Stratigraphy and Depositional History of the Onondaga Limestone in Eastern New York

by

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INTRODUCTION

The Onondaga Limestone is prominently exposed in hundreds of outcrops over a 550+ km. belt extending from Buffalo to the Helderbergs and south to Port Jervis. The formation is a coarse- to fine-grained limestone, which ranges in thickness from 21.5 m. in the central New York type sections, to 49+ m. in the eastern and western parts of the state. The Onondaga Formation was deposited during Eifielian time (Middle Devonian) (Rickard, 1975), prior to and during the initial clastic influxes from the Acadian Orogeny. The unit represents the last extensive carbonate and reef-building phase in the region.

In modern times the stratigraphy, petrology, and bioherms of the formation have been extensively studied. Most investigations have neglected the formation's non-biohermal paleoecology, as well as its eastern sections. Whole formation studies have concentrated in western and central New York, and have extended typical characteristics to eastern areas. The purpose of this trip is to examine the lithologic and faunal characteristics of the Onondaga Formation in an area extending form the Helderbergs to Catskill, and to consider the paleoenvironmental conditions under which the unit developed.

STRATIGRAPHY

The Onondaga Limestone consists of five members. Member thicknesses at several localities across the state are given in Table 1. Characteristics of each member in type section and in eastern New York are presented below.

Edgecliff Member

The type locality is Edgecliff Park, southwest of Syracuse (Oliver, 1954). The Edgecliff is a light-gray, coarse-grained, crinoidal limestone, with beds up to 1 m. thick. Crinoidal columnals 2 cm. or more in diameter are characteristic of the unit. The Edgecliff bears a rich and abundant fauna of rugose corals

MEMBERS	Buffalo	Leroy	Syracuse	Richfield Springs	Cherry Valley	Cobleskill	Helder- bergs	Catskill- Leeds	Saugerties
Seneca	12.3	9.2	7.8	2.1	2.1	-	-	-	-
Moore- house	17	17	7	22.9	22.9	21.3	21.3	11.3	30.5
Nedrow	Uncer- tain	Uncer- tain	4.3	3.7	3.7	4	4.6	13.1	10.4
Clarence	13.8	13.8	-	-	-	-	-	-	-
Edge- cliff	0-5-3.1	3.1	6.1	6.1	7	9.1	9.1	10.7	11
Total	45.7+	42.7	25	34.7	35.7	34.4	35	35	51.8

Locations

Table 1. Average thickness of the Onondaga Limestone throughout N.Y.S. The units are meters.

and tabulates. Brachiopods, bryozoans, gastropods, and trilobites are present, though not usually abundant. In places the Edgecliff swells into coral bioherms. Typical Edgecliff fauna and lithology can be traced to the Helderbergs. South of the Helderbergs, in the areas of Leeds, Saugerties, and Kingston, the unit gradually becomes fine-grained, darker, and less fossiliferous. Further south, at Wawarsing, the Edgecliff can be recognized only by the presence of crinoid columnals.

The Edgecilff has been divided into three faunal zones; descriptions of these can be found in Oliver (1954, 1956a). Zone A is the basal unit at many localities west of Richfield Springs. This is a brachiopod dominated unit with quartz sand and silt scattered about in the limestone. The abundance of quartz decreases upwards with the lowermost bed occasionally containing sufficient quartz to be referred to as a sandstone. Zone A ranges in thickness from less than 2 cm. to 1.2 m. Zone B is a discontinuous, coral dominated, limestone which exists only in western New York and Ontario. This zone has been found to consist of erosional remnants of the Early Devonian Bois Blanc Formation and has been removed from the Onondaga Formation (Oliver, 1966, 1967). Zone C is the predominant and typical Edgecliff faunal zone. This zone is dominated by rugose corals and tabulates and is the "coral biostrome" of Oliver (1954, p. 635), as well as the "great coral-bearing limestone" of Hall (1879, p. 140). The coral fauna and coarse-grained texture of this unit can be traced from Buffalo to Leeds, a distance of about 490 km. East of West Winfield two subzones, designated C_1 and C_2, can be recognized. C_1, the lower of the two, is a medium-gray, fine-grained limestone with a non-prolific coral (dominant) and brachiopod fauna. C_2 is the typical and predominant coarse-grained, light-gray, coraliferous Edgecliff Member.

Clarence Member

The Clarence Member of the Onondaga Formation (Ozol, 1963) is that portion of the formation in western New York which overlies the Edgecliff Member and consists of 40-75% chert. The Clarence roughly corresponds to the Cornitiferous Limestone of Hall (1841) and the Nedrow black chert facies of Oliver (1954). The member is a medium- to dark-gray, non-argillaceous, fine-grained, limestone with such an abundance of chert that the limestone is often found only as small "islands" floating in the chert. Fossils are typically absent or very scarce, though occasionally the lower beds bear a fauna similar to that of the underlying Edgecliff Member. The Clarence is approximately 13.8 m. thick over most of its extent. It can be traced from

Ontario, though its type locality in Clarence, New York, to Avon, New York. East of Avon it pinches out.

Nedrow Member

The Nedrow is typically a thin-bedded, very fine-grained, argillaceous limestone. Clay content can range up to about 25%. At its type locality, Indian Reservation Quarry south of Nedrow, New York, the unit measures 4.6 m. with an abrupt base and a gradational upper contact (Oliver, 1954). The member maintains its typical thickness over most of its range, except in the vicinities of Leeds and Saugerties where it is 13 m. and 10.5 m., respectively.

Despite lithologic variations, the Nedrow frequently bears a distinct fauna. The base of the member often consists of a 0.6-1.5 m. zone of thinly bedded, argillaceous limestone containing <u>Heliophyllum</u> <u>halli</u> and <u>Amplexophyllum</u> <u>hamiltonae</u>, as well as a diversity of platycerid gastropods and brachiopods. This unit, designated Zone D of the formation (Oliver, 1954), contains very few corals other than those mentioned above, and is widespread thoughout the state. Zone E, which succeeds the latter, has thicker beds, is less argillaceous, and bears a low diversity, high density, brachiopod fauna to the exclusion of most other taxa.

Moorehouse Member

The type locality of the Moorehouse Member is the Onondaga County Prison Quarry at Jamesville, New York (Oliver, 1954). Here the unit is a limestone, 6.3-7.7 m. thick, medium-gray, and very fine-grained with numerous shaley partings. Bedding ranges from 0.6-1.5 m. Chert is found throughout, but is most abundant in the upper half of the member. The Moorehouse increases in thickness both east and west of its type area.

In central New York, two Onondaga faunal zones can be recognized in the Moorehouse (Oliver,1954). Zone F is a sparsely fossiliferous unit with a low diversity, brachiopod dominated fauna. Zone G, which is gradational with the subjacent Zone F, has the most diverse fauna of any unit in the formation. Numerous brachiopods, gastropods, cephalopods, and trilobites are present , and often abundant. Several mollusc species are characteristic of this zone. The two zones are not continuous throughout the state. They lose their typical faunas and lithologies both east and west of central New York.

Seneca Member

The Seneca Member was first described by Vanuxum (1839) from several exposures in Seneca County. Oliver (1954) established the type section at Union Springs, in Cayuga County, where the member is fully exposed. Here it measures 7.8 m. It is a fine-grained limestone which becomes darker and less fossiliferous upward as it grades from a Moorehouse-like lithology to a shale above. The Seneca and Moorehouse members are separated by the Tioga Bentonite. Thinning eastward from its type area, the Seneca passes out of existence beyond Cherry Valley. Its position in eastern New York is physically and temporally occupied by the Marcellus Shale.

Eastern Stratigraphy

Onondaga facies changes across the state have resulted in several lithologic variations form the formation's type localities. East of Cherry Valley the Seneca Member pinches out. Cherry Valley also marks the easternmost extent of the typically shaley Nedrow Member. In eastern New York, the Nedrow and Moorehouse members are lithologically similar to the Edgecliff Member. These facies changes cause some difficulties in the application of central New York nomenclature ot eastern lithologies.

Oliver (personal communication, 1973) first subdivided the Onondaga Formation in the Helderberg area. He described the Edgecliff Member as 7 m. of coarse-grained, coraliferous, limestone bearing the characteristic crinoid columnals. Initially the top of the Edgecliff was placed at the top of a 4 m. section of limestone containing light chert. The 22 m. of Onondaga which supersede the Edgecliff were originally left unpartitioned because of a lack of lithologic discriminators. These rocks, referred to only as the "Upper Onondaga" (Oliver, 1954), were described as a succession of 4-5.6 m. chert-free limestone, 6.1-7.6 m. dark-chert-bearing limestone, and 9.1 m. chert-free limestone,

In 1956, Oliver reconsidered Onondaga stratigraphy in the Helderberg area. He assigned the formation's lower 8.2-9.1 m. to the Edgecliff Member, based on faunal characteristics. The Edgecliff's top was now placed several feet below the uppermost light-colored chert nodules. Succeeding the Edgecliff, 4.6 m. of section were assigned to the Nedrow Member. The criteria used in this case were the presence of thin beds and platycerid gastropods. The strata extending from here to the formation's top, about 21.3 m., were assigned to the Moorehouse Member. The base of the Moorehouse was placed about 2 m. above the uppermost light-colored chert. This unit includes about 3.9 m. of chert-free limestone, 7.6 m. of limestone containing dark-colored chert, and 9.1 m. of chert-free limestone. The Nedrow-Moorehouse contact was placed in a chert-free section, at the upper limit of beds containing platycerid gastropods. Oliver (1963) retained this partitioning of the formation. With some reservations it is retained here.

The author's reservations regarding Onondaga subdivisions in eastern New York arise from the fact that members, lithostratigraphic units, are discriminated on faunal criteria. Intermember boundaries correspond to faunal variations and not to lithologic breaks, or even to arbitrary horizons within gradational lithologic sequences. Consider the typically shaley Nedrow, which can contain up to 25% clay. In the Helderbergs, the Nedrow has a mean argillaceous content of 2%. This compares with 5-6% in the typically cleaner Edgecliff and Moorehouse members. In fact, maximum clay content and maximum gastropod abundance do not occur in the Nedrow but rather in the dark-chert section of the Moorehouse. Again, the typically spar-free Nedrow has a mean spar abundance of 8% in the Helderbergs where the other members have a comparable 10%. Nontypical lithologic characteristics also extend to fossil and calcisiltite abundances. These nontypical characteristics extend at least from the Helderbergs to Leeds. The eastern subdivisions arose from a "layer cake" approach to stratigraphy. Though they are not actual lithostratigraphic units, the eastern New York member designations are firmly entrenched in the literature (Rickard, 1975) and can be used if an awareness of their faunal origins is maintained.

ONONDAGA BIOHERMS

Historic Review

The presence of organically constructed mounds within the Onondaga Formation was first recorded by Hall (1859). It appears that he was impressed by the abundance of coral in the Edgecliff Member and referred to the entire unit as "reef". Grabau (1903, 1906) was the first to examine a discrete organic mound. He described an 11 m. domal mound with flanking beds dipping outward at about 10°. Oliver (1954, 1956b, pers. comm., 1975) located over thirty mounds outcropping in eastern and western New York. Lindemann and Simmonds (1977) noted the presence of an additional bioherm near Syracuse, thus bridging the gap between eastern and western occurrences. Subsurface bioherms (Cumings, 1932) were discovered in 1967 in south western New York (Waters, 1972) and in adjacent Pennsylvania (Piotrowski, 1976).

Subsurface Bioherms

The faunal and lithologic characteristics of five subsurface bioherms were recently summarized by Kissling and Coughlin (1979). The following is primarily derived from the above source and from D. L. Kissling (pers. comm.,1979).

Several large Onondaga bioherms have been located in the subsurface of western New York and northern Pennsylvania, at depths exceeding 4600 feet (Piotrowski, 1976). With heights up to 63 m. and diameters up to 3200 m., these structures are morphologically more akin to an inverted pie than to a multitiered wedding cake. Therefore, these mounds are classed as bank structures rather than pinnacle reefs. Despite their large diameters, the subsurface bioherms stood as prominent elevations on the sea floor. They project above the remainder of the formation by in excess of 50 m., and extend above the interreef portions of the Edgecliff Member by as much as 60 m.

These bioherms show an orderly succession of faunas and associated lithologies. They began with a pioneer community dominated by thickets of *Acinophyllum* and *Cladopora*. The thickets grew on slightly elevated areas of the sea floor, during a transgression of the sea. The corals trapped lime mud and grew rapidly upward. A second stage of biohermal growth is shown by a more diverse assemblage of *Cladopora*, *Cylindrophyllum*, *Acinophyllum*, and various solitary rugosans. *Cylindrophyllum* dominates the central areas of the mound, while *Cladopora* dominates the periphery. These corals grew in dense thickets which accumulated a matrix of crinoid sand, which occasionally inundated them. Crinoids, which provided the matrix sediment, lived amongst the thickets. The quantity of lime mud deposited during this growth stage is conspicuously small. A terminal growth stage is marked by a *Cladopora* and *Cystiphylloides* dominated assemblage. *Acinophyllum* and *Syringopora* are present, though not numerous. The fauna of this stage is enclosed by a matrix of lime mud and fenestrate bryozoans.

Bioherms of the subsurface are believed to have grown on a hinge between a rapidly subsiding basin to the south and a slowly subsiding, relatively shallow, shelf lagoon to the north (Warters, 1972). Kissling and Coughlin (1979) cite evidence for this conclusion. First, both outcrop and subsurface bioherms were initiated and grew under similar environmental conditions. This is shown by similarities in their faunal and lithologic successions. Despite similarities, the subsurface bioherms attained thicknesses of about 50 m. in excess of their outcrop counterparts. Second, the subsurface bioherms have the steepest slopes on their southern, basinward, sides. Regardless of their proximity to a deep basin, it is apparent that biohermal growth in southern New York was influenced by a rate and extent of subsidence which far outstripped that of areas to the north.

An Ontario Bioherm

A mound of similar morphology to the subsurface bioherms, crops out west of Port Colborne, Ontario (Cassa, 1979). The mound is 15 m. vertically and about 1 km. horizontally. Unlike other Onondaga bioherms, this one is bedded rather than massive and contains chert. This mound qualifies as a bioherm because it swells above the combined thicknesses of the Edgecliff and Clarence members by about 9 m. and must have stood erect on the sea floor.

Successional development of the Port Colborne bioherm is comparable to that of other Onondaga bioherms (Cassa, 1979). The pioneer community is dominated by Acinophyllum which trapped a shale and calcisiltite matrix. Acinophyllum grades upward into argillaceous, crinoidal wackestones and packstones. The second stage fauna is more diverse than the first, including solitary rugosans, laminar stromatoporoids, massive tabulates, and branching tabulates. The mound is capped by a third stage fauna of solitary and colonial rugosans, Fistulipora, and large crinoids. The lithology of this stage includes non-argillaceous, crinoidal packstones and grainstones. Crinoid holdfasts found in this unit indicate that the crinoidal debris was deposited near its life site. The mound is gradationally terminated and superceded by a cherty, argillaceous wackestone.

The Port Colborne bioherm is believed to have developed as a bank in a shallow lagoon (Cassa, 1979). Though evidence of turbidity fluctuations is present, the prime environmental controls on the bank's development were the differential rates of subsidence and coral growth. The subsidence rate, possibly coupled with a rise in sea level, exceeded coral growth and growth of the bank was terminated.

Bioherms in Outcrop

The majority of Edgecliff bioherms seen in outcrop differ greatly in size and shape from the Port Colborne and subsurface mounds. The former are round or ovoid in plan view, with diameters or lengths rarely exceeding 400 m. In cross section these bioherms are domal or lensoid, with heights of 1-22 m. (Oliver, 1954,1956b).

Five, of the approximately thirty, outcrop bioherms have been studied to characterize successional lithologies and faunas and to interpret paleoenvironmental conditions (Mecarini, 1964; Bamford, 1966; Poore, 1969; Collins, 1978; Williams, 1979).

These studies have shown the bioherms to be generally similar in their respective successional patterns. Four facies common to most bioherms of the Onondaga Formation are described below.

1) Basal Facies

This unát is found wherever the base of a bioherm is exposed. 30-70% of the facies' volume is occupied by Acinophyllum corallites. This branching coral acted as a baffle, trapping a matrix of terrigenous mud, calcisiltite, and quartz silt. Exclusive of intracorallite spaces, pore filling spar is absent.

This facies represents a pioneer community, strongly dominated by a single opportunistic genus. The unit corresponds to the "quiet water stage" of Lowenstam (1957) and to the "basal stabilization zone" of Walker and Alberstadt (1975). Finks and Lamster (1979) have reported Acinophyllum growth rates of 5-15mm./year. Therefore, through rapid lateral and upward growth, Acinophyllum was able to colonize a portion of the sea floor and raise a stabilized platform for further coral growth.

2) Core Facies

This unit, comprising the main body of the mound, overlies and is gradational with the basal facies. The fauna is diverse and spatially distributed in more or less complex patterns. The more abundant genera include Cylindrophyllum, Cladopora, Favosites, Emmonsia, Cystiphylloides, and Acinophyllum. The core of each bioherm is unique in relative abundances of the above taxa and in their spatial segregations and integrations. Faunal complexity precludes a biologic generalization of whole--core of sub-core characteristics.

Despite complex faunal characters, the core facies contain lithologic traits which can be generalized. Sediment grain-size and spar abundance increase upward and terrigenous mud and calcisiltite diminish upward. Core lithologies grade from calcisiltite to sparry calcarenite and/or calcirudite. Lindemann and Simmonds (1977) related subdivisions of the core facies to the "overlying colonization zone" and the "diversification zone" of Walker and Alberstadt (1975).

3) Cap Facies

Where exposure permits, a third facies can be seen overlying the core. This cap is usually dominated by Heliophyllum. In what appears to be the cap of an eastern New York bioherm, Finks and Lamster (1979) report a single Heliophyllum corallum which exceeds 12 m. in diameter. Cystiphylloides may also be abundant, present, or absent. Lithologically, this facies is a sparry calcarenite or calcirudite. Instances have been reported where calcisiltite is totally absent except within the coralla of dendritic and phaceloid corals (Bamford, 1966).

4) Flank Facies

Many bioherms have beds of bioclastic sediment which dip from the unbedded coral mounds at angles of 10-15°. These beds are primarily composed of bioherm derived detritus. Flank lithologies are highly variable. Spar dominates the proximal and upper beds, while calcisiltite dominates the distal and lower areas. The flank beds grade distally from biohermal to normal Edgecliff lithologies and faunas (Bamford, 1966).

Paleoenvironmental Interpretations

Among those who have studied them, there is general agreement that Onondaga bioherms developed on a shallowly submerged lagoonal shelf (Bamford, 1966; Collins, 1978; Kissling and Coughlin, 1979; Ossa, 1979). Biohermal growth began in relatively quiet water, as Acinophyllum thickets stabilized the substrate and raised a slight mound above the surrounding sea floor. This pioneer community was succeeded by the diverse fauna of the core facies, which continued to accumulate sediment and build the mound upward. During upward growth, the mound was exposed to ever increasing water agitation. This is shown by the upward abundance trend of pore filling spar cement. With the onset of agitation, some bioclastic sediments were swept from the mound and deposited as flanking beds. Eventually, the mound reached a position at which water turbulence and shifting sediments exceeded faunal tolerances and coral growth was extinguished (Mecarini, 1964; Bamford, 1966; Poore, 1969). Contrary to the statements of Turner (1977) and Lindemann and Simmonds (1977) there is no evidence that an influx of clastic mud eradicated the corals.

It has been suggested that biohermal growth was terminated because of the inability of coral growth to keep pace with the rate of subsidence (Kissling and Coughlin, 1979; Ossa, 1979). This may hold true for the subsurface bioherms, but it contradicts the faunal succession and spar abundance trends of most outcrop bioherms. Furthermore, if the 5-15 mm./cycle growth rate of Acinophyllum reported by Finks and Lamster (1979) is in fact an annual rate, it is difficult to imagine a subsidence rate exceeding this on a persistent regional basis. To date, the environmental condition or conditions which terminated the bioherms are uncertain.

Banks of Reefs?

During the past two decades, there have been so many semantic manipulations of the term "reef" that it has become a term of dubious meaning. The problem has partly arisen from divergent perceptions and purposes of the various geolinguists. It has also partly arisen form the use of "reef" as a genetic rather than a descriptive term, requiring interpretation rather than strict observation. In this paper the author uses Heckel's (1974, p. 96) definition of reef. In accord with this definition, a reef must be an organically constructed buildup that displays:

(1) Evidence of (a) potential wave resistance, or
(b) growth in turbulent water, which implies wave resistance;
(2) evidence of control on the surrounding environment.

According to Heckel, wave resistence can be evidenced by:

(1) Inorganic spar cementation
(2) Organic construction of rigid skeletal frameworks
(3) Sediment binding by encrusting organisms
(4) Sediment binding by rooted organisms
(5) Coherence of lime mud
(6) Inertia of large skeletons

To evaluate the degree to which Onondaga bioherms conform to the above characteristics, we will examine the characteristics of these bioherms as compared to those of a Florida patch reef. The patch reef described here is Clearwater Reef, which lies in the back reef lagoonal area of the Florida Reef Tract. Clearwater Reef is within the bounds of Biscayne National Monument, a short distance from Margot Fish Shoal. The fauna and sediments of this reef were mapped and studied in detail by Black and others (1975).

Clearwater Reef is a coral bioherm which rises from a depth of 5 m. to within 0.2 m. of sea level at low tide. This reef is approximately 50 m. long, ovoid in plan view, and oriented parallel to prevailing currents. Sediments on the reef surface are reef derived bioclasts with approximate grain--size frequencies of 15% mud, 65% sand, and 20% gravel. The gravel fraction consists primarily of fragmented reef corals. The sediment is mobile and sometimes inundates corals growing on the reef surface.

A sediment apron surrounds and flanks Clearwater Reef. It extends from the reef surface to a water depth in excess of 5 m. Apron sediments are reef derived bioclasts, with an average mud content of 20%. No discernable grain-size frequency trends were found in radial transects around the reef. Due to bioturbation, the sediment apron shows no laminations.

Onondaga bioherms are coraliferous mounds of ovoid shape. They are believed to have been oriented parallel to prevailing currents (Poore, 1969; Kissling and Coughlin, 1979; Williams, 1979; R. M. Finks, pers. comm. 1979). Mud content in the core and cap facies ranges from 40% (Mecarini, 1964) to nearly 0% (Bamford, 1966). Higher calcisiltite contents are found within the coralla of branching corals. Sediments of flanking beds are mound derived bioclasts containing as little as 8% calcisiltite and as much as 35% pore filling spar cement (Bamford, 1966).

From the foregoing, it can be seen that Onondaga bioherms satisfy Heckel's (1974) reef definition as well as, or more fully than, a Holocene patch reef. Onondaga reefs have less surface mud than Clearwater Reef. Where calcisiltite is abundant in Onondaga reef cores it is confined to coralla interiors, imparting stability and wave resistance. The lime mud in these bioherms may have resulted from original skeletal framework material that underwent biodegradation and subsequent cementation by cryptocrystalline cement (G. M. Friedman, pers. comm., 1979). This process is common just below the living surfaces of post-Paleozoic reefs and masks the original rigid structure of the reef (Friedman, 1978). The flanks of Onondaga reefs contain less mud than the cores and show more biohermal control of nearby sediments than do the flanks of Clearwater Reef. Onondaga reefs contain an abundance of spar cement. The flanks also were life sites of organisms, such as crinoids, which served to stabilize the sediments. Submarine cementation which occurs in Holocene reef just beneath the living surface could also have helped in sediment stabilization (Friedman, Amiel, Schneidermann, 1974). Unless we are willing to claim that Holocene Patch reefs, of proven wave resistant abilities, are not reefs then we must consider Onondaga bioherms to be reefs also. The main area of remaining doubt lies with the degree of "agitation" or "turbulence" they could or did withstand.

Albrights Reef

Albrights Reef is a crinoidal coraliferous mound measuring 305 m. x 250 m. x 6 m. A regional dip of 20°W and irregular exposure obscure the original dimensions. Bamford (1966) divided the reef core into two subfacies and recognized a total of five reef facies. The spatial relationships of the facies, as seen in an east-west transect through the reef center, are shown in Figure 1. The facies descriptions of Bamford (1966) are given below.

1) Acinophyllum Facies

This facies is bedded to massive limestone consisting of 50+% Acinophyllum corallites with a matrix of calcisiltite and fossil fragments. Matrix bioclasts include crinoids, ostracods, brachiopods, and bryozoans. Spar cement is present only within the corallites. Bedding is best developed at the unit's base, and is gradually obscured upward. These deposits represent the basal facies and the poineer community of the reef.

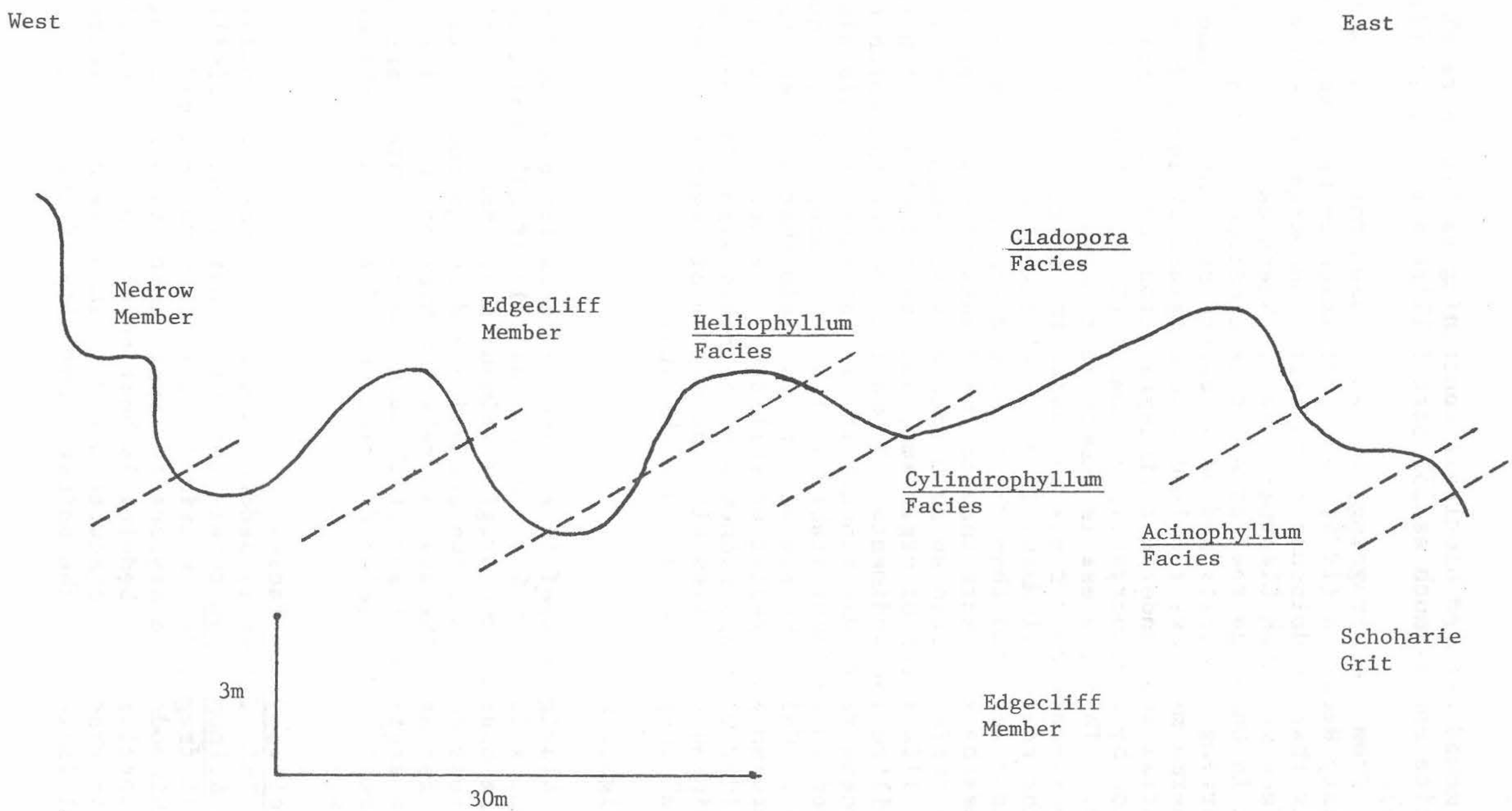

Figure 1. Schematic cross section of Albrights Reef. (From Bamford, 1966)

2) Cylindrophyllum Facies

This unit is characterized by an abundance of Cylindrophyllum, with an accessory fauna of Acinophyllum, Cladopora, and Favosites. The facies is lithologically variable. Calcisiltite ranges to 50%, and spar ineceasrs upward from nearly 0% to a maximum of 35%. Massive Favosites coralla are most abundeant in the sparry sediments. The branching corals are best developed in muddy areas. The sediments are primarily crinoidal. Crinoid holdfasts helped stabilize the sediment substrate. During this facies' development, water agitation increased to the extent that calcisiltite was swept from the mound.

3) Cladopora Facies

Here the reef attains maximum diversity. The coral fauna includes Cladopora, Syringopora, Siphonophrentis, Heliophyllum, Favosites, and Emmonsia. The latter two are large, massive, and hemispheric. Crinoids attain maximum abundance in the sediments where they are found with ostracods, brachiopods, bryozoans, platycerid gastropods, and coral fragments. This unit is a packstone or grainstone. Calcisiltite is virtually absent. Development occurred as the reef grew into an environment of ever increasing water agitation.

4) Heliophyllum Facies

This unit is a grainstone or biosparite. Bioclasts are larger and better sorted than elsewhere on the reef. This facies caps the reef and grades laterally into reef flank beds. The unit was deposited in agitated water which precludes the growth of most corals other than large colonies of Heliophyllum . Bamford (1966) concluded that high levels of agitation prevented further reef growth.

5) Flank Facies

A series of graded beds, which flank the reef, were deposited lateral to and gradational with the reef cap. These deposits become finer distally and grade into normal Edgecliff fauna and lithologies.

Roberts Hill Reef

In outcrop, this reef is a mound 370 m. long with a 9 m. east facing cliff exposure. The hill formed by the reef is elongate in a north-south direction. The reef's true orientation and dimensions have been obscured by erosion and by a regional dip of 24° W. Collins (1978) studied the reef facies to determine their spatial relationships and depositional environments. Five facies were recognized, four belonging to

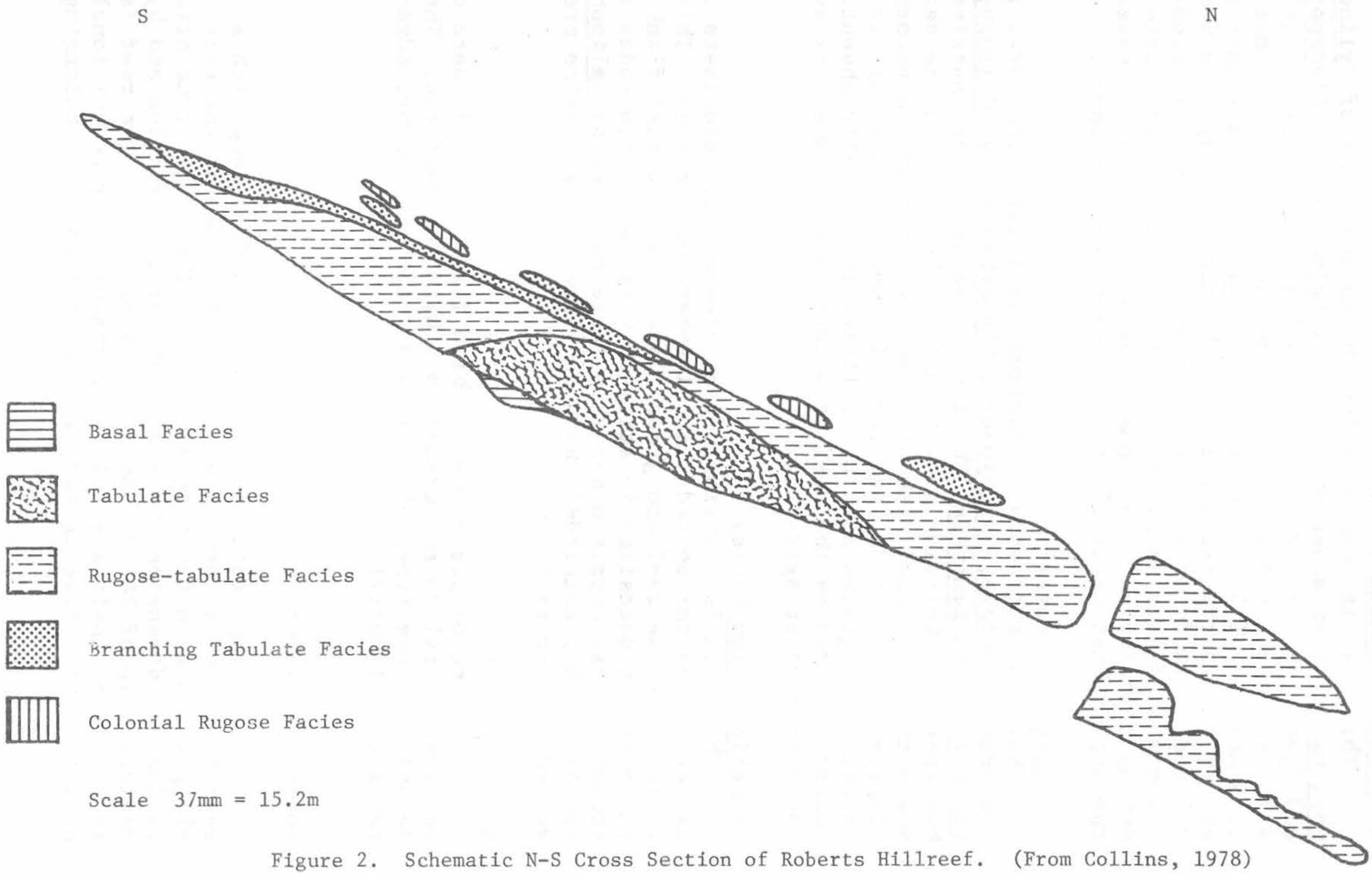

Figure 2. Schematic N-S Cross Section of Roberts Hillreef. (From Collins, 1978)

the reef core. The remaining facies is either part of the core or is an unusual cap facies. Figure 2 is a schematic of the facies exposed in the east facing cliff. Characteristics of the facies (Collins, 1978) are given below.

1) Basal Facies

The lowermost section of Roberts Hill Reef is overwhelmingly dominated by Acinophyllum. The corallites are enclosed in a matrix of calcisiltite. Additional bioclastic particles include fragments of crinoids, ostracods, gastropods, and bryozoans.

2) Tabulate Facies

The fauna of this unit is dominated by large, hemispheric or laminar, coralla of Favosites and Emmonsia. Acinophyllum is also present. The corals are enclosed by a sediment of large, well-washed, crinoid columnals, some of which attain lengths of 15+ cm. Crinoidal sediments appear to have occasionally smothered the corals. Calcisiltite is rare or absent in most samples. These sediment characteristics indicate deposition in agitated waters which may have been influenced by storm waves.

3) Rugose-Tabulate Facies

The fauna of this unit is dominated by Cylindrophyllum, Cystiphylloides, Favosites, and Emmonsia. The latter two genera are neither as large nor as numerous as in the previous facies, and are restricted to areas where well-washed bioclastic sands and gravels predominate. The regose coralla, some of which attain heights of 3 m., contain a calcisiltite matrix. In outcrop, this facies extends up and around the tabulate facies. It has been postulated that Cylindrophyllum formed a wave baffle, partly protecting the interior tabulate facies. This pattern is exactly opposite to that seen at Thompsons Lake Reef (Williams, 1979) where tabulates are peripheral to Cylindrophyllum. A definitive determination of the relationships between these two facies was not accomplished at Roberts Hill due to insufficient exposure.

4) Branching Tabulate Facies

This unit contains the reef's greatest faunal diversity. Thirteen coral genera occur here, dominated by Thamnopora, Coenites, Favosites, and Acinophyllum. The branching coralla are loosely arranged in a calcarenite matrix. Calcisiltite content is generally low except in proximity to the superceding unit.

5) Colonial Rugose Facies

Corallites of Cylindrophyllum and Acinophyllum occupy over 40% of the volume of this unit. These corals, along with Coenites and Heliophyllum, trapped a calcisiltite matrix. The sediments also include ostracods and Styliolina. As far as can be determined from the exposures on hand, this facies caps the bioherm. Judging from the successive faunas and lithologies of other Onondaga bioherms, it may be concluded that the actual cap facies is not exposed. If this is not the case, this facies represents an atypical cap developed under abnormal reef termination conditions.

ONONDAGA LITHOLOGIES

Discrimination of carbonate petrologic types within the Onondaga Formation is based on relatively few lithologic variables. The only particle types or textures of sufficient abundance to be quantitatively significant are lime mud, biogenic grains, spar cement, and terrigenous mud. Quartz sand is abundant in very few samples and absent in almost all others. Small quantities of silt-size subhedra and euhedra of quartz and dolomite are nearly ubiquitous and are the result of syntaxial overgrowths on 1-5 micron eolian grains of corresponding composition (Lindholm, 1969b). Volumetric increase due to grain-growth and the pervasive low-abundance distribution of these grains preclude their use as discriminators in the reconstruction of primary petrologic characteristics. Pyrite is also present in small quantities (1-3%) in almost all samples. Ooids and glauconite are present in less than 2% of the samples studied. Oncolites and pellets are absent. Sedimentary structures, such as ripples or crossbeds, are extremely scarce. The result is that Onondaga lithologies are relatively simple compared to many other carbonate formations in New York.

The formation is volumetrically dominated by silt-size carbonate particles with diameters in excess of 5 microns (Lindholm, 1967, 1969a). The remaining volume is occupied by recognizable fossil material, in various states of degradation, and by pore-filling spar cement. Spar cement is often common or abundant where fossil abundance is high. Where carbonate silt is abundant, spar is rare or absent. Noting that the carbonate mud particles are poorly-sorted, of variable shape, and frequently associated with terrigenous mud, Lindholm (1969a) concluded that they are of detrital origin and not the result of micrite neomorphism (Folk, 1962). Lindholm further concluded that the carbonate mud was produced by the mechanical and organic breakdown of invertebrate shell material. This conclusion is supported by the presende of recognizable organic microstructures in some of the silt-size grains and by the fact that they often form a continuum between unrecognizable minute grains and larger identifiable fossil grains. For these resons Folk's (1962) lithologies which contain the term "micrite" are not applicable to this formation. Calcisiltite is the term most appropriate for the carbonate mud of the Onondaga Limestone.

Lindholm (1967) recognized four limestone types in the Onondaga Formation. They are discriminatdd by abundances of biogenic allochems (fossil grains), calcisiltite, and spar cement. The lithologies are described below.

1) Fossiliferous Calcisiltite - Composed of 1-10% fossil material in a calcisiltite matrix. Terrigenous mud can comprise up to 20% of the rock volume and quartz silt less than 7%.
2) Sparse Biocalcisiltite - Composed of 10-50% fossil material in a calcisiltite matrix. Quartz silt, dolomite, biotite, and terrigenous mud are common, though not abundant.
3) Packed Biocalcisiltite - Composed of fossil material in excess of 50% with a calcisiltite matrix. Terrigenous materials are scarce or absent.
4) Biosparite - Composed of fossil material with sparry calcite cement. Calcisiltite and terrigenous mud are usualły absent.

The above lithologies are primarily divided at arbitrarily selected fossil abundance thresholds. This type of classification strategy is excellent for pigeonholing samples. However, it attempts to subdivide the continua of several rock components based on abundance subdivisions of one component. This obscures the rock's true characteristics, because the component grains or textures do not vary in a totally inclusive/exclusive manner. To avoid the nebulization of potentially significant rock component intergradations, Q-mode cluster analysis has been used in accord with the classification strategy of Park (1974).

Q-mode cluster analysis relates samples according to the joint consideration of the abundances of all their variables. In this case the lithologic variables used in the classification included calcisiltite, fossils, spar cement, argillaceous material, pyrite, quartz silt, quartz sand, and glauconite grains. The latter three failed to contribute to the clustering process and were deleted. Pyrite, being extremely common but always scarce, did not significantly effect the classification. It is, however, retained because its abundance is inversely related to spar abundance, and , therefore, environmentally significant. Thus, Onondaga lithologies were discriminated on the relative abundances of only the first four lithologic variables listed above.

The Q-mode cluster dendrogram of 87 lithologic samples collected in eastern New York segregates six lithologic types. Though the dendrogram is not included here, the variable percent

I	II	III	IV	V	VI	Lithologic Types
42-81	72-84	29-44	6-30	13-26	3-14	Calcisiltite Extremes
53	78	36	22	19	9	Calcisiltite Means
10-43	4-18	39-68	55-73	45-53	59-68	Fossil Extremes
35	14	54	61	50	62	Fossil Means
0-5	0-1	0-14	9-16	21-31	20-30	Spar Extremes
2	1	5	11	25	25	Spar Means
1-13	3-10	2-12	1-10	1-11	1-9	Mud Extremes
8	6	5	4	5	4	Mud Means
1-4	0-3	0-2	0-2	0-1	0-1	Pyrite Extremes
1	1	1	1	1	1	Pyrite Means

Table 2 - Compositions of the six lithologic types. Numbers are expressed as persents.

abundance extremes and means for each of the lithologic types is provided in Table 2. Examination of this table reveals that the lithologic types are not segregated by abundances of a single variable, but by the relative abundances of calcisiltite, fossils, and spar. All of the lithologic types, except I and II, are discretely segregated by their component abundances. Though I and II haveoverlaps in the four prime variables, the mean abundances of these variables differ greatly and they do plot as discrete clusters in the Q-mode dendrogram. Thus is developed a quantitatively based classification of limestone types within the formation. The paleoenvironmental significances of the six lithologic types will be considered in ane other section.

FOSSIL ASSEMBLAGES

Fossil assemblages were discriminated by Q-mode cluster analysis performed on faunal census samples. Faunal variables used in assemblage identification are in the form of percent abundance data and include more than thirty taxa, most extending to the generic level. The fossil assemblages identified here are recurrent groups of samples having similar faunal structures. They correspond to the "quantitatively defined communities" of Kauffman and Scott (1974, p. 11) and, therefore, are not considered to be paleocommunities as defined by Kauffman (1974, p. 12.3). The term assemblage which "consists of organisms derived from more than one community" (Kauffman and Scott, 1974, p.18) is more appropriate in that the faunal units identified here have similar but not "identical" structure. No attempt is made here to discern faunal interactions.

Table 3 provides information on the percent abundances of organisms in each of the fossil assemblages. The abundances of organisms in a group of samples, designated assemblage X, which

Faunal Assemblages	Solitary Rugose	*Acinophyllum*	Massive Tabulates	Branching Tabulates	Auloporids	Trilobites	Gastropods	Brachiopods	Fenestrate Bryozoans	Ramose Bryozoans	Massive Bryozoans
A	18	2	15	4	1	16	1	34	1	8	1
B	2	1	1	1	2	6	2	12	7	64	2
C_1	8	1	1	1	18	9	1	29	11	20	1
C_2	5	-	-	-	22	14	1	48	1	8	1
C_3	2	1	1	1	7	15	1	37	11	23	1
D	32	1	1	-	1	6	5	50	1	2	1
E	31	11	26	2	1	1	1	25	1	1	-
X	3	-	-	1	-	9	-	84	1	1	1

Faunal Elements

Table 3 - Percent abundances of faunal elements in fossil assemblages

clustered at very low levels are also recorded. Data have been reduced to high level taxa or to growth forms within taxa. The faunal integrity of the fossil sssemblages is retained even at these levels. The numbers in Table 3 were calculated by dividing the total number of occurrences of a taxon by the total number of fossil specimens in that assemblage. Taxa which comprise less than one percent of the total fauna within the assemblage are scaled to one percent to facilitate presentation. Thus data from all samples within an assemblage are combined to present a typical or average characterization of the assemblage. In this way faunal abundances are based on large samples which are more representative of the whole assemblage than any small sample could be (de Caprariis, Lindemann, and Collins, 1976). This approach partially circumvents the problems of sampling patchy fossil distributions (de Caprariis and Lindemann, 1978).

PALEOENVIRONMENTS AND DEPOSITIONAL HISTORY

Context From Previous Investigations

Oliver (1954, 1956a) and Lindholm (1967, 1969a), in their respective studies of Onondaga stratigraphy and petrology, came to similar conclusions regarding paleoenvironmental conditions of deposition. Their interpretations are summarized below.

The Onondaga Limestone was deposited in an initially shallow, subsiding, epicontinental sea, within 30° of the equator. A westward transgression is indicated by the formation's base. In eastern New York the base is gradational with the subjacent Schoharie Formation. Between Sharon Springs and Richfield Springs the gradational contact is marked by phosphate nodules and glauconite, indicating a slight unconformity. Westward from Richfield Springs the Onondaga rests unconformibly on successively older formations. At many localities the contact is erosional and the lowermost Onondaga beds contain quartz sand reworked from the underlying formations. This transgression follows the pattern of the lower Devonian transgressions during which the Helderberg Group was deposited. Unfortunately in the case of the Onondaga ravinements (Anderson, 1971) have removed all of the supratidal and intertidal deposits created by the transgression.

During and immediately following the initial Middle Devonian transgression, the Edgecliff Member was deposited in shallow, wave-affected waters (Laporte, 1971). This is shown by the lithologic characteristics of the member and its coral reefs. Fluctuations in water turbidity and agitation were the major environmental factors controlling deposition throughout the eastern half of the Edgecliff Member (Lindemann, 1974). Continued subsidence carried the sea bottom beneath the effects of waves and set the stage for deposition of the succeeding members (Laporte, 1971).

Deposition of the Nedrow Member took place during an influx of clastic mud which was synchronous with a pulse of subsidence. The Nedrow was most typically deposited in a topographic depression which developed in central New York (Oliver, 1954; Lindholm, 1969a). This trough extended from beyond the northern erosional limit of the formation, southward into southern New York. The Nedrow contains several cycles of turbid water/clean water sedimentation. The turbid cycles have sharp bases and gradational tops. These cycles may be the result of pulses of subsidence followed by rapid deposition (E. J. Anderson, pers. comm., 1977).

The Moorehouse Member marks a return to relatively nonturbid conditions. Deposition took place in quiet water. The Seneca Member was deposited in quiet water during the westward progradation of the Marcellus Shale. The gradational and interbedded relationship between the Seneca Limestone and the Marcellus Shale indicates that turbidity levels waxed and waned for a time prior to the inundation of the sea by clastic mud. This argillaceous influx took place in eastern New York prior to deposition of the Seneca in central and western New York. Therefore, in the eastern area the uppermost member of the Onondaga Formation is the Moorehouse Member.

Lithologic Paleoenviroamental Interpretations

The Onondaga Limestone of eastern New York differs significantly form its type characteristics of central New York. Therefore, because regional depositional environments were originally interpreted from central New York sections, a reexamination is in order. This was partly accomplished by use of a Q-mode ordination analysis performed on the lithologic samples previously used in the identification of lithologic types. For references and a description of the ordination technique see Park (1968, 1974). Briefly, ordination arrays samples along orthogonal axes based upon their mutual similarities or dissimilarities. The axes represent environmental gradients which influenced sample characteristics. This type of analysis has been successfully applied to the interpretation of sedimentary environments (Ali, Lindemann, and Feldhausen, 1976) and in paleoecologic studies of fossil assemblages (Park, 1968; Lindemann, 1974). The ordination array itself is only an arrangement of samples in a space, it does not provide its own interpretation.

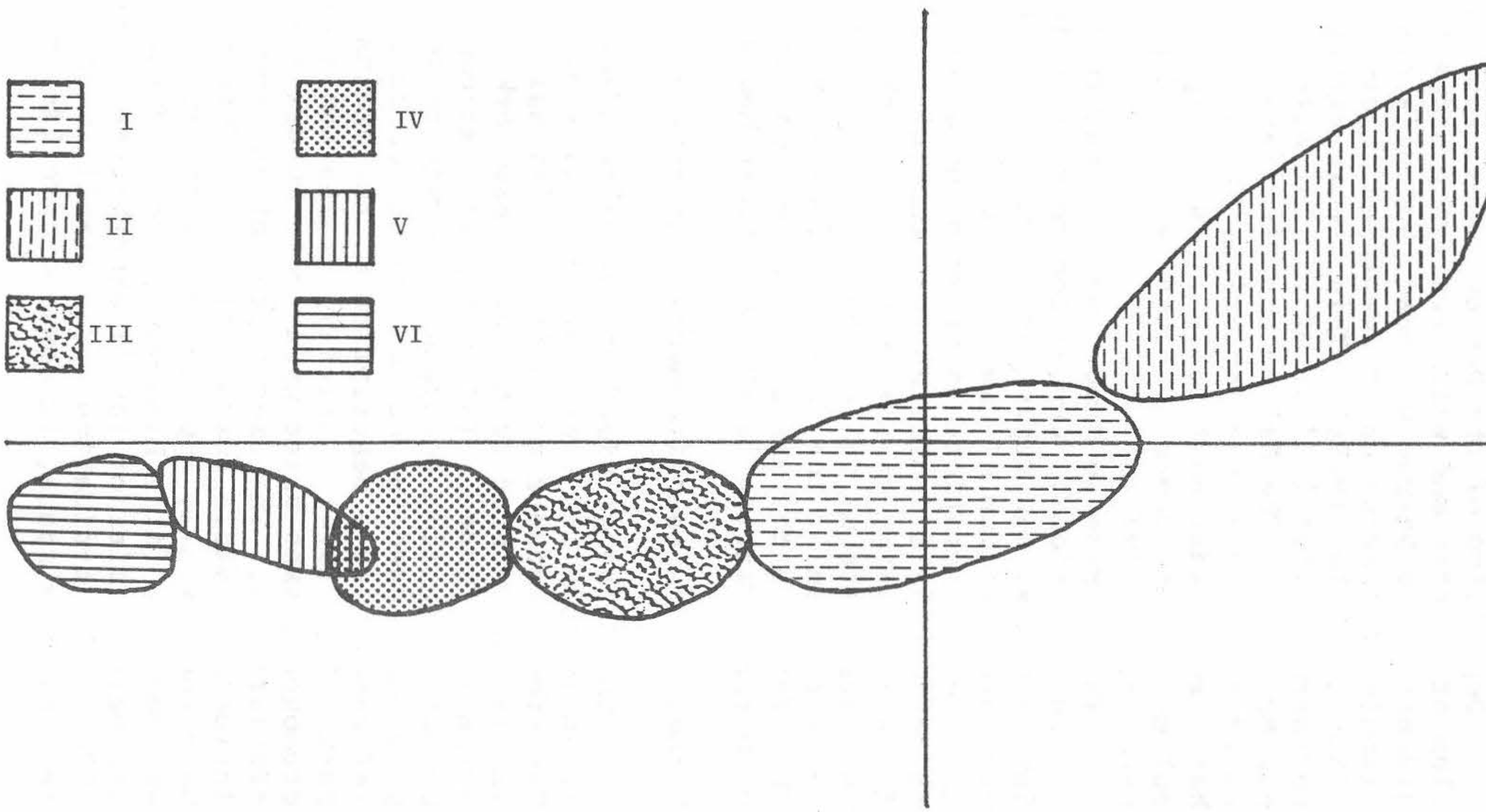

Figure 3. Q-mode ordination of lithologic data showing distribution of lithologic types.

A plot of Onondaga lithologic types on a Q-mode ordination model of lithologic samples (Fig. 3) shows their mutual arrangement with respect to the axes. The trends of mean variable values of the lithologic types on the ordination diagram (Fig. 3) reveal that the two axes correspond to gradient complexes. What is shown is the opposition of sparry fossiliferous sediments and spar-free calcisiltite rich deposits. Because spar requires an abundance of pore creating fossils, its abundance is dependent on fossil abundance. Therefore, the separate natures of the two axes are indistinguishable. It appears that these axes, which account for 74% of the dissimilarity between samples, represent an overall gradient of water agitation. Agitated water is indicated to the left while quiet-water sedimentation is indicated in the central and upper right of the diagram. This is evidenced by the opposed positions of abundant spar and abundant calcisiltite.

A third ordination axis was anticipated to reveal fluctuations in turbidity levels. It failed to show any trend whatsoever. Therefore, it is concluded that mud influxes in the area had little effect on deposition, and must have been eigther minimal or constant.

Examination of formational lithologic successions at Leeds and in the Helderbergs (Figs. 4,5) reveal paleoenvironmental dissimilitaries. In both areas, deposition began in slightly turbid, quiet water. At Leeds, turbidity and agitation generally remained low. Fluctuations in environmental conditions were small and lacked a real trend in any direction. The only prolonged divergence from the quiet water condition is shown by lithologic type V. This indicates a long period of water agitation and, presumable, shallower water. Unfortunately, the top of the formation is not exposed at Leeds and the Marcellus Shale contact is unavailable for study.

In the Helderbergs, the water remained rather clean and agitated throughout much of Onondaga deposition. The dark chert section of the Moorehouse Member reveals increased turbidity levels, and may be correlative with the shaley Nedrow Member of central New York. Following a period of relatively turbid, quiet-water conditions, a time of cleaner more agitated water prevailed. Agitation fluctuations eventually gave way to persistent quiet-water conditions until the termination of limestone deposition in the area. Despite the examination of beds at and very near the formations top, evidence of gradually increasing turbidity was not seen.

The quiet-water, relatively anoxic conditions of lithologic type II were not evident in the eastern sections. Lithologic type II is present further west in the Cobleskill and Cherry Valley areas.

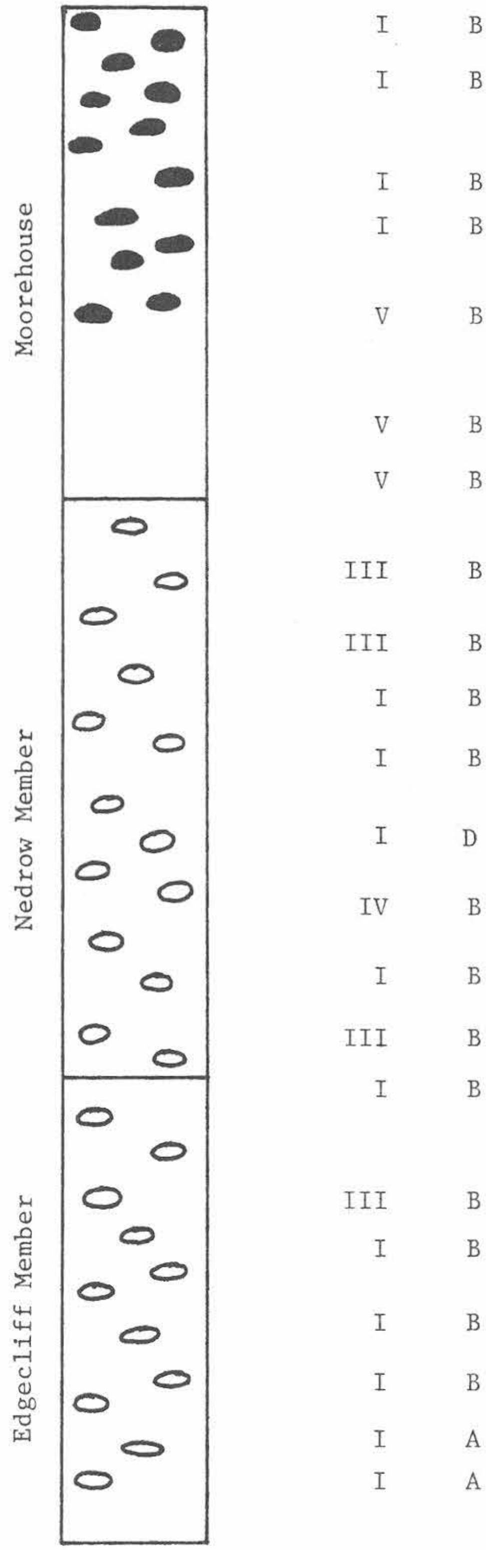

Figure 4. The Onondaga section at Leeds showing distribution of faunal assemblages and lithologic types. Scale 6mm = 1m. The ovals represent chert.

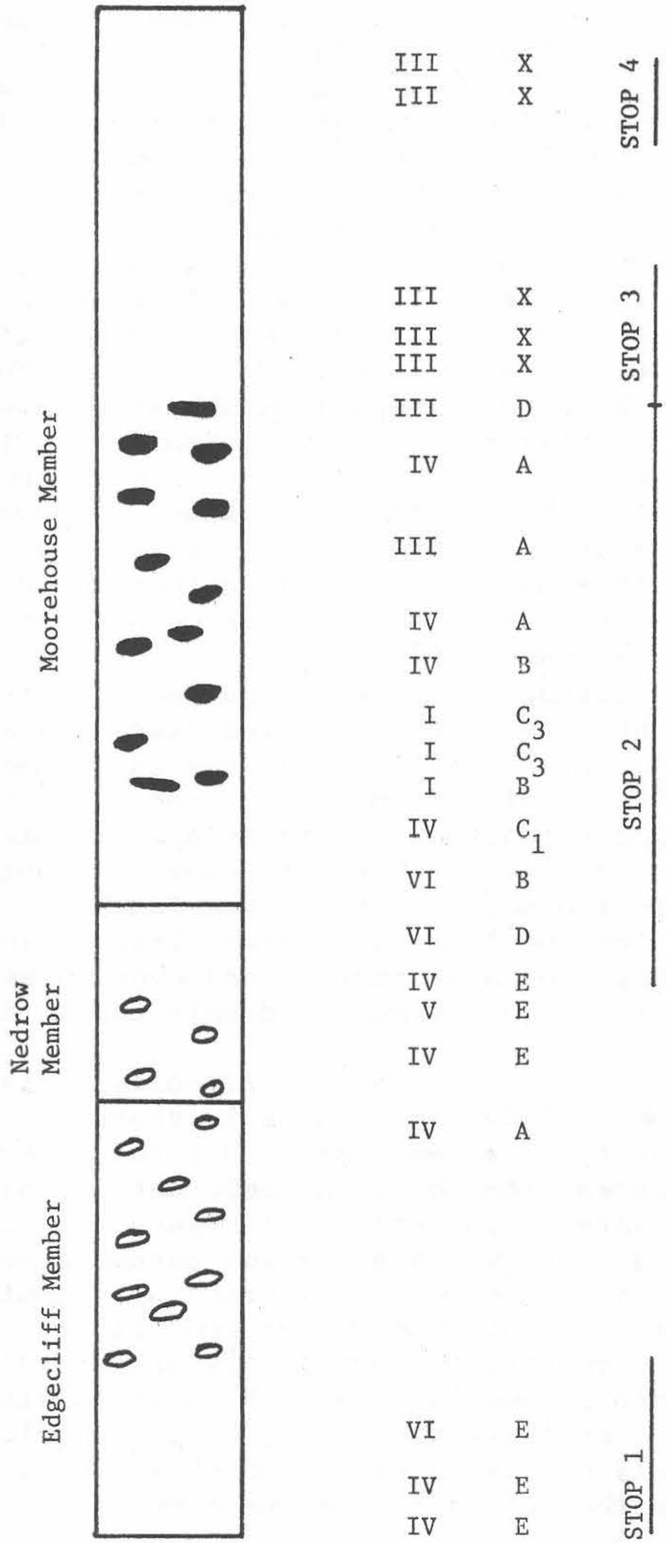

Figure 5. Idealized Onondaga section in the Helderberg area showing trip stops, distribution of fossil assemblages and lithologic types. Scale 6mm = 1m. The ovals represent chert.

Biologic Paleoenvironmental Interpretations

A Q-mode ordination performed on faunal census data failed to account for sufficiently reliable dissimilarity between samples. This is often the case with biologic data, where many environmentally significant variables are complexed resulting in a large amount of variability between samples. The biologic ordination model did show a general trend of coral-rich opposed to brachiopod dominated faunas.

Several of the faunal assemblages can be interpreted by classic paleoecologic generalizations. An abundance of massive tabulates such as found in assemblages A and E indicate life in clean, shallow, agitated waters (Wells, 1967; Philcox, 1970; Crowley, 1973). Because *Acinophyllum* prefered somewhat calm waters (Mecarini, 1964; Bamford, 1966; Poore, 1969) assemblage E probably developed in slightly deeper or more protected areas than assemblage A. The abundance of ramose bryozoans in assemblage B indicates an environment of relatively calm water near wave-base (Anderson, 1971). The auloporids of assemblages C_1 and C_2 are indicative of quiet water conditions (Lowenstam, 1957; Vopni and Lerbekmo, 1972).

From the preceding, it appears that water agitation and corresponding substrate characteristics exerted a great deal of control on the fauna. Unfortunately, the environmental tolerances of most Paleozoic organisms are not sufficiently understood to be diagnostic of most environmental parameters. However, by plotting the occurrences of faunal assemblages on the lithologic ordination model (Fig, 6) the fauna can be more fully interpreted and the model tested. Assemblages X and D are not included in Figure 6 because their areas overlap several other assemblages and their inclusion would only result in visual confusion.

Bear in mind that Figure 6 is a lithologically based ordination model on which independent data is plotted. It is evident, from the array of assemblages, that the water agitation gradient interpreted from the lithologic data is also reflected in the biologic data. Examination of Figure 6 in conjunction with Table 3 clearly shows coraliferous assemblages of clean, well-agitated waters opposed to the brachiopod dominated assemblages of more turbid, quiet-water conditions. Thus the paleoecologic interpretations previously arrived at and the lithologic paleoenvironmentas are both substantiated. Through the formal use of both types of data, biolithologic paleoenvironmental models can be developed which are more sensitive to change than either of their component models.

PALEOENVIRONMENTAL RECONSTRUCTION

Figures 4 and 5 show successional faunal and lithologic changes in the Onondaga Formation at Leeds and in the Helderbergs. They show less than perfect correspondence between the faunal assemblages and the lithologic types. This lack of correspondence has several possible sources. 1) A thin-section may not fully represent the bed from which it is taken, eeen if that bed is fairly homogeneous. 2) A bed may contain several different lithologies resulting from synchronous deposition in microenvironments. 3) Organisms respond to different environmental stimuli than do sediments. A small change in water agitation could significantly alter the character of a sediment and be totally unrepresented in the fauna if that fauna was not extremely sensetive to water agitation. The opposite also holds true. A slight change in water chemistry or temperature could affect the fauna and leave no trace in the sediment. This may explain the demise of the Edgecliff reefs in rocks which haven't revealed a reason for their termination. Despite their imperfedt correspondences, the lithologic and faunal successions within the formation do not present any paleoenvironmental contradictions.

The Leeds section shows little environmental change throughout the formation's development. Deposition began with a coral dominated, brachiopod rich fauna, which lived in relatively calm, slightly turbid water. Following this, the ramose bryozoan domianted assemblage B colonized the area, and remained throughout the formation's duration. The lithology does not record an environmental change accompanying the change to assemblage B. The ordination models and the persistence of assemblage B indicate that the ramose bryozoans were tolerant of fluctuations in turbidity and water agitation. The fauna flourished and caused the deposition of more limestone in less time than in the western part of the formation. Therefore, despite its calm character, water circualtion must have been sufficient of supply the nutritional requirements for such prolific growth.

The only major environmental change in the Leeds section is shown by three successive occurrences of lithologic type V. This indicates deposition in clean, agitated water and may represent shallower than normal conditions. Even this change is not reflected by a faunal change. The upper Leeds samples mark a return to calm water conditions. Despite lack of exposure of the uppermost Onondaga in this area, it is interesting that those upper beds which are present show no evidence of increasing turbidity in preparation for deposition of the Marcellus Shale.

Unlike Leeds, the Helderberg section shows significant environmental changes during Onondaga deposition, The lowermost beds contain a coraliferous assemblage which lived in clean, shallow, agitated waters. These conditions are indicated by spar-rich, calcisiltite-poor lithologies and by the prolific fauna of massive tabulates and rugose corals. This environment endured through the deposition of those beds assigned to the Nedrow Member and represents a significant deviation from type area conditions. Moorehouse deposition began in agitated water, but under conditions which favored a brachiopod-ramose bryozoan assemblage rather than a coraliferous one. The black chert section of the Moorehouse is coincident with a mud maximum in this area and fossil assemblages rich in brachiopods, bryozoans, and detritus feeding trilobites. This fauna indicates deposition in quiet, slightly turbid waters. If there is any "shaley Nedrow" of Oliver (1954) in the Helderberg area, the black-chert section is most likely it. Near the top of the black-chert section, and above, turbidity decreased and water agitation increased somewhat...Brachiopods remained common and corals returned to the area. The bryozoans, which appear to have been selectively favored by turbidity, decreased in abundance. Following a period of mildly agitated conditions, the waters near the sea bottom once again calmed. The water did not become turbid. Relatively calm, nonturbid conditions favoring an overwhelmingly brachiopod dominated fauna continued until termination of the limestone deposition.

Paleoenvironmental Significance of the Missing Mud

The eastern New York sections of the Onondaga Formation do not reflect the prominent argillaceous influxes which caused Nedrow deposition in central New York. Within the Nedrow Member the abundance of terrigenous mud decreases east and west of the type area. If the mud originated as an early pulse from the Acadian Orogeny, it would have had to be transported westward across the eastern areas where it should have left its mark. Environmental conditions were correct for mud deposition, but very little was deposited. It is unreasonable to expect the mud to bypass eastern New York where calcisiltite was deposited in quantity. Therefore, the mud must have originated to the north and was transported southward into central New York through the trough postulated by Oliver (1954) and Lindholm (1967). Paleocurrent direction measurments taken in central New York support this conclusion.

Another significant aspect of mud deposition in the eastern Onondaga is seen in the contact of the limestone with the overlying shale formation. In central New York the contact is gradational over a sequence of beds measuring a meter or more.

Eastward the gradational sequence thins until it is so abrupt in eastern New York that it was once believed to be an erosional surface (Chadwick, 1944). The uppermost Onondaga beds in the Helderberg and Cobleskill areas show no significant argillaceous increase except in the very top limestone bed. Also absent from the east are the shale/limestone cycles of the type area. It is concluded that the mud influx which terminated Onondaga deposition inundated the eastern area very rapidly and then slowly prograded across the remainder of the state. This rapid influx could have been initiated by a subsidence pulse of the P.A.C. variety (Anderson, Goodwin, and Cameron, 1978), by infilling and overrunning an eastern basin (Oliver, 1954), or by a rapid uplift of the Acadian Orogeny. In the final analysis the latter mechanism caused the termination of Onondaga deposition as the limestone's fauna was choked and buried beneath the thickening muds of the Marcellus Shale.

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REFERENCES

Ali, S. A., Lindemann, R. H., and Feldhausen, P. H., 1976, A multivariate sedimentary environmental analysis of Great South Bay and South Oyster Bay, New York: Math. Geol., v. 8, p. 283-304.

Anderson, E. J., 1971, Environmental models for Paleozoic communities: Lethaia, v. 4, p. 287-302.

Anderson, E. J., Goodwin, P. W., and Cameron, B., 1978, Punctuated aggradational cycles (PACS) in Middle Ordovician and Lower Devonian swquences: in Merriam, D. F., ed., New York State Geol. Assoc. 50th Ann. Meeting, Guidebook, p. 204-224.

Bamford, Ross, 1966, Paleoecology of the Albright's reef Onondaga Limestone (Devonian), eastern New York: unpub. M.S. Thesis, University of Nebraska, 56p.

Black, R. A., and others, 1975, A study of the organisms and their relationship to the sediments on a Florida patch reef: unpub. report, University of Miami, 130p.

Cassa, M. R., 1979, A Middle Devonian bioherm, Niagara Peninsula, Ontario: Geol. Soc. Amer., N.E. Section, Programs with Abstracts, v. 11, n. 1, p. 7.

Chadwick, G. H., 1944, Geology of the Catskill and Katterskill quadrangles. Pt. 2: N.Y. State Mus. Bull. 336, 251p.

Collins, C. M., 1978, Quantitative analysis of the faunal and sediment associations on Roberts Hill reef, Onondaga Formation, in eastern New York: unpub. M.S. Thesis, R.P.I., 76p.

Cumings, E. R., 1932, Reefs or bioherms?: Geol. Soc. Amer., Bull., v. 43, p. 331-352.

Crowley, D. J., 1973, Middle Silurian patch reefs in Gasport Member (Lockport Formation), New York: Am. Assoc. Pet. Geol., Bull., v. 52, p. 283-300.

de Caprariis, Pascal, Lindemann, R. H., and Collins, C. M., 1976, A method for determining optimum sample size in species diversity studies: Math. Geol., v. 8, p. 575-581.

de Caprariis, Pascal, and Lindemann, R. H., 1978, Species richness in patchy environments: Math. Geol., v. 10, p. 73-90.

Finks, R. M., and Lamster, I. B., 1979, Differential coral growth in Onondaga Reefs: Geol. Soc. Amer., N.E. Section, Programs with Abstracts, v. 11, n. 1, p. 13.

Folk, R. L., 1962, Spectral subdivision of limestone types: in Ham, W. E., ed., Classification of carbonate rocks. Am. Assoc. Pet. Geol., Mem. 1, p. 62-84.

Friedman, G. M., 1978, Recognition of post-Paleozoic reefs: an experience in frustration, in Friedman, G. M., ed., International Association of Sedimentologists, Abstracts, v. 1, p. 220.

Friedman, G. M., Amiel, A. J., and Schneidermann, N., 1974, Submarine cementation in reefs: example from the Red Sea: Jour. Sed. Pet., v. 44, p. 816-825.

Grabau, A. W., 1903, Paleozoic coral reefs: Geol. Soc. Amer., Bull., v. 14, p. 337-352.

_____, 1906, Guide to the geology and paleontology of the Schoharie Valley in eastern New York: N.Y. State Mus. Bull., 92, p. 77-386.

Hall, James, 1841, Fifth annual report of the fourth geological district, N.Y. Geol. Survey Ann. Rept. 5, p. 149-179.

_____, 1859, Descriptions and fugures of the organic remains of the lower Helderberg Group and the Oriskany Sandstone: Natural History of N.Y., Paleontology, Part 6, v.3, 532p.

Heckel, P. H., 1974, Carbonate buildups in the geologic record: in Reefs in time and space. Soc. Econ. Paleontologists and Mineralogists, Spec. Pub. 18, p. 90-154.

Kauffman, E. G., 1974, Cretaceous assemblages, communities, and associations: western interior United States and Caribbean Islands, in Principles of benthic community analysis. Sedimenta IV, Comp. Sed Lab., Univ. Miami, p. 12.1-12.27.

Kauffman, E. G., and Scott, R. W., 1974, Basic concepts of community ecology and paleoecology: in Structure and classification of paleocommunities. Dowden, Hutchinson, and Ross, Inc., Stroudsburg, PA, p. 1-28.

Kissling, D. L., and Coughlin, R. M., 1979, Succession of faunas and frameworks in Middle Devonian pinnacle reefs of south-central New York: Geol. Soc. Amer., N.E. Section, Programs with Abstracts, v. 11, n. 1, p. 19.

Laporte, L. F., 1971, Paleozoic carbonate facies of the central appalachian shelf: Jour, Sed. Pet., v. 41, p. 724-740.

Lindemann, R. H., 1974, Quantitative paleoecology of the Edgecliff biostrome, Onondaga Formation, in eastern New York: unpub. M.S. Thesis. R.P.I., 71p.

Lindemann, R. H., and Simmonds, R. T., 1977, Physical and biostratigraphy of the Onondaga Limestone in Otsego County, New York: in Wilson, P. C., ed., N.Y.S. Geol. Assoc., 49th. Ann. Meeting, Guidebook, B-9.1-B-9.29.

Lindholm, R. C., 1967 Petrology of the Onondaga Limestone (Middle Devonian), New York: Ph.D. thesis, Johns Hopkins Univ., 188p.

_____,1979a, Carbonate petrology of the Onondaga Limestone (Middle Devonian), New York: A case for calcisiltite: Jour. Sed. Pet., v. 39, p. 268-275.

_____, 1969b, Detrital dolomite in Onondaga Limestone (Middle Devonian) of New York: its implications to the "dolomite question": Am. Assoc. Pet. Geol., v. 53, p. 1035-1042.

Lowenstam, H. A., 1957, Niagaran reef in ghe Great Lakes area: Geol. Soc. America, Mem. 67, v.2, p. 215-248.

Mecarini, Gino, 1964, Ecological succession in a Middle Devonian Bioherm: unpub. M.S. Thesis, Brown Univ., 48p.

Oliver, W. A., Jr., 1954, Stratigraphy of the Onondaga Limestone (Devonian) in central New York: Geol. Soc. Amer., Bull., v. 65, p. 621-652.

_____, 1956a, Stratigraphy of the Onondaga Limestone in eastern New York: Geol. Soc. Amer., Bull., v.67, p. 1441-1474.

_____, 1956b, Biostromes and bioherms of the Onondaga Limestone in eastern New York: N.Y.S. Mus. and Sci. Service Circ. 45, 23p.

_____, 1963, The Onondaga Limestone: in Geol. Soc. Amer. Guidebook, 76th. Ann. Meeting, Mohawk and Central Hudson Valley p. 11-16.

Ozol, M. A., 1963, Alkali reactivitu of cherts and stratigraphy and petrology of cherts and associated limestones of the Onondaga Formation of central and western New York: Ph.D. Thesis, R.P.I., 228p.

Park, R. A., 1968, Paleoecology of Venericardia sensu lato (Pelecypoda) in the Atlantic and Gulf Coastal Province: an application of paleosynecologic methods: Jour. Paleo., v. 42, p. 955-987.

_____, 1974, A multivariate analytical strategy for classifying paleoenvironments: Math. Geol., v. 6, p. 333-352.

Philcox, 1970, Geometry and evolution of the Palisades reef complex, Silurian of Iowa: Jour. Sed. Pet., v. 40, p. 171-183.

Piotrowski, R. G., 1976, Reef hunting in McKean County continues: Pennsylvania Geology, v. 7, n. 1, p. 2-3.

Poore, R. Z., 1969, The Leroy bioherm: Onondaga Limestone (Middle Devonian) western New York: unpub. M.S., Thesis, Brown Univ., 69p.

Rickard, L. V., 1975, Correlation of the Silurian and Devonian rocks in New York State: N.Y.S. Mus. and Science Service, Map and Chart Ser. No. 24.

Vanuxum, L., 1839, Third annual report of the geological survey of the third district: N.Y. Geol. Survey Ann. Rept. 3, p. 241-385.

Vopni, L. K., and Lerbekmo, J. F., 1972, Sedimentology and ecology of the Horn Plateau Formation: Middle Devonian reef, Northwest Territories, Canada: Geologische Rundschau, v. 61, p. 626-646.

Walker, K. R., and Alberstadt, L. P., 1975, Ecological succession as an aspect of structure in fossil communities: Paleobiology, v.1, p.238-257.

Warters, H. R., 1972, Pinnacle reefs of Middle Devonian Onondaga Limestone, upstate New York and Northern Pennsylvania: Am. Assoc. Pet. Geol. Bull., v. 56, p. 660.

Wells, J. W., 1967, Corals as bathometers: Marine Geology, v. 5, p. 349-365.

Williams, L. A., 1979, Physical and biotic controls in Thompsons Lake reef (Middle Devonian, New York State) and their relevance to community succession models: Geol. Soc. Amer., N.E. Section, Programs with Abstracts, v. 11, n. 1, p. 59-60.

ROAD LOG

Miles Total	Miles from last point	Route Description
0	0	Leave RPI Field House by way of Peoples Ave.
0.7	0.7	Left on 8th Street.
1.2	0.5	Right on Rt. 7, cross bridge over Hudson River.
2.1	0.9	Right on Rt. 32.
2.3	0.2	Through intersection and right onto 787.
7.3	4.8	Exit right onto Rt. 90.
11.3	4.0	Take Exit 4 onto NY 85, south.
23.6	12.3	Right onto Rt. 157, toward Thacher Park.
23.8	0.2	Take the first left onto Indian Ledge Rd.
24.5	0.7	STOP 1. Park on right side of road just above lowermost limestone outcrop. Figure 5.
25.3	0.8	Return to 157 and hang a right.
25.5	0.2	Right on 85 W.
27.6	2.1	Left on Rt. 443 through Clarksville.
30.6	3.0	Right onto Flat Rock Road.
31.7	1.1	Right onto Rt. 32.
32.5	0.8	STOP 2. Park in pulloff on left above limestone outcrops. Fig. 5.
37.4	4.9	Return to intersection of 443 and 85, take a left on the latter.
37.6	0.2	STOP 3. Park in abandoned quarry on right side of road. Fig. 5.
38.5	0.9	Return to 157 and turn left, continue through Thacher Park.
44.2	5.7	Left on Beaver Dam Road.
45.7	1.5	STOP 4. Park on right side of road and walk down old lane in Thacher Park until you come to an abandoned quarry. Figure 5.
47.0	1.3	Continue west on Beaver Dam Road and turn right onto 157.
53.8	6.8	Follow previous directions to 32, turn right.
59.1	5.3	Turn left onto Rt. 143 toward Revena.
68.1	9.0	In Revena turn right onto Rt. 9W.
70.2	2.1	Turn right onto Green County Road 51.
71.3	1.1	Turn left onto Roberts Hill Road
71.4	0.1	STOP 5. Figure 1.
71.9	0.5	Right onto Green County Road 54.
72.6	0.7	Left on Highmont Road.

Miles Total	Miles from last point	Route Description
72.9	0.3	Left on Reservoir Road.
73.1	0.2	Right on Limekiln Road.
73.6	0.5	STOP 6. Park in wide area on left side of road. Figure 2.
75.3	1.7	Continue south on Limekiln Road and turn left on Titus Mill Road.
76.9	1.6	Turn left on Rt. 81.
80.0	3.1	Turn right on Rt. 9W.
89.5	9.5	Take a right onto entrance ramp to and a left onto Rt. 32, just north of Catskill.
90.9	1.4	Take the first exit off of 23 onto Green County Road 23B towards Leeds.
92.8	1.6	In Leeds turn left on Gilfeather Park Road just past Gilfeather's Silgo Hotel.
92.9	0.1	STOP 7. Park at the end of Gilfeather Park Road. Walk down into creek to lowermost limestone outcrop. Figure 4.
		Return to RPI.

TRIP B-11

Field guide to the Chatham and Greylock slices of the Taconic allochthon in western Massachusetts and their relationship to the Hoosac-Rowe sequence

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Introduction

Since 1977, the U.S. Geological Survey has been engaged in a program to compile data for a new bedrock geologic map of Massachusetts. Don Potter and I have been mapping in northwestern Massachusetts in an attempt to fill in unmapped or poorly understood areas. Detailed mapping of the Pittsfield West, Hancock, Berlin, Williamstown, and parts of the North Adams and Cheshire quadrangles have been completed.

This trip will deal principally with new data regarding the age, distribution, and structural characteristics of major boundaries at the soles of the Taconic allochthons (Fig. 1) exposed along the New York State line, and on Mount Greylock. An attempt will be made to relate these features to rocks exposed in the eugeoclinal Hoosac-Rowe belt east of the Berkshire massif. Time constraints for movement of these materials across the autochthon and the modes of emplacement will be discussed.

Stratigraphy

The stratigraphic columns can be divided into three sequences: 1) allochthonous group of rocks exposed in thrust slices resting on 2) the lower autochthonous miogeoclinal sequence, and 3) a tectonically separate eugeoclinal sequence east of the Berkshire massif.

A correlation chart is given in Figure 2.

The base of the autochthonous section, the dark albitic Hoosac Formation, interfingers with basal beds of the Dalton Formation on Hoosac Mountain (Herz, 1961). This relationship has been confirmed in remapping of the Hoosac belt. The Dalton consists of feldspathic quartzite schist and conglomerate. It rests unconformably on basement gneiss of the Berkshire massif or of the Green Mountains. Quartzites assigned to the Dalton contain Ollenelus fragments near North Adams (Walcott, 1888). Chesire Quartzite, vitreous quartzite, and the Stockbridge Formation form a continuous sequence of shallow water quartzite and carbonate rocks deposited as a miogeoclinal wedge that is the time equivalent of deeper water slope and rise sediments in the Taconic allochthons.

A major sedimentary break occurs in the Middle Ordovician, where an unconformity is recognized beneath the Walloomsac Formation. Taconic allochthons commonly rest on a cushion of Walloomsac.

The Taconic allochthonous rocks have been assigned to four slices, after Zen (1967), for discussion here: the Giddings Brook, Chatham, Rensselaer Plateau, Everett, and Greylock slices. These slices overlap eastward. In addition, several slices of distinctly Taconic-like rock are exposed east of the main Taconic allochthons. The diagrammatic

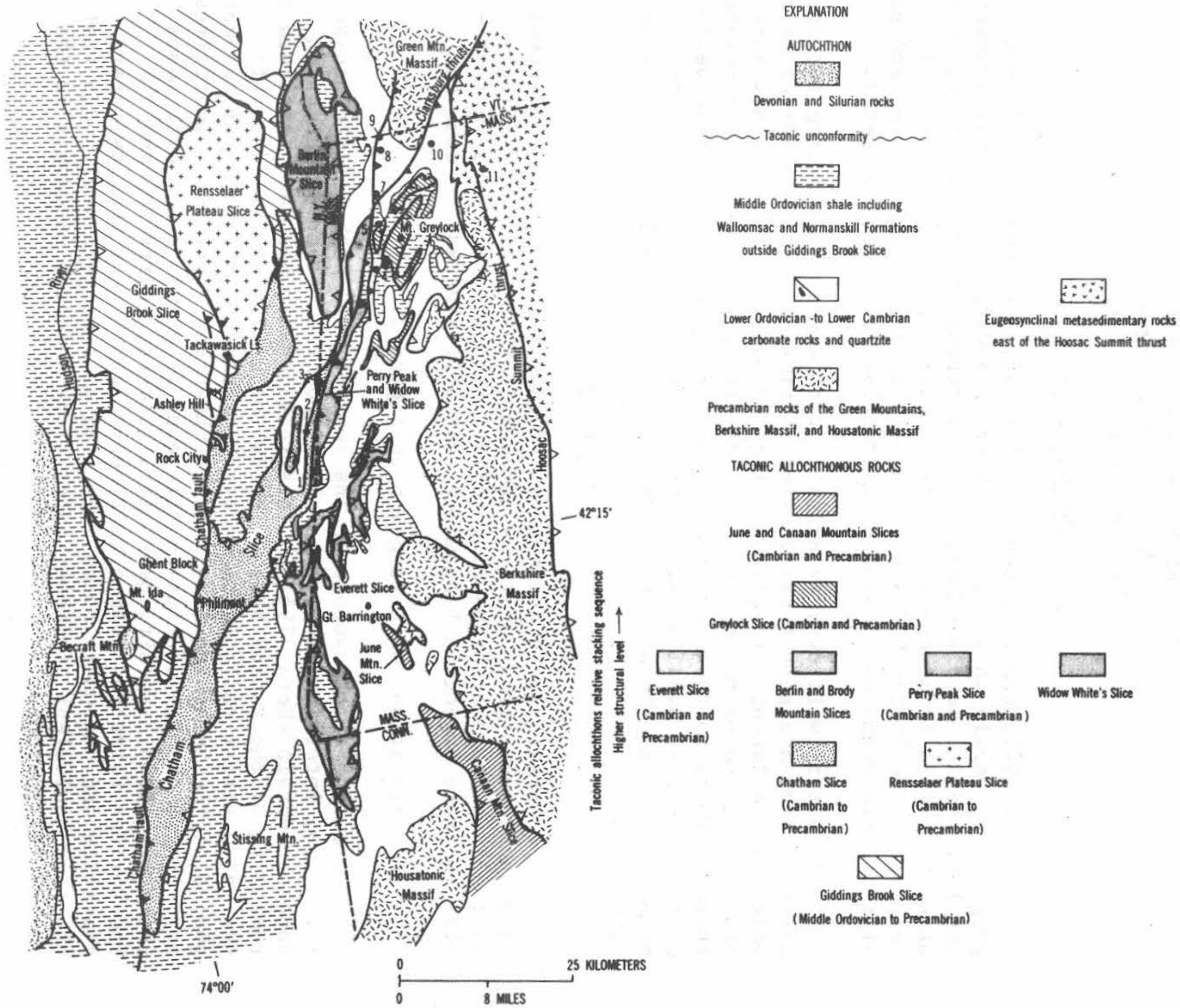


Figure 1. Regional geologic map showing generalized slices of the Taconic allochthon modified from Zen (1967), based on data in Ratcliffe (1947a), Ratcliffe and Bahrami (1975), and Potter (1972). The Chatham fault and other Acadian faults are shown with solid triangles. For relationships among Rensselaer Plateau and other slices in New York and adjacent Vermont, see Potter, this guide book

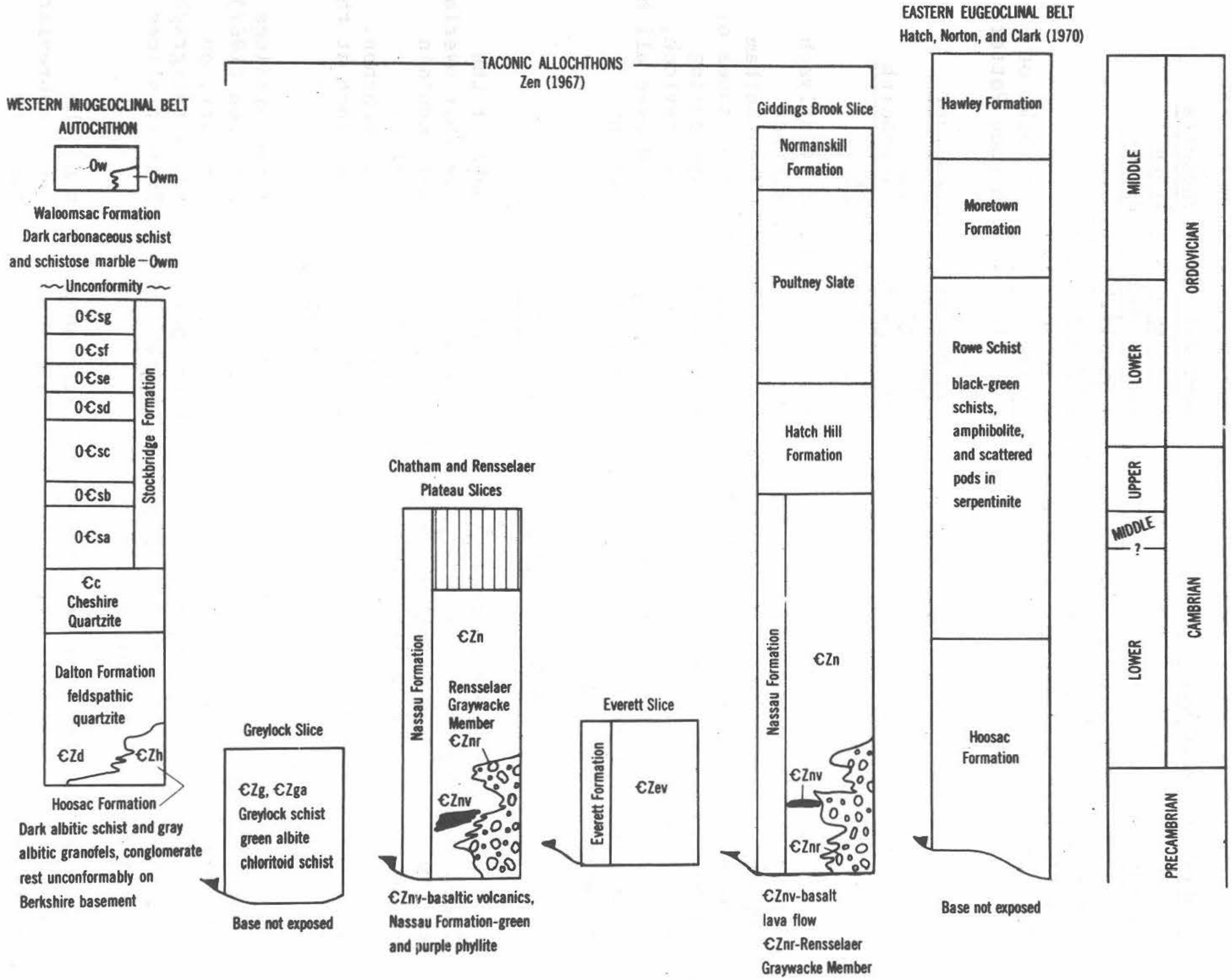


Figure 2. Correlation of major rock units in the autochthon, Taconic allochthons, and eastern eugeoclinal belt

listing below shows the general stacking sequence, and possible correlation between slices exposed in the southwest and northwest of Figure 1.

Southwest Massachusetts	Northwest Massachusetts Eastern N.Y. State
Canaan Mountain Slices (Harwood, 1975)	Hoosac east of Hoosac Summit thrust
Rocks on June Mountain (Ratcliffe and others, 1975)	Greylock slice
Everett slice	Everett slice(?) = rocks on Berlin Mountain (see Potter, this guidebook)
Chatham slice	Chatham slice=Rensselaer Plateau slice(?)
Giddings Brook slice	Giddings Brook slice=North Petersburg slice

The lowest slice contains dated Middle Ordovician wildflysch deposits along the sole, suggesting soft rock or near surface emplacement of the Giddings Brook slice (Zen, 1967). The Chatham Rensselaer Plateau and Everett slices are marked locally by zones of carbonate blocks ripped from the autochthonous Stockbridge during emplacement as hard rock slices. The highest slices, the Greylock, Canaan Mountain and June Mountain, and the allochthonous Hoosac all have synmetamorphic fold-thrust emplacement fabrics that cross cut metamorphic foliation in the slices.

A brief survey of Taconic allochthons

Zen (1967) has proposed that the allochthonous rocks of the Taconic belong to six or seven discrete structural slices that overlap eastward so that the highest structural level, the Dorset Mountain slice, in western Massachusetts is known as the Everett slice (Ratcliffe, 1969) and crops out at the east edge of the allochthon. Rocks of the Everett slice constitute the high Taconic sequence at this latitude and are presumed to have been emplaced last.

The low Taconics here are represented by rocks of the Giddings Brook, Chatham, and Rensselaer Plateau slices according to Zen (1967). The distinction between high and low Taconic is based, in part, on topographic expression, relative structural position, and stratigraphic considerations. For this discussion the Greylock and Everett slices are considered high Taconic slices.

Zen further proposed that the stratigraphic range of the individual slices is greatest in the lowest slices and most abbreviated in higher slices, which contain rocks largely of inferred late Precambrian (Proterozoic Z) age (Zen, 1967). The lowmost and westernmost slices, the Giddings Brook and Sunset Lake (in Vermont), were emplaced by gravity gliding in the Middle Ordovician, contemporaneously with wildflysch-like (Forbes Hill Conglomerate of Zen, 1961) material that contains fossiliferous and nonfossiliferous fragments of the allochthon itself. Graptolites of Zone 13 (Berry, 1962, p. 715) in the matrix of the wildflysch-like conglomerate that underlies the Giddings Brook (North Petersburg slice of Potter, 1972) and Sunset Lake slices date the time of submarine emplacement (Zen, 1967; Bird, 1969). Graptolites of Zone 12 (Berry, 1962) have been collected from the Walloomsac Formation which underlies wildflysch-like conglomerate at the eastern (trailing) edge of the Giddings Brook slice (North Petersburg slice) at Whipstock Hill (Potter, 1972). This suggests that the Giddings Brook slice was emplaced during the timespan represented by Zones 12 and 13, although the lack of fossils in the matrix at Whipstock Hill precludes proof of this point.

The Chatham slice overrides the Giddings Brook slice along the Chatham fault of Craddock (1957) (Fig. 1). The fault zone contains slivers of carbonate and other rocks and appears to be an Acadian fault (Ratcliffe and Bahrami, 1976). To the east, the Chatham slice is overlain by the Everett slice at the sole of which are distinctive tectonic breccias that consist of complex mixtures of fragments of all the shelf sequence carbonate rocks, and Walloomsac and Everett, lithologies concentrated along the soles of imbricate slices (Zen and Ratcliffe, 1966; Ratcliffe, 1969, 1974a).

Chatham slice and the Chatham fault

The rocks of the Chatham slice were studied previously by Craddock (1957) and Weaver (1957), who did not map detailed stratigraphy within the slice. Thus, Zen in his 1967 compilation had only limited data available bearing on Chatham slice stratigraphy. Rocks assigned to the Chatham slice extend northward along the New York-Massachusetts State line (Fig. 1).

Rocks of the Chatham slice resemble closely gray-green and purple slate (Mettawee), Rensselaer Graywacke, and other rocks of the Nassau Formation (Bird, 1962a) in the Giddings Brook and Rensselaer Plateau slices. The Chatham slice sedimentary rocks (Nassau) probably also are pre-Olenellus in age. Distinctive but sporatically developed diabasic basalts, pillow lavas, and pyroclastic volcanic rocks are spatially associated with the base of the Rensselaer facies in all three slices (Balk, 1953; Potter, 1972; Ratcliffe, 1974a).

Massive quartzites, similar to those of the Zion Hill Member of the Bull Formation of Zen (1961) and the Curtis Mountain Quartzite of Fisher (1962), which crop out in the Chatham slice are not clearly one horizon but underlie coarse Rensselaer-type Graywacke in many areas. One of these quartzites that has a polymict basal conglomerate (Ratcliffe and others, 1975, stop 5) contains angular fragments of basaltic or andesitic scoria. This suggests that the relatively thin subgraywackes and quartzites exposed in the western part of the Chatham slice may be tongues of Rensselaer-like material that extended westward into the sedimentary basin.

Importantly, the Rensselaer-like graywacke of the Chatham slice in the Austerlitz outlier and in the State Line quadrangle overlies a considerable thickness (300-1,000 m) of purple and green slate, siltstone, and laminated green slate typical of the Nassau elsewhere. However, Rensselaer Graywacke of the Giddings Brook and Rensselaer Plateau slices appears at or near the base of the preserved stratigraphic succession. The stratigraphic position of the Rensselaer within the original (as opposed to the allochthonous) sequence is really moot, because the original sequence is nowhere preserved intact, and we do not know at present if the Chatham slice relationships are the rule rather than the exception.

Internal structure of the Chatham slice is complex and folds of pre-emplacement age have been recognized in many areas, however, the regional Taconic slaty cleavage crosscuts the thrust contacts. Locally, a wildflysch-like zone is preserved at the sole (Ratcliffe and others, 1975, p. 82) and Stop 3 of this trip. In addition, evidence for interleaved fault slivers of autochthnous carbonate and allochthonous rocks as well as localized recumbent folding in the autochthon are recognized (see Stop 1).

Everett slice

Rocks of the Everett Formation that form the high Taconic Everett slice are greenish-gray, green, and locally purplish slate containing relatively minor amounts of interbedded Rensselaer-like graywacke. In general, the Everett Formation resembles rock of the lower part of the Nassau Formation when the effect of increased metamorphic grade is considered. Zen and Hartshorn (1966), Zen and Ratcliffe (1966), and Ratcliffe (1969, 1974a, 1974b) consider the Everett rocks to be as old or older than rocks of the western slices. No fossils have ever been found within rocks of the Everett slice, and are not likely to be, so that the age problem may never be completely resolved. The Everett slice is about 12 km wide and probably originated from a depositional site at least this wide. Internal structure within the Everett slice, however, is poorly known, owing to the lack of coherent stratigraphy, the possibility of stacked slices of material that all rooted from the same zone could reduce this 12 km estimate for the original sedimentary width.

The contact relationships of the Everett and Chatham slices are complicated because the leading edge of the Everett slice is a zone of intense imbrication involving both allochthonous and locally detached autochthonous rocks. A belt of parautochthonous Walloomsac commonly separates the two slices (Fig. 1). Locally slivers several kilometers long of purple and green slates typical of Chatham slice rocks are found incorporated in the parautochthonous belt of Walloomsac. In addition, at least two imbricate slices of Everett rocks are found above the Walloomsac sliver and above the slivers of Chatham slice rocks (Ratcliffe, 1974a).

The contact of parautochthonous Walloomsac on the Chatham slice and between the Everett and all other rocks is marked locally by an intensely developed tectonic breccia composed of inclusions of Stockbridge Formation. These breccias mark tectonic movement zones that differ from conventional fault zones in one important aspect. The carbonate clasts in the highly imbricated slate matrix are exotic blocks, not derived from the present hanging wall or foot wall, but from the autochthonous Stockbridge belt, and thus are considered tectonic inclusions transported within the movement zone from some site to the east. The tectonic breccia is evidence for a thrust beneath the Everett slice, which is independent of the regional stratigraphic arguments (Zen and Ratcliffe, 1966). These breccias have been mapped throughout southwestern Massachusetts (Zen and Ratcliffe, 1966; Ratcliffe, 1974a, 1974b) and are found in east and west dipping contacts as well as along the noses of plunging folds of the thrust contacts. The emplacement of the breccias predated the first regional metamorphism and the penetrative foliation that crosscut the contact of the thrust slices with the autochthon. Emplacement of the Everett slice resulted in brittle deformation (plucking) of the carbonate rocks, indicating that the carbonate rocks were lithified at the time of thrusting. Similar brittle deformation of the pelitic rocks is not recognized, although an abnormally strong phyllitic foliation has been noted by Zen (1969) immediately adjacent to the carbonate slivers. Clearly, emplacement of the Everett slice involved hard rather than unconsolidated sediments. The age of emplacement of the Everett slice is unknown, but on the basis of geometric relationships, its final movements postdated emplacement of the Chatham slice in the Middle Ordovician and predated formation of the regional slaty cleavage that probably is Late Ordovician in age.

Structural Geology of the Greylock slice

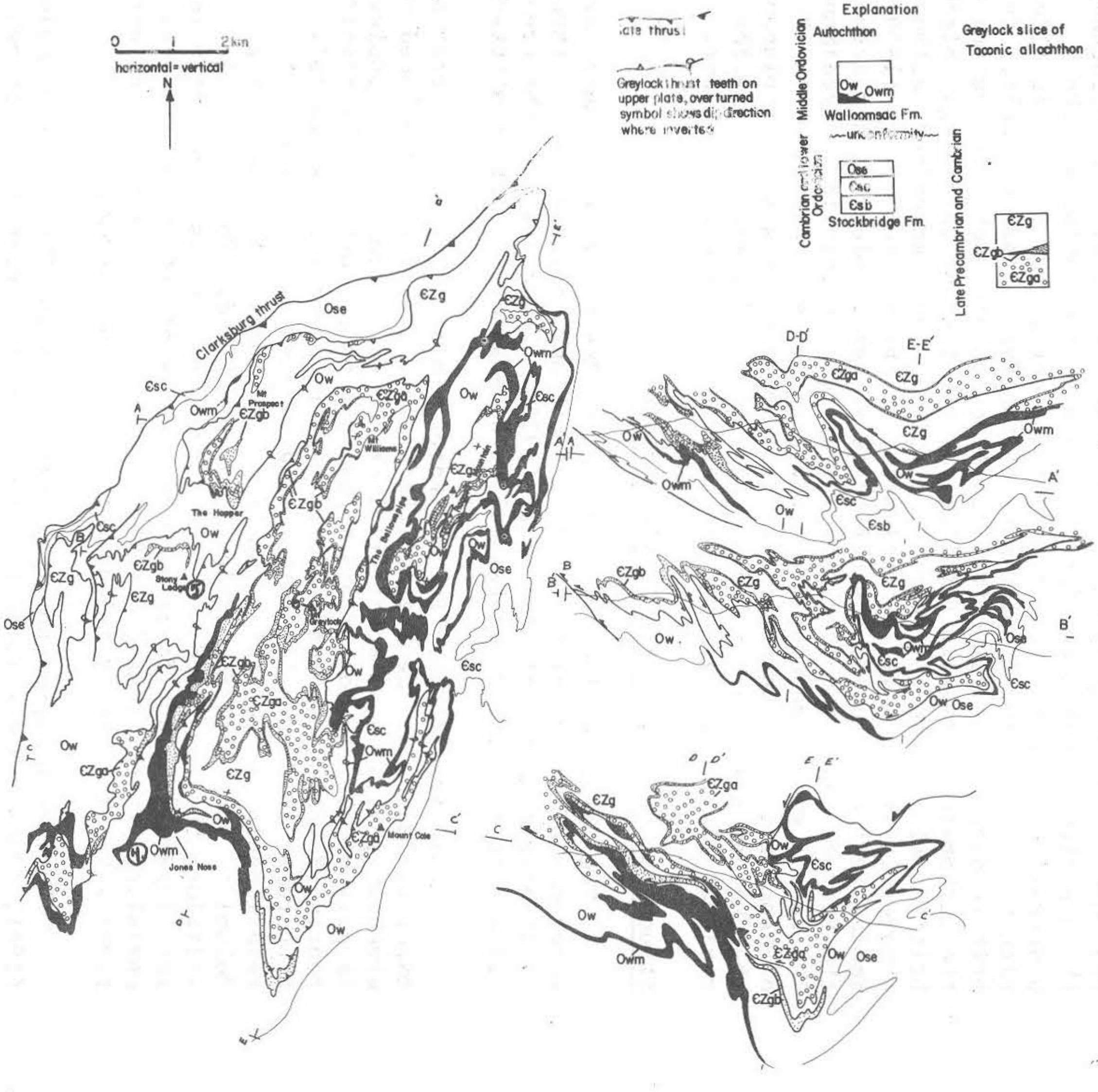
The Greylock slice was proposed by Zen (1967) to account for the occurence of Taconic-like rocks on the Mount Greylock above a thrust fault mapped by Prindle and Knopf (1932). Zen correlated the Greylock slice with the Dorset Mountain slice because of similar stratigraphy.

Louis Prindle and Eleanora Knopf, in an extremely penetrating paper (1932) on the geology of the Taconic quadrangle, concluded that greenish phyllites on Mount Greylock formed a thrust sheet consisting largely of albitic Hoosac and minor amounts of lustrous chloritoid-bearing Rowe Schist (rocks of the eastern eugeoclinal sequence). Structural arguments for a thrust were based on the discordant relationship of the schists on Mount Greylock to the underlying dolomitic and calcitic marble of the Stockbridge Formation. In addition, they proposed that the thrust contact was recumbently folded into large recumbent folds with amplitudes of approximately 6 km. They envisioned a complicated movement history in which the already emplaced rocks were recumbently folded during continued movement.

Norm Herz mapped both the North Adams and Chesire quadrangles (1961, 1958) and concluded that the Greylock Schist was conformable with the underlying Walloomsac Formation of Middle Ordovician age.

Results of remapping Mount Greylock in the Williamstown, North Adams, and Cheshire quadrangles are shown in Figure 3 and Figure 4. The contact of the Greylock Schist with the autochthon is shown as a highly

GEOLOGIC MAP AND CROSS SECTIONS OF THE MOUNT GREYLOCK ALLOCHTHON



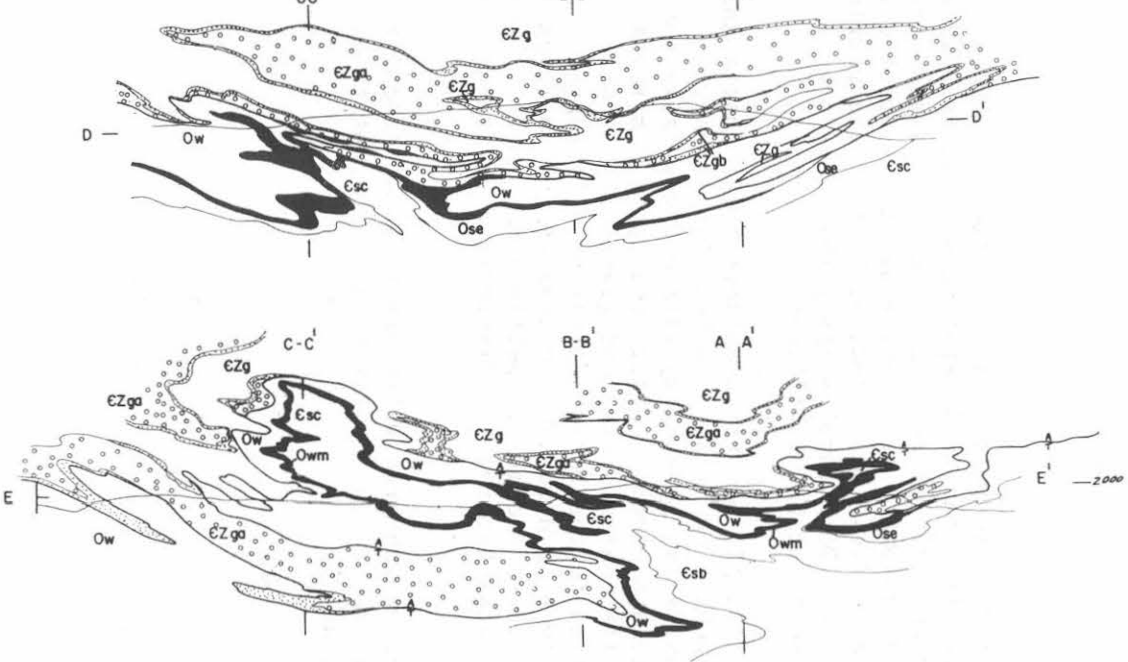
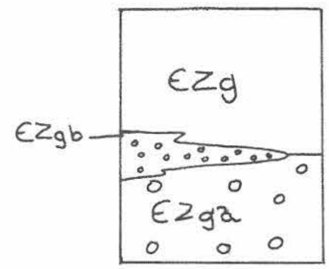


Figure 3. Geologic map and cross sections of the Greylock allochthon. Designs are intended to correlate with designs used on Fig. 5.

Figure 4. Stratigraphy of the Greylock slice. Compare with descriptions of Hoosac Formation on Fig. 5.

Greylock Schist



€Zg Predominantly light-green to pale yellowish-green, lustrous chloritoid quartz phyllite with minor beds spotted with white albite. Local well-laminated gray and gray-green phyllite contain discontinuous beds 1 to 2 cm of quartzose dolomite or of white quartzite spotted with brown weathering pits of ankerite. Minor beds of blue quartz pebble conglomerate up to 1 m thick. Dark purplish-gray phyllites are interlayered in roadcuts south of Mount Williams.

€Zgb Black, dark-gray chloritoid or stilpnomelane-albite quartz knotted schist and quartz pebble schist or metagraywacke. Lenses of gray pinstriped feldspathic quartzite, green-gray vitreous quartzite and quartz pebble conglomerate are common. Minor beds up to 5 m thick of white-spotted biotite albite quartz schist and granulite resemble closely the more albite beds in the Hoosac Formation. On Ragged Mountain, salmon pink weathering dolostone in beds up to 1 m thick is interbedded with albitic schist and quartz pebble conglomerate. Unit grades into more albitic rocks lacking the distinctive quartz-knotted appearance.

€Zga Dark-gray to light-greenish-gray white-albite studded schist chlorite granulite, either massive or poorly bedded. Magnetite or ilmenite locally is abundant.

folded thrust fault. Internal structures within the allochthon are discordant to the fault. Various units of the autochthon terminate against the contact as Prindle and Knopf described.

The general structure of Mount Greylock is a doubly plunging synformal mass produced by crossfolding of older structures by two episodes of late northeast trending folds with northwest overturned to upright axial surfaces. The late folds are given expression by a strong crenulation cleavage or spaced slip cleavage in the axial planes.

Older foliated structures are complexly folded and appear to consist of at least two recumbent to strongly westward to southwesterly overturned structures. One of these nearly recumbent fold phases folds the thrust contact in two large scale recumbent and reclined folds as shown in the cross sections A-A′ through E-E′. Fold styles are isooclinal with fold axes commonly inclined at high rake angles to the northeast, east, southwest, and west, depending upon the dip direction of the axial surfaces. These folds postdate metamorphic foliate structures both in the autochthon and the allochthon. Near thrust contacts, minor recumbent folding of foliation is especially intense. This indicates that thrusting postdated some metamorphism in both autochthon and allochthon alike.

Folds older than the allochthon's emplacement are found within the allochthon. Detailed tracing of the three part stratigrahy within the Greylock allochthon reveals a major recumbent fold repetition, with the plane of symmetry passing through the outcrop belt of the chloritoid-rich phyllite unit on Mount Greylock. This recumbent structure is judged to be a west-facing syncline. This assumption is based on correlation of the albitic member with the observed lower units of the Hoosac Formation on Hoosac Mountain. The fold closure shown in section D-D′ is not observed on the ground and is conjectural. However, early hinge lines in the northern part of Mount Greylock plunge southwest to west and could intersect section line D-D′ in the air north of Mount Williams. The large lefthand digitation on the lower limb in the area of A-A′ crossing of section D-D′ suggest that the closure may be expected where drawn.

The interpretation of the Greylock structure presented here differs from that of Prindle and Knopf. Previous workers (Pumpelly, Wolf, and Dale, 1894) showed carbonate rocks (Bellowspipe Limestone) encircling Greylock Schist on Mount Greylock. This belt formed the axial portion of the large recumbent anticline shown by Prindle and Knopf. Remapping shows that this limestone belt is not continuous around the north end of the mountain but is traceable into the main belt of autochthonous Stockbridge in the Adams area, as Prindle and Knopf show. Rather than encircling Mount Greylock, this belt of Stockbridge is interpreted as a recumbent anticline cored by unit C of the Stockbridge, that is downfolded in the main synform on Mount Greylock. The lower limb of this structure is exposed in the hooklike bend west of Mount Cole where carbonate rocks and the Walloomsac Formation overlie Greylock Schist in northeast-plunging folds.

The model for emplacement of the Greylock slice differs little from that outlined by Prindle and Knopf. Emplacement took place under metamorphic conditions and involved folding of older metamorphic structures. Metamorphism outlasted thrusting as chloritoid and albite clearly are imprinted on the foliated fault fabric. Folding accompanied thrusting presumably as a result of large scale westward transport of higher slices of the tectonic cover, that consists of the Berkshire massif and the eastern eugeoclinal sequence.

Cross folds and slip cleavages of two different orientations are superposed on all of the Taconic fabrics.

Relationship of Greylock slice to the Hoosac Formation, the Hoosac summit thrust and root zone of the allochthons

Stratigraphic comparisons of Greylock stratigraphy with that of the Hoosac Formation show striking similarities. Prindle and Knopf correlated the two sequences but suggested that the more aluminous green phyllite on the Greylock represented Rowe rather than Hoosac. Remapping of the Hoosac belt has shown that a major fault exists within the type Hoosac along the Hoosac summit thrust (Fig. 5). East of this fault, the Hoosac contains interbedded green aluminous phyllite identical to the chloritoid phyllite unit on Mount Greylock. Albitic units also are present as shown in Figure 5.

The green chloritoid unit on Mount Greylock is interbedded with albitic rock and is more coarsely crystalline than the type Rowe exposed east of the Hoosac belt. This unit compares more favorably with the green unit mapped within the Hoosac Formation (Fig. 5).

Coarse green or grey albitic rocks similar to those at the base of the eastern Hoosac sequence are found in the Greylock schist but are lacking in the Rowe. The Hoosac, however, contains distinctive beds of Dalton-like units and appears to have been deposited on the Berkshire 1 b.y.- old basement.

These relations suggest that the Greylock slice was derived from the Hoosac belt east of the Hoosac summit thrust but west of the Rowe depositional position. Examination of the rock fabric near the Hoosac summit thrust shows phyllonitic rock with green spears of chlorite and ultrafine grained paragonite-sericite matrix. Isoclinal recumbent folds with reclined axes are formed by folding of an older schistosity within the Hoosac of both plates. In addition, the allochthonous Hoosac contains recumbent fold repetitions that predate the Hoosac summit thrust. Post-thrust metamorphism, however, has produced static albite and garnet overgrowths of probable Acadian age on this fault fabric.

The structural characteristics of the Hoosac summit thrust and the sole of the Greylock slice are, therefore, quite similar.

A comparison of the stratigraphy, internal structure, and emplacement fabrics of the Greylock allochthons with the Hoosac allochthon suggests a common tectonic history. The final emplacement of

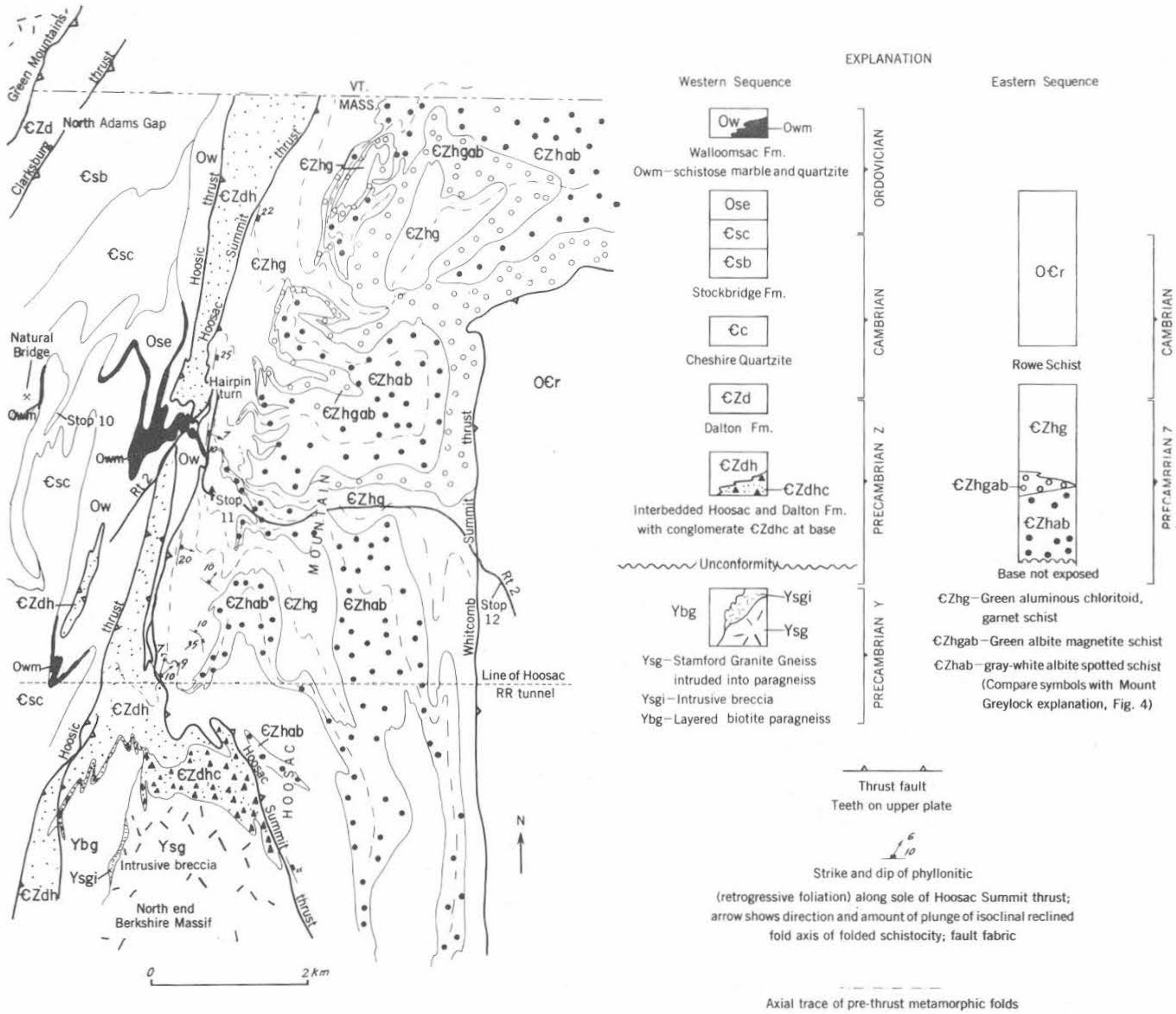


Figure 5. Geologic map of Hoosac Mountain based on mapping by N. Ratcliffe and R. Stanley, 1976-77, (unpub. data).

the Greylock allochthon across the miogeoclinal sedimentary rocks was a hard rock thrust that may have been rooted within the Hoosac belt east of the Berkshire massif. Possibly the root zone for the Greylock and other slices is buried beneath the Whitcomb summit thrust (Stanley, 1977) at the base of the Rowe sequence (Ratcliffe and others, 1975, p. 64).

Comparison of the Greylock Slice with other Taconic allochtonons

No other allochton in the main Taconic range from Vermont south to Connecticut has the structural characteristics of the Greylock allochthon. The combination of recumbent folding of the thrust surface and relatively late emplacement is remarkable and is unlike the relationships at the sole of the Giddings Brook, Chatham, or Everett slices. Allochthonous rocks similar to the Hoosac Formation on June Mountain (Ratcliffe and others, 1975) and in the Canaan Mountain allochthons (Harwood, 1979, 1975) do have these structural attributes and may have been transported westward with the Berkshire massif following initial upthrusting onto the massif rocks.

Evidently the Taconic allochthons of Giddings Brook, Rensselaer Plateau, Chatham and Everett slices were ejected from a root or slide zone east of the Hoosac summit slice prior to dynamothermal metamorphism. The Greylock and Canaan Mountain fault slices, however, did not escape metamorphism prior to emplacement. The fold style and character of emplacement fabrics for these rocks resembles closely that found in the remobilized and thrust-faulted basement rocks of the Berkshire massif itself. Large scale westward overthrusting and compressional tectonics were responsible for final emplacement of the Greylock slice probably during the event in which the Berkshire massif was thrust over the miogeocline.

The relative stacking order requires that the Everett, Greylock, and June Mountain slices were emplaced last under conditions of increasingly greater metamorphic intensity and under greater tectonic cover than the western slices. Williams (1975), on the other hand, suggested that the Newfoundland Taconic allochthons had been preassembled, with movement first occurring on eastern and highest metamorphic slices and associated ophiolite. The prestacked assemblage moved last on the melange at the floor of the lowest and nonmetamorphic slice.

Although the stacking sequence within the Chatham and higher slices is similar to that proposed by Williams, the higher slices here are all marked by evidence of tectonic slivering of local autochthonous carbonate rocks between the slices. Such a relationship requires that the present stacking order of Taconic slices with interleaving of fault slivers of the autochthon cannot be the same as a preassembled one that formed in a tectonic staging area prior to ejection of the allochthons.

The very close stratigraphic and structural similarity between the Hoosac and Greylock allochthons suggest that the higher slices in Massachusetts traveled the shortest relative distance and that the structurally lowermost and now westernmost Giddings Brook, Chatham and other slices traveled the farthest from more oceanward realms. It is important to appreciate that these arguments involve relative positions of allochthons in their restored pre-thrust positions east of the restored position of the Berkshire massif and its Hoosac cover.

These observations suggest that the Taconic allochthons in Massachusetts were emplaced sequentially. This process may have involved 1) unroofing or unpeeling from a single source (diverticulation) or, 2) hardrock thrusting and westward driving of successive fault slivers, which are samples of different paleogeographic areas in the Taconic sedimentary realm. Zen (1967) preferred the first hypothesis in order to explain the existence of older rocks and abbreviated sections found in the higher slices. He suggested that the Giddings Brook slice was detached from the upper part of the sediment column for which the high Taconic sequences formed the base.

Because of the rather extensive overlap in the stratigraphy of the oldest (Nassau) and of Nassau-like rocks of each allochthon, and the unique association of basaltic volcanics with only certain of these slices, it appears that the basal stratigraphic units (in the separate allochthons) actually are samples of quite different geographic realms. Therefore, the various allochthons probably did not all occupy the same paleogeographic position but represent lateral correlatives.

I prefer hypothesis (2) with hardrock thrust faulting produced by the accumulation of westward driven slices. Movement was initiated in the easternmost parts of the Taconic basin first (west of the palinspastic site of the Rowe). The most proximal (closest to the western craton) thrust slices (Greylock Schist) were driven out last and moved westward under a growing tectonic overburden. Tectonic imbrication within the autochthon then became important and the basement massif rocks were mobilized and thrust westward. The present stacking sequence is a result of large scale imbrication of slablike wedges of basement gneiss and cover rocks that have been thrust westward over the earlier emplaced Taconic slices.

Original depositional basin of Taconic allochthonous rocks

The original depositional basin of the Taconic allochthon rocks at this latitude, on the basis of the admittedly insecure arguments above, should have been more than 70 km wide. Palinspastic reconstruction of the Berkshire massif (see Ratcliffe and others, 1975) suggests that the Precambrian (Proterozoic) crystalline rocks of the Berkshire massif, in the Middle Ordovician, were very likely about 60 km wide and located at least 21 km farther east than their present position with respect to the miogeocline. The entire Taconic sequence could not likely have been deposited on the "basement" that was to become the Berkshire massif, as has generally been suggested (for example, Zen 1967) because rocks of the Dalton-Cheshire-Stockbridge shelf sequence were deposited on at least the western 30 km of the gneiss. Bird and Dewey (1970) suggested that much of the sequence was deposited to the east of the Grenville basement. The Taconic depositional basin probably was located largely to the east of the rocks making up the present Berkshire massif, and east of the Hoosac facies. This argument suggests that the root zone of the allochthon lies somewhere within the vicinity of the Hoosac-Rowe boundary east of the Berkshire massif. The Taconic rocks were probably deposited (initially) in an ensialic basin, with graben and horst structure and basaltic volcanism (Bird and Dewey, 1970; Bird, 1975). This basin may have evolved into a true oceanic basin with some sediment deposited on oceanic crust; however, clear evidence of this is lacking. Grenville gneissic detritus in these Taconic rocks may have been derived largely from intrabasinal sources, as the spatial relationships of the Giddings Brook-Chatham and Rensselaer Plateau slices cited earlier require. If such a model is true, and the comparison with Triassic and Jurassic(?) rift basins is valid, the Rensselaer (border conglomerate) may have been deposited throughout a considerable period of time and may not be the oldest rocks of the allochthon as commonly assumed.

Metamorphic and tectonic events in the central Taconic area of New York and Massachusetts

Figure 6 (reproduced from Ratcliffe and Harwood, 1975) presents the major tectonic features recognized in a 50 km east-west belt extending fom Mt. Ida and the Giddings Brook slice eastward into the core of the Berkshire massif.

Structures associated with emplacement of the allochthon D_1 - Phase A of Taconic orogeny

Large recumbent folds, such as Zen (1961) reported from the northern region of the allochthon, have not been generally found in the central Taconic region. Potter (1972) presents data indicating that rocks of the Giddings Brook slice are locally overturned as if on the brow of a nappe. However, broad areas of lower limb (inverted rocks) are not present in the areas mapped by Potter. Zen and Ratcliffe (1971), and Ratcliffe (1969, 1974a, 1974b), report the existence of prefoliation minor folds both in the autochthon and allochthon. Through recent mapping in the Chatham slice, Ratcliffe and Bahrami (1976) have noted that a wide range of bedding-cleavage intersections are found within individual outcrops. Steeply plunging, almost reclined, axes of major and minor folds are characteristic of both autochthonous and allochthonous rocks. The Giddings Brook slice reveals similar steeply plunging F_2 fold structures. No evidence for truly recumbent folds has been found. Wildflysch-like conglomerates are found at the sole of the Giddings Brook (Zen, 1967) and Chatham slices (Ratcliffe and others, 1975). However, the Chatham slice also contains intercalated fault slivers of autochthon near the sole but locally shows intensely developed recumbent folding of Stockbridge units beneath the thrust. These relationships suggest near surface emplacement of coherent rock rather than of unconsolidated sediments. (Stop 1).

Phase B of Taconic orogeny

Emplacement of the Everett slice (high Taconics) was marked by tectonic breccia zones that are distributed along the Everett-Walloomsac

Deformational event	Number of fold system	Types of folds and areal extent	Important tectonic features	Metamorphic event	Important crystalloblastic and other structures	Igneous intrusion	Probable age of rocks in figure 1	Orogeny
D_6	F_6	North-south open folds of foliation locally recognized in Stockbridge valley	Northwest- and north-trending normal faults		Hematite-cemented breccias		Uncertain (Middle Devonian to Late Triassic)	
D_5	F_5	N. 25°–40° E.-trending upright to northwest overturned folds of foliation, with axial planar slip or crenulation cleavage. Folds recognized throughout area of figure 1 west to Mount Ida in SW corner of Kinderhook 15-minute quadrangle, N. Y., where Taconic unconformity is folded by N. 40° E. upright folds	Refolds thrust sheets and blastomylonitic foliation	M_2 Thermal maximum Acadian metamorphism	Crenulation of sillimanite alined in axial surface of F_4 folds; granulation of garnet and staurolite that includes F_4 foliation		Middle to Late Devonian (Ratcliffe 1969a, b, 1972)	Acadian orogeny
D_4	F_4	Northwest-trending upright to southwest-overturned folds with axial planar slip, crenulation, and flow cleavage. Folds recognized throughout area of figure 1, west to Chatham, N. Y., in center of Kinderhook 15-minute quadrangle	Folds thrust sheets and blastomylonitic foliation resulting in local overturning of thrusts; northwest-trending high-angle reverse faults		Muscovite, biotite realined and recrystallized in axial surface foliation; coarse sillimanite crystallized in foliation. Garnet, staurolite include folded F_2 fabric, and blastomylonitic foliation		Middle to Late Devonian (Ratcliffe 1969a, b, 1972)	
			Granite crosscuts thrust fault and blastomylonitic foliation	Thermal maximum Taconic metamorphism	Granite lacks blastomylonitic foliation in country rocks	Granite stock, South Sandisfield quadrangle	Late Ordovician(?) (Harwood, 1972)	Phases of Taconic orogeny
D_3	F_3	Northwest-trending recumbent to strongly southwest-overturned folds of basement gneiss and large-scale southwestward thrusting of Precambrian rocks of Berkshire massif across autochthon. Fold and thrust style recognized from Windsor quadrangle, Massachusetts (Norton, 1969), south to Norfolk quadrangle, Connecticut (Harwood, unpub. data), along west front of Berkshire massif	Faulted recumbent folds and nappes, mylonite gneiss, blastomylonite associated with major thrusts Thrust sheets at June and Canaan Mountains transported with Berkshire massif		Alaskite has weakly developed blastomylonitic foliation but intrudes more highly cataclastic rock in fault zones; mylonite gneiss, blastomylonite has muscovite, biotite, hornblende with lepidoblastic texture, cataclasis of F_2 foliation, thrusting synmetamorphic	Alaskite sills in faults and magnetite mineralization	Synchronous with latest movements or thrusts (Late Ordovician?) Thrusting probably late Ordovician based on age of crosscutting granite	D
D_2	F_2	Isoclinal northeast-trending northwest-overturned to nearly recumbent folds with strong axial planar foliation which is dominant foliation in most autochthonous and allochthonous (Taconic) rocks, but not clearly present in Paleozoic rocks attached to Berkshire massif. Folds extend west to Mount Ida where unconformable beneath lowermost Devonian	Folding of Taconic thrust contacts, regional foliation and refolding of slump or soft-rock folds in Taconic allochthonous rocks	M_1	Lepidoblastic muscovite, chlorite, biotite, and ilmenite in foliation; chloritoid, albite include foliation but are linked by F_4 structures		Middle to Late Ordovician(?)	C Taconic orogeny
D_2(?)	F_2(?)	Folding and metamorphism of Lower Cambrian metasedimentary rocks attached to Berkshire massif and in independent thrust slices at June and Canaan Mountains	Coarse foliation or schistosity formed	M_1(?)	Muscovite, biotite lepidoblastic in schistosity		Time of metamorphism very uncertain depending upon original position of these rocks, and timing of tectonic events at that site (Middle Ordovician to Cambrian?)	(C)?
D_1	F_1	Intrafolial minor folds associated with Taconic thrust contacts. Soft rock or slump folds in Taconic allochthonous rocks; scale of pre-F_2 folds not determined but widespread, area shown in figure 1, west to Mount Ida	Emplacement of upper Taconic slices (here, Chatham and Everett slices) Emplacement of lower Taconic slices to west of area shown in figure 1	No metamorphism recognized	Tectonic breccias with inclusions of Stockbridge Formation along thrusts (Zen and Ratcliffe, 1971) Wild-flysch-like sedimentary rocks along base of thrusts		Uncertain (Middle Ordovician?) Middle Ordovician (Zen, 1972b, table 1)	B A Degree of basement participation and intensity of deformation increasing with time
		Middle Ordovician unconformity						
D_0		Warping of Lower Cambrian to Lower Ordovician carbonate shelf sequence; locally dips near vertical (Ratcliffe, 1969a); possible block faulting					Late Early to Middle Ordovician (Zen, 1972b, table 1)	Pre-Taconic disturbance
		Pre-Dalton unconformity						
$D_{p€}$	$F_{p€}$	Isoclinal east-west-trending folds with generally steeply dipping axial surfaces and strong axial planar foliation; deformation of all Precambrian rocks including granitic intrusions such as Tyringham Gneiss	Gneissosity in Precambrian rocks of Berkshire massif	$M_{p€}$	Diopside, sillimanite, hornblende, microcline, perthite formed in dynamothermal event	Granodiorite-quartz monzonite intrusions such as Tyringham Gneiss, syntectonic	Dynamothermal event and granite intrusion approximately 1.04 b.y. (Ratcliffe and Zartman, 1971)	Grenville orogeny
		Pre-Tyringham foliation						

Figure 6. Chronology of tectonic events recognized in southwestern Massachusetts, northwestern Connecticut, and adjacent eastern New York (reproduced from Ratcliffe and Harwood, 1975).

contact and locally between the Everett and Chatham slices (Stop 3). The emplacement of all of the Taconic slices at this latitude was premetamorphic, and no firm evidence is known in support of Bird and Dewey's (1970) suggestion that the Rensselaer Plateau and higher slices might have been metamorphosed prior to emplacement in the Ordovician. The offset chloritoid isograd shown by Potter, 1972, at the west edge of his Berlin Mountain slice might be used to support this argument; however, similar relationships could be produced by offset on Acadian thrust faults.

Phase C of Taconic orogeny (D 2 and M 1 Taconic metamorphism)

Following emplacement of all slices, regional dynamothermal metamorphism took place, and a slaty cleavage or true axial planar foliation (S_2) formed in the rocks from the vicinity of Mt. Ida eastward into the area of the Berkshire massif and presumably beyond. In the low-grade rocks, fine-grained sericite, chlorite, and lenticular quartz define the slaty cleavage. Small, round blebs of chlorite that has a 001 cleavage subparallel to bedding are ubiquitous in the low-grade rock and may be retrograded detrital biotite or diagenetic chlorite. However, lepidoblastic grains are not developed parallel to beds. Sandstone and siltstone dikes have not been found parallel to S_2, and no evidence, thus far, indicates that tectonic dewatering was an important mechanism in the formation of the Taconic slaty cleavage. Large finite strain is indicated by flattened pebbles within the slaty cleavage. Locally, intense transposition structures are developed, and false bedding is common, particularly in laminated slates and some quartzites. Taconic thrust contacts of the Giddings Brook slice (Zen, 1961; Potter, 1972), Chatham (Ratcliffe, 1974a; Ratcliffe and Bahrami, 1976), and Everett slices (Zen and Ratcliffe, 1966; Ratcliffe, 1969, 1974a, 1974b) were cross foliated and folded during the D_2-M_1 metamorphic event to produce F_2 Taconic folds on a regional scale.

Phase D of the Taconic orogeny

Emplacement of the slices of the Berkshire massif and large-scale, westward overthrusting was concommitant with metamorphism. Recumbent folds formed both in the autochthon and in gneissic rocks (see Trips B-2 and B-6 of the 1975 N.E.I.G.C. guide book for further information).

Acadian orogeny

Post-Taconic foliation structures are common throughout this belt and increase both in intensity and degree of concommitant mineral growth eastward. By using inclusion textures, we may delimit the approximate extent and character of the post-Taconic metamorphic imprint. East of the biotite isograd, approximately at the New York State line, post-S_2 mineral textures are abundant, indicating that the Acadian thermal overprint produced new mineral growth of muscovite (second generation with decussate texture), albite, chloritoid, biotite, garnet, and staurolite. The prominent mineral zonation is almost certainly composite (polymetamorphic) and is dominantly controlled by the Acadian overprint in areas east of the biotite isograd.

F_4 and F_5 folds are inconsistently developed and show contradiction of relative ages from place to place. In eastern areas, the northeast-trending refolds are the F_5 folds, whereas in the low Taconics east to the Stockbridge valley, the northwest-trending refolds are the later folds.

The Chatham fault formed during the northeast-trending refolding episode, for it is refolded by northwest crenulation folds north of Chatham (Ratcliffe and Bahrami, 1976). Locally, thrust faults with mylonitization of pre-existing foliation and chlorite-quartz-albite mineralization formed in sections of the Chatham slice containing massive quartzite and graywacke. The contact between the Chatham slice and the overlying Everett(?) slice in the area of this trip also is such a late, presumably Acadian fault.

<u>K-Ar age data from Taconic phyllites</u>

Two new K-Ar age dates on muscovite concentrates from phyllite in the vicinity of this field trip have been obtained. A sample of lepidoblastic fine-grained muscovite aligned in the regional foliation (M_1) event of Figure 6 yielded a K-Ar age of 434 ± 16 m.y. (442 ± 16 m.y. using newer decay constants).* The dated sample of purple Nassau phyllite collected 2 km southwest of Stop 1 contains well-developed F_2 folds (Fig. 6). This age confirms the Taconic age of the regional schistosity and is the first such Taconic age from the central part of the Taconic area.

A second sample of phyllonitic (retrograded) muscovite-rich phyllite was collected from the imbricate thrust zone at the contact between the Chatham and Everett(?) slices, 1 km north of Stop 2. K-Ar age of this phyllonitic muscovite is 367 ± 13 m.y. (374 ± 13 m.y.).* Thin-section examination and field observations show that the muscovite dated is aligned in the phyllonitic fabric that is subparallel to the imbricate faults. These faults cross cut and produce folds of older F_2 foliation. This age determination suggests that the imbrication and cataclasis marking the contact between the Chatham slice and the higher Everett(?) slice in the area of Stop 3 is an Acadian fault.

Regional metamorphism during the Acadian, M_2 event of Figure 6, has overprinted wide areas of western Massachusetts and the general lack of K-Ar or Rb/Sr mineral ages east of the biotite isograd probably reflects this overprinting.

* values in parentheses are ages using new decay constants

Field trip stops will be in the following quadrangles: Canaan, N.Y.-Mass., Pittsfield West, Cheshire, Williamstown, and North Adams, Mass. Stops 1, 2, and 3 are located in the published geologic map (Ratcliffe, 1978).

Road log for N.E.I.G.C. '79 field trip

Log starts at large parking lot on Rt. 22, 0.3 mi. north of Interstate 90 (Berkshire spur of N.Y. State Thruway).

Cumulative mileage

0 Proceed north on Rt. 22 over RR tracks.

1.2 Turn left on Tunnel Hill Road.

2.1 Park at bend in road before RR tunnel.

Stop 1. Contact between rocks of Chatham Slice of Taconic allochthon and autochton. Canaan 7-1/2 quadrangle.

Walk down the slopes to the east to the railroad tracks. Caution! This railroad is operating and trains may come from either direction. There is enough room in the center island or along the rock walls, should a train arrive.

The geologic relationships at this stop are shown in Figures 7 and 8. The exposures in the tunnel are on the east flank of a doubly plunging anticline that produces a semi-window exposing unit g of the Stockbridge Formation. The allochthonous rocks of the Chatham slice consist of purple and green slate, green silty slate, massive beds of Rensselaer Graywacke, and basaltic volcanic rocks believed to be flows and tuffs.

The outcrops in the railroad cut expose the contact between the Chatham slice and the autochthon. A folded thrust contact (T_1, Fig. 9) between overlying green phyllite of the Nassau Formation can be seen. Note the truncation of beds in the limestone by the contact. In addition, a fault sliver of green phyllite underlies an inverted sequence of Stockbridge and Walloomsac in a small anticline nearer the portal (thrust T_2, Fig. 9).

Fault T_1 traces out of the cut and forms the western limit of the allochthon. Fault T_2 is not exposed again in recognizable form.

The relationship here suggests that the autochthon and allochthon were tectonically mixed during emplacement. In this model, rocks of the autochthon were overturned during thrusting and overridden by younger thrusts as illustrated below in Model 1 of Figure 9.

An alternative model involves recumbent folding of the thrust contact as shown in Model 2.

Return to Rt. 22 via Tunnel Hill Rd.

3.0 Turn left on Rt. 22 north.

4.0 Outcrops of unit c of the Stockbridge.

4.5 Intersection of Rt. 295. Proceed North.

4.7 Outcrops of unit c of the Stockbridge.

6.1 Turn into Berkshire Farm for Boys. Stop and ask permission to drive through property. Road log resumes on leaving Farm area.

Drive through the property and bear left on the drive leading

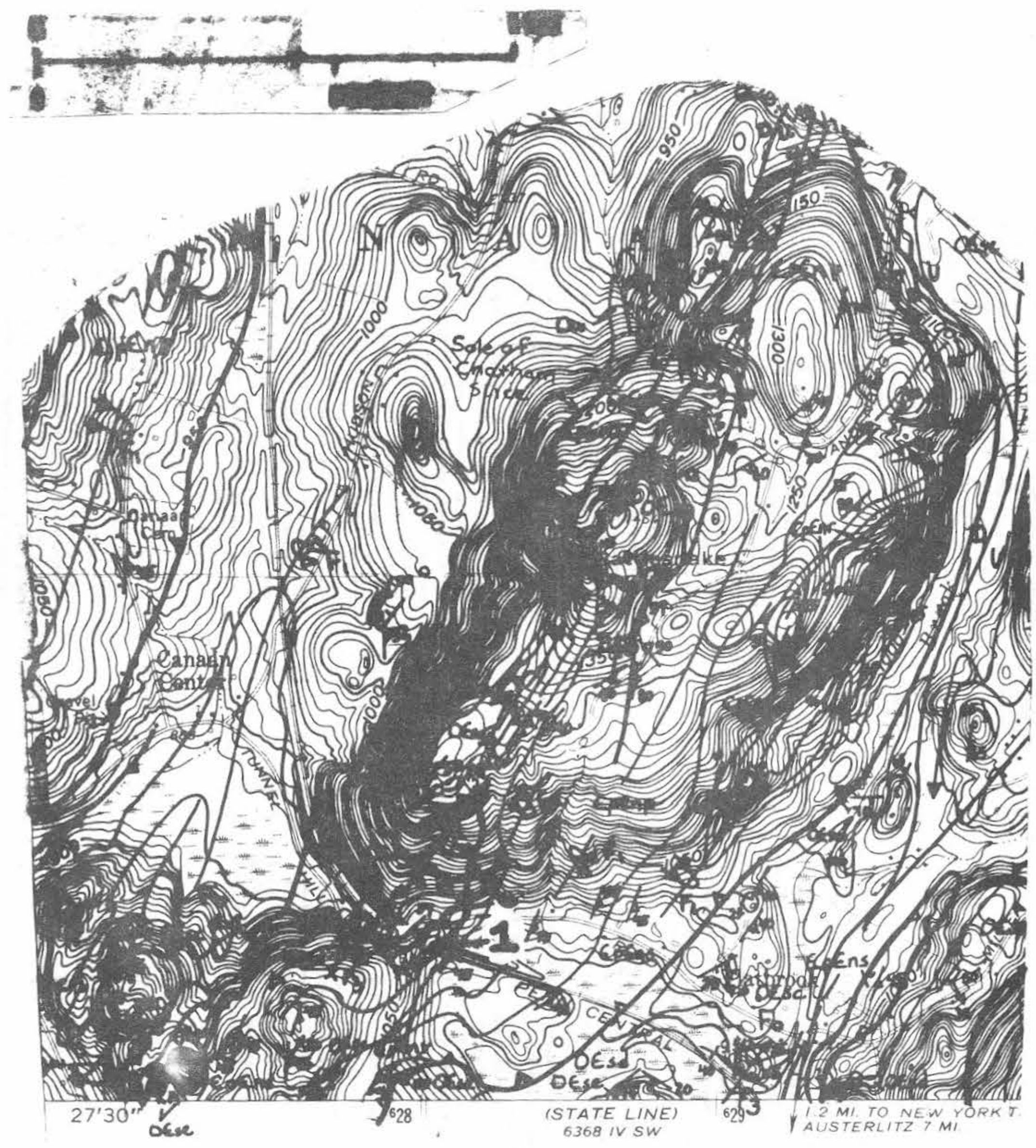


Figure 7. Geologic map of area around Stop 1. (Reproduced from Ratcliffe, 1978). See following pages for explanation

Figure 7. cont'd.

DESCRIPTION OF MAP UNITS
(Major minerals are listed in order of increasing abundance)

BEDROCK OF THE AUTOCHTHON

WALLOOMSAC FORMATION (UPPER? AND MIDDLE ORDOVICIAN)

Ow

Dark-gray to black, carbonaceous, sooty-gray-weathering, fissile phyllite or schist containing minor punky-weathering limy phyllite, schistose marble, and calcite marble. Biotite, plagioclase, and garnet developed in more highly metamorphosed rocks exposed to the east. Recognized as higher metamorphic grades by dark color, coarse black biotite metacrysts, and punky-weathering limy layers. Distinguished from Everett Formation by being less garnetiferous, much richer in biotite, and lacking chloritoid, ilmenite/magnetite, and green iron-rich chlorite

Owl

Dark-blue-gray crystalline, discontinuous basal limestone and limestone conglomerate containing pelmatazoan, bryozoan, algae, gastropod, and rugose coral remains; weathers to buff gray. Rugose corals from fossil locality xF1, 0.75 km southwest of Shaker Village, are no older than Black Riverian and may be Trentonian (Zen and Hartshorn, 1966). At this locality the unit (Owl) rests on the uppermost unit (O€sg) of Stockbridge Formation; thus the fossil date establishes the minimum age of the Stockbridge Formation. Pelmatazoan, bryozoan, and possible brachiopod fragments are found in the unit (Owl) west of Queechy Lake, fossil locality xF2 and at the west end of the Penn Central Railroad tunnel, fossil locality xF3

Owm

Impure feldspathic, schistose calcite marble studded with metacrysts of black biotite and black albite forms the basal limestone in the eastern part of the map area. The feldspar component probably was derived from erosion of the Precambrian gneisses and Dalton Formation during Middle Ordovician time. This unit thickens eastward, being 50-70 m thick in the Stockbridge quadrangles (Ratcliffe, 1974b), and passes gradationally through interbedding of schist in dark-gray to black calcitic biotite schist

STOCKBRIDGE FORMATION (LOWER ORDOVICIAN TO LOWER CAMBRIAN)

O€sg

Medium- to dark-gray calcite marble; massive, white, coarsely crystalline calcite marble; light-gray, fine-grained, phyllitic marble; and bluish-gray and white-mottled calcite marble that weathers to a smooth glistening surface. Subordinate beds of cream- to beige-weathering dolostone commonly are boudinaged

O€sf

Predominantly tan- to gray-weathering, massive, sandy-textured dolostone to calcitic dolostone; more calcite-rich layers are punky-weathered and reddish; weathered surfaces commonly "wood grained" resulting from weathering of fine, quartz-rich sandy laminae less than 1 mm thick; local cross-bedding abundant; beds of light-tan vitreous quartzite as thick as 0.75 m are locally found

O€se

Coarsely crystalline, white to light-gray, blue-gray and white-mottled, bluish-gray and white-layered, or massive white calcite marble in Massachusetts. Excellent exposures at the abandoned quarries north and south of Richmond Pond are exceptionally pure and coarsely crystalline. Gray to bluish-gray, pale-blue-gray-weathering, finely layered calcite marble interlayered with dark calcite dolostone, and light-gray and white crystalline calcite marble characterize the unit in New York State at lower metamorphic grade

O€sd

In Massachusetts, beige-weathering sandy dolostone, reddish-weathering calcitic sandstone, and gray sandy-textured calcitic marble. Minor white vitreous quartzite 1 to 3 cm thick and thin interbedded black phyllite are characteristic. In New York State, at lower metamorphic rank massive gray- to light-tan-weathered calcitic dolostone with sandy crossbedded laminae of positive relief and intense orangish-tan-weathered, massively bedded, dark-blue-gray dolostone with punky-weathering, crossbedded calcitic metaquartzite beds several centimeters thick

O€sc

Massive, light-gray-weathering, steel-gray, very fine grained calcitic dolostone and gray and light-gray, layered calcitic dolostone with scattered silvery-gray phyllitic partings. Massive white calcitic dolostone and dull-gray-weathering, fissile calcitic dolostone with milky-white quartz knots and vuggy cavities 1 to 2 cm thick (possible metachert nodules) common near top of the unit. West of the Massachusetts State Line nodules of black chert as much as 2 cm in diameter are common

O€sb

Beige- to light-cream-weathering, gray to dark-gray, non-calcitic dolostone with punky-weathering quartzites, white vitreous quartzites 1 to 2 cm thick, rare blue quartz-pebble conglomerates, black pyritiferous calcareous schist, and green and reddish phyllite beds. Silvery-gray partings of phlogopite, black phyllitic partings or discontinuous quartz chaining several millimeters thick are common except in the middle of the unit which contains about 50 meters of dark-blue-gray-weathering, dark-gray, fine-grained dolostone, and light-powder-blue-gray-weathering, gray nonsiliceous dolostone

BEDROCK OF THE ALLOCHTHON

Allochthonous rocks are tentatively assigned to four structural slices on the basis of stratigraphic uniqueness of some rocks and on the basis of stuctural position. Owing to the high degree of imbrication along relatively late thrust faults the original boundaries of the slices have been modified.

Rocks assigned to Chatham slice

NASSAU FORMATION (LOWER CAMBRIAN? AND (OR) UPPER PRECAMBRIAN?) [Lithic subdivisions may appear at different stratigraphic levels in different areas owing to widespread interfingering]

ꞒpꞒns Massive greenish-gray to gray metasiltstone or chlorite-sericite-rich phyllite locally containing 1- to 3-cm beds of greenish metaquartzite and olive-drab, fine-grained metasiltstone; pale-greenish, fine-grained chlorite-sericite phyllite with minor quartz metaconglomerate lenses and, more rarely, granitic gneiss-boulder metaconglomerate beds or dark-green graywacke beds as much as 10 cm thick

ꞒpꞒnr Rensselaer(?) Graywacke Member--Massive, bedded, dark- to pale-green metagraywacke or metasubgraywacke containing minor blue-quartz pebble-, coarse gneiss boulder-, and gneiss pebble-conglomerate layers. Unit interfingers with and grades into massive green-gray to gray metasiltstone unit (ꞒpꞒns) having many lenses of graywacke (ꞒpꞒnr) too small to be shown on map

ꞒpꞒnvt Metavolcanic rock, dark-brownish-green-weathering, ilmenite-leucoxene-amphibole-stilpnomelane-epidote-plagioclase metatuff(?) having fragmental relict plagioclase phenocrysts as much as 2 mm long; unit is fine grained and passes gradationally upwards into metasiltstone unit (ꞒpꞒns) or into metavolcanic rock (ꞒpꞒnv)

ꞒpꞒnv Metavolcanic rock, dark-green to yellowish-green, ilmenite-leucoxene-chlorite-actinolite-hornblende-epidote-plagioclase greenstone forming conformable, massively layered units as much as 10 m thick. Individual layers commonly show relict intersertal igneous texture grading toward a finer grained, strongly foliated rock with scattered akeritic amygdaloidal(?) fillings at lower contacts with the metasiltstone unit (ꞒpꞒns) and the Rensselaer(?) Graywacke Member (ꞒpꞒnr). Well exposed in the Knob, on a slope southwest of Queechy Lake, and on the slopes east of the Berkshire Farm for Boys. A contact zone at the base is marked by alternating layers of metasedimentary rock and thin layers of volcanic rock. Locally the volcanic rocks interlayer with plagioclase-rich Rensselaer(?) Graywacke Member (ꞒpꞒnr), while at other localities they appear to discordantly overlie the graywacke

ꞒpꞒnp Dark-maroon phyllite, green and purple laminated or mottled sericite-hematite-quartz phyllite including red or pale-green shale-chip metaconglomerate layers, and many light-green or purple-tinted metaquartzites 10 cm thick that commonly are crossbedded. Beds of soft yellowish-green to gray paper-thin phyllite are interlayered with the predominantly purple and green mottled rocks

ꞒpꞒnq Massive-bedded, light-green to gray-weathering metaquartzite or metasubgraywacke as much as 20 m thick forms lenticular bodies near the top of the phyllite unit (ꞒpꞒnp) in the western part of the Chatham slice

Rocks assigned to the Perry Peak slice

ROCKS NEAR PERRY PEAK (LOWER CAMBRIAN? AND (OR) UPPER PRECAMBRIAN?)

ꞒpꞒs Light-greenish-gray to gray metasiltstone, chlorite-rich siliceous phyllite or dark-green gritty metagraywacke with 1- to 3-cm beds of vitreous metaquartzite. Overall this unit is quartz rich and well bedded, and it closely resembles the metasiltstone unit of the Nassau Formation (ꞒpꞒns) of the Chatham slice, although volcanic rocks, abundant in rocks of the Chatham slice, are absent

ꞒpꞒp Light-yellowish-green and deep-purple variegated phyllite, dark-purplish-gray phyllite, and soft yellow-green lustrous quartz-chlorite-paragonite-muscovite phyllite. Minor interbeds of purplish-gray to dark-green metagraywacke 1 to 2 cm thick are widely distributed. This unit resembles closely the phyllite unit of the Nassau Formation (ꞒpꞒnp) of the Chatham slice, although the purplish coloration is less intense in areas of higher metamorphic grade to the east where dark-gray phyllite with a subtle but distinctive purple cast is found

Rocks assigned to the Widow Whites slice

(The Widow Whites slice is named for occurrences on Widow Whites Peak in the southern part of the Hancock quadrangle.)

ROCKS NEAR WIDOW WHITES PEAK (LOWER CAMBRIAN? AND (OR) UPPER PRECAMBRIAN?)

€p€g

Dark-gray to black lustrous chloritoid-quartz-muscovite phyllite interlayered with gray-green quartzose albitic phyllite. Distinctive dark-gray beds rich in ilmenite and chloritoid have a sparkling luster

Rocks assigned to the Everett slice

EVERETT FORMATION (LOWER CAMBRIAN? AND (OR) UPPER PRECAMBRIAN?)

€p€ev

Green to greenish-gray and lustrous, silvery-gray sericite-(muscovite)-quartz-chlorite phyllite or schist in which chloritoid, paragonite, ilmenite, and garnet may be present; gray-green and dark-gray laminated phyllite; gritty-textured, white sodic plagioclase-spotted, greenish phyllite that weathers to a distinctive pitted surface and occurs in single beds as much as 7 m thick; and greenish-gray phyllite with quartzite layers 0.5 to 1 cm thick grading into a quartzose, gray-green phyllite with abundant pea-sized magnetite metacrysts. The bulk of the rocks assigned to the Everett Formation on Lenox Mountain are dark-gray to gray-green with abundant dark flecks of chloritoid, deep-red garnet, and black biotite. With allowances for higher metamorphic grade and the consequent elimination of the abundant chlorite present in the lower grade rocks, the Everett may be compositionally equivalent to the metasiltstone unit of the Nassau Formation (€p€ns) or to the metasiltstone unit near Perry Peak (€p€s). It resembles most closely the dark albitic phyllite of Widow Whites slice (€p€g).

BEDROCK OF THE MOVEMENT ZONE

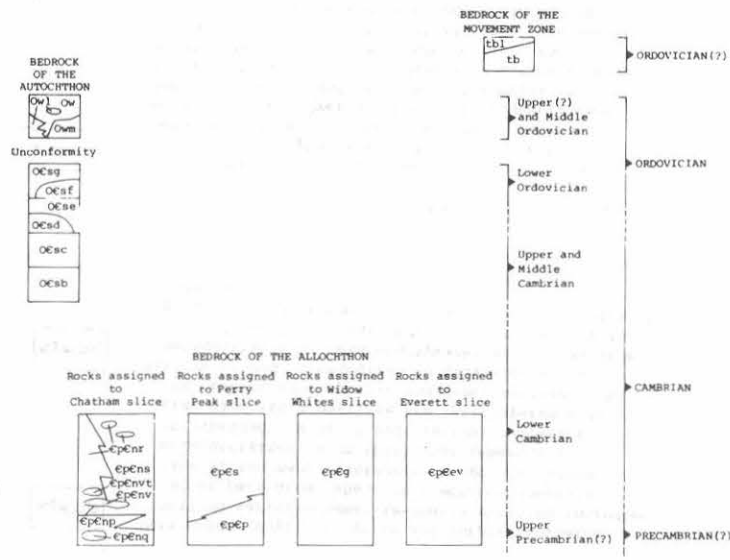
tbl

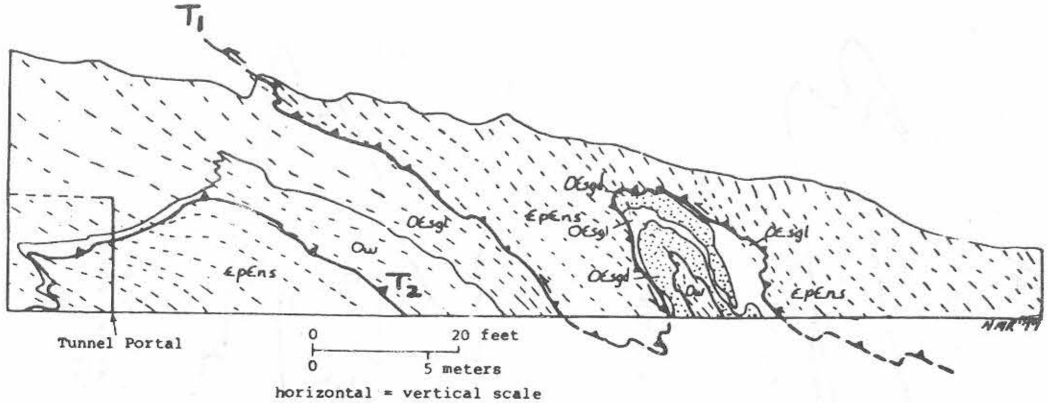
TECTONIC BRECCIA (ORDOVICIAN?)--A zone rich in inclusions of carbonate rock of the Stockbridge Formation in a polymict tectonic breccia interleaved tectonically during Ordovician(?) time with slices of black Walloomsac and greenish-gray metasiltstone unit near Perry Peak (€p€s) at or near the sole of overriding plates of allochthonous rock. Because of the fine scale of imbrication of rocks in the movement zone, separate rock types in the breccia are not distinguished. Where individual fault slices are of sufficient size and are composed of coherent strata, they are mapped as standard fault slices and the units identified. These breccias mark tectonic movement zones that differ from conventional fault zones in one important aspect. The carbonate clasts in the imbricated phyllite matrix are exotic blocks not derived from the hanging wall or the footwall but from the autochthonous carbonate rock of the Stockbridge Formation; carbonate clasts are considered tectonic inclusions transported within the movement zone from some site to the east

tb

Completely intermixed zone of black Walloomsac and greenish phyllite with irregular inclusions of either rock type in a matrix of the other. Zones exhibit variation of rock types on a scale of centimeters to tens of meters. The distribution and shape of inclusions indicates that incorporation predated formation of the regional slaty cleavage (S_1). Breccias of this type found in zones below the sole of the Perry Peak slice may represent highly disarticulated and "kneaded out" remnants of earlier fault slices dismembered during overthrusting of the Perry Peak slice. The presence of these breccias within the body of the Walloomsac Formation suggests that fault displacements of considerable magnitude may exist within the "autochthonous" belt of Walloomsac underlying the Perry Peak and Widow Whites slices.

CORRELATION OF MAP UNITS





EXPLANATION

Ow	Phyllite and schist of the Walloomsac Formation	Ordovician
OЄsgl	Limestone beds in the medium- to dark-gray calcite marble (OЄsg) of the Stockbridge Formation	Ordovician and Cambrian
OЄsgd	Dolostone beds in the medium- to dark-gray calcite marble (OЄsg) of the Stockbridge Formation	
ЄpЄns	Green phyllite of the Naussau Formation. Part of the Chatham slice	Cambrian and (or) Precambrian

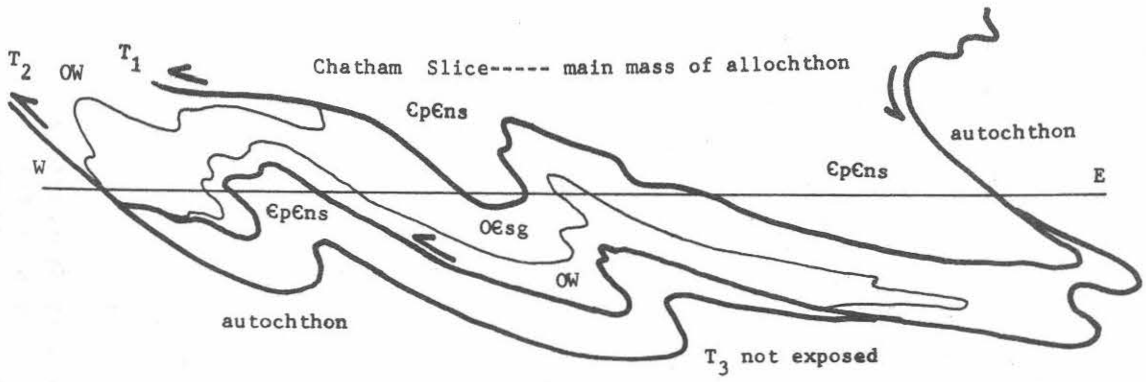
——— Contact

———▲——— Thrust contact of green phyllite of the Nassau Formation (ЄpЄns) on autochthon. Teeth on overthrust plate; dashed where inferred; contact is discordant to beds in autochthon

- - - - - S_1 foliation crosscuts thrust faults as axial plane of F_1 folds

Figure 8. Contact relationships of the Chatham slice with Stockbridge and Walloomsac Formations exposed in north face of Penn Central railroad cut, east portal of tunnel at the southern edge of the Canaan quadrangle, showing interleaved units in complex tectonic breccia (tb).

MODEL 1: TWO FAULT SLIVERS OF TACONIC ALLOCHTHON WITH INTERNAL RECUMBENTLY FOLDED SLIVER OF STOCKBRIDGE AND WALLOOMSAC. THRUSTS LATER FOLDED BY UPRIGHT ACADIAN FOLDS.



MODEL 2: RECUMBENT FOLDING OF A SINGLE FAULT FOLLOWING OR DURING EMPLACEMENT, FOLLOWED BY ACADIAN FOLDING.

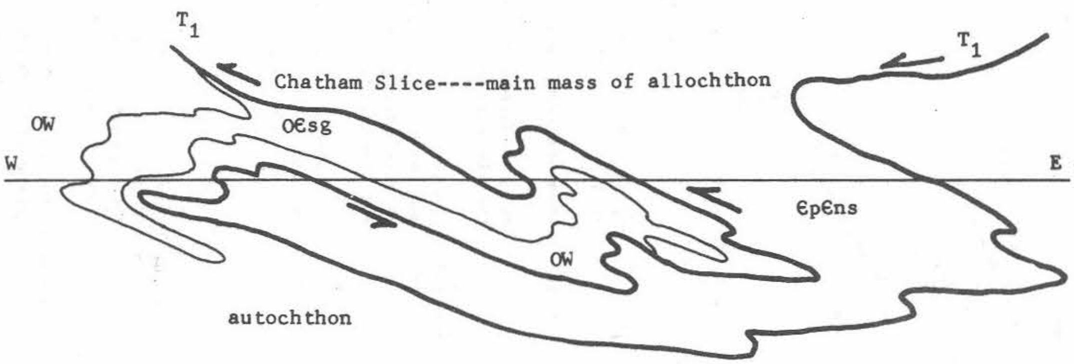


Figure 9. Sketch of alternate models for explanation of fault relationships seen at Stop 1. On map, Fig. 7, area is shown as tectonic breccia (tb), and model 1 is the preferred interpretation. T_1, T_2, and T_3 refer to faults identified in Fig. 8.

to the home of the director. Park at edge of large field before house. Walk east-northeast across field to slopes.

Figure 10 shows the location of Stop 2. Basaltic volcanic rocks of the Chatham slice are well exposed on the slopes above the Boys Farm. This zone associated with graywacke has been traced continuously from the State Line quadrangle north to the Jiminy Peak area for a distance of 19 km. Similar occurrences are on the Knob across the valley to the west. This eastern belt of volcanic rocks is better laminated than those on the knob and at Fog Hill in the State Line quadrangle. The associated graywackes contain fragments of coarse-grained granite containing oligoclase and perthitic K feldspar.

The association of coarse, continentally derived clastic rocks with basaltic volcanics and probable fluviatile red beds suggests rift facies rocks similar to Triassic and Jurassic rocks of the Atlantic passive continental margin. These rocks are restricted Chatham and Rensselaer Plateau slices.

Return to Rt. 22 and turn right.

9.5 Turn right on Rt. 20 at New Lebanon

10.2 Rt. 20 branches right. Follow it toward Pittsfield.

12.9 Stop 3, just past curve in road. Large roadcut of purple and green phyllite.

Excellent roadcuts expose purple and green laminated phyllites of the next higher Taconic slice, the Perry Peak slice that is correlated with the Everett slice on Figure 1. Strong secondary slip cleavage and microfaults, accentuated by pods of bull quartz are common. The contact between the Chatham slice and the higher Perry Peak slice is a late thrust probably of Acadian age. From this contact zone eastward to the carbonate belt, a complex system of late thrust faults have been mapped in which slivers of carbonate, Walloomsac, and Taconic rocks are found. To the north, this thrust zone roots in the Precambrian (Proterozoic) rocks of Clarksburg Mountain.

Volcanic rocks and thick graywacke are not recognized in the higher Taconic slices. Is the absence of these rocks of stratigraphic significance, or are the volcanic rocks absent because of faulting? This is a serious problem that has not been resolved. Certainly the Greylock and Hoosac Mountain rocks are different stratigraphically than either the Chatham or Perry Peak rocks, as will be discussed at Stop 4.

From the roadcut, walk southwestward down to the old road and then follow the new wood road up the hill westward past green phyllite of the Perry Peak slice.

At 1550 ft elevation on bench a sliver of dolostone may be seen. The fault sliver occupies a position between the Perry Peak slice and Walloomsac of the autochthon below in a late fault.

This anticlinal belt of Walloomsac, however, underlies the Chatham slice rocks seen to the south at Stop 2.

From the carbonate rocks, walk northeastward down the slopes toward the brook, where inclusions of greenish-gray Taconic rocks may be found imbedded in the Walloomsac matrix.

This chaotic zone of wildflysch-like material evidently underlies the Chatham slice. Similar rocks are exposed to the south

Figure 10. Geologic map of the Chatham and Perry Peak Slices of the Taconic allochthon on the east side of the Canaan Valley showing the location of Stops 3 and 4. For explanation, see Fig. 7

beneath the Chatham slice (Ratcliffe, Bird, and Bahrami, 1975, p. 82).

When the observations at Stop 1 and 3 are compared, the relations suggest that the Chatham slice was emplaced by a mechanism that involved recumbent folding, slivering of the autochthon, plucking of carbonate blocks signifying deformation of consolidated rocks. On the other hand, more ductile behavior with turbulent mixing of materials from autochthon and allochthon is necessary to account for the wildflysch-like breccias seen. A gravity induced spreading model to allow for creep in the autochthon may be the best explanation of these relationships. However, there is little evidence in favor of soft rock gravity sliding. Continue east on Rt. 20 toward Pittsfield.

14.5 Side road leads to fossil locality in the basal part of the Walloomsac.

15.0 Hancock Shaker Village and intersection with Rt. 41. Continue on Rt. 20.

16.5 Y branch in road. Follow Rt. 20 left.

20.0 Intersect Rt. 7, Pittsfield. Turn left. Follow around the circle in Pittsfield 90° and follow Rt. 7 and 9 north. Left turn at light, and follow Rt. 7 north out of Pittsfield.

25.7 Town of Lanesboro.

27.1 Turn right onto entrance road to Mount Greylock (N. Main St.).

28.0 Turn right. Follow signs to Greylock Reservation.

28.4 Take left Y, Rockwell Rd.

28.9 Entrance to Reservation.

31.9 Rounds Rock. Excellent cliffs west of road of green albitic Greylock Schist (optional stop).

32.9 Jones Nose. Stop 4.

Walk up the Meadow along the Appalachian Trail across limey albitic schist and schistose marble of the Walloomsac Formation here exposed by breaching of a refolded antiform that has a north-dipping axial surface. This carbonate-rich unit is found at or near the base of the Walloomsac Formation regionally and on Mount Greylock, thus suggesting that older rocks are coming to the surface here. To the north, farther up the slope, a complementary, nearly recumbent syncline and anticline pair are exposed, also with north-dipping refolded axial surfaces. Section C C′ of Mount Greylock, Figure 3.

33.9 Ashfort Rd.

34.7 Entrance to campground. Turn left to Stony Ledge.

The bench the road follows is located on the contact between Greylock Schist and Walloomsac of the autochthon.
Follow dirt road to Stony Ledge.

36.4 Stony Ledge. Stop 5.

Magnificant view to north and east into the Hopper. The ridge to the north is underlain by Greylock Schist, the valley by Walloomsac. To the east, the peak of Greylock and the steep west slope can be seen. Recent mapping suggests that there is double section of Greylock Schist exposed on Greylock with the axial surface of a large recumbent fold separating the mountain into lower-limb and upper limb structures.

The Bellowspipe Limestone shown by Prindle and Knopf (1932) does not exist as a throughgoing unit but is sporadically developed.

Excellent late crenulation cleavage can be seen in the ledges of green phyllite. These folds are postmetamorphic and correlate with the Green Mountain uplift folds or folds of F_5 regional structures.

39.8 Return to Rockwell Rd. Turn left and climb road past dark albitic and green albitic schist of the lower limb into fine lustrous phyllites forming the core of the recumbent synform.

41.3 Large cuts of albitic Greylock and associated quartzite repeat sequence into upper overturned limb of recumbent synform. Turn right to top.

42.0 Park in lot at top. Stop 6.
Excellent view all around, weather permitting. Discussion of the regional geologic relationships.

At crest of Greylock, we are standing on chloritoid-rich phyllite in the axial part of a large westward topping syncline. Darker albitic phyllite, quartzite and graywacke form a rather continuous marker horizon that separates the more chloritoid-rich rocks from green and gray-white albite spotted granulites. Locally, this transition zone contains lenses of salmon pink dolostone in boudins, magnetite-rich schist, and gneissic pebble conglomerate. Albitic schists crowded with albite in massive exposures can be seen on the crest of Ragged Mountain to the north, on Cole Mountain to the south, Rounds Rock, Jones Nose, and on Mount Prospect to the west.

The structural interpretation of Mount Greylock shown in Figure 3 calls for a thrust sheet that has been recumbently folded. Plunges are to the north and south into the sections. In addition to recumbent folds that involve the thrust contact, internal structures are shown which are also recumbent but which are truncated by the thrust contacts. This suggests that rocks of the Greylock slice were folded prior to emplacement on the Stockbridge-Walloomsac sequence.

Relationship of the Greylock to the Hoosac Schist

The ridge seen to the east is the Hoosac Mountain underlain by Precambrian (Proterozoic) gneiss to the south and Hoosac Schist to the north. The carbonate valley narrows to the north in the North Adams gap where the Hoosac Mountain approaches the Green Mountains. A longitudinal sketch section is shown in Figure 11. It identifies two major thrusts on Hoosac Mountain: a lower thrust places Precambrian (Proterozoic) gneiss with its unconformable cover (Hoosac Schist and interfingering Dalton Formation) over carbonate rocks of the autochthon. This fault is essentially the fault mapped as "Hoosic thrust" by Herz (1961). A higher thrust places Hoosac with a more easterly facies above the western Hoosac belt. This fault is termed the Hoosac summit thrust. The eastern Hoosac sequence contains green albitic schist, dark albitic and garnet-bearing schist and light green chloritoid schist. This sequence resembles closely Greylock Schist units, although a greater development of albite-rich rocks is found on

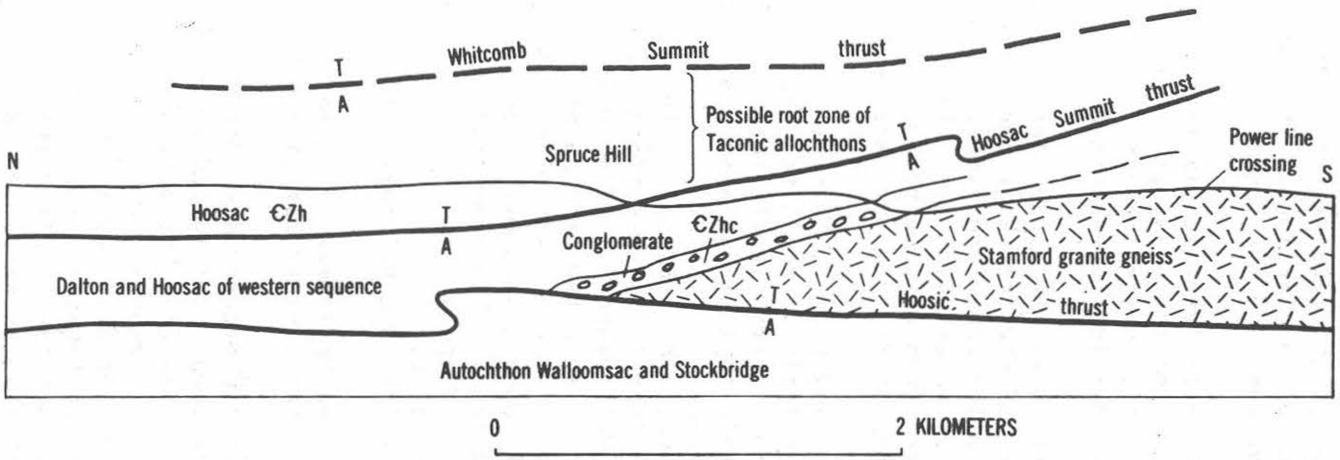


Figure 11. Sketch of Hoosac Mountain looking east from Mount Greylock, showing position of Hoosic thrust with wedge of Berkshire Massif with unconformable Hoosac cover and higher thrusts. T=Toward A=Away, show movement on fault

Hoosac Mountain than on Mount Greylock. Stratigraphic units defining recumbent metamorphic fold structures within the eastern Hoosac are truncated by the Hoosac summit thrust.

East of the Hoosac belt is the Rowe Schist, a pale-green, fine-grained chloritoid phyllite with interbedded dark carbonaceous phyllite and amphibolite all of presumed Cambrian through Ordovician age. Stanley has recently suggested (1978) that a major thrust fault (Whitcomb summit thrust) separates the Rowe from the Hoosac on the basis of recognized truncation of units in the footwall and hanging wall (see Fig. 5). The Rowe also contains abundant pods of ultramafic rock. This unit may mark the locus of crustal convergence in the Cambrian and Ordovician (Stanley, 1978).

On the basis on the stratigraphic similarities between the eastern Hoosac sequence and the Greylock section, and the lack of similarity of Greylock with western Hoosac, the root zone of the Greylock slice lies either in the eastern Hoosac belt or between belt and the Rowe.

The root zone of the Greylock slice lies east of exposed Precambrian (Proterozoic) of the Berkshire massif at this latitude. Because the Greylock slice shows the closest affinity to cover rocks of the massif, it seems clear that the lower slices of the allochthon were derived from more easterly sites.

Return via Rockwell Rd. to Rt. 7 along same route taken up mountain.

52.4 Turn right, north on Rt. 7 from N. Main St.

53.4 Small crops of green chloritoid phyllite, with strong late fault fabric. Crops continue intermittently for 0.7 mile. Dark Walloomsac biotite grade phyllite is seen near north end of crops.

54.4 Large crop of Walloomsac.

54.8 Crop of Walloomsac (Owl) and isoclinal F_1 folds.

55.3 Road to Jiminy Peak. Turn left for optional stop. Stop at large crops of black Walloomsac 1.1 miles west. Cataclastically deformed phyllite can be seen faulted against green phyllite of the Nassau Formation. Characteristic plication of foliation, slickensides, and bull quartz pods mark a series of late faults in the Buxton Hill fault zone. This zone projects east of the late fault zone seen at Stop 3.

Return to Rt. 7. Turn left.

56.1 Crops of Stockbridge (O€sg) and Walloomsac (Ow), Brodie Mountain ski area.

56.4 Large outcrops of sheared Walloomsac in a late thrust zone. Slickensides, sulfide quartz mineralization mark faults in Buxton Hill zone.

56.9 Stop 7. Large roadcuts in Stockbridge (O€sg). Excellent recumbent folds can be seen, crossfolded by folds in N. 20° W. 22° NE dipping axial planes. The origin of the cross folds is not known but may be an expression of compression of a faulted sliver within the Buxton Hill fault zone.

57.1 Large crop of Stockbridge (O€sg) on east. Thin zone 1 m thick of Walloomsac in a slickensided fault sliver is traceable for 75 m in this outcrop. This is an excellent example of character of the fault slivering within this late fault zone.

58.0 Large crops of fault slivered and cataclastic Walloomsac opposite the Mill on the Floss.

The valley to the north at South Williamstown to the east are highly fault intercalated zones of green albitic and nonalbitic phyllite of the Greylock slice. To the west and north are green phyllites of Brody Mountain and Deer Hill. Exact placement of the boundary of the Greylock slice has not been possible. Allochthonous rocks on the northern end of Brody Mountain, however, are separated from the autochthon by a complexly intermixed zone of Walloomsac and green phyllite similar to the zone of mixed rocks seen below the Chatham slice at Stop 3, according to work by Don Potter. Because the Greylock slice does not have similar emplacement breccias and because the Greylock stratigraphy contains a greater abundance of albitic rocks than the Brody Mountain slice, the two slices are believed to be discretely different allochthons.

62.1 Intersection Rt. 43. Continue north on Rt. 7.
At crest of hill, excellent view of Mount Greylock and the Hopper, and the Green Mountains to the north. The broad carbonate valley extending from Williamstown east to North Adams consists of complexly folded rocks that appear to be detached from the Green Mountains by a major thrust fault that roots in the Hoosac Gap area. This fault slice, the Stone Hill slice, is named for exposure of the faulted rocks on Stone Hill south of Williamstown.

63.4 Turn right on Scott Hill Rd. and shortly turn left on Stone Hill Rd.

63.7 At dirt road, continue to north if permission to drive in is available. If not, we will walk in.
The Stone Hill section can be traced north to Buxton Hill where slivers of Stamford Granite Gneiss, Dalton and Cheshire thrust across units b and c of the Stockbridge in a complex thrust zone (see Stop 9).

The Stone Hill slice is believed to be a thin flap of autochthon originally rooted in the Adams or North Adams area that was thrust westward out of the North Adams gap area during an early deformation stage.

East-west trending hinge lines for early folds dominate the structure within the slice eastward to Mount Greylock where a late fault, the Clarksburg thrust, truncates structures in the slice.

64.5 <u>Stop 8</u>. Stone Hill slice.

One of the most complete stratigraphic sections of the Dalton Formation through unit C of the Stockbridge Formation is exposed in the area of Stone Hill. Thin Cheshire Quartzite 10-0 m thick is interlayered with underlying black quartzose phyllite of the Dalton Formation. Dolostone of unit a of the Stockbridge overlies the Cheshire. Lithostratigraphic units of the Stockbridge here match closely rock section of the Vermont Valley sequence to the north.

The most anomalous features of the Stone Hill section are the very thin Cheshire and the dissimilarity of this section to the Dalton-Cheshire section exposed on the Green Mountains to the north. Dark quartzose schists with a very thin quartzite is also a characteristic of the more easterly belt of Dalton and Cheshire mapped along the base of Hoosac Mountain and in the North Adams gap area.

Excellent recumbent folds can be seen in the cliffs above the road. The contact between the Cheshire and dolostone of unit a of the Stockbridge can be seen at the base of the quartzite cliffs to the north.

65.0 Return to Rt. 7. Turn right.

66.9 Just after Bee Hill Rd. turn left on Thornliebank Rd. Crops to right are Chesire Quartzite.

67.0 Turn right on Hawthorn St. Park at east edge of field.

Stop 9. Buxton Hill - Outlier of Stamford Granite Gneiss and blastomylonite and the sole of the Stone Hill slice.

The ridges to the south consist of blastomylonitic gneiss and Dalton with excellent recumbent fold thrust style. Syntectonic biotite from the blastomylonite yielded a K-Ar age of 387 $\pm$ 14 m.y. (394 $\pm$ 14 m.y.).* The biotite age is interpreted as a cooling age and a minimum for emplacement.

Crops around the north and east end of the hill are dolostone of the autochthon. Continue straight down Hawthorn St. to intersection. Turn left on Buxton Hill Rd. Follow Buxton Hill to next intersection.

67.6 Turn right at end of Buxton Hill Rd.

67.9 Intersect Rt. 7 and 2. Bear right around island and follow Rt. 2 to east. Williams College.

69.7 Luce Rd. east of Williamstown. Follow Rt. 2 east to North Adams and beyond on Rt. 2 and 8.

79.0 Just past mills, east of North Adams, turn left on Rt. 8 north.

79.6 Just after Red Mill and entrance to natural bridge, stop.

Stop 10. Hoosic thrust.

Herz (1961) drew the Hoosic thrust at the base of the exposures which he assigned to the Hoosac Formation. Reexamination indicates that these dark albitic, graphitic schists quartzites and limy schist should be assigned to the Walloomsac. Faulting is indicated by the shallow dipping spaced crenulation cleavage.

These exposures are on the north-plunging lower limb of a large recumbent syncline cored by Walloomsac that is refolded by north-trending late folds.

If time allows, we will drive into the quarry at natural bridge where an excellent recumbent anticline in unit e of the Stockbridge can be seen. This structure is on the lower limb of the major synclinal structure.

Those wishing to examine the excellent solution features in Mr. Elder's natural bridge, may do so for a slight fee.

Those wishing to disband at this point, feel free to do so. Hard core elements may continue on to Hoosac Mountain!

Turn right at Rt. 8, where road log resumes.

79.9 Turn left on Rt. 2.

At the base of Hoosac Mountain are small crops of Dalton Formation interfingering with dark albitic schists of the western

* age calculated using new decay constant.

autochthonous Hoosac.

82.8 At the hairpin turn, green muscovitic chloritoid and garnet-bearing schist of the eastern Hoosac may be seen. Excellent recumbent folds outlined by thin vitreous quartzite. This section is separated from the autochthonous Hoosac by the Hoosac summit thrust.

83.3 Top of Hoosac Mountain by observation tower. Stop 11.
Excellent view back at Greylock, Taconic Range beyond, Green Mountains, and the North Adams Gap. Summary discussion of the relationships of the Hoosac-Greylock-Taconic belts and paleogeographic reconstructions.

- End of trip -

References

Balk, R.M., 1953, Structure of graywacke areas and Taconic Range, east of Troy, New York: Geol. Soc. America Bull, v. 64, p. 811-864.

Berry, W.B.M., 1962, Stratigraphy, zonation, and age of Schaghticoke, Deepkill, and Normanskill shales, eastern New York: Geol. Soc. America Bull., v. 73, p. 695-718.

Bird, J.M., 1961, Age and origin of the Rensselaer greywackes, Nassau quadrangle, south-central Rensselaer County, New York, p. 135-136 in the Geol. Soc. America Abs. for 1961: Geol. Soc. America Sp. Paper 68, 322 p.

_____, 1962a, Geology of the Nassau quadrangle, Rensselaer County, New York: Rensselaer Polytech. Inst., Troy, New York, Ph.D. thesis, 204 p.

_____, 1969, Middle Ordovician gravity sliding in the Taconic region, in North Atlantic - geology and continental drift: Am. Assoc. Petroleum Geologists Mem. 12, p. 670-686.

_____, 1975, Late Precambrian graben facies of the Northern Appalachians [abs.]: Geol. Soc. America Abs. with Programs, v. 7, no. 1, p. 27.

Bird, J.M., and Dewey, J.F., 1970, Lithosphere plate - continental margin tectonics and the evolution of the Appalachian orogen: Geol. Soc. America Bull., v. 81, p. 1031-1060.

Craddock J.C., 1957, Stratigraphy and structure of the Kinderhook quadrangle, New York, and the "Taconic Klippe": Geol. Soc. America Bull., v. 68, p. 675-724.

Fisher, D.W., 1962, Correlation of the Cambrian rocks in New York State: N.Y. State Mus. and Sci. Service Map and Chart Ser.: no. 2.

Fisher, D.W., Isachsen, Y.W., and Rickard, L.V., 1970, Geologic map of New York State: New York State Mus. and Sci. Service Map and Chart Ser. no. 15, scale 1:250,000.

Harwood, D.S., 1975, Fold-thrust tectonism in the southern Berkshire massif, Connecticut and Massachusetts, in New England Intercollegiate Geological Conference, 67th Annual meeting, Great Barrington, Mass., Oct. 10-12, 1975, Guidebook for fieldtrips in western Massachusetts, northern Connecticut and adjacent areas of New York: New York City College of C.U.N.Y., Dept. Earth and Planetary Sci., p. 122-143.

Harwood, D.S., in press, Bedrock geologic map of the Norfolk quadrangle, Connecticut and Massachusetts: U.S. Geol. Survey Geol. Quad. Map.

Herz, N.C., 1958, Bedrock geologic map of the Cheshire quadrangle, Mass.: U.S. Geol. Survey Geol. Quad. Map GQ-108.

_____, 1961, Bedrock geology of the North Adams quadrangle, Massachusetts and Vermont: U.S. Geol. Survey Geol. Quad. Map GQ-139.

Norton, S.F., 1969, Unconformities at the northern end of the Berkshire Highland, in New England Intercollegiate Geological Conference, 61st Annual Meeting, Albany, N.Y., 1969, Guidebook for field trips in New York, Massachusetts, and Vermont, p. 21-1 to 21-20.

Potter, D.B., 1972, Stratigraphy and structure of the Hoosick Falls Area, New York - Vermont, East-Central Taconics: N.Y. State Mus. and Sci. Service Map and Chart Ser. no. 19, 71 p.

Prindle, L.M., and Knopf, E.B., 1932, Geology of the Taconic quadrangle: Am. Jour. Sci., 5th ser., v. 24, p. 257-302.

Pumpelly, Raphael, Wolff, J.E., and Dale, T.N., 1894, Geology of the Green Mountains in Massachusetts: U.S. Geol. Survey Mem. 23, 192 p.

Ratcliffe, N.M., 1969, Stratigraphy and deformational history of rocks of the Taconic Range near Great Barrington, Massachusetts, in New England Intercollegiate Geological Conference, 61st Annual Meeting, Albany, N.Y., 1969, Guidebook for field trips in New York, Massachusetts and Vermont, p. 2-1 to 2-23.

_____, 1974a, Bedrock geologic map of the State Line quadrangle, New York-Massachusetts: U.S. Geol. Survey Geol. Quad. Map GQ-1143.

_____, 1974b, Bedrock geologic map of the Stockbridge quadrangle, Massachusetts, U.S. Geol. Survey Geol. Quad. Map GQ-1142.

_____, 1978, Reconnaissance bedrock geologic map of the Pittsfield West quadrangle and part of the Canaan quadrangle, Berkshire County, Massachusetts, and Columbia County, New York: U.S. Geological Survey Miscellaneous Field Studies Map MF-980.

Ratcliffe, N.M., Bird, J.M, and Bahrami, Beshid, 1975, Structural and stratigraphic chronology of the Taconide and Acadian polydeformational belt of the central Taconics of New York State and Massachusetts: in New England Intercollegiate Geological Conference, 67th Annual Meeting, Great Barrington, Mass., Oct. 10-12, 1975, Guidebook for field trips in western Massachusetts, northern Connecticut and adjacent areas of New York: New York City College of C.U.N.Y., Dept. Earth and Planetary Sci., p. 55-86.

Ratcliffe, N.M., and Bahrami, Beshid, 1976, The Chatham fault: a reinterpretation of the contact relationships between the Giddings Brook and Chatham slices of the Taconic allochthon in New York State; Geology, v. 4, no. 1, p. 56-60.

Ratcliffe, N.M., and Burger, H.R., 1975, Preliminary bedrock geologic map of the Ashley Falls quadrangle, Massachusetts and Connecticut: U.S. Geol. Survey open-file report 75-148.

Ratcliffe, N.M., and Harwood, D.S., 1975, Blastomylonites associated with recumbent folds and overthrusts at the western edge of the Berkshire massif, Connecticut and Massachusetts - a preliminary report, in Tectonic studies of the Berkshire massif, western Massachusetts, Connecticut and Vermont: U.S. Geol. Survey Prof. Paper 888-A, p. 1-19.

Stanley, R.S., 1978, Bedrock geology between the Triassic and Jurassic basin and the East Flank of the Berkshire massif, Massachusetts: [abs.] in Geol. Soc. America Abs. with Programs, v. 10, no. 2, p. 87.

Williams, Harold, 1975, Structural succession, nomenclature, and interpretation of transported rocks in western Newfoundland: Canadian Jour. Earth Sci., no. 12, p. 1874-1894.

Walcott, C.D., 1888, The Taconic system of Emmons; Am. Jour. Sci., 3rd ser., v. 35, p. 229-242, 307-327,394-401.

Weaver, J.D., 1957, Stratigraphy and structure of the Copake quadrangle, New York: Geol. Soc. America Bull., v.68, p. 725-762.

Zen, E-an, 1961, Stratigraphy and structure at the north end of the Taconic range in west-central Vermont: Geol. Soc. America Bull., v. 72, p. 293-338.

_____, 1964a, Taconic stratigraphic names: definitions and synonymies: U.S. Geol. Survey Bull. 1174, 95 p.

_____, 1964b, Stratigraphy and structure of a portion of the Castleton quadrangle, Vermont: Vermont Geol. Survey Bull., no. 25, 70 p.

_____, 1967, Time and space relationships of the Taconic allochthon and autochthon: Geol. Soc. America Spec. Paper 97, 107 p.

_____, 1969, Stratigraphy, structure, and metamorphism of the Taconic allochthon and surrounding autochthon in Bashbish Falls and Egremont quadrangles and adjacent areas, in New England Intercollegiate Geological Conference, 61st Annual Meeting, Albany, N.Y., 1969, Guidebook for field trips in New York, Massachusetts, and Vermont, p. 3-1 to 3-41.

_____, 1972, The Taconide Zone and the Taconic orogeny in the western part of the northern Appalachian orogen: Geol. Soc. America Spec. Paper 135, 72 p.

Zen, E-an, and Hartshorn, J.H., 1966, Geologic map of the Bashbish Falls quadrangle, Massachusetts, Connecticut and New York: U.S. Geol. Survey Geol. Quad. Map GQ-507.

Zen, E-an, and Ratcliffe, N.M., 1966, A possible breccia in southwestern Massachusetts and adjoining areas, and its bearing on the existence of the Taconic allochthon: U.S. Geol. Survey Prof. Paper 550-D, p. 36-46.

_____, 1971, Bedrock geologic map of the Egremont Quadrangle and adjoining areas, Berkshire County, Massachusetts, and Columbia County, New York: U.S. Geol. Survey Misc. Geol. Inv. Map I-628.