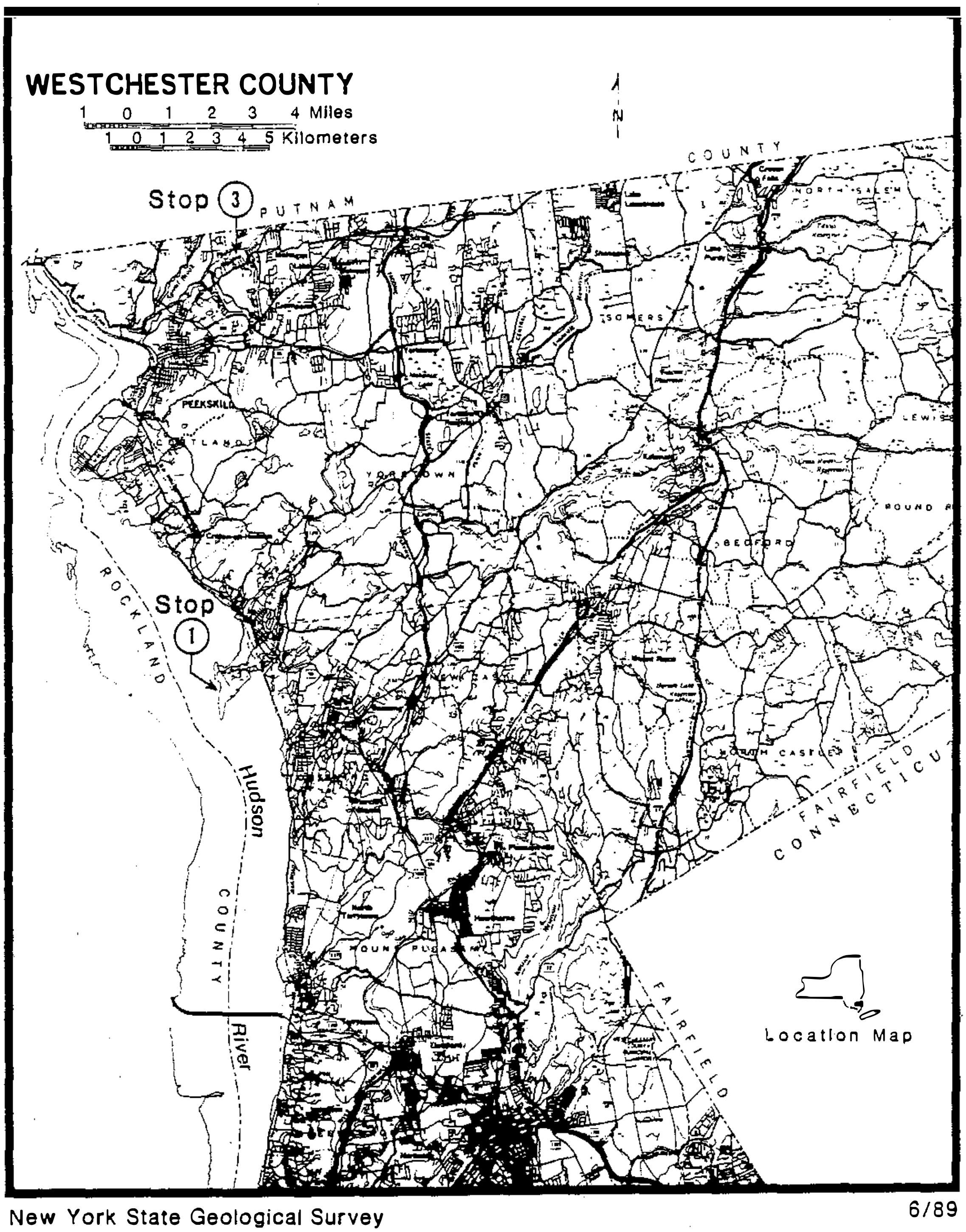

Figure 3. Generalized street map of northern Westchester County with the location of Fieldtrip STOPS 1 and 3.

18.7	3.4	Exit Rt. 9 to 9A at Montrose exit
19.2	0.5	Follow 9A southbound to the south side of the bridge over the railroad.

<u>STOP 2.</u> <u>PEEKSKILL HOLLOW CREEK: CLAY BEDS</u>. Park at the entrance to the new railroad station construction site. Walk northeastward to stream bed. Examination and discussion of clay beds exposed in the stream banks.

Return to the cars.

19.7	0.5	Return to Rt. 9 north
24.8	5.1	Follow Rt 9 to the Bear Mountain Parkway, exit to eastbound lane.
26.4	1.6	Exit at Division Road; turn left on Oregon Road, the northeast continuation of Division Rd.
28.0	1.6	Entrance to Dam-Fino Construction Company on left.

<u>STOP 3.</u> <u>PEEKSKILL HOLLOW CREEK: OUTWASH AND KAME DEPOSITS.</u> Walk along the northbound lane in the yard, bear right at fork between trailer home and concrete building, follow lane to gravel pit. This pit has been excavated into one of the many outwash and kame deposits in Peekskill Hollow. Some kames show ice contact features. At this site a kame delta has been overtopped by flood gravels suggesting a change from deltaic to high energy fluvial deposition. Return to cars and proceed south on Division St.

Return to cars.

29.6	1.6	Junction Bear Mountain Parkway, turn right, proceed west.
29.8	0.2	Bear right toward Route 9
30.5	0.7	Junction Route 9, proceed north
32.0	1.5	Enter Putnam Co.
42.1	10.1	Enter Dutchess Co.
42.4	0.3	Enter sand and gravel operation on left

<u>STOP 4.</u> <u>KAME DELTA.</u> Proceed into gravel pit. This is a large kame delta complex constructed by meltwater flow through the Highlands during development of the Shenandoah Moraine.

End of field trip. Return to Route 9.

THE STRATIGRAPHIC RELEVANCE AND ARCHAEOLOGICAL POTENTIAL OF THE CHERT-BEARING CARBONATES WITHIN THE KITTATINNY SUPERGROUP

PHILIP C. LA PORTA
Queens College and Department of Earth and Environmental Sciences
C.U.N.Y.

INTRODUCTION

All of the formations to be visited on this field trip are Cambro-Ordovician in age and were deposited in very shallow waters (shelf settings) (Figure 1). A few of the formations contain shaley members but are almost entirely composed of dolomites and lesser amounts of limestone. The Cambrian formations are remarkably persistent both laterally and vertically and this leads to extensive facies continuities, covering distances as great at 50 to 100 miles (Read, 1985). The Lower Ordovician formations (capped by the Knox Unconformity; Sloss, 1963) have a more heterogeneous character, still with the same lateral persistence, but with considerably greater variations in facies composition.

The Cambrian cherts appear to be facies-controlled (porous oolitic layers, sandy layers, etc.) except for the top of the Allentown Formation. The upper Allentown appears to be a surface of unconformity, but is not marked by deep incisions, just shallow channel cuts. The zone of chertification follows this surface and occurs as a series of complex rubble zones, and silicification along joints.

The cherts within the Rickenbach Formation (Lower Ordovician) are facies-controlled and are widely persistent forming a distinct marker horizon. The lower member exhibits an intrastratal breccia, but there is no evidence of an associated unconformity. The Epler Formation contains chert which is facies-controlled, but the most obvious varieties of chert are associated with paleokarst features. Some of the paleokarst features were filled by argillite which has also suffered later silicification. The Epler Formation is overlain by the Ontelaunee, a fossiliferous carbonate, a good portion of which has been strongly dolomitized. Chert is ubiquitous to both members of the Ontelaunee Formation and the Harmonyvale (uppermost member) is strongly cut by erosion features. This eroision surface is very variable and in some places has incised all the way down into the Rickenbach Formation.

The great variety and volume of chert extant within the chert-bearing carbonates of the Kittatinny Supergroup has been known for some time (Markewicz, 1967). The use of chert beds as stratigraphic marker horizons or key beds was first implemented in this region by Markewicz (1967) and Markewicz and Dalton (1973). Their formation and member level subdivision (Fig. 1) will be used for the field trip. The chert appears to be restricted to specific units within members of the formations of the Kittatinny Supergroup. The chert-bearing facies are laterally persistent for substantial distances. (The host rock for the silica are often highly porous beds, unconformities and paleokarst fractures.) An extension of the chert stratigraphy has been the discovery, over the last two decades, of more than 200 small and large sites of prehistoric mining operations sites. The chert stratigraphy is the focus of this field trip and has been specifically designed to also aid the archaeologist conducting provenance studies in the field.

PREVIOUS WORK

Markewicz (1967) and Markewicz and Dalton (1973) were the first to use the lower chert-bearing unit of the Hope Member of the Rickenbach Formation as a stratigraphic marker horizon.The presence of the Hope Member has since become a stratigraphic aid, as a key bed, for field workers in the complex fold and thrust belt. The refined member-level stratigraphy for the Kittatinny Supergroup (Fig. 1) was presented by Markewicz and Dalton (1973, 1974, 1978). The member level subdivisions and stratigraphic section descriptions of Markewicz and Dalton (1977), represented the next logical step towards understanding the depositional environments in a region where both detailed lithostratigraphy and invertebrate fossils are rare. More recently, further refinement of the members into distinct stratigraphic units has been compiled during the mapping of the Hamburg Quadrangle by Canace (1988). To date, most depositional and paleoecological interpretations are drawn from detailed studies in contiguous states. Presently, the tectonic map for the Great Valley Sequence is being assembled by Herman and Monteverde (pers.comm. 1989).

Structural and stratigraphic studies published in recent years (Offield, 1967, Baker and Buddington, 1970) mention but do not discuss the chert within the Beekmantown Group. Fortunately, structural and stratigraphic studies conducted in contiguous states have helped to establish the chert stratigraphy. The studies of Wherry (1909) and Weller (1900) Wilson (1962) both document the stratigraphic position of chert as beds and nodules. The highly refined stratigraphic work of Zadnik and Carozzi (1963) within the Carpentersville, N.J. section is also directly applicable to studies further north.

The stratigraphic work of Hobson (1963) has aided the present research. Hobson noted the positions of chert-bearing units within the Cambrian System, within the base of the Rickenbach Formation, at the base as well as higher in the Epler Formation, and most particularly within the Ontelaunee Formation.

Professional archaeology within the Wallkill River Valley in New Jersey, is almost non-existent. At the turn of the century, Schrabisch (1909) conducted a site survey in Sussex, N.J. which noted the locations of two potential prehistoric chert quarries. The first was at Wildcat Rock, between Hamburg and Franklin, N.J. where it was said that blue flint was abundant in the fields. The second was at the base of the Wallkill Pond, in the town of Franklin, N.J. There was also a small excavation at the Hamburg Rockshelter by Philhower (personal collected notes and letters of Louis M. Haggerty 1930-1979). During the WPA (Works Progress Administration) days, Cross (1940) excavated a rock shelter at the Todd Estate which produced a trait list of prehistoric implements. Ritchie (1965) noted the great concentration of fluted projectile points in the vicinity of Pine Island, N.Y., and made mention of the archaeological potential of the valley. The current study efforts of the Orange County Chapter of the New York State Archaeological Society have still not been synthesized into any specific research design. To date, there is no cultural stratigraphy for the Wallkill River Valley in New Jersey, but there are countless stone tool collections in cigar boxes and tin cans, tucked away in garages and stored on work shelves awaiting analysis, and mute testimony to the potential wealth of information that awaits the researcher.

THE FIELD TRIP AREA

The field trip area lies within the rocks of the Great Valley Sequence. The focus of the present trip is the carbonates within the Beekmantown Group and the stratigraphically lower Cambrian Age formations. The last field stop occurs along the southwest flowing Paulins Kill. The area is bounded on the east by Precambrian igneous and meta-sedimentary rocks, and to the west by the Martinsburg Formation. A belt of Grenville-Age marble underlies the eastern portion of the valley bottom. The carbonates to be discussed and observed occupy the NE-SW trending axis of the Wallkill River Valley.

THE ARCHAEOLOGICAL APPLICATION OF THE CHERT STRATIGRAPHY

A major outcome from this writer's chert stratigraphy studies in the area has been the creation of a well structured framework for lithic provenance studies in archaeology. The decades of the sixties and seventies produced only a handful of prehistoric quarry sites in the field trip area. As the chert stratigraphy was refined the number of prehistoric quarry discoveries grew. The period between 1979 and 1984 witnessed the discovery of a few dozen small quarries. The last five years of work have had the cooperation of Frank Markewicz and most recently several members of the New Jersey Geological Survey. The member-level subdivision of the Kittatiny Supergroup combined with the more recent chert stratigraphy has led to the discovery of nearly 150 new quarries of various types. A full spectrum of new problems now confronts us. The quarries are currently being categorized into large open cut quarries, such as the location along the Paulins Kill, smaller outcrop quarries, screes, workshops, exploration pits, conical shafts, and failed explorations. Furthermore, the quarries cluster into districts each of which is associated with a complex of open-air sites. Most recently, a number of outcrop quarries have been discovered, which occur in the centers of large open air sites. Many of these quarries appear in woodlots between furrowed fields. The chert stratigraphy has allowed the writer to move quickly over the landscape, marking member and formation contacts, then proceeding along strike to discover new quarries. The method allows one to gather both geological and archaeological data or focus more specifically on one particular task, for example attitude measurements using the chert beds where bedding is difficult to discern.

The knowledge of the exact location of prehistoric lithic resources should allow the archaeologist to locate the place of origin of chert. The outcrop to outcrop study has led to the creation of a chert classification

	Formation Name Used on N.J. Geol. Map	Formations Recognized by H. B. Kummel and Others	Formations Recognized by A. A. Drake and F. J. Markewicz	Current Stratigraphy As Used by F. J. Markewicz and R. F. Dalton		
LOWER ORDOVICIAN	KITTATINNY LIMESTONE	Beekmantown	Epler (Oe)	Ontelaunee Formation	Harmonyvale mbr. OO_2	
					Beaver Run mbr. OO_1	
				Epler Formation	Lafayette mbr. Oe_3	
					Big Springs mbr. Oe_2	
					Branchville mbr. Oe_1	
			Rickenbach (Or)	Rickenbach Formation	Hope mbr. Or_2	Crooked Swamp Dolomite Facies Or_3
					Lower mbr. Or_1	
CAMBRIAN		Allentown	Allentown (Ca)	Allentown Formation	Upper mbr. Ca_2	
					Limeport mbr. Ca_1	
		Tomstown	Leithsville (Cl)	Leithsville Formation	Walkill mbr. Cl_3	
					Hamburg mbr. Cl_2	
					Califon mbr. Cl_1	

Fig. 1 Correlation Chart for the Beekmantown Group and Cambrian System borrowed from Markewitz and Dalton, 1977. The figure depicts the more recent member level stratigraphy employed in this research.

which was designed with the archaeologist in mind. The use of the stratigraphy requires no special training or analytical procedure. The following is meant to be purely descriptive which also enforces the beliefs of the writer, that what is needed first in archaeological geology is not analytical techniques, but solid field data. The research documents the inter-outcrop variation of the chert within a single member. This aids in focusing the provenance study to a specific outcrop. Many of the chert-bearing members are laterally persistent for miles. To describe a chert as originating in the Leithsville Formation is useless as the Leithsville Formation occurs in New York, Pennsylvania and New Jersey. This has been the downfall of most provenance studies, along with the lack of field data and the heavy reliance on analytical methods. An example of the refinement of the study is as follows: an open air site in the Vernon Valley, near DeKay's Hamlet produces an assemblage of Early Archaic stone tools. The diagnostic artifacts appear to have been manufactured from the distinctive blue-gray brecciated and laminated chert of the Upper Allentown member which occurs as an outcrop quarry at the site. Along with the diagnostic artifacts are two large cores of which one has been heat treated. Both cores are manufactured from the chert pods within the Wallkill Member of the Leithsville Formation. This portion of the Vernon Valley lacks outcrops of the Wallkill member. The microfossil residue, closely spaced unsealed fractures, and the large size of the core aids in fingerprinting the core as having originated at the outcrop quarry at the Rudinski Farm, some 3 miles to the west. The smaller heat-treated core shows concentric banding, glassy to waxy luster after heat treatment, and is roughly fist size. Pods of this size, with concentric algal banding are common at a small outcrop quarry along Lake Road in Hamburg, 5 miles to the southwest. The glassy texture is the result of heat-treatment. Both specimens are pods, but that is the common characteristic of the Wallkill member chert at all locations. The other characteristics aid in distinguishing quarries within the same member. Each chert type possesses characteristics which are common to all specimens within that member. The other characteristics vary and aid to distinguish explicit quarries.

Oftentimes, a number of quarries will occur in a small geographic area. In addition, some stratigraphic members such as the Branchville Member of the Epler Formation offer a wide assortment of potential lithic resources. Because of this close association, entire assemblages can be accurately traced back to a small ravine or a complex of small outcrop quarries. Such is the case at Haven's Estate, where quarries occur in the Upper Allentown member, the Lower Rickenbach and the Branchville Member of the Epler Formation. On first inspection the diversity of lithic types occurring on a site may lead to some confusion. On close inspection it is revealed that all the diverse lithic types could be acquired with little effort, in one small ravine.

Occasionally, a chert bed is so distinctive that artifacts can be traced for many miles and related to a particular bed within a member. Such is the case for a group of stage 2 bifaces found in Highland Lakes, N.J. and fingerprinted to a distinct mottled blue bed of chert in Rock Island, N.J. ten or so miles away. The same can be done for the green chert within the Big Springs member of the Epler Formation and much of the white and lavender chert within the Branchville Member.

The potential for this type of study is obvious to the archaeologist. Thus the chert stratigraphy has become a research tool. The present study employs a large archived surface-gathered artifact collection which has accumulated for 60 years. Preliminary examination has already revealed a cultural preference for distinct lithic types, and lithic preference for certain functional types. An example of the former would be the great number of Lamoka-like projectile points fashioned from the white and lavender chert from the Branchville Member. An example of the latter would be the great number of unprepared cores fashioned from the welded brecciated chert of the Lower Rickenbach Member. The writer would like to approach questions concerning the role of human cognition in resource selection, prehistoric land utilization, and the variation of the fundamental strength within each chert type. These are a few of the many problems to be considered on this trip.

DESCRIPTION OF CHERT TYPES

Leithsville Formation (C1)

Califon Member (C1l). The chert within the Califon Member occurs as rounded pods and colloform masses. Some pods are circular in crosssection, while others are elongate. The pods may be as large as 3 feet to less than 6 inches in diameter. They are usually limonite encrusted and produce a rank, fetid odor when struck by a hammer. The bottom portions of the pods possess an unusual coxcomb structure, which is vuggy and may contain druses of quartz. The centers and bottoms have a fibrous texture, while the outer portions of the pods are homogeneous, waxy and never fibrous. Unlike the core, the outer portions are usually darker in color, and possess a pitchy to waxy luster. The colors range from black, to gray, to blue gray, but orange and black mottled varieties are not rare. The chert within the Califon Member usually occurs in the lower unit associated with sulfides.

A unique aspect of the Califon Member chert is its variation in strength or consistency from the core to the outer portions of the nodules. The centers of the pods are fibrous and impossible to work into stone tools. When found on sites, they are generally mistaken for early stage cores, when actually they should be treated as exhausted and discarded cores. The outer darker portions bear a subconchoidal to conchoidal fracture, and even the thinnest flakes are opaque along the edges. The outer areas also contain concentric open fractures, which often control initial reduction. The unsealed fractures cause the chert to break away from the fibrous core as concentric shells. Artifacts of Califon chert are more common along the southwest flowing drainages within the Great Valley Sequence in New Jersey. Discarded artifacts are common at sites along the Piedmont and Coastal Plain as this chert occurs in quantity within the glacial drift. Outcrops are uncommon because the Califon Member is easily eroded and usually underlies valleys occupied by ponds. Therefore quarries in this member are rare. Most Califon chert was gathered from glacial drift or from residual soil within these same solution valleys. Chert masses occurring as a residual can be seen near Califon, New Jersey. Because Califon chert is extremely tough, reduction sites often produce as many hammerstone spalls as chert flakes. Artifacts of Califon chert are uncommon in the Wallkill

River Valley. Recovered artifacts are usually diagnostic prepared and unprepared cores, and large bifaces. Artifacts examined in archived collections bear a waxy patination, and the surface colors can approach olive green. The artifacts are always limonite encrusted from the oxidation of minute evenly dispersed sulfides within the chert, and the artifact may illustrate any number of the physical characteristics described above.

Hamburg Member (Cl_2). The Hamburg Member of the Leithsville Formation is best exposed along Wildcat Rd. in Franklin N.J., on a series of hillsides south of Rt. 23 in Hamburg, N.J., and north of Rt. 23 along Scenic Lake Rd., in Hamburg, N.J.

The chert occurs as a series of continuous and discontinuous, closely-spaced beds, from 1 inch to 6 inches in thickness, occurring between alternating dark green, pink, and gray argillic, finely-laminated dolomites and infrequent shale beds. The chert is characteristically black, but weathers to light blue and blue-black on the surface. Unique to the Hamburg Member chert are veinlets of white quartz, as well as open vugs lined with microcrystalline quartz. When the white quartz veinlets are widely dispersed, the chert appears to bear phenocrysts. When the quartz venation is closely spaced, the chert appears banded or dendritic. Sulfide inclusions are ubiquitous in this chert. Hamburg chert has a vitreous to waxy luster and is translucent only when the white quartz veinlets are concentrated. Associated with this member is a lower unit of silicified dolomite, which is homogeneous, sparkly, and occurs as thin flaggy beds.

Small outcrop quarries and screes are commonly developed within this member and good exposure above river valleys is part of the reason. Beds of the Hamburg Member can be steeply inclined, creating easy access by simple prying techniques. Hamburg chert is very dense as is attested to by the great numbers of chert hammerstones fashioned from this material, which appear in archived collections. Where thick beds of this chert occur, they are usually intersected by sets of open and sealed fractures. Open fractures appear to aid in the quarrying process, but create some of the limitations within the reduction sequence. Often breakage in bifaces occurs along quartz lined vugs or unsealed fractures. Discarded artifacts in the early stages of reduction, the result of failure along fractures, are very common.

The Wallkill Member (Cl_3). The base of the Wallkill Member is marked by a rather distinctive chert zone. The chert occurs as beds up to 2 feet in thickness, and contains chertized algal beds, pisoids, and silicified dolomite. Broken pieces commonly exhibit a shardy appearance, which may represent a fossil coquina concentrateor a rubbly, brecciated chert zone. The nature of the shardy clasts is yet undecided. Fresh surfaces are light gray and black where the chert is concentrated, but weathered surfaces are brown.

A second chert zone occurs in the upper unit which is a sparkly coarse grained dolomite. Here, thin stringers and wispy beds of an opaque black chert, sometimes associated with sulfides, are common. Above this occurs a zone of unique pods of chert. The chert pods occur as concretionary masses, from 1 to 20 inches in diameter, and may be elongate, ellipsoidal, or tube-like in form. The texture ranges from glassy to saccharoidal, and the textural types are distinctly zoned within each mass. Waxy, highly lustrous pieces are very common. The colors range from dark blue-gray to light gray with patches of white, and thin pieces are often highly translucent. Concentric color banding and claystone cores are common, as well as ankerite skeletal casts, which conform to the concentric layering of the chert. These pods are best developed on the Rudinski Farm, south of Pine Island, New York. Polished sections bear evidence of burrowing, and the outer surfaces of the pods have a hide-like leathery appearance. Fairly disseminated rod-like and spherical microfossil material is occasionally found in these pods. Conjugate shear fractures, unsealed in most cases, penetrate each pod.

Quarries within the Wallkill Member occur on the the west facing slopes along the eastern margin of the Wallkill River Valley. Quarries occur as outcrop workings, screes, and conical shafts sunk into swampy soil at the base of outcrops. The Wallkill Member usually occurs as a terrace above swamp areas, therefore access is limited. There is strong evidence at archaeological sites suggesting that Wallkill chert is heat-treated during the critical reduction process.

Allentown Formation (Ca)

Limeport Member (Ca_1). The chert within the Limeport Member is most conspicuous as chertized oolite sequences. Beds are usually thin, less than 3 or 4 inches, but infrequent concentrations of 8 or 9 inches do occur. Silicification appears to follow the highly porous oolite sequences, and when this facies is laterally persistent, the chert beds can be traced for some distance. The contacts between chert-rich sequences and the adjacent dolomite is often gradational, and in many instances the silica content decreases outwards from a central zone of concentration. The algal-rich sequences of the lower and middle units are often silicified, and the silicification appears to be determined once again, by the amount of available porosity or open spaces within the algal structure (Bathurst 1981). The middle and upper units often contain thin stringers of chert associated with sulfides. The chert is usually black, becoming gray to blue-gray northward near Warwick, N.Y. Broken outcrop samples glisten from cleaved oolites, but the matrix chert is usually waxy or resinous, depending upon the concentration of oolites. Thin slivers are usually highly translucent to transparent, revealing the rounded form of the oolites, but when the oolites are concentrated, the chert may appear nearly opaque. Chertized oncoid and pelletal facies are very common in the lower two units of the Limeport Member.

Chert-rich sequences within the Limeport Member are common, but chert infrequently occurs in the concentration to be quarried. Also, much of Limeport Member occupies valley floors within the Wallkill River Valley. Quarries in this member are more common along the southwest flowing drainages of the Paulins Kill, Musconetcong and Pequest Rivers. A series of small quarries occur in Wantage Township, near Hamburg, New Jersey. Great quantities of Limeport chert occur at sites in the Wallkill River Valley, indicating that larger quarries or great numbers of smaller outcrop workings do exist.

Upper Allentown Member (Ca_2). The lower unit within the Upper Allentown Member, a micritic dolomite which weathers light blue, often contains black chert as thin beds and clots. The contacts between the chert beds and surrounding dolomite are sharp boundaries, and beds are very thin, concave upwards, and may occur as a series of alternating ribbons and clots. The chert is homogeneous vitreous, and black. Oftentimes sealed fractures criss-cross the chert, and these fracture surfaces are sometimes coated with white micro-crystalline quartz.

The upper unit contains thin beds and seams of oolitic chert. This chert is often blue-gray, and possesses a sugary to resinous luster. Fresh outcrop samples illustrate an unusual step-like outer surface along bedding surfaces. Within 200-250 feet of the upper contact of the Upper Allentown Member, there is a zone of silicified algal structures (Aitken, 1967). These domal algal forms are banded white and gray chert, and serve as excellent facing indicators. The algal-rich beds are visible at the Havens' Estate as well as other localities not scheduled for this field trip. Within 50 feet of the upper contact with the overlying Rickenbach Formation there occurs a series of thin discontinuous beds and rubble zones comprised of ribbon chert and chert occurring as irregular masses and clots. The rubble zones mark the beginning of an unconformity which can be traced from north of Warwick, N.Y. south to Phillipsburg, N.J. These rubble chert zones are rather extensive, the chert ranging in color from dull black to light blue and white convoluted masses. The clasts are always angular, vuggy, and interstices in the surrounding dolomite are filled with a druses of white quartz crystals, and fine grained white quartz coats many of the chert clasts. Above the rubble zone are a series of thin to thick quartzite beds. Where the quartzite is absent, there is occasionally an ash-gray porcelain-like chert, which occurs as irregular masses.

In the Vernon Valley, there occurs a rather unusual chert within the Upper Allentown Member. The chert is light blue, blue-gray, or ash gray and occurs as finely laminated masses. The laminations may represent relict bedding features. The texture varies from glassy to saccharoidal within a single hand sample. The glassy layers contain vacuoles where euhedral dolomite rhombs have been dislodged by weathering. The chert masses are always brecciated and the dolomite which fills interstices between chert masses is very coarse grained. Individual beds may be greater than 12 inches in thickness, and a myriad of different colors and textures can be obtained from a single outcrop. These breccias may be intrastratal and genetically related to doline formation during the development of the Knox-Beekmantown Unconformity, or related to the unconformity at the top of the Upper Allentown Member (Kerens, 1988). A similar chert, finely laminated, brecciated, ox blood red in color occurs near the upper surface of the Upper Allentown Formation at the Delaware River bridge on I-78 south of Easton, Pennsylvania.

The unique chert occurring within the Upper Allentown Member in Vernon Valley appears to have supported a great number of small outcrop quarries. The presence of fine laminations will aid in distinguishing this chert from that of the Harmonyvale Member of the Ontelaunee Formation, which can also appear light powder blue. There also appears to be a cultural preference for the thin beds of black chert occurring in the lower unit, as great numbers of projectile points, which represent the Brewerton manifestations are fashioned from this chert. Virtually all of the rubble zones are worked as outcrop quarries. Hammerstones and chert debitage litter these outcrop surfaces. It is difficult to delimit the boundaries of a quarry within the rubble zone because the chert is discontinuous and workings may occur over a great distance along strike.

The Rickenbach Formation (Or)

Lower Member (Or_1). The chert within the lower member always occurs as tightly welded brecciated masses of angular to sub-angular clasts. The chert beds are discontinuous masses up to 100 feet in length and may attain thicknesses of up to 3 feet. The chert itself occurs as transparent to opaque black clasts welded into a matrix of translucent to opaque light blue to dark blue gray chert matrix. Occasionally, entire chert beds are lodged within the brecciated masses, and broken beds and clasts can often be re-fitted. In some instances, the welded fragments, appear to blend into each other causing a color gradation without grain boundaries. A simple hand sample of this latter variety may exhibit color variation from light gray to black. More often, the matrix chert is dull opaque gray and the chert clasts are vitreous opaque black masses. The physical attributes of the chert breccia varies from outcrop to outcrop, and the most spectacular examples occur in DeKay's Hamlet, Vernon Valley, New Jersey. At this location the chert clasts are usually embedded in a coarse dolomite.

At DeKay's Hamlet, single chert clasts were apparently separated from the dolomite matrix before manufacturing into stone tools. At other locations, this initial sorting process is not necessary. The chert from the lower member makes its presence in the archaeological records as great quantities of unprepared cores. When bifacially prepared pieces are found, they usually illustrate the brecciated form, and the welded clasts appear to offer little problem to the manufacturer. Most artifacts exhibit a mottled appearance.

The Hope Member (Or_2). The chert within the Hope Member occurs in two distinct zones. The lower chert zone or 7 inch cherts is a series of discontinuous beds 3 to 6 feet in diameter and up to 6 inches in thickness. At some locations, between 3 to 5 or as many as 7 chert zones are present. The beds are usually convex upwards and are black to dark blue in color. Hand samples are highly vitreous, translucent and may contain fine laminations or ooids. An algal origin for this chert has been suggested (pers. comm. N.J. Geological Survey, 1988). Above this is yet another chert zone comprised of thin wispy convex upward beds, as well as clots, of chert. This chert is dark blue-gray, translucent to transparent in thin slivers and will exhibit very fine dark colored laminations. Silicified domal algal structures are also associated with this member.

Actual quarries in the Hope Member are uncommon, as this member is easily eroded and often occurs in low swampy areas. The thinness of the beds and low relief lead to the occurrence of small outcrop quarries developed along strike.

The Crooked Swamp Dolomite Facies (Or_3). The sugary light blue dolomite of the Crooked Swamp Dolomite Facies contains chert in several different forms. At some locations great masses of silicified dolomite occur. The chert appears to disrupt bedding and is often cross-hatched with microcrystalline quartz occurring in the fracture spaces. These chert masses can be several feet in diameter and thickness, and can be easily traced along strike. The chert is light gray to blue gray, its texture is vitreous, and individual seams are rarely greater than 1 or 2 inches in thickness. The whole mass has a gnarled appearance.

A second variety occurs as thin wispy beds, clots and algal replacements. Occasionally thick beds containing chertified oolites will occur. Thin splinters of the bedded chert often will be highly translucent to transparent and exhibit finely layered dark laminations.

Chert within the Crooked Swamp Facies rarely occurs in such concentration as to warrant the development of quarries. Small outcrop quarries and screes can be traced along strike.

The Epler Formation (Oe)

Branchville Member (Oe_1). A thick chert sequence occurs at the base of the Branchville Member. At the Lake Illiff section (Markewicz and Dalton, 1977) the lowest chert beds are dark blue to blue-black and contain a fine venation of lighter blue chert. The beds attain a thickness of up to 1 foot and may be laterally continuous over substantial distances. At the nearby Harmonyvale section, the lowest chert beds are opaque black with a brown venation, up to 1 foot thick, and laterally discontinuous. Above this zone, there are sporadic occurences of a series of ash gray porcelain-like chert beds, which are finely laminated, opaque, may contain sulfides, and are present as discontinuous beds up to 6 inches thick. This chert also occurs as convoluted masses and infillings within solution features, where it attains a gleaming white porcelain-like luster and color. Above this zone occur beds and lenses of a distinctive orchid to lavender colored chert, which may contain oolites, and is often associated with algal-rich beds. The chert is highly translucent and brecciated masses are commonly encountered. In some instances boudins of lavender chert occur, and bedding plane slabs illustrate soft-sediment boudinage structures. Commonly, the lavender chert will grade laterally into a white translucent, waxy chert, which contains vacuoles where dolomite euhedra have weathered out. This variety can attain extreme whiteness, and near transparency, or occur as brecciated masses along with lavender fragments. Occasionally lavender and white brecciated masses are found lightly welded, with coarse dolomite and sphalerite infillings between clasts. Above this zone occur intermittent dark maroon laminated shale beds, which appear to mark the transition into the Big Springs Member. The maroon shales and much of the lower unit of the Branchville Member can be highly siliceous and break with a conchoidal fracture.

All the varieties of chert described above are quarried in the Wallkill River Valley. At some archaeological sites in Sussex, N.J., entire assemblages of artifacts can be precisely provenanced to a series of outcrop quarries in a single ravine or hillside. At the Haven's Estate along the Beaver Run Creek, all the lithic resources mentioned above are mined in one small ravine. Countless scattered small outcrop quarries occur along strike, each marking the location of a concentration of white or lavender chert. Archived collections have shown that a cultural preference has been toward the white and lavender chert. These collections exhibit a Lamoka-like projectile point form which is characteristically fashioned from the white chert. At the Harmonyvale section, a one-half mile traverse along strike in the Branchville Member has allowed the discovery of more than 20 small outcrop quarries. Along quarry walls, in the porcelain-like chert beds, there is indication of heat treatment of the chert, which turns the chert a dark brown, tan, carmel to nearly maroon.

Big Springs Member (Oe_2). The previously described lavender and white chert beds can also be found in the Big Springs Member, where they usually occur as brecciated masses. At the Harmonyvale section, an alternating maroon and white chert occurs. This chert is vitreous, slabby, and may represent the silicified equivalent of the upper maroon shale in the Branchville Member. The beds are nearly 1 foot thick in places. Above the maroon chert is a unique jade green chert which occurs as thick and thin beds, up to 8 or 9 inches thick and it contains numerous vacuoles where dolomite euhedra have been dislodged. Much of the Big Springs Member is highly silicified.

As previously described, wherever the white and lavender chert occurs it is mined out, and the Big Springs Member contains many small quarries. The maroon chert can be easily confused with heat-treated Pennsylvania jasper from the Hardyston Formation. Projectile points of the Broadspear Traditions, fashioned from the green chert beds have been located in archived collections. Once this chert is seen in hand sample, it should not be confused with the Normanskill chert.

Lafayette Member (Oe_3). Discontinuous beds of a steel blue-gray chert occur at the top of the Lafayette Member. The chert is finely laminated, highly lustrous and occurs in beds up to 8 to 10 inches thick. Thin splinters are translucent on the edge. Occasionally, vacuoles are present in this chert, where dolomite rhombs have weathered out. On Skull Island, south of Pine Island, New York, a light blue and gray mottled chert occurs in the Lafayette Member. The colors appear convoluted and may represent original bioturbation structures.

Quarries within the Lafayette Member are rare. Dislodged beds are noticeable when one works along strike. Outcrop workings are usually small and artifacts fashioned from the gray chert will exhibit fine laminations. The blue-gray chert bed on Skull Island is so distinctive that Stage II bifaces have been provenanced from 10-15 miles away back to this one particular bed of chert.

Ontelaunee Formation (Oo)

Beaver Run Member (Oo_1). The greatest concentration of chert in the region occurs in the Ontelaunee Formation and particularly in the Beaver Run member. At the Harmonyvale section the first variety of chert occurs approximately 75 feet above the base of the formation. This chert occurs as great masses up to 3 or 4 feet in thickness, which appear to disrupt bedding. The chert beds are discontinuous, and can be traced along strike as a series of hummocky ledges. The chert is dull opaque black to translucent gray, highly porous and rugose, appearing much like the lower chert unit of the Crooked Swamp Dolomite Facies. Occasional microcrystalline quartz is visible in vugs and pockets, and brecciated, lightly welded masses are common. The texture of the chert is further modified by penetrative cleavage which criss-crosses the chert, and creates quartz-lined fractures. This chert in general is described as gnarled.

A second zone of chert occurs in the middle unit as a series of distinct beds which can be 50 feet in thickness at the Harmonyvale section. At the Parlins' Kill Lake section these beds are up to 2 feet thick and 150 feet long and at Beaver Run, they are up to 6-8 feet thick. The beds appear to pinch and swell, occasionally occur as convex upwards pods, are always dull gray to black, possess a pitchy, dull to lustrous texture, and may occur as mottled and resinous varieties. This chert bears a peculiar petroleum-like or sulphurous odor when struck, as does the surrounding dolomite. Organic halos and dark splotches are commonly present in the chert of the middle unit. The upper unit is usually devoid of chert beds, but when they do occur, they are usually lighter in color, and possess more of the physical attributes of the Harmonyvale Member chert.

The archaeologist should pay particularly close attention to the presence of the Ontelaunee Formation. By far, the largest quarries in the Wallkill River Valley are developed in the Beaver Run Member of the Ontelaunee Formation. Quarries along the Paulins' Kill, Beaver Run Creek, and the Wallkill River may be 300 feet in length. The quarries at the Harmonyvale section are a string of smaller workings, as many as 50 in the Beaver Run Member, which stretch for 2,000 feet along strike, wherever the chert concentrates. The quarries themselves are characterized by countless conical pits and excavations parallel to beds. Oftentimes, large beds are entirely dislodged, and tree falls in the quarries indicate that the debitage occurs to a considerable depth. The ground is generally littered with quartzite hammerstones and anvils, some of which weigh up to 75 to 100 pounds. Chert debitage, evidence of very early stage reduction is everywhere. Very infrequently, late stage bifaces and failed objects occur along the quarry face, along with pottery.

Harmonyvale Member (Oo_2). The transition into the Harmonyvale Member is marked by the appearance of lighter-colored varieties of chert; first dark blue grading upwards to a series of rhythmic beds of light blue chert. Occasionally, dark and light gray with red mottled varities are present but the light blue predominates over all. The chert is usually homogeneous, lacking vacuoles, laminations or any type of inclusions. The blue color is consistent throughout and beds are up to 8-10 inches thick. When brecciated masses occur, the clasts are generally very light colored and nearly the color of the matrix chert. Organic halos and splotches may be present and cleavage may penetrate the chert beds, rendering them fractured along distinct flat planes. Sometimes, the infilling along cleavage planes is lighter colored. At the Sarepta Quarry, near Hope, New Jersey, the Harmonyvale chert occurs as convex upward pods, of black chert, 5-8 feet in diameter, 6 inches thick, and contain floating quartz grains.

Mottled dark and light blue varieties are common in the lower portion of the member.

The Harmonyvale Member was apparently highly sought after, possibly because of its general consistency. In locations where both the Beaver Run and Harmonyvale Members are present, the Beaver Run Member is often neglected in favor of the Harmonyvale Member. Because the chert of the Harmonyvale Member is restricted to fewer and thinner beds, the quarries in the Beaver Run Member appear to be more extensive. In locations where both members exist, such as at Beaver Run Creek and Harmonyvale, the primary focus of quarry activity appears to be the Harmonyvale Member. The two members are adjacent to each other and therefore the quarries generally include both members. Chert of the Harmonyvale Member is apparently preferred for the Laurentian Tradition of stone tool making. At the Phillips Estate, a small quarry, approximately 75 feet by 30 feet, was developed in the Harmonyvale Member. The site includes mounds of tailings and a rock-cut shaft following a six inch thick chert bed, which dips along with bedding approximately 47° NW. The Harmonyvale section also includes some extensive workings in the Harmonyvale Member. The Ontelaunee Formation with its numerous chert beds often occurs in areas of higher relief. Because the Harmonyvale Member occurs stratigraphically further up section, it usually has received more erosion, and hillsides are often littered with tailings, rendering this member very noticeable in the field.

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ROAD LOG FOR THE STRATIGRAPHIC RELEVANCE AND ARCHAEOLOGICAL POTENTIAL OF THE CHERT-BEARING CARBONATES WITHIN THE KITTATINNY SUPERGROUP

CUMULATIVE MILEAGE	MILES FROM LAST POINT	ROUTE DESCRIPTION
0.0	0.0	From Chester, New York, take Route 94 south to DeKay Road, in Vernon Valley, New Jersey
19.4	19.4	Make a right turn onto DeKay Rd.
20.3	0.9	Make a left turn onto Price Rd.
21.1	0.8	STOP 1 is the Paleokarst feature on right.

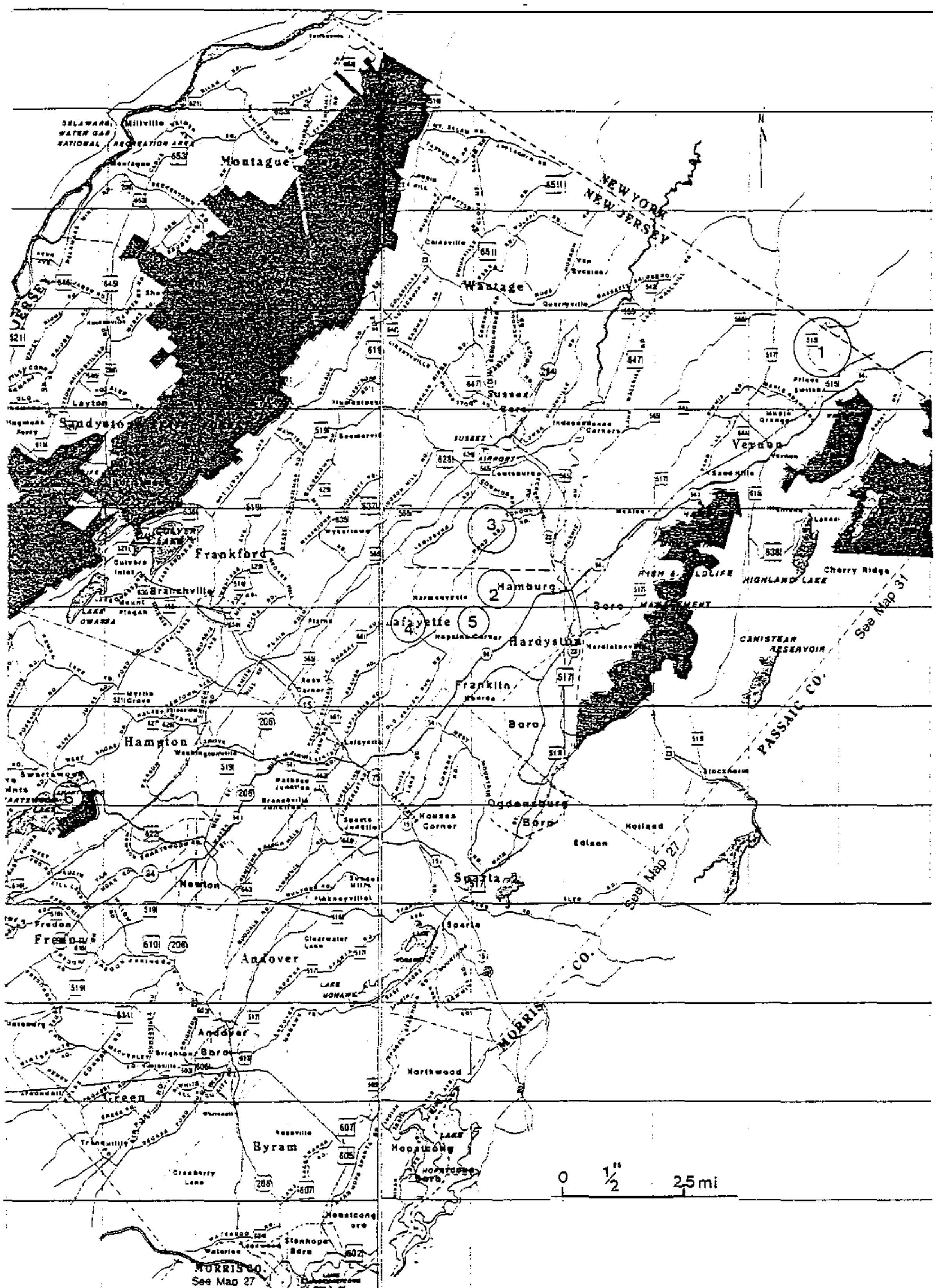


Fig. 2 Map of the General Study Area depicting field trip localities in open circles.

STOP 1. PALEOKARST FEATURES AT DeKAY'S HAMLET (Fig. 3).

The paleokarst features at DeKay's Hamlet are unique for several reasons. They are clearly visible on the Wawayanda topographic atlas sheet, and are outlined in Fig (3). The single hole-like depressions are of great areal extent, and the likelihood of discovering more in the vicinity is very good. The infilling within the inverted depressions contains much more chert than is seen at other locations. The feature rests on the Upper Member of the Allentown Formation. To the writer's knowledge, this represents the deepest penetration of the Knox-Beekmantown erosional event in this area.

Along Price Road, the Allentown Formation is poorly exposed and is visible as thickly bedded coarse grained blue dolomite with thin beds and wisps of black oolite chert. Capping the Upper Allentown is a chert sequence of variable thickness. Sawed slabs of chert from the Price Road outcrops have yielded both algal rich beds and slabs with a structureless form. The chert sequence capping the Allentown at DeKay's Hamlet may represent algal beds in some locations, while in others the chert may represent a silcrete (Khalaf, 1988). Above this zone is the extensive rubble of the paleokarst feature. Chert clasts can be very large and the refitting of chert clasts, blocks, and beds within a single block of karst infilling dolomite is easily accomplished. The infilling between chert clasts and individual dolomite blocks is a very coarse grained buff dolomite.

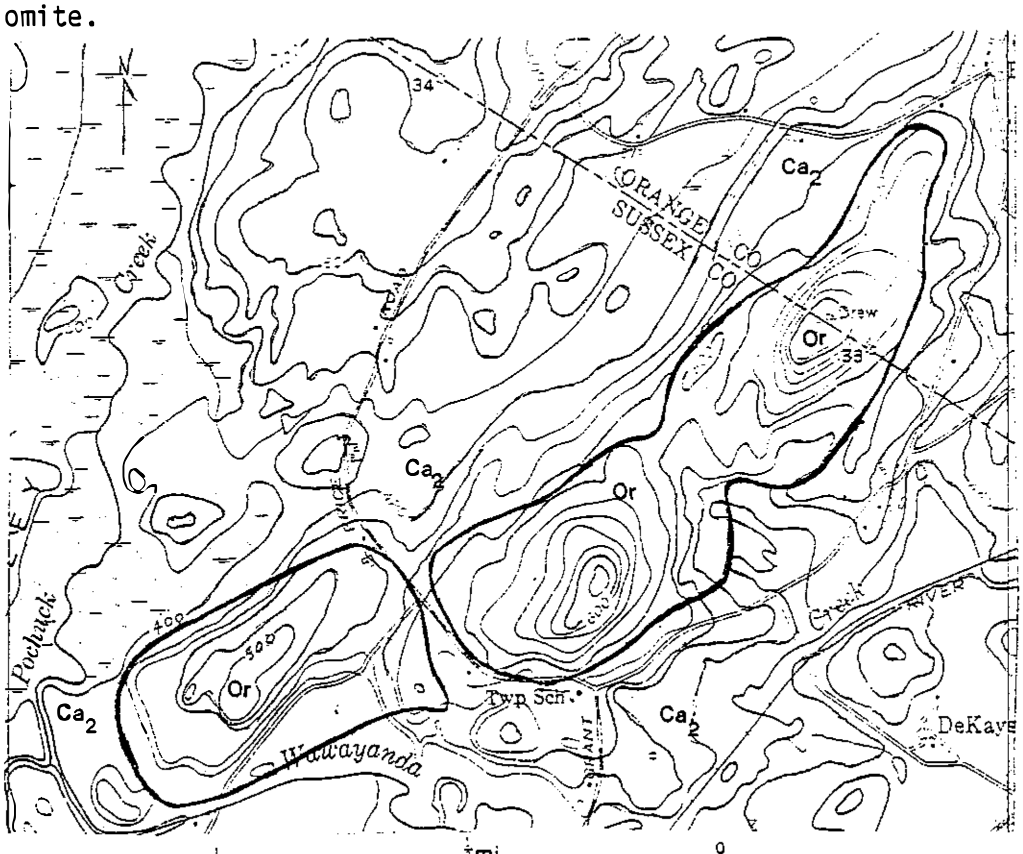


Fig.3 Outline of Paleokarst Feature, DeKay's Hamlet, Wawayanda Quandrangle

The chert clasts appear to be representative of the Hope Member and the Crooked Swamp Dolomite Facies of the Rickenbach Formation. At a nearby location along Owen's Station Rd. in Wantage Township, a similar chert breccia contains the preserved remains of cystoid-like organisms.

In the surrounding valley, the outcrops of the Upper Member include a distinct intrastratal breccia zone with siliceous infilling. The chert horizons may represent the remains of a vast doline system which penetrates the Allentown Formation and creates a plexus of breccia zones which may be genetically related to the vast subaerial feature being viewed at the first stop. This unique chert breccia and silicified infilling has not been encountered elsewhere in the valley, and recent field evidence suggests that in the Vernon Valley, the karstification may proceed down as far as the Limeport Member of the Allentown Formation. Virtually all of the farm fields in DeKay's Hamlet show exposed beds of the breccia and the laminated infilling described under the Upper Allentown Member. Similar laminated chert beds and brecciated zones are visible in the Big Springs Member of the Epler Formation but these do not bear the color characteristics of the Upper Allentown Member chert. Therefore it is logical to assume that the process of silicification and brecciation is similar in both cases.

The paleokarst feature appears to have provided quantities of chert to Archaic as well as Woodland Age groups of hunter-gatherers. The workshops and screes have a different character about them because the quarrying process is unlike other locations. Here, large angular chert clasts are worked free from the surrounding dolomite, instead of mining a single continuous bed of chert as at other locations. In addition, the archived collections from the area exhibit diagnostic projectile points whose overall morphologies are statistically different from culturally similar stone tool inventories from neighboring areas. This measurable difference in physical attributes may be linked to the individual artisan or to the physical properties and limitations of the chert. The writer feels the latter reason may be more important at this location.

The breccia zones described above crop out on nearby farm fields and all have been quarried for the intrastratal breccias and silicious infillings.

22.8	1.7	Retrace route back to Rt. 94, make right and continue south to Beaver Run Rd.
32.8	10.0	Make right turn onto Beaver Run Rd.
33.6	0.8	Travel west to the Paige Farm on the right. STOP 2 is the paleokarst feature along Beaver Run Rd. and a chert quarry in the Ontelaunee Formation involving a traverse ending 0.2 mi. west of the well house at the bend in Beaver Run Rd.

STOP 2. THE PALEOKARST AT THE BEAVER RUN.

The traverse at Beaver Run begins in the Hope Member of the Rickenbach Formation, with the Leithsville and Allentown Formations behind you to the east (Fig. 4). Prominent within the Hope Member is a silicified intraformational conglomerate, in which both rounded clasts as well as matrix have been silicified. This intraformational conglomerate is also visible along the eastern edge of the Haven's Estate, and in the Richter Estate (see Fig. 4). The beds are inclined to the northwest and the intraformational conglomerate has been interpreted by some to represent the base of an extensive paleokarst system. The paleokarst is first visible within the Branchville Member of the Epler Formation, where the dolomite is brecciated and re-cemented in-situ. When this occurs within the Big Springs and Lafayette Members of the Epler Formation, the solution crevices contain a laminated green argillic infilling, which occasionally bears euhedral pyrite. Farther up-section the topographically inverted karst infilling is visible to your right. The feature stands 180 feet in elevation above the roadside, and is prominent at the 700 ft. contour interval in Fig. 4. The eastern half of the feature is developed within the Epler Formation. The dislodged and overturned masses of dolomite in the eastern portion of the feature are derived from the Big Springs and Lafayette Members.

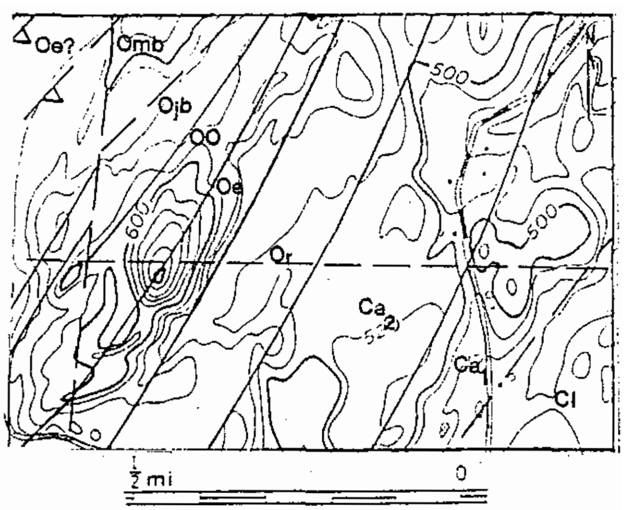


Fig. 4 The Paleokarst Feature and Prehistoric Chert Quarry along the Beaver Run Road Traverse

Midway through the paleokarst feature, large blocks of chert-bearing dolomite, derived from the Beaver Run Member are visible as a stack of disorientated re-cemented blocks. The slabs may represent the collapsed roof structure of the karst system. The paleokarst is best developed within the Lafayette Member. The evidence for karstification within the Epler Formation is visible along strike to the north into the Blair Estate Traverse and the Haven's Island #1 traverse, for greater than 6,000 ft. The feature continues to the south across Beaver Run Road more than 4,000 ft. to the Richter-Phillips Traverse (see Fig. 7).

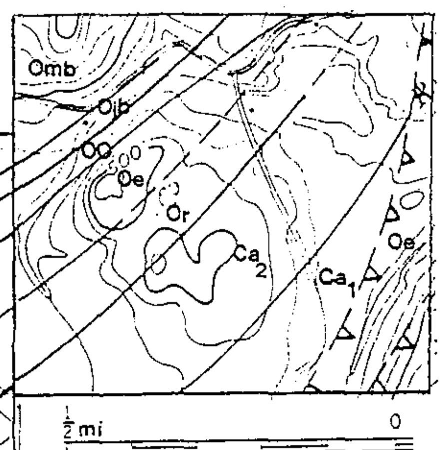
The thickened Ontelaunee Formation section may be the result of faulting (see Fig. 4), and as you continue up-section, you encounter the Jacksonburg Formation (Ojb). From this point, proceed north along strike in the Beaver Run member. Here, along the western edge of the traverse are a splay of intrusive dikes, thought to be upper-Lower Ordovician. Approximately 1,000 feet north, along strike and southeast of the Dagmar Dale Estate, is a prehistoric chert quarry developed within the Beaver Run Member. This quarry and associated scree is typical of those developed within the Beaver Run Member. There is an enormous quantity of broken chert, all representing the earliest stages of the lithic reduction process.

If time and interest exist, traverse downsection on the Blair Estate, of which you are presently on the western edge. Visible will be paleokarst within the Epler formation, chert quarries within the chert breccia zones of the Lower member of the Rickenbach Formation, and a series of convex upwards silicified algal beds present in the Upper Allentown Formation. The algal structures are present on the Haven's Estate and elsewhere within the Hamburg Quadrangle and may represent a distinct marker horizon.

33.8	0.2	Cont. south on Beaver Run Rd. to De Kort Rd.
35.6	1.8	Make right on DeKort Rd. and proceed north to Pond School Rd.
37.5	1.9	STOP on right across from Kimble Farm. STOP 3 is the Chert quarry at the Beaver Run Creek.

STOP 3. THE PREHISTORIC QUARRIES AT BEAVER RUN CREEK.

The traverse begins on the south side of Pond School Road, 700 ft. east of the road, in the Lower member of the Rickenbach Formation. The rubble zone marking the unconformity is at the top of the Upper Allentown Member. The chert breccia of the Lower Rickenbach is present here (Fig. 5). Above the chert breccia is the lower chert horizon of the Hope Member. Proceed up section into the Branchvile member of the Epler Formation. Visible from this point is yet another type of paleokarst infilling. Here are large cemented masses of a very coarse grained dolomite. Throughout the dolomite are clasts of lavender and blue chert. The clasts represent the broken and disrupted chert bearing beds of all three members of the Epler Formation. Two hundred feet to the south is a series of mound-like structures which represent a prehistoric quarry operation within the Branchville Member. The chert present is translucent white with lavender brecciated masses. This breccia is tightly welded, and the quarry is represented by buried tailings.



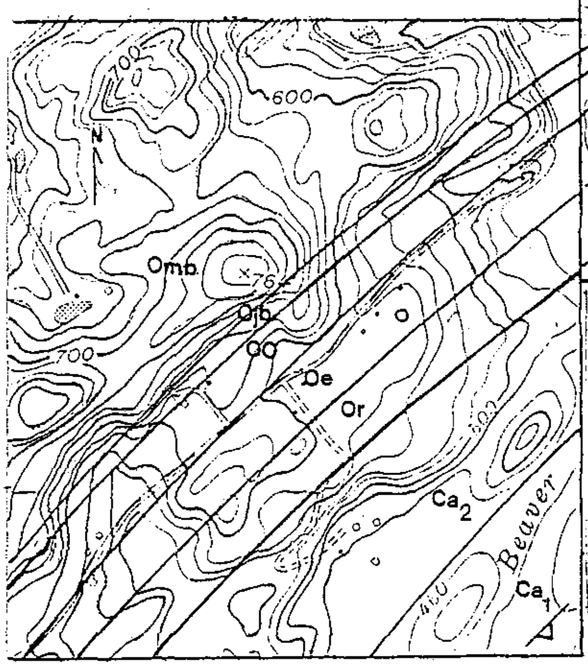


Fig. 5 Prehistoric Chert Quarries in the Ontelaunee Formation, Beaver Run Quarry Traverse and Alternate Stop the Haven's Estate, Hamburg Quadrangle.

Across School Pond Road to the west 300 feet is the prehistoric quarry at Beaver Run. This is an excellent location because the quarry has been acted upon by erosion and the prehistoric workings are visible.

The traverse begins 200 feet upsection in the Beaver Run member of the Ontelaunee Formation. A massive black chert zone is present in the Beaver Run member. Evidence of the reduction process is to be found at the quarry face. Many of the quartzite boulders and cobbles lying in the rubble represent the broken remains of discarded hammer and anvilstones. Approximately 75 feet above the thick chert sequence is the base of the Harmonyvale Member. There are a great number of conical depressions along strike in the many chert beds of the Harmonyvale Member. This pattern may represent the differential weathering of the solution cleavage within the Harmonyvale Member.

Both members of the Ontelaunee Formation are included in this quarry, and the workings and prospects extend an area of 450 feet by 300 feet. In this quarry, the Harmonyvale Member appears to be the preferred ore, as most of the deeper prospects are developed within this member. There is more archaeological visibility within the lower Beaver Run member, than in the Harmonyvale Member. The Beaver Run chert is usually conspicuous on the landscape, forming benches and topographic high areas. The prehistoric workings are usually extensive and the quarries can be developed continuously along strike. The tough ash-gray dolomite of the Harmonyvale Member, oftentimes caps the crests of hillisdes and is covered by vegetation. Prehistoric workings within the Harmonyvale member are much less visible to the observer. Usually, a single light blue chert bed will be worked continuously along strike for several hundred feet as at this location. Because Harmonyvale chert occurs as thinner beds, the conical pits and workings along strike exhibit less visibility. Chert beds within the Beaver Run member can attain thicknesses of 8 or 10 feet, which render the quarries conspicuous. The writer feels that when both members are present in one quarry complex, the Beaver Run Member may be a second choice of lesser grade ore. The more highly sought after ore is the light blue chert of the Harmonyvale Member. This seems to be substantiated through the examination of archived collections.

Great quantities of quartzite hammerstones are present at this site. One large pounding stone or anvil near the Beaver Run member weighs more than 100 pounds. A collection of very diagnostic hammerstones was made at this quarry. These are always fashioned from quartzite, perfectly round and flat along their top sides, much like a wheel. Many instruments of this kind have been collected as well as hammerstones which appear to approach this diagnostic form. The hammerstones in the quarry represent the continuum of mining instruments, from expedient tools to instruments which are midway or at some stage of refinement towards being diagnostic. The final stage of processing appears to be represented by the spherical form with the flattened edges, or achieving the ideal pre-determined form.

Most of the quarry workings represent the initial reduction process. Chert from the quarry is found as unprepared cores, prepared cores, and diagnostic tools throughout the valley. It appears that much of the reduction process took place away from the quarry site.

39.4	1.9	Return south along DeKort Rd. to intersection with Beaver Run and Fox Hill Rd.
39.5	0.1	Continue south on Beaver Run Rd. to private drive on right.
39.8	0.3	Follow drive to Perry Farm and Pond at Harmonyvale. STOP 4 is the stratigraphic section and prehistoric quarries at Harmonyvale.

STOP 4. THE HARMONYVALE TRAVERSE.

The Harmonyvale traverse begins at the base of the Branchville member of the Epler Formation (Fig. 6). The ravine to your east is occupied by the Hope member of the Rickenbach Formation, and the eastern edge of the summit is occupied by the Crooked Swamp Facies. Present is the sugary blue dolomite, which has a sulfurous odor on freshly broken surfaces, clots of varied-colored black and blue mottled chert, and silicified algal structures.

Proceeding up-section, the base of the Branchville Member exhibits discontinuous black chert beds, up to 8 or 10 inches thick. Small chert outcrop quarries are present within this horizon. Up-section and to the south, one encounters the white chert of the Branchville member. The outcrops are littered with prehistoric workshop debitage and 30-40 small outcrop quarries exist along strike to the south in this member.

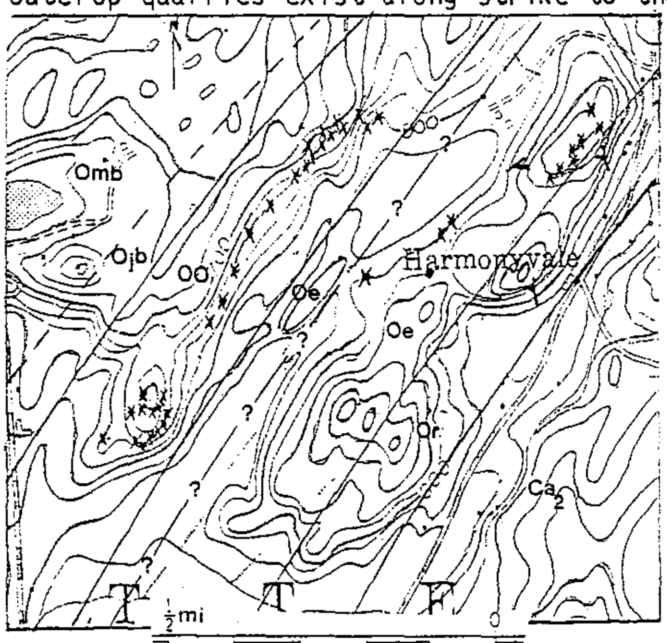


Fig. 6 The Harmonyvale Section in the Branchville Quadrangle clearly depicts the continuation of the Beekmantown Group to the south and west of the Hamburg Quadrangle. The areas marked X indicate the locations of prehistoric quarries. The quarries mark the traces of contacts between members.

The Big Springs member crops out within the woodlot near the eastern shore of the pond. The lower unit contains a unique variety of chert which is maroon and white, and slabby. It represents a silicified shale unit which is stratigraphically similar to the maroon shales within the Upper Branchville member, at Haven's Island #4 Traverse. Above this unit are a series of jade green and white chert beds. The green chert beds may represent the silicified equivalent of the argillaceous karst infilling ubequitous to the Big Springs Member. The white chert achieves an acme in form along this series of outcrops, where it is highly translucent, milky to waxy white to gray, highly lustrous, and occurs as both beds and clots. The laminated procelain-like gray chert is also present at these outcrops.

Close inspection will reveal prehistoric outcrop quarries, and a large quarry is buried under tailings at this location. The dolomite and chert blocks and rubble are part of the quarry tailings. Quartzite hammerstones, and chert debitage are abundant near the surface. Proceed along strike to the southeastern edge of the pond and back downsection 300 feet to the base of the Branchville Member, and the location of yet another prehistoric quarry. The principal ore here is the white chert and the porcelain-like chert. A small surface clearing has yielded hundreds of hammerstones and countless chert flakes. This series of quarries clearly illustrates the difficulty encountered when attempting to locate prehistoric chert mines. The ancient mining operations are usually buried in old tailings and fallen vegetation. This fact has also served to preserve their potential wealth of knowledge.

Now traverse up-section through the Big Springs and Lafayette members across the southern edge of the pond, and into the base of the Beaver Run member. Approximately 100 feet upsection occur the first series of dolomite and associated chert beds. The chert is vuggy, rugose and occurs in concentration as discontinuous swells which create topography to the south along strike. Each of these swells marks the location of very small chert outcrop quarries. Up-section is a second zone of chert which occurs as a series of beds which pinch and swell. The chert is homogeneous vitreous, and black, whereas the associated dolomite beds are thick and billowy. These chert beds are extensively quarried along the southwestern portion of the pond. The base of the steep ledge created by the resistant chert beds is covered with broken dolomite and chert beds, which have been pried from their ledges.

Transitional into the Harmonyvale Member is a dark and light blue mottled chert bed which is laterally continuous and eventually grades into a dolomite bed. The dolomite here is resistant, crosshatched, weathers to a white color and marks the lower Harmonyvale Member. Prehistoric mining activity is extensive here, but the precipitous climb and lush vegetation hinder close inspection.

Traverse along strike in the Beaver Run Member, passing the swells of rugose chert, and continue 1,500 feet to the 700 feet contour interval and a large scree developed along one bed of Harmonyvale chert. The workings in this light blue bed are continuous for 90 feet Below this scree is a large quarry face developed in the upper unit of the Beaver Run Member. Inspect the proliferation of chert. To the immediate west lies the Jacksonburg Formation. Turn south, and traverse to the pond along strike in the Epler Formation. End of the Traverse.

40.2	0.4	Take Beaver Run Rd. north to Ross Rd. Make right on Ross Rd.
40.7	0.5	Travel to the Phillips Estate. Park in drive on left by duck pond. STOP 5 is the Richter-Richter-Phillips Traverse.

STOP 5. THE RICHTER AND PHILLIPS ESTATES TRAVERSE.

The traverse begins in the rubble zone of the Upper Allentown member.

Proceed up section through a thick chert breccia zone of the Lower Member of the Rickenbach Formation (Fig. 7). The ravine to your west is underlain by the Hope Member of the Rickenbach Formation and a distinct chert horizon is visible to the south along strike. The bedding dips to the northwest (see Fig. 7). Traverse through the open field which is underlain by the Crooked Swamp Facies, further south along strike. The step cliff overlooking the ice pond is composed of the Hope Member and Crooked Swamp Facies. The distinct chert horizon is visible in this ravine. The wall of the cliff face contains a partially silicified intraformational conglomerate. Above this is a zone of sugary coarse grained blue dolomite containing wisps, beds and clots of a blue-gray translucent chert. This outcrop bears great resemblance to the Crooked Swamp Facies. The crest of this hill contains the Branchville member of the Epler Formation. The characteristic lavender and white brecciated chert zone is at the base of the member. Small prehistoric outcrop quarries are abundant along strike to the north. The ravine is occupied by the Lafayette Member and is much similar to the Lafayette Member ravine at Harmonyvale. You will cross this ravine to a hillside underlain by the Beaver Run Member of the Ontelaunee Formation. This is the Richter Estate.

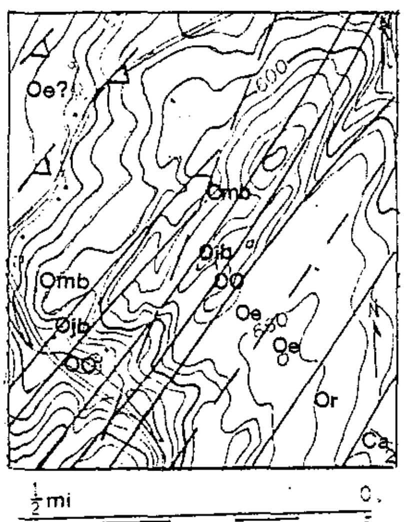


Fig. 7 The Richter-Phillips Traverse, Hamburg Quadrangle

The Beaver Run Member is covered by vegetation at this location. If you traverse north towards Beaver Run Road, you will cross over into the Harmonyvale Member, when the stratigraphic section thickens. Here are a number of small outcrop quarries and screes developed in the Harmonyvale Member. This light blue chert is very similar to the chert exposed in the quarry at the Phillips Estate, just to your west, and very similar to the chert at the Harmonyvale and the Beaver Run Quarries. The west facing slope is underlain by very fossiliferous Jacksonburg limestone. As you cross the Jacksonburg, you pass the location of a proposed fault, (see Fig. 7) and thereby re-enter the Beaver Run Member. Here the Beaver Run contains the upper chert-bearing unit, and a small prehistoric outcrop quarry and scree is present along the summit of the hill.

Passing westward, down the steep west-facing slope and into the Harmonyvale Member, you encounter a unique prehistoric quarry in the Harmonyvale Member. Vegetation-covered mounds of quarry tailings are present, along with the characteristic quartzite hammerstones. The unique feature of this quarry is what appears to be a rock-cut shaft which is about 10' x 10' and filled with tailings. A 6 inch thick light blue chert bed is inclined into the shaft at an inclination of 47° NW. The writer does not know how deep the shaft is or how extensive are the workings. As in the instance of the other quarries, all the workings are deeply buried and only a thin veneer of cultural debitage marks the locations of the ancient prospects. Erosion along steep slopes has exposed the workings and created a window into the past. Similar workings are present throughout this ravine which is to the north on the Phillips Estate. Just to your west is the Jacksonburg Formation repeated after the fault, and the Bushkill Member of the Martinsburg Formation (Omb) which comprises the western hillside and the termination of the traverse.

		Follow Ross Rd. east (0.9 mi) to Bunn Rd.
41.6	0.9	Make right on Bunn Rd. and proceed to Rt. 94.
42.3	0.7	Make right on Rt.94 and travel south to junction with Rt.206.
50.8	8.5	Follow Rt. 206 south to Mill St.
53.1	2.3	Make right on Mill St. Travel west
53.7	0.6	and make right onto Old Swartzwood Rd.
56.0	2.3	Proceed SW to Newton-Swartzwood Rd.
56.7	0.7	Merge to the right and proceed SW until you reach the T in the road. Make sharp left onto Dove Island Rd. and follow for 0.8 mi.
57.5	0.8	Park along roadside. STOP 6 is the chert quarry in the Ontelawnee Formation.

STOP 6. THE CHERT QUARRY ALONG THE PAULINS KILL.

The traverse along the Paulins Kill was suggested by Don Monteverde and Greg Herman of the New Jersey Geological Survey. They have noted large concentrations of chert within the Ontelaunee Formation. The prehistoric chert quarry on this traverse was located within the first day of research in the area.

The quarry is developed within the Beaver Run member of the Ontelaunee Formation, and has a length greater than 200 feet (Fig. 8). The chert beds pinch out to the north and south. To the writer's knowledge, the Harmonyvale Member is not present along this traverse. A distinct rubble zone occurs at the top of the section, and marks the incision of the Knox-Beekmantown Unconformity which removed the Harmonyvale Member at this location. The Jacksonburg Formation (Ojb) lies to our east under Paulins Kill Lake. The steep rise to your east is comprised of the Beaver Run Member and across Dove Island Road to your west lies the Epler and Richenbach Formations. If the stratigraphic formations contacts are interpreted correctly, the chert quarry is developed along the eastern limb of an antiform.

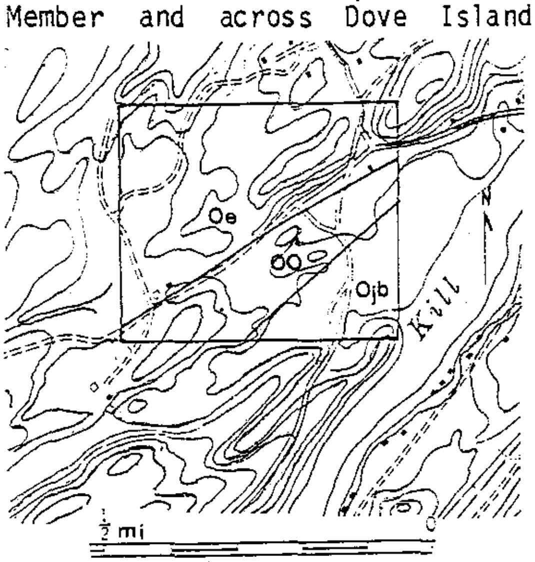


Fig. 8 Paulin's Kill Traverse, Newton East Quadrangle

The series of black and gray chert beds have been extensively worked. The gently inclined beds allowed easy access to the chert. This factor often determines to what degree the chert beds will be exploited. Tailings from the initial stages of reduction are abundant and this is one of the few quarries where structure and erosion are combined for excellent viewing. The chert beds are thick and homogeneous, with well-spaced fractures, creating a valuable ore of the black chert.

		Retrace route east back to Mill St. in Newton.
61.3	3.8	Follow Mill St. east.
61.9	0.6	Make left onto Rt. 206.
		To get back to starting point in New York State,
62.9	1.0	Follow Rt. 206 north to Rt. 94 north.
64.2	2.3	Proceed on Rt. 94 north to Chester, N.Y. to exit from research area to the south for New York City vicinity.
77.5	14.6	Make right on Mill St. south, right onto High St. left onto Spring St. right turn onto Main St. proceed to Rt. 206S.
		Follow Rt. 206S to Interstate 80.

NOTES

THE NYACK SECTION OF THE PALISADES: FACIES, CONTACTS AND LAVA DOMES

Jeffrey C. Steiner
Dept. of Earth and Planetary Sciences
The City College of New York
New York, NY 10031

INTRODUCTION

The Palisades is a large, internally layered tholeiite intrusion which is exposed for approximately fifty miles along the west side of the Hudson River (figure 1). This trip examines pigeonite and augite dolerite sites on an Upper Nyack to Valley Cottage, New York traverse (figure 2).

The data previously collected at these sites will be used to address the issue of identifying the number of magma pulses which produced these and related facies. Peripherally, the relationship of the internal layering to the pulses will also be explored.

The Palisades has widely been considered to have arisen from at least two magma pulses (Walker, 1969). These pulses were defined petrographically by Walker (1969) and may be synchronous with the First Watchung and Ladentown-Union Hill Basalts, respectively (Puffer, et al., 1982, and this volume). Recently, Shirley (1987) identified three to four poorly defined pulses, none of which were characterized petrographically. The pigeonite and augite dolerites of present interest document a pulse distinct from those identified by Shirley (1987), but petrographically similar to Walker's Pulse 1 at Haverstraw Quarry.

INTERNAL STRUCTURE

Walker (1969) defines Pulse 1 as comprised of the chilled dolerite at Kings Bluff, N. J., the early dolerite at Englewood Cliff, N. J., and an early pigeonite dolerite stage at Haverstraw, N. Y. This suite is overlain by the early, middle and late fractionation stages of his Pulse 2. The juncture between these pulses is marked in the southern Palisades by the olivine layer (fractionation stage 2, hyalosiderite) which was considered by Walker (1969) and Walker, F (1940) not to be a separate picritic intrusion, though Husch (1989) provides recent comment to the contrary. Walker (1969) proposed that the hyalosiderite facies is not a cumulus zone (Walker, F., 1940), but rather an unusual facies produced as a contact effect during a quiescent period just subsequent to the second pulse.

The olivine layer occurs 20 to 60 feet above the base of the intrusion in the southern Palisades. It extends to a suggested terminus in Nyack Beach State Park (Walker, 1969). The northern continuation of the inter-pulse boundary was considered to be

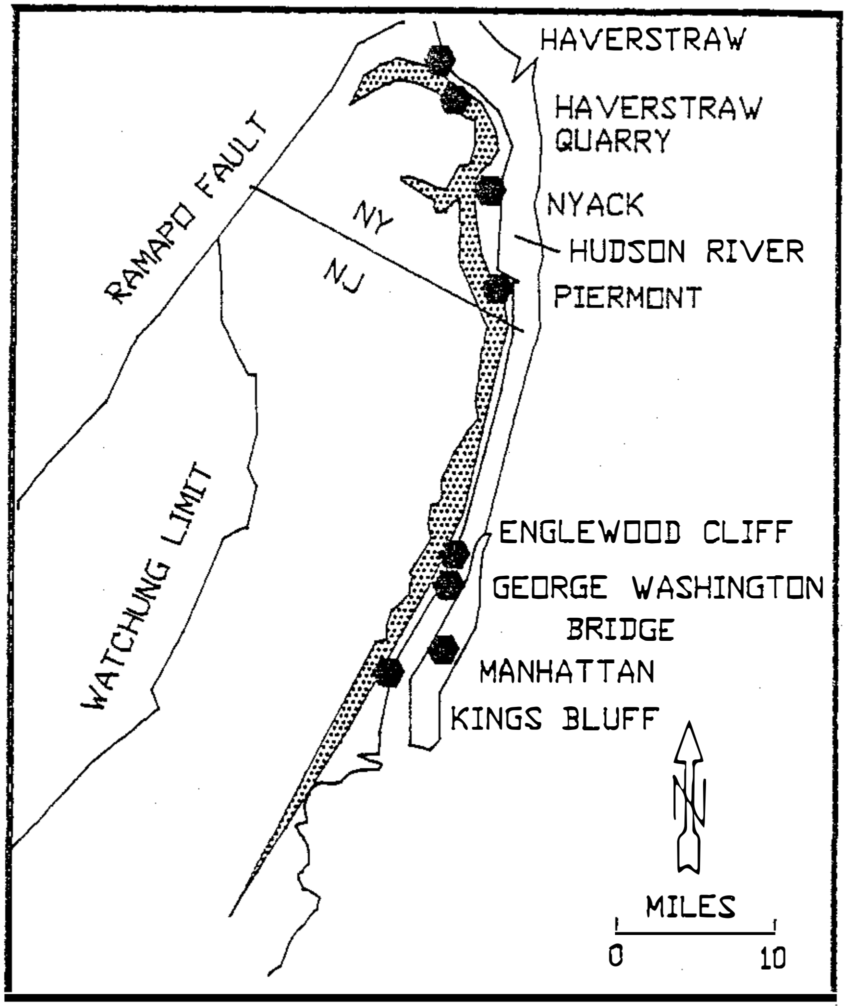


Figure 1. Generalized geologic map after Fisher, et al. (1970) showing the outline of the Palisades.

represented by the (early) pigeonite dolerite of Haverstraw Quarry by virtue of (1) an apparent reversal in pyroxene crystallization, (2) the presence of interstitial glassy mesostasis thought to be otherwise absent in the intrusion and (3) its appropriate location at 105 feet above the base of the intrusion. The early bronzite dolerite encountered by Walker (1969) at the 150 foot level was considered to mark the partially commingled base of the second pulse.

Recent investigations by Steiner, et al. (1989a,b) describe occurrences of pigeonite dolerite south of Nyack Beach State Park and in areas overlooking the Park which are remarkably similar to the early pigeonite facies (Pulse 1, Walker, 1969) at Haverstraw.

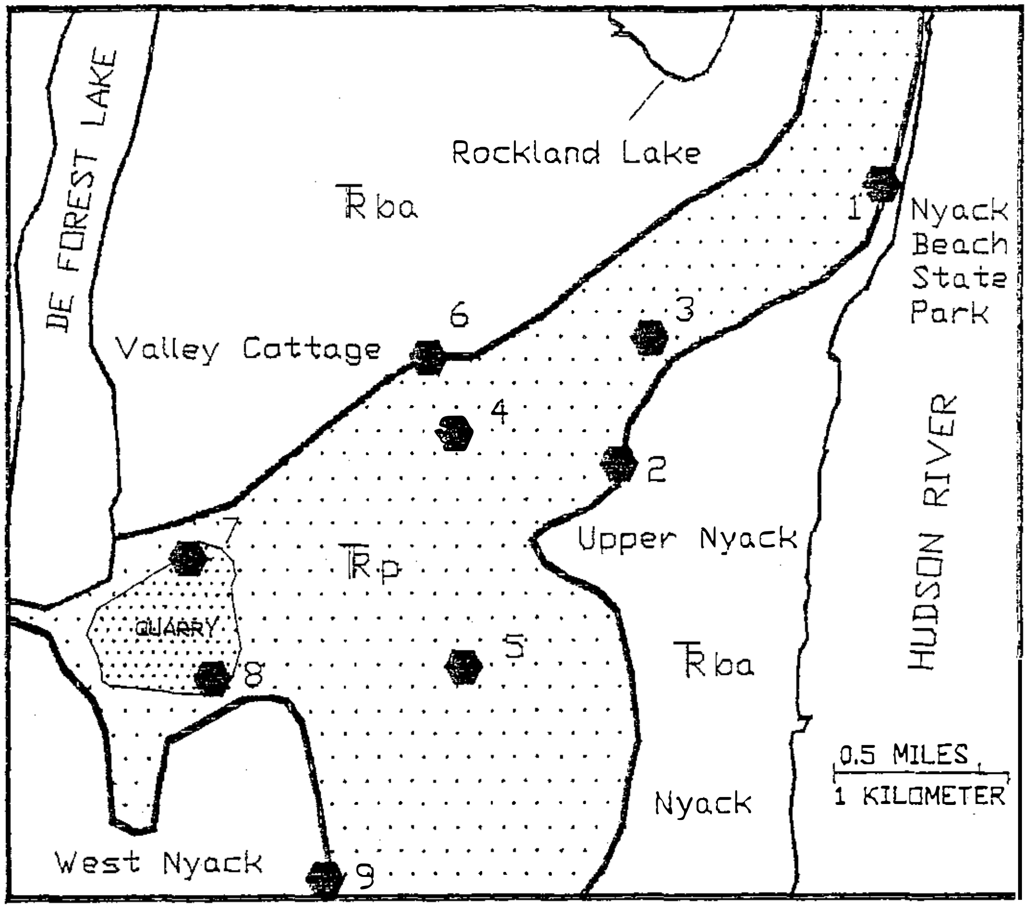


Figure 2. Enlargement showing field trip stop numbers; the Palisades (Trp, stipled) invades mudstone, siltstone and arkose of the Brunswick Formation (Trba, Fisher, et al., 1970).

PIGEONITE DOLERITE

According to Walker (1969), the early pigeonite dolerite and overlying bronzite dolerite are characterized by (1) the absence of reaction rims (presumably on bronzite), (2) scattered oscillatory zoning in minerals of the bronzite dolerite, (3) the presence of augite and a few pigeonite grains on somewhat corroded bronzite microphenocrysts of the bronzite dolerite, (4) numerous patches of glassy mesostasis, (5) independent primary pigeonite crystals, and (6) serpentine after olivine in both dolerites (ibid, p.144). Of these, 4 through 6 apply specifically to the pigeonite dolerite, and the presence of independent pigeonite and "glassy mesostasis" are particularly distinctive and unambiguous.

Walker's (1969) chilled dolerite and early dolerite possess primary pigeonite throughout. In the northern Palisades the texture tends to be intersertal, with pools of dark basaltic glass

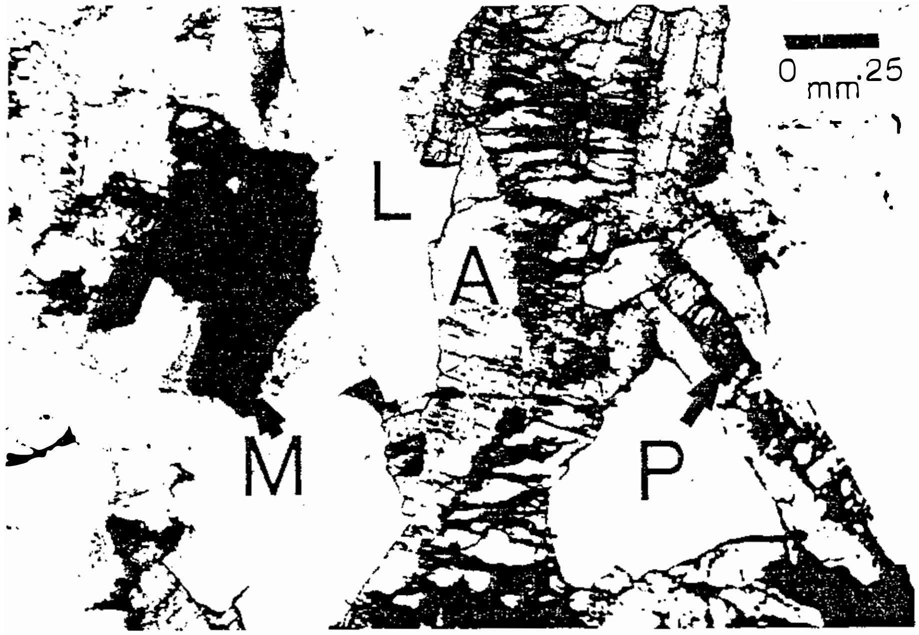


Figure 3. Photomicrograph of coarse pigeonite dolerite (stop 3); augite (A) laths with pigeonite (P) cores subophitically intergrown with labradorite (L); accessory titanomagnetite (M); 0.25mm scale.

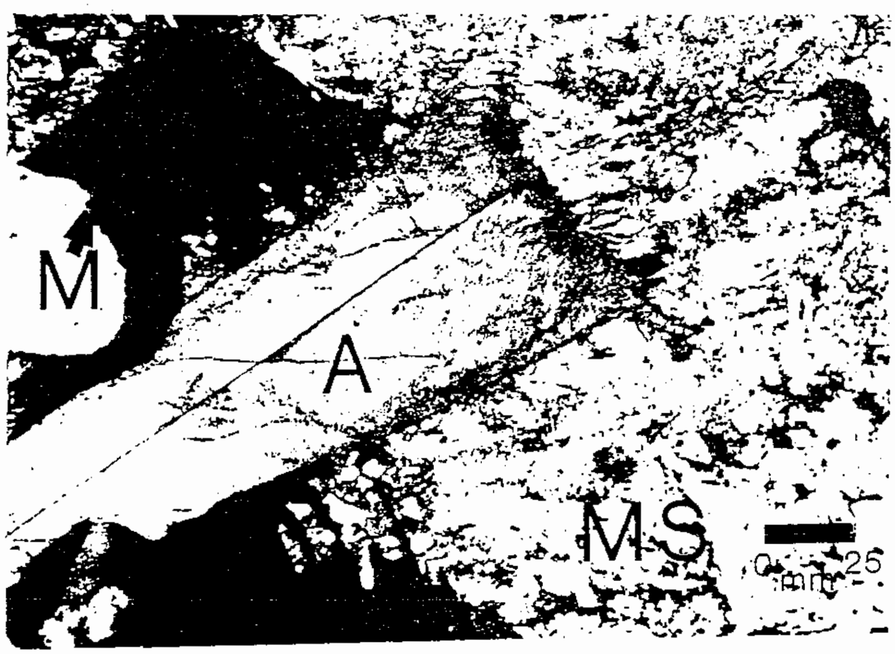


Figure 4. Photomicrograph of augite dolerite (stop 4); augite phenocryst bracketed by dendritic titanomagnetite extending into devitrified mesostasis (MS); 0.25 mm scale.

separating patches of subophitically intergrown pyroxenes and labradorite (Stop 2, figure 7 and 8). Primary pigeonite in the bronzite dolerite (Walker's stage 3i) is absent in bronzite dolerite (Walker's stage 3ii, above 110 feet). Early pigeonite dolerite of the Middle Stages (Walker's stage 5i) contains free pigeonite at the 365 foot level. Pigeonite occurs with augite as principle phases exclusive of orthopyroxene only in Middle Stage 5i and in the first pulse at Haverstraw.

Though there is considerable textural and chemical variation throughout (Steiner, et al., 1989b), the present preliminary survey found that pigeonite occurs as free grains or coarse intergrowths (not restricted to exsolution bands) over the entire traverse exclusive of the augite dolerite (described below). Since these pigeonite facies retain patches of variously recrystallized glassy mesostasis (highly visible in figures 7, 8 and 9), and since intervening orthopyroxene facies have yet to be defined, the pigeonite facies is tentatively assigned petrographically to Walker's Pulse 1 of Haverstraw. This necessitates, at the minimum, a westward revision of the interslab boundary in the northern section.

AUGITE DOLERITE

In the Upper Nyack Section, distinctive iron- and REE-enriched, chromium-poor augite facies with little or no pigeonite is enveloped by pigeonite-bearing facies on the east and a chromium-richer (100 to 250 ppm) pigeonite-poor facies on the west (Stop 6). The facies can be subdivided texturally into pegmatitic and other varieties but is distinctive in (1) in being a facies containing little or no low calcium pyroxene, and (2) retaining the glassy mesostasis patches of the pigeonite dolerite. The texture is locally porphyritic (figure 4), but appears to reflect an increased amount of interstitial mesostasis relative to subophiticly intergrown augite-plagioclase clusters. This gives a mottled appearance to some samples. The tendency for mesostasis to become concentrated away from clusters may produce gabbroic-aphanitic phase separations, as noted by Clay (1988) for a horizon in Haverstraw Quarry.

Even though certain facies of the Palisades become as coarse as plutonic varieties (grain size exceeding 1 mm), the tendency has been to retain the dolerite designation, as in the historical usage of the term pegmatite dolerite.

REE CHEMISTRY

The incompatible elements, inclusive of the rare earth elements, constitute a useful base line comparison technique for magma series. Figure 5 shows that the upper and lower contact facies appear to be derived from essentially identical basaltic magma.

This magma is REE enriched relative to the lower 125 meters of the Palisades at George Washington Bridge (figure 6; GWB data from

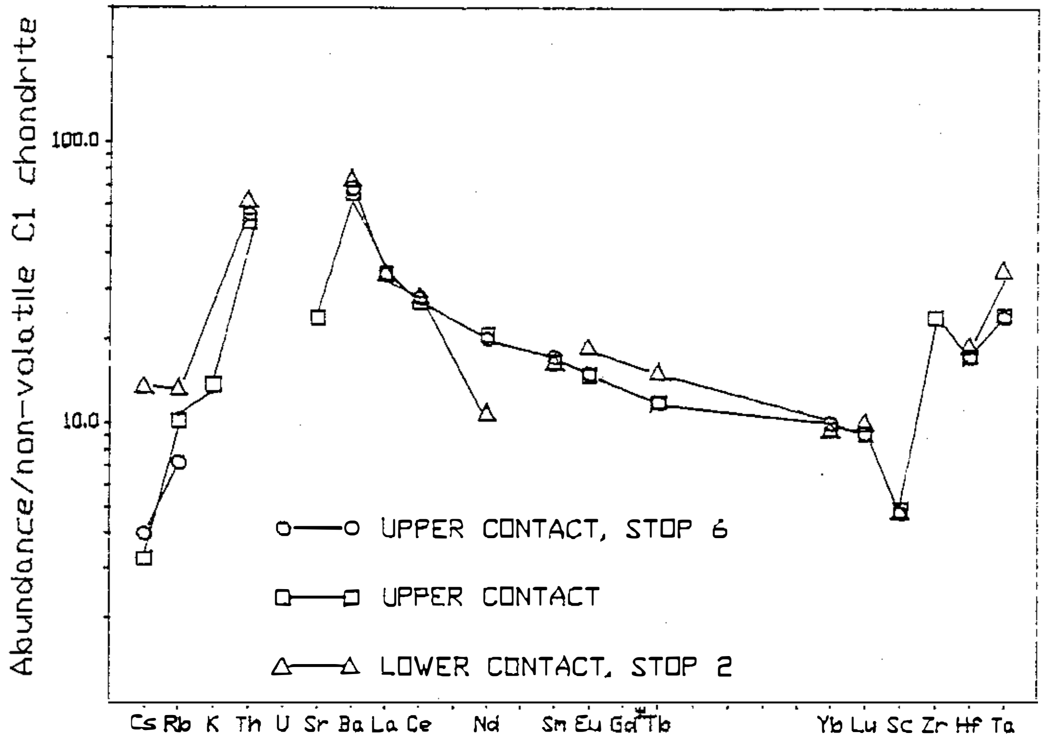


Figure 5. Incompatible element map of quench dolerite facies showing conformable profiles, except for Nd of the lower contact.

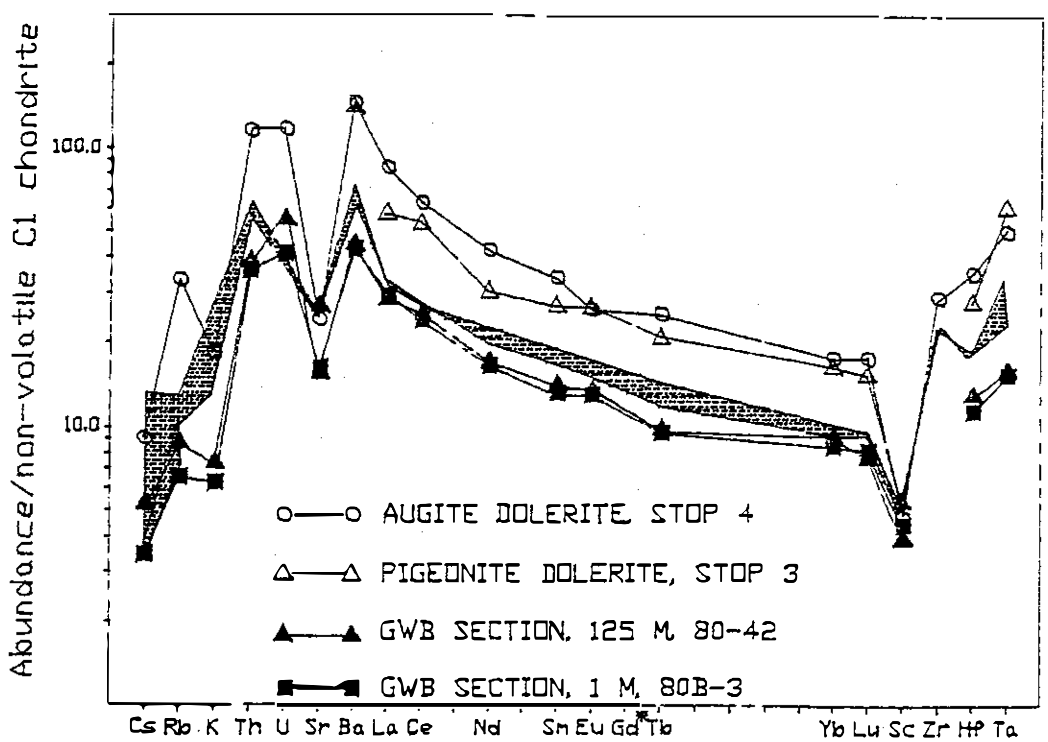


Figure 6. Trends of Nyack quench basalt (shaded) against local dolerites and George Washington Bridge samples at 1 and 125 meters.

Shirley. 1987). Thus, the REE chemistry of the Nyack magma is clearly inconsistent with quench magma of Walker's Pulse 1 or Pulse 2 in the southern Palisades.

The REE patterns of individual rocks in a fractionated suite are systematically offset from one another and comprise an array of non-intersecting curves. The inferred magma is in this sense a possible lateral derivative of magmas at GWB, or a more evolved representative of fractionation processes which gave rise to both varieties. Since both series possess essentially parallel trends, crustal processes are likely to have exerted the dominant control in the latter case (garnet fractionation in the mantle, for example, would generate crossing trend lines).

The Nyack quench basalts also lack the slight positive Eu anomaly of GWB contact lithologies which is suggestive of plagioclase settling at higher levels (Shirley, 1987), or a feature suggestive of pre-emplacement fractionation. The apparent lack of enhancement of the Europium anomaly and the consistency of the pattern over 125 meters indicates that Pulses 1 and 2 of Walker (1969) are indeed very similar chemically at the Bridge, and that additional plagioclase accumulation is not pronounced in the lower levels of Pulse 2 (Shirley, 1987).

ACKNOWLEDGMENTS

The author is indebted to T. Olson, R. Warner and D. Weiss for their constructive comments.

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ROAD LOG FOR FACIES AND CONTACTS OF THE UPPER NYACK SECTION

Cumul. Milage	Miles from Last Point	ROUTE DESCRIPTION
0.0	0.0	Exit left onto Route 59 West from Exit 11 of New York State Thruway (I-287).
0.4	0.4	Turn left at traffic light onto Route 9W North.
2.1	1.7	Turn right at traffic light onto Christian Herald Road (East).
2.4	0.3	Turn left at bottom of hill (no light) onto Old Mountain Road (East).
2.7	0.3	Turn left on Broadway.
3.6	0.9	Enter Nyack Beach State Park, curve around the toll booth and follow the paved incline up the hill to the left. Turn left at the top of the hill approximately 100 yards to the circular turn around and park cars.

STOP 1 NYACK BEACH STATE PARK - OLIVINE LAYER

The escarpment at the eastern edge of Nyack Beach State Park represents the western wall of an infilled stone quarry. On this rock face, at approximately 40 feet above the presently obscured basal contact, a subhorizontal "rotten" zone (first horizontal depression above tree line) can be observed. This zone presumably represents the last vestige of the olivine layer referred to by Walker (1969). It appears to fade northward of the present location.

A closeup view of this structure, which involves a certain risk due to the fractured nature of the rock infrastructure, reveals that the basal part of this horizon is undulose and uneven. It is marked by nodules of basalt enclosed in a weathered granular matrix. Samples of both material show that the composition of the matrix is nearly identical to the composition of early dolerite at thirty feet above the lower contact at Englewood Cliff (Steiner, in preparation). An analysis of one of the olivine bronzite nodules yield 10 percent MgO. This percentage is high for most facies of the Palisades, but substantially less than the 19 percent reported by Walker (1969) for the hyalosiderite facies.

Here, as elsewhere, it appears that cooling cracks are reasonably continuous through the rotten zone. This indicates that the rotten zone belongs to the same cooling unit as the overlying dolerite.

5.1	1.5	Retrace path to intersection of Christian Herald and Route 9W; turn right on Route 9W North.
5.3	0.2	Park in Piccolo Foreign Car Repair lot. Walk about 50 yards North to small escarpment behind trailers.

STOP 2 UPPER NYACK - LOWER CONTACT OF UPPER NYACK SECTION

The basal contact is well exposed along a small escarpment. Here, small, infilled cavities can be observed in the Triassic baked shales indicating that the sediments were not fully lithified at the time of intrusion, and the intrusion was relatively shallow.

A scramble up the escarpment for forty feet reveals small (3 cm) nodule-like protrusions in an otherwise uniform dolerite. These analyze to normal basalt (Steiner, in preparation). An olivine layer has yet to be discovered at this locality.

Backscatter electron imaging of the quench rock (figures 7 and 8) shows subophitic pyroxene and labradorite clusters surrounded partially by pools of interstitial glass. Titanomagnetite and minor plagioclase, sanidine, quartz, and residual glass populate the interstitial spaces (figures 7 and 8).

6.0	0.7	Continue North to top of hill and pull off the road. Be careful of the traffic. Interest and time permitting, drivers will proceed North to Rockland Lake South Lot.

STOP 3 UPPER NYACK - COARSE PIGEONITE DOLERITE

This coarse pigeonite facies is characterized by elongate (1 cm) augite laths with pigeonite cores or attached plates which run parallel to the augite prisms (figure 3). This facies is substantially enriched in REE relative to the contact facies, and shows a slight positive Europium anomaly suggestive of plagioclase accumulation (figure 6).

Time permitting, proceed 25 yards south on Route 9W to intersect a well marked service trail which drops sharply down the east bank. Samples have been taken from approximately 30 yards down the trail to the summit of Hook Mountain. From Hook Mountain, the tabular form of the Palisades is visible against the southern skyline. At Piermont, the topography changes dramatically in apparent response to a stepwise increase in the angle of westerly dip of the Palisades. Haverstraw Quarry is visible to the north. The rolling hills of the Triassic basin are framed by the Hudson Highlands to the east.

6.2	0.2	Proceed north to traffic light at entrance to

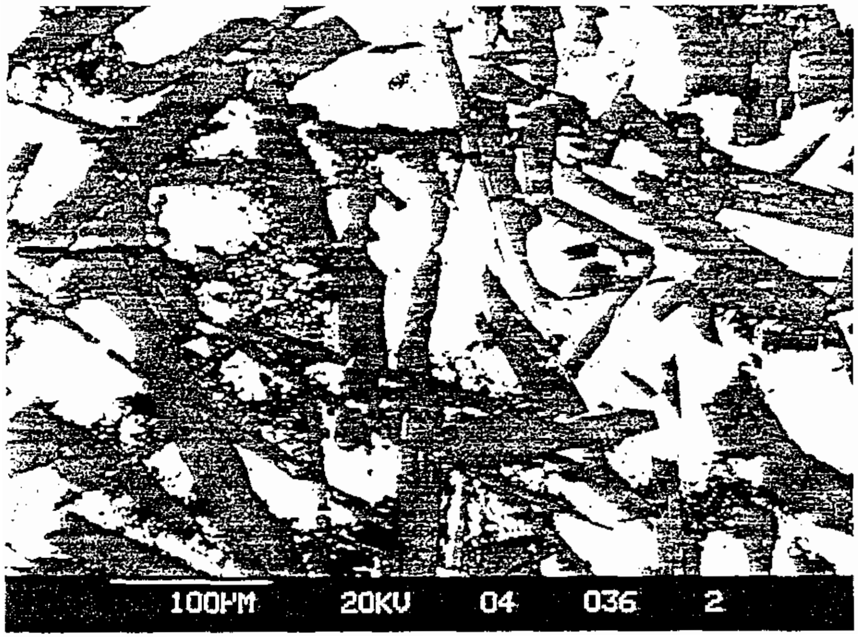


Figure 7. Backscatter electron image of quench dolerite, stop 2. Intersertal texture: subophitic intergrowths of pyroxene (light) and plagioclase (dark); interstitial pools of basaltic glass;

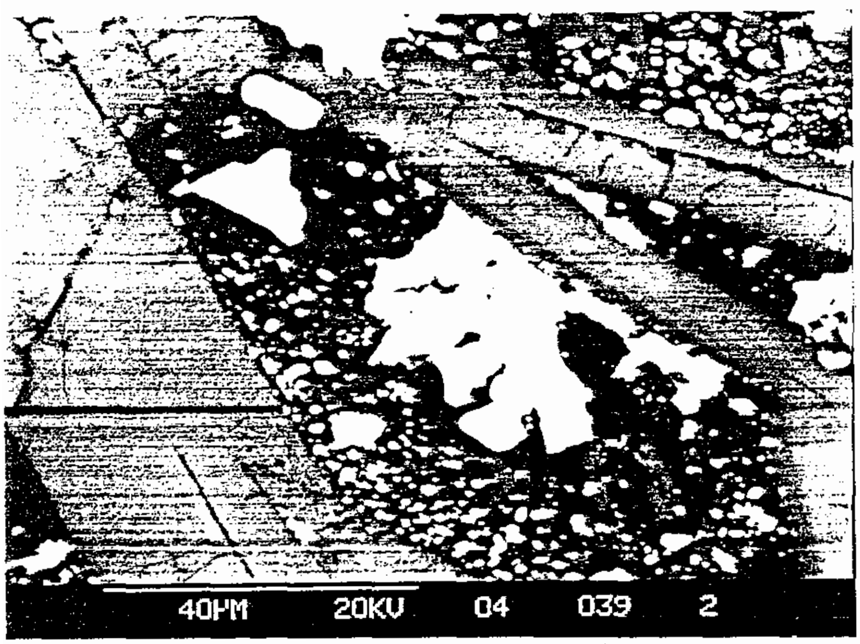


Figure 8. Enlargement showing interstitial area populated with titanomagnetite (white globular), plagioclase, quartz and glass.

		Rockland Lake Park, turn left onto Lake Road.
8.0	1.8	Turn sharply left and uphill onto Christian Herald Road just prior to the traffic light at the intersection with Route 303. Five streets meet at this intersection.
8.4	0.4	Turn left onto Herald Court and pull over to the curb at the first house on the right.

STOP 4 VALLEY COTTAGE - COARSE AUGITE DOLERITE

The best developed porphyritic texture (figure 3) in coarse augite dolerite occurs in the landscaped outcrop at the corner of Christian Herald and Herald Court. Slabs of samples taken prior to landscaping will be shown. PLEASE DO NOT DEFACE THE PROPERTY. DO NOT VISIT WITHOUT THE PERMISSION OF THE OWNER.

The iron-enrichment of the Palisades at this stop is due to abundant titanomagnetite in the groundmass. Titanomagnetite is often dendritic, and is considered to occur as both a primary and a quench feature. There is little indication that the magnetite was emplaced secondarily via an aqueous vapor or solution. Chemically, and to a certain extent petrographically (mesostasis replaces micropegmatite) this facies appears to be equivalent to the late stage ferrodolerite, fayalite granophyre and other facies of Walker (1969), except that the iron enrichment is not accompanied by iron rich ferromagnesian minerals, such as ferroaugite or fayalite. Ferroaugite rims may occur, but the bulk clinopyroxene is clearly augite.

The REE patterns (figure 5 and 6) are consistent with the derivation of this facies from the pigeonite facies through crystal liquid fractionation.

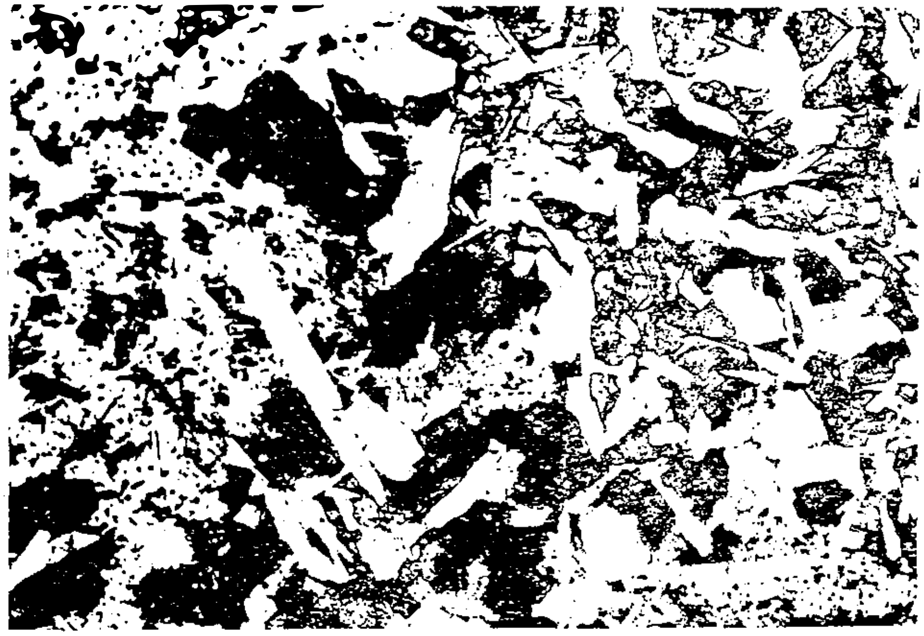
9.1	0.7	Turn left from Herald Court on Christian Herald and turn right onto Mountainview Road.
9.6	0.5	Turn right into asphalt turn around for houses set back from road. Leave cars and walk back to rocks exposed at last curve.

STOP 5 VALLEY COTTAGE - WEATHERED COARSE AUGITE DOLERITE

This is one of the few localities not in someone's yard which shows weathered coarse augite dolerite. Other localities occur sporadically along Mountainview Road.

10.2	1.6	Return to intersection of Lake Road and Route 303. Probably best to park behind the Chemical Bank. Walk across Lake Road to the Convenience Market on the Southeast corner.

STOP 6 VALLEY COTTAGE - UPPER CONTACT RELATIONS



0 .25 mm

Figure 9. Dolerite at stop 6 showing subophitic texture right of center grading into glassy mesostasis on the far left; 0.25 mm scale.

0 cm 1

Figure 10. Photograph of polished slab showing quench texture comprised of plagioclase and augite growing normal to the interface with pigeonite dolerite (lower quarter); 1 cm scale.

From the intersection of Route 303 and Lake Road, walk 20 yards north along the eastern side of Route 303 to examine the contact between diabase and sedimentary rock which dips northerly at about 45 degrees. The contact is very sharp at about 10 feet up from the sidewalk alongside the telephone pole. It is otherwise somewhat hard to observe the trend do to the tendency for the dolerite to reduce the iron in the sedimentary rock producing a black baked sandy shale.

As observed at Stop 2, the intrusive dolerite has vesiculated the sedimentary rock. At this stop the cavities are somewhat sharper and lack infilling. The baked zone buttresses the east-west trending slope which marks the position of the Palisades contact as it extends toward Rockland Lake.

The texture of the contact rocks is microporphyritic with augite microphenocrysts. Within five feet of the contact the texture is ophitic, and beginning perhaps fifteen feet south, occasional recrystallized patches of interstitial mesostasis appear (figure 9).

Proceeding South, the diabase appears to plunge beneath the sedimentary cover. About 0.5 miles north of Lake Road, the contact is rediscovered on the upper potions of the exposed slope which is inset 60 yards east of route 303. The dolerite dips roughly west at about 45 degrees underneath a veneer of backed shales. The intrusive contact is therefore complex, wending in a curvilinear fashion along its margin in the general form of poorly exposed coalesced domes or localized arches.

To see quarried rock of horizon 1 (Steiner, et al. 1989a,b) drive west on Lake Road 0.3 mi to the Valley Cottage Fire Station and park along the West edge of the lot. The large blocks along the stream bank contain pegmatite dolerite clusters with sprays of augite locally reaching 6 cm.

10.6	0.4	Proceed west on Lake Road to traffic light on Kings Highway and turn left.
11.2	0.6	Turn right at stop sign on Crusher Road.
11.7	0.5	Park at company office at base of hill.

STOP 7 WEST NYACK - COARSE PIGEONITE DOLERITE, INTRUSIVE RELATIONS AND LAVA DOME

The suggestion of a domal configuration at many points along the western contact is strongly supported by the lava dome situated southwest of the main quarry buildings. This dome was well exposed in the Fall of 1988, but has since been partially buried by crushed quarry rock.

Several facies are transected when walking or driving to Stop

8. In particular, trondhjemitic and granophyric varieties of the coarse dolerites, ultramafic segregation veins (Steiner, in preparation) and secondary "flowers" of chalcocite against malachite ovals along fracture systems may be visible. Unfortunately, many of these features have been completely removed by the quarrying operation.

STOP 8 WEST NYACK QUARRY - COARSE PIGEONITE DOLERITE WITH PEGMATITIC DOLERITE (COARSE AUGITE DOLERITE) PODS

Along the south end of the westernmost wall at the second level from the top, a coarse intrusive facies (H1, Steiner, et al., 1989a,b) invades diabases comparable to those exposed at Stop 6 along a narrow 12" dike. The dike shows an approximately 1" gap through which the magma welled up into adjacent rocks creating a balloon-like profile. This dike apparently represents an offshoot of the major contact which runs subparallel to the intersecting north wall. Pegmatite dolerite, comparable in many respects to the coarse augite dolerite of Stops 4 and 5 characterizes the general zone separating the two lithologies.

OPTIONAL STOP 9 WEST NYACK - SECOND UPPER CONTACT AND REDUCED ZONE

0.0	0.0	Intersection of Routes 303 and 59.
0.4	0.4	Turn left onto dirt road just past Fesco Fence, over find Greenbush Ave. off Route 59. Park and walk up slope to prominent outcrop.

Here, a dip slope of the Triassic arkose (fossil fish locality) which is visible below abuts the Palisades. Portions of the metamorphosed sedimentary rock form a veneer over and are partially crosscut by offshoots of the dolerite. The tendency for the reduction of iron is again visible in the sedimentary rock.

Return to Route 303 and proceed north to intersection with N.Y.S. Thruway.

ROCK SLOPE STABILITY: DESIGN, CONSTRUCTION AND REMEDIAL TREATMENT

CLAYTON L. BOLTON JR.
Engineering Geology Section
Soil Mechanics Bureau
New York State Department of Transportation

INTRODUCTION

The design and construction of rock slopes with emphasis on stability began with the ability to drill and blast on an alignment whereby the final slope product was sound and durable and free of potential rock fall hazards. The first attempt at constructing a planar surface in rock was not for a highway but for construction of the sides of an intake channel for the "Niagara Power Project" in Niagara Falls, New York. The extra effort to develop a vertical planar surface for the intake channel was not to prevent potential rock falls but to establish a surface whereby the concrete channel wall liners could be built with a minimum overrun in concrete quantities. (When one considers that the majority of the listed 38,846,000 cubic yards of excavation was in rock, and that several million square feet of rock face had to be prepared within a tolerance of six (6) inches, it is easy to see that quite a problem was presented to the various contractors to keep from overbreaking beyond the payline, or "B" line, as it was called, with conventional methods of drilling and blasting). The method for constructing the planar face was named "presplitting".

The results of "presplitting" in constructing the vertical planar rock wall were so good that it soon was considered to be worthy of utilization in establishing minimum maintenance, hazard free rock faces for highways, tunnels, building foundation excavations and other locations where rock stability is a major concern. New York State Department of Transportation incorporated "presplitting" in its EXCAVATION SPECIFICATIONS on August 27, 1970 by inclusion in Addenda 49 to the January 2, 1962 PUBLIC WORKS SPECIFICATIONS. Once the "presplitting" tool was in place for constructing a stable rock slope, the Engineering Geologist could then put his knowledge to work to establish a suitable rock slope design. This method of construction leading to proper design is another example of empirical knowledge's ability to make theory, fact.

ROCK SLOPE DESIGN

Engineering geologists must consider two major items in designing slopes for a proposed rock cut, stability and cost. The engineering geologist must also use every method available to obtain the structural discontinuities present in the bedrock before designing the slope. The most reliable information may be obtained from outcroppings present in the area of the design. Often however, the outcrops are one dimensional and without a cross section view of substantial extent, the structural information obtained may be misleading. Oriented rock cores from strategically located drill holes, video taped information from television borehole cameras, refraction seismic data, remote sensing equipment and electrical resistivity are methods which can be employed in confirming surficial structural detail and/or portraying attitudinal changes which must be known by the designer to establish a structurally sound slope design.

Plotting the joint, bedding, shear zones and cleavage planes on a stereo-net will enable the designer to detail a stable slope for any particular rock cut.

The cost of excavation for the alignment of any highway project in rock is usually more expensive than for any other type of material. The present bid prices received in New York State for rock excavation average $30/cubic yard whereas soil excavation averages $6.50/cubic yard. Many other items such as right-of-way costs, environmentally protected area (swamp) replacement and special cases (i.e. slum evasion) take precedence in corridor selection over avoidance of rock excavation. Excavation and embankment balance is not as important a factor in highway corridor selections as it once was. The material balance is however a prime consideration in Westchester County since no soil borrow areas are available.

Rock slope designs in New York State range from vertical (rarely) to a one vertical on one and one quarter horizontal (38°-40') which approaches the stable angle of repose of talus (37°-30'). Common slope angles of three vertical on one horizontal (71°-34'), two vertical on one horizontal (60°-15'), three vertical on two horizontal (56°-19') and one vertical on one horizontal (45°) are utilized to enable construction of accurate templates for the control of presplit drill alignment.

Vertical drill holes are rarely used and never in slopes greater than 15 feet in height since they create an illusion of toppling causing drivers to inadvertently drift away from the rock face and into oncoming traffic. Vertical design in shallow limestone cuts is sometimes used to avoid intersection of the standard vertical jointing by presplit drill holes. Explosively generated gasses intersecting the vertical joints provides pressures which cause sliding of limestone blocks along its bedding planes toward the open rock cut face. The resulting product is an unstable slope rock.

The one vertical on one horizontal slope is used for stable rock cuts in competent shales, siltstones or bedrock containing structures which dip toward the free face on a 45° inclination.

Rock slopes are constructed on the flatter talus slope if the material to be excavated is deeply weathered and/or fractured. Poor quality horizontally bedded shales will maintain stability on the talus slope. Both the deeply weathered rock and the poor quality shales will achieve total stability on the talus slope angle provided seeding is included following completion of construction. Pre-splitting is not required for the one vertical on one and one quarter horizontal slope since 1) it is extremely difficult to drill on such a flat an angle; 2) the rock can usually be broken by mechanical ripping or light explosive loads in vertical drill holes before final grading with a dozer on the slight incline and 3) seeding covers any discontinuities while greatly increasing the stability and enhancing the esthetics.

The intermediate slope inclinations (3V on 1H, 2V on 1H, 3V on 2H) are used based upon their stability relationship with the inherent structures present in the bedrock to be excavated.

ROCK SLOPE CONSTRUCTION

Two major types of rock slope construction are performed on highway projects, new locations and trim cuts. Trim cutting is for stabilizing previously constructed slopes.

The slopes for rock cuts on new locations are more difficult to design since it involves programming a field evaluation of the site; establishment of a drilling program to delineate the pertinent structures present in the bedrock; plotting the structural features on a stereo-net and finally, establishing the final slope inclination. The design of trim cuts usually is quite straight forward since virtually all of the structural features are available for direct measurement. Trim cuts are made to remove unstable blocks of fractured rock from the slope. This unstable condition was created during the original rock cut blasting construction by excessive explosive gasses generated by standard fragmentation blasting methods employed prior to the advent of "presplitting".

The second most important consideration to be included in trim slope design (slope inclination is first) is to insure that the slab of rock to be removed from the face of the existing slope be a minimum of five (5') feet in width. The control point for the slope face slab offset location is a point a minimum of five feet in back of a point on the slope which is the furthest distance from the centerline of the roadway along the rock slope cut. This point becomes the continuous parallel slope offset throughout the length of the rock cut.

The five foot thickness is required to assure that the presplitting charges break between the three foot spacing of the parallel presplit drill holes rather than towards a closer free face on the slope. This will provide a uniform planar final slope. It also greatly reduces the potential for fly rock.

Construction begins once the contractor is selected. Usually excavation is one of the first stages of the contract work to be performed following the removal of vegetation and topsoil. Rock excavation operations are not seasonal since the work is not limited by temperature restrictions.

Prior to the commencement of drilling operations on any contract, a preblasting meeting is essential. The purpose of the meeting is to:

1 - Learn the date of drilling equipment mobilization.

2 - Establish the procedure for controlling the drill steel alignment in drilling the presplit holes on the design slope inclination. Drill alignment is the most important consideration since the stability of the slope is totally dependent on achievement of the proper slope inclination.

3 - Determine the type and weight per linear foot of the presplitting explosive.

4 - Locate the types of utilities and structures nearest to the blast site.

5 - Specify the maximum pounds of explosives to be detonated per 25 millisecond delay period to insure that blasting induced ground vibrations do not damage any of the utilities and/or structures in the vicinity of the blast site. (Explosive loading limits guide-line follows NYSDOT Standard Specifications for Rock Excavation.)

6 - Determine if a potential for fly rock is is present and if blasting mats are required to protect personnel and property in the blast area.

7 - Insure that the blaster is in compliance with all Local, State and Federal permits, codes, laws and licenses required.

8 - Review the project specifications to enable the blaster and contractor be aware of what they are required to do to construct minimum maintenance, hazard free rock slopes. (Current NYSDOT Standard Specifications follow preblasting meeting subject description.)

9 - Agree upon the location and limits of the rock slope test section area.

10 - Establish the location and types of blasting warning signs, the preblast warning signals and the placement of flag personnel for traffic control.

NYS DOT - STANDARD SPECIFICATIONS of January 2, 1981

as amended by ADDENDUM NO. 1 of December 9, 1982

Section 200

EARTHWORK

§203-3.05 Rock Excavation

Attention is directed to § 107-05, SAFETY AND HEALTH REQUIREMENTS, concerning rock drilling and blasting work.

Presplitting is required where the design rock slope is one vertical on one horizontal or steeper and the vertical height of the exposed rock slope exceeds five feet. Ripping will not be allowed within ten feet of a slope that requires presplitting. Test sections will be required at the outset of presplit drilling and blasting operations for the evaluation of the presplit rock slopes by a Departmental Engineering Geologist. The Contractor will be required to completely expose the presplit rock face in the test section for evaluation prior to any further presplit drilling.

All rock slopes shall be thoroughly scaled to the satisfaction of the Engineer. For rock excavations involving multiple lifts, scaling of upper lifts shall be completed prior to drilling and fragmenting of lower lifts. Scaled rock slopes shall be stable and free from possible hazards of falling rocks or rock slides that endanger public safety. If, after proper scaling, such conditions still exist, a determination of the cause will be made by a Departmental Engineering Geologist and if it is determined that the conditions are the result of poor workmanship or improper methods employed by the Contractor, the Contractor shall provide approved remedial treatment, at no expense to the State. Such treatment may include, but is not necessarily limited to, laying back the slope, rock bolting, or shotcreting. In no case shall the subgrade be trimmed prior to the completion of the scaling operation at any location.

A. Presplitting. Prior to drilling presplitting holes, the overburden shall be completely removed to expose the rock surface along the presplitting line. The methods of collaring the holes to achieve proper inclination and alignment shall be approved by the Engineer.

The presplitting holes shall be a maximum four inches in diameter, spaced not more than three feet center to center along the slope, and drilled at the designed slope inclination for a maximum slope distance of 60 feet. When excavation operations are conducted in multiple lifts, the presplitting holes for successive lifts may be offset a distance of not more than three feet for a design slope of one vertical on one horizontal and not more than one foot for slopes of steeper design; however, a presplitting hole shall not be started inside the payment line. If presplitting is conducted in lifts, each lift shall be approximately of equal depth. All presplitting holes shall be checked and cleared of obstructions immediately prior to loading any holes in a round. All presplitting holes shall be loaded with a continuous column charge manufactured especially for presplitting which contains not more than 0.35 pounds of explosive per foot. The top of the charge shall be located not more than three feet below the top of rock. A bottom charge of not more than three pounds of packaged explosive may be used; however, no portion of any bottom charge shall be placed against a proposed finished slope. Each presplitting hole shall be filled with No. 1A Crushed Stone Stemming meeting the gradation requirements of § 703-02, Coarse Aggregates. The presplitting charges shall be fired with detonating cord extending the full depth of each hole and attached to a trunk line at the surface. Detonation of the trunk line shall be with blasting cap(s) and shall precede the detonation of fragmentation charges within the section by a minimum of 25 milliseconds. Pre-splitting shall extend for a minimum distance equal to the burden plus three feet beyond the limits of fragmentation blasting within the section.

B. Fragmentation Blasting. Fragmentation holes, or portions thereof; shall not be drilled closer than four feet to the proposed finished slope. Where presplitting is required, fragmentation holes adjacent to the pre-splitting holes shall be drilled parallel to the presplitting holes for the full depth of the production lift at a spacing not exceeding the spacing of the production pattern. Only packaged explosives shall be used ten feet or less from a design slope which requires presplitting regardless of the construction sequence.

Fragmentation charges shall be detonated by properly sequenced millisecond delay blasting caps.

EXPLOSIVE LOADING LIMITS

In the absence of more stringent requirements, the maximum quantity of explosives allowed per blast shall be based on a maximum particle velocity of 1.92 inches per second at the nearest structure to be protected. In the absence of seismic monitoring equipment, the following explosive loading limits shall apply:

<u>Distance Equal to or Less than Two Hundred and Twelve (212) Feet from the Nearest Structure</u>

1. When the distance from the proposed blasting area to the nearest structure to be protected is six (6) linear feet or less, no blasting will be allowed.

2. When the distance between the blasting area and the nearest structure to be protected is greater than six (6) and equal to or less than fifteen (15) linear feet, a maximum of one quarter pound of explosive per delay period* blasting cap shall be allowed.

3. When the distance between the blast area and the nearest structure to be protected is greater than fifteen (15) and equal to or less than two hundred and twelve (212) linear feet, a Scaled Distance of thirty (30) shall be utilized to determine the maximum pounds of explosive allowed per delay period* blasting cap. The Scaled Distance Formula is as described below:

$$SD = \frac{D}{\sqrt{E_{Max}}}$$

OR

$$E_{MAX} = \left(\frac{D}{SD}\right)^2$$

Where:

SD = Scaled Distance

D = Distance from blast area to nearest structure to be protected in linear feet

E_{MAX} = Maximum pounds of explosive per delay period* blasting cap

<u>Distance Greater than Two Hundred and Twelve (212) Feet from the Nearest Structure</u>

4. When the blaster elects to utilize more than fifty (50) pounds of explosive per delay period* blasting cap, a seismograph shall be employed to monitor the blasting vibrations generated. The initial loading shall be computed using a Scaled Distance of thirty (30). The resulting particle velocity measured by the seismograph shall be evaluated by a Department Engineering Geologist. His evaluation shall be the basis for adjusting the Scaled Distance.

No separate payment will be made for this work. The cost shall be included in the appropriate excavation item. The above requirements shall in no way relieve the Contractor of liability for any damage incurred as a result of his blasting operations.

*a delay period shall be a minimum of twenty five (25) milliseconds.

PRESPLIT DRILLING, BLASTING AND TEST SECTION REVIEW

A. Presplit Drilling: The Engineering Geologist must be present at the rock slope test section area on the date of mobilization. It is very important that the Contractor's driller and the project inspector learn from the start exactly what is expected of them to insure that drilling alignment control agreed upon at the preblasting meeting is performed.

The following drilling check list is provided to assist the driller and inspector in achieving the final slope as designed. Establish that:

1 - Overburden is stripped from bedrock along the top of the presplit line. Insure that the bedrock surface is not overexcavated as in the case of weak shales.

2 - The drill steel is straight and in satisfactory condition.

3 - The plumb line for orienting the drill steel alignment is correctly located on a line parallel to the presplit line. (Preblast meeting agreement.)

4 - The slope inclination template is the proper dimension and that a minimum two foot long carpenters level is attached to the template. (Preblast meeting agreement.)

5 - The drillers assistant has achieved the proper drill steel alignment as the drill bit is collared by the bedrock surface. (The alignment can only be assured at this time since once the drill progresses into the rock it is literally on its own.)

6 - The drill hole is of the proper depth (including sub-drilling) for each hole.

7 - The presplit drill holes are located on three foot centers.

8 - The driller is using a carbide insert cross bit rather than button bits and a solid drill steel rather than spiral drill steel. The button bits and spiral drill steel are not specifically banned from use but are strongly suspect to cause wander in deep (30 feet plus) drill holes. Drill steel wander is checked for on exposure of the test section face.

9 - The closest row of production (fragmentation) holes to the presplit line is drilled no closer than four feet to and on the same angle as the presplit slope holes.

B. Blasting: The following check list is presented to enable the blasting inspector to determine that the initial test section blast and all future presplit blasting operations are conducted in a manner whereby the best possible rock slope alignment may be achieved.

The inspector shall check:

1 - The depth of each presplit hole to insure that no blockages are present prior to explosive loading. If explosives are loaded following hole check and the adjacent hole is blocked, you would be unable to bring in a drill to clear the hole.

2 - The presplit explosive weight to insure that it is not heavier than the specified maximum weight of 0.35 pounds per linear foot. It is recommended that the inspector count the number sticks of explosive, multiply by the standard length of each cartridge to obtain the total cartridge length of the box and divide the box weight by the cartridge length. This is a good way to check the manufacturers quality control and to prevent manufacturers errors from the disrupting the proper construction of the presplit slope. The explosives is also very important for determining the maximum pounds of explosives per delay period blasting cap when blasting vibration control is a consideration.

3 - To insure that the presplit line is loaded a minimum distance of nine feet in advance of the closest loaded production hole in the section. This insures that the greater quantity or gas generated by the larger production explosives does not follow an open joint, fracture or bedding plane to a point behind the intended presplit slope and thereby disrupt the slope continuity.

4 - To see that the earliest sequenced delay detonator is affixed to the presplit trunk line detonating card. This will insure that the presplit slope is blasted prior to any production hole by a minimum of 25 milliseconds.

5 - For use of free flowing explosives (ANFO, Prills or Water gels) in the production blasting operations and insist that none be used in any production holes located within 20 feet of the presplit slope.

6 - That the stemming material to be used is a #1A crushed stone rather than crushed gravel. Crushed gravel has rounded edges and shotguns out of the hole rather than locking together to keep the presplit explosive gasses in the hole to split the bedrock.

C. <u>Test Section</u>: The establishment of the test section is of enormous value to the Engineering Geologist.

1 - The test section exposes all discontinuities present in the bedrock. Since even the most advanced design exploration methods cannot reveal every feature

present, the test section will enable the Engineering Geologist to determine if the slope will be stable as designed. If it is determined upon evaluation of test section that the slope is unstable, the Engineering Geologist can change the slope design to one which will be stable. The contractor will not be detained for more than a day and can than proceed with the rock slope construction on the new slope inclination.

2 - The test section examination will also reveal if the presplitting explosive is too heavy or too light as evidenced by the presplit hole drill butt trace which remains.

3 - The drill butt trace also provides direct evidence if drill attitude at the time of initial drill set up and any drill wander as it occurs from the top of the slope to the bottom.

Direct measure at the top indicates if the initial alignment was correct. Drill butt traces which curve in parallel paths can be attributed to spiral drill steel. Diverging or crossing drill butt traces can be blamed on use of button bits or initial improper drill alignment. Diving drill steel which causes an over steepened unstable rock slope can be related to alternating horizontal hard-soft beds (i.e. Sandstone - Shale); excessive down pressure on the drills by the drill head causes drill steel to knuckle downward in a hard bedrock medium and finally gravity can be found to be the culprit in soft shales and siltstones. A diving drill steel can be corrected by the driller commencing the drill at a flatter angle at the surface to achieve the proper slope at the toe.

Rarely does a drill kick out to cause a flatter slope. If this does occur, it is usually due to improper driller setup. Occasionally a drill bit will intersect an open void whereby the drill strikes the bottom surface of the void on an acute angle which dips toward the roadway.

ROCK SLOPE STABILIZATION

We have all seen signs along highways throughout the country which state "Caution - Falling Rock Zone". Falling rock is basically a seasonal phenomena. Rock which is cyclically moved toward a free face by frost wedging, ice buildup, heavy rain fall and snow melt occurs usually in late autumn and early spring when its center of gravity moves over the edge of its adjacent precipice. This is only one small incident which undoubtedly will miss any auto on the roadway and if it does hit a vehicle, the operator treats it as an "Act of God" and complains to no one but his insurance company. The fact that there is a minimum of complaints registered with highway agencies is why little is done to correct them.

A recent (January, 1988) rock slide on the New York State Thruway at Elmsford resulted in the death of one woman and injury to another. This event and the publicity generated by the news media prompted political pressure to be heaped on the New York State Thruway Authority and the New York State Department of Transportation to delineate unstable rock slope problem areas and stabilize them.

The New York State Thruway Authority budgeted $30 million for rock slope stabilization operations to be conducted over the next several years. A great amount of this type of work has been completed in the 20 month period since the accident.

There are six major types of rock slope failures and each has occurred in New York State to various degrees. The remainder of this text will describe the failure types, the occurrence location(s) and the recommended remedial treatment which will correct the situation.

1 - <u>Plane Failure</u>:

Plane failure occurs when a geological discontinuity, such as a bedding or foliation plane strikes parallel to the slope face and dips into the excavation at an angle greater than the angle of friction.

A plane failure on a bedding plane is found along Route 52 west of Ellenville on Shawangunk Mountain in the Shawangunk quartz pebble conglomerates and metamorphic quartzite grits. (Stop 7) A plane failure on a foliation plane in gneiss along Route 4, between Fort Ann and Comstock in Washington County.

The most economical remedial treatment for stabilization of this condition is rock bolting. The most recent innovation in the rock bolting arena is the introduction of a resin rock bolt anchor system. A hole capable of receiving a three quarter inch to one and one half inch diameter grade 150 reinforcing steel bolt threaded on one end is drilled to a depth three feet beyond the failure plane.

Cartridges of a fast set (three to five minute set time) are loaded in the bottom of the hole followed by slower setting cartridge of resin (15 to 20 minute set time). The bolt is inserted and spun with an impact hammer to mix the two component resin cartridges for a full minute mix time. The pull tester is then attached to the rock bolt and after the fast set resin is hardened, the bolt is pretensioned to the specified strain pressure and maintained until completion of the slow set time of hardening. The pull tester is then released and removed and appropriate plate, beveled washer and nut are attached and snug tightened. The strain resistance of the resin has not been determined to date. They have been on the market for approximately ten years and no failures to date have been reported. It is however the best anchorage in rock devised to date. An epoxy resin was tested previously but was found to be unable to maintain anchorage in wet holes.

The purpose of the rock bolt is to increase the coefficient of friction along the shear plane thus preventing initial movement of the block to be retained. Should the block move, the bolts could not hold it and would either pull out or snap like toothpicks.

2 - Wedge Failure:

When two discontinuities strike obliquely across the slope face and their line of intersection daylights in the slope face, the wedge of rock resting on these discontinuities will slide down the line of intersection, provided that the line of intersection of this line is significantly greater than the angle of friction.

A major wedge failure on Interstate Route 84 near the top of Hosner Mountain in southeastern Dutchess County on April 15, 1982 closed the three eastbound lanes of the highway. An estimated 2,500 cubic yards of rock was subsequently removed. The cleanup was completed by the State and two contractors in a three week time period. It was determined that the slide was caused by heavy rain fall during the week prior to the slide. The rain decreased the coefficient of friction on the slide plane and created excessively high hydrostatic pressures which triggered the failure. (Stop 6)

3 - Circular Failure:

When the material is very weak, as in a soil slope, or where the rock mass is very heavily jointed or broken as in a rock fill, the failure will not be defined by a single discontinuity surface but will tend to follow a circular failure path.

Circular failures are found in the clays of Troy and Central Albany County; in the talus slopes in the high peaks region of the Adirondack Mountains and in the diorite talus along the west side of Route 9W in Haverstraw, Rockland County.

Circular soil failures are often balanced with a counter weight berm and soil anchors and weep drains to move the water away from the failure plane. Talus failures are counteracted with retaining walls.

Circular failure in a deeply weathered gneiss will be seen at West Point at the intersection of Route 9W and 218. (Stop 2) At the West Point failure area, a gabion catch wall constructed along the front edge of a bench is used to catch rather than correct the problem.

4 - Critical Slope Height versus Slope Angle Relationship:

A high steep slope is certainly much more unstable than a low flat slope. The factor of hydrostatic pressure and frost wedging created when rain fall lands on the high steep slopes causes major rock falls and slides to be initiated. The deep tension joints created by stress release following Pleistocene glacial retreat left major repositories for water buildup and resulting hydrostatic and frost wedging to move enormous blocks of rock from the top of the high steep rock faces.

A major example of this condition may be found along Route 218 - Old Storm King Highway in eastern Orange County between West Point and Cornwall along the west bank of the Hudson River. Trim blasting at the top of slope and rock bolts have been used in an attempt to reduce the major falls but small spalls of rock from the face nearly covered the highway when it was closed for the limited rock slope remedial treatment. Rock falls of such enormous magnitude located as far as 1000 linear feet west of Route 218 fall from the tops of cliffs over five hundred feet in height into a huge talus pile. Desk sized talus splash material has reached Route 218. No treatment or catch scheme is available to deal with this situation.

5 - Toppling Failure:

A toppling failure occurs when a block of rock which is higher than it is wide is positioned on a slope which is intersected by one or more joint planes which dip steeply into the slope of the rock face. The coefficient of friction is approaching zero along its base. Water penetrating the steeply dipping joint plane will easily dislodge the block and cause it to topple. Normally the joint plane is located in a relatively parallel plane with a whole family of joints. Once the initial block topples, a domino effect is established.

A toppling failure condition is to be found along Route 97 at Sparrow Bush in southwestern Orange County.

Rock bolting of substantial length is recommended to increase the width of the base of the block to exceed the toppling moment. Bolt lengths would increase proportionately with the block height.

6 - Ravelling Slopes:

Small spalls of rock which are found at the base of all rock slopes are caused by two prime factors, cyclical wetting and drying and/or freezing and thawing and deterioration of the cement matrix due to weathering processes.

This is not a major problem for highway slopes as it rarely causes more damage than a small dent or scratches in vehicles. Vehicles grind them to powder and snow plows sweep them off the highway into the ditch. Larger sized spalls of the size of baseballs to basketballs may be restrained and prevented from reaching the highway by a wire mesh screen draped over the slope. The screen is strung from cables attached to rock bolts. This method of correction has been applied by the New York State Thruway Authority at several sites located between Newburgh and Albany. The first application of the wire mesh draped over the slope was used on Route 52 east of Ellenville in Ulster County in 1975. (Stop 8)

Several types of rock catch methods have been utilized to contain or deflect problem rock falls without correcting it. The most common methods are a collection ditch at the toe of slope, rock catch fences and double corrugated beam guide railing.

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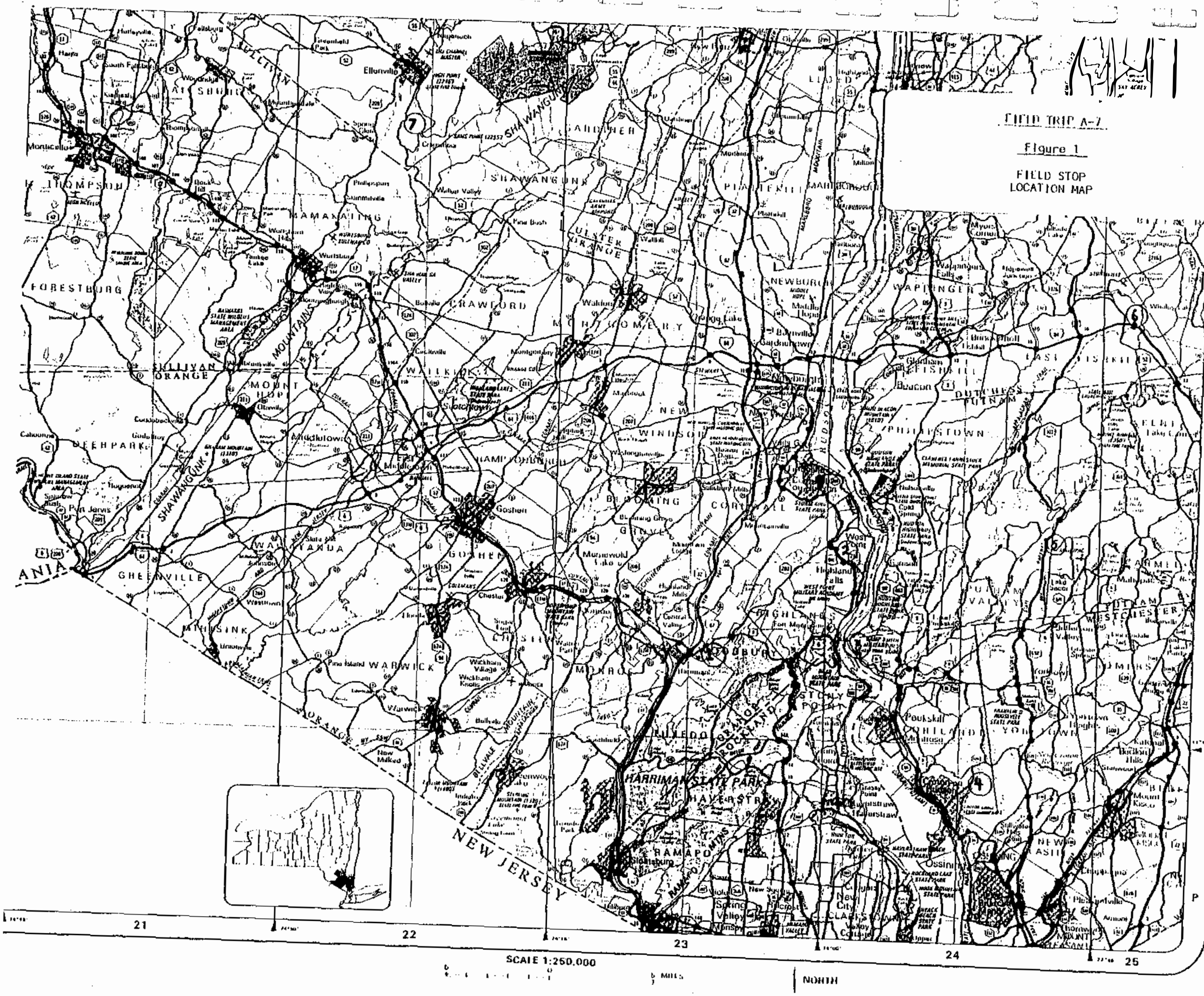
ROAD LOG FOR ROCK-SLOPE STABILITY: DESIGN, CONSTRUCTION, AND REMEDIAL TREATMENT TRIP A-7

	Cumulative Mileage	Miles From Last Point
Left onto Bennett Street from OCCC parking lot.	0.0	
Right at stop sign onto Route 17M east.	0.2	0.2
Right onto ramp for Route I-84 east.	1.9	1.7
Take Exit 4E to Route 17 east.	5.5	3.6
Take Exit 131 to Routes 6, 17 and 32.	22.8	17.5
Turn right at light at end of ramp.	23.1	0.3
Turn left at first light onto Route 6 east.	23.3	0.2
Enter first parking area on left.	25.2	1.9



FIELD TRIP A-7

Figure 1

FIELD STOP LOCATION MAP

Location Marker

STOP 1: Slope has two (2) lifts. A few bolts are in the top lift across from the west end of the parking area. There are more bolts near the location marker.

6
8301
2362

Leave parking area and continue east on Route 6.	25.3	0.1
Bear left on Route 292 north.	26.4	1.1
Bear right onto Route 9W south.	33.3	6.9
Park on shoulder just past exit for for Route 218.	34.5	1.2
Walk back to the exit ramp.		

Location Marker

STOP 2: Slope has two (2) lifts. A gabion wall is installed on the bench. There are numerous bolts in the upper face. A large dike cuts through the center of the face.

9W
8302
1047

Continue south on Route 9W to traffic circle.	38.9	4.4
Take third exit off circle to Route 6 east (Bear Mt. Bridge).	39.0	0.1
Pay Bridge Toll.	39.1	0.1
Turn right at east end of Bridge to stay on Route 6 east	39.6	0.5
Pull into parking area on right.	40.4	0.8
Walk back toward Bridge.		

Location Marker

STOP 3: There are several areas between the the parking area and the Bridge which have rock catchment fences, rock bolts and/or buttresses.

6
8703
1009

Continue east on Route 6 to traffic circle.	43.2	2.8
Take first exit off circle towards Routes 6 east, 9 south & 202 south.		

Turn right at end of bridge onto Route 9 south toward Peekskill.	43.4	0.2
Continue south on Route 9 to exit for Routes 9A & 129 to Croton-on-Hudson.	51.0	7.6
Turn left at bottom of ramp onto Route 9A south.	51.2	0.2
Turn left at second (2nd) light onto Route 129 east.	51.4	0.2
Intersection of Route 129 and Old Post Road.	51.8	0.4
Downtown Croton-on-Hudson on left. (Pit Stop?)		
Continue east on Route 129 and park at entrance to Croton Gorge Park.	53.5	1.7
Walk east on Route 129 to work zone on westbound side of road.		

<u>Location Marker</u>

<u>STOP 4</u>: Rock Slope Stabilization Project. Slope is being trimmed to an inclination of 1 vertical on 1 horizontal.

129 8701 1022

Continue east on Route 129.		
Cross new bridge over New Croton Reservoir.	55.2	1.7
Turn left at sign for Taconic State Parkway (T.S.P.) (Underhill Ave.).	56.8	1.6
Go under T.S.P. and turn left onto ramp for T.S.P. Northbound.	57.7	0.9
Turn right onto Bryant Pond Road.	66.8	9.1
Take first left onto Wood Street.	67.0	0.2
Bear left at Bullethole Road sign.	68.3	1.3
Stop at road work sign just before T.S.P.	68.5	0.2
Walk north on T.S.P. behind jersey barriers to work zone.		

<u>Location Marker</u>

<u>STOP 5</u>: Shoulder is being widened and rock slope is being trimmed to an inclination of 2 vertical on 1 horizontal.

TSP 8402 1045

Continue north on T.S.P. Turn right onto ramp for Route I-84 east.	81.1	12.6
Go east on Route I-84 to Exit 17 (Ludingtonville).	87.2	6.1
Turn left at end of exit ramp.	87.5	0.3
Go under I-84 and turn left onto ramp for I-84 west.	87.6	0.1
Go west on I-84 to first rest area.	90.9	3.3
Walk east on I-84 to rock cut.		

<u>STOP 6</u>: Vertical slopes on both sides of highway and in median. Massive wedge failure scar on south slope. (Fig. 2).

<u>Location Marker</u>

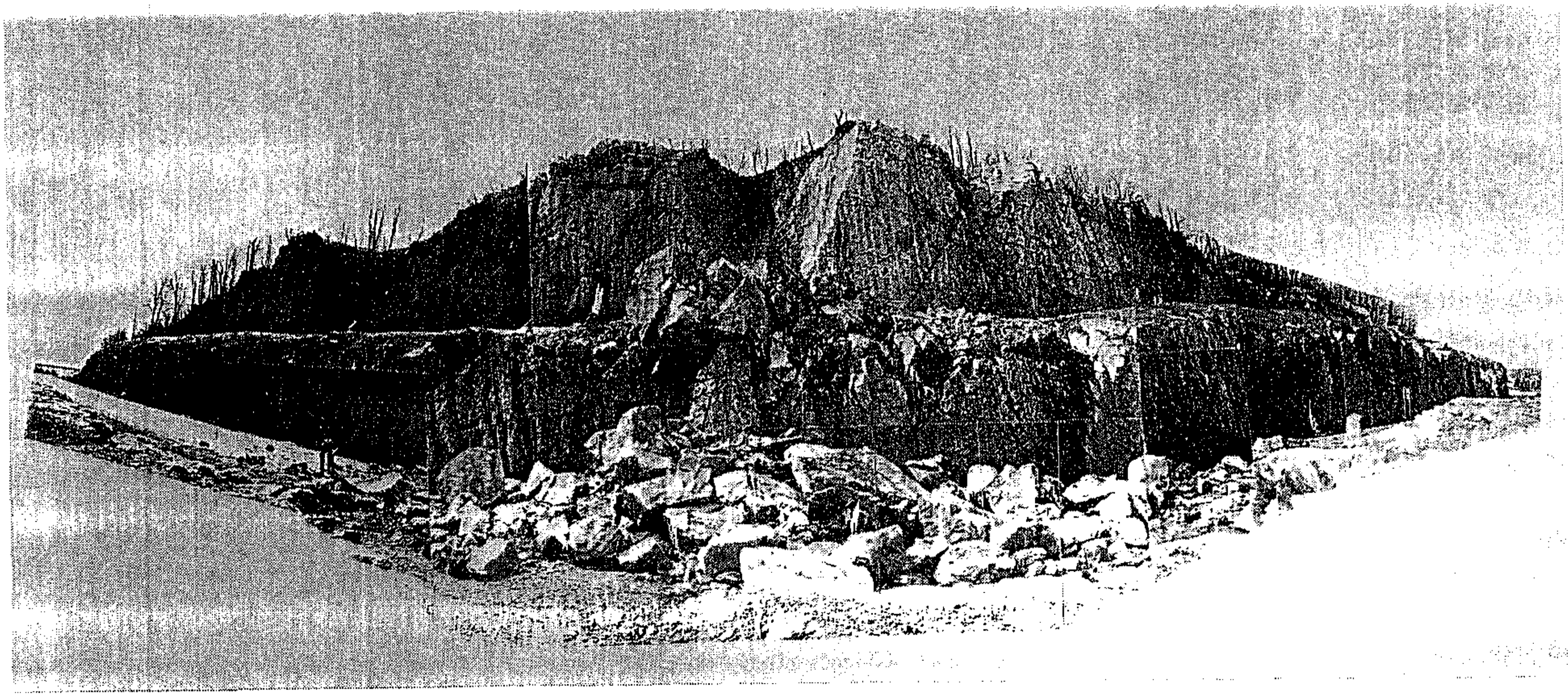
841
8202
1154

Leave rest area and continue west on I-84.	91.3	0.4
Cross Hudson River and continue west on I-84 to Exit 8 (Route 52).	109.2	17.9
Turn right at end of exit ramp onto Route 52 west.	109.3	0.1
Turn left at light in Walden to stay on 52 west.	118.1	8.8
Park on wide shoulder on east-bound side of 52.		
Walk down hill along Route 52.		

<u>Location Marker</u>

52
8602
1127

<u>STOP 7</u>: Several areas showing thick mylonite zones, rock bolts & wedge failures. Catchment mesh at mile 134.9	134.9	16.8
Continue west on Route 52		
Turn left onto Route 209 south in Ellenville.	137.3	2.4
Turn left onto ramp to Route 17 east.	150.8	13.5
Park on shoulder at top of mountain.	152.8	2.0



FIELD TRIP A-7

Figure 2

ROUTE I-84 ROCKSLIDE
Dutchess County, N. Y.
April 15, 1982

STOP 8: Rock slopes on both sides of highway. There is a wide bench on the south side with a guard rail installed at the edge to act as a catchment.

Continue east on Route 17 to Exit 118A (Route 17M - Fair Oaks).	159.5	6.7
Turn left at end of ramp onto Route 17M East.	159.6	0.1
Turn right on Wickham Ave. in Middletown to stay on 17M East (Route 211 West).	163.6	4.0
Turn left on Monhagen Ave. to stay on 17M East (211 West turns right).	164.4	0.8
Turn right on Academy St. to stay on 17M East.	165.2	0.8
Route 17M East bears right onto Dolson Ave.	165.4	0.2
Turn right on Bennett St.	165.6	0.2
Turn right into O.C.C.C. parking lot. TOTAL	165.8	0.2