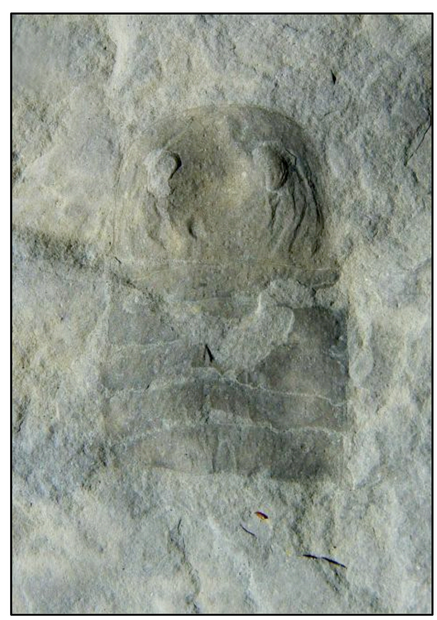
New York State Geological Association
83rd Annual Meeting

Field Trip Guidebook

October 14-16, 2011

Co-Hosted by
Central New York Association
Of Professional Geologists
&
SYRACUSE UNIVERSITY
Department of Earth Sciences

COVER IMAGE CREDITS

Left Photo : Ver Straeten et. al. - Photos of the Marcellus *thick, synorogenic clastic facies association*. d) interbedded sandy mudstones and thin sandstones (Mount Marion Fm., Cobleskill, NY).

Right Photo : Samuel Ciura, Jr - The eurypterid found in the Canoga Waterlime is a type of *Eurypterus* sp. and detailed comparisons with eurypterids in other horizons have not been made.

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Trip A-1

EFFECTS OF ADIRONDACK BASEMENT UPLIFT ON JOINTS AND FAULT DEVELOPMENT IN THE APPALACHIAN BASIN MARGIN, NEW YORK

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Introduction

Recent development of unconventional shale gas plays in the Appalachian Basin has drawn considerable attention to basement structures, and joints and fracture zones in the Paleozoic cover rocks (Figure 1), and nowhere is this more apparent than in the region of central Pennsylvania through western and central New York State (Engelder and Lash, 2009; Engelder, 2008; Jacobi, 2002; Jacobi and Fountain, 1996). Two ubiquitous joint sets occur throughout this region (*J1 and J2 of Engelder*), in addition to significant domains of fracture intensification related to faults (Jacobi, 2002). It was demonstrated that the regional joint sets are largely the result of hydrocarbon maturation fluid pressure under late Paleozoic Alleghanian orogenic stress (Lash and Engelder, 2009), however domains of fracture intensification have been linked to faults in the underlying crystalline basement (Jacobi and Fountain, 1996; Jacobi, 2002). Basement structure controlled the distribution and orientation of brittle faults in the overlying Paleozoic strata (Jacobi, 2002), and in some cases the displacement on basement faults had a substantial influence on patterns of Paleozoic sedimentation (Jacobi and Mitchell, 2002).

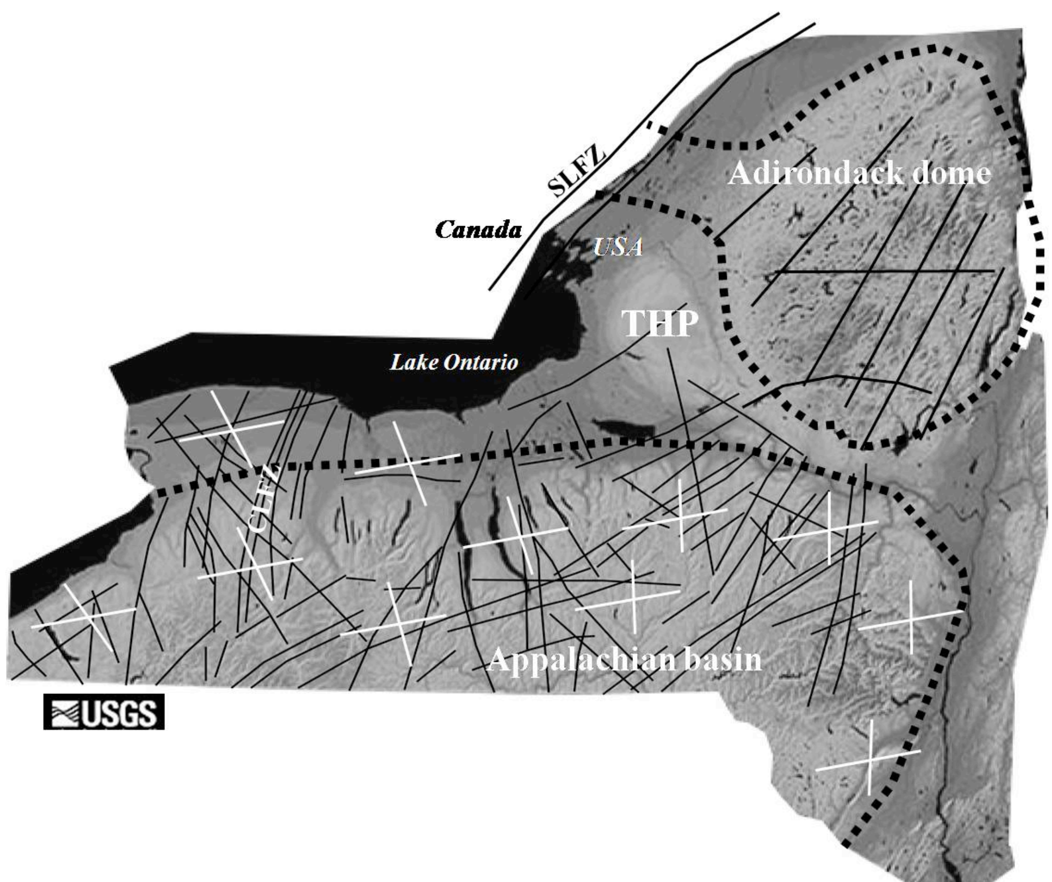


Figure 1 - Physiographic map of New York State (USGS) with generalized faults, fracture zones and joint patterns. The Appalachian basin and Adirondack dome are outlined by heavy dashed lines. Within the Appalachian basin, the dark lines represent inferred faults, and the white lines represent the strike direction

of steeply dipping joints (Jacobi, 2002; Jacobi and Fountain 2002; Jacobi and Mitchell 2002; Lash and Engelder 2009; Engelder and Geiser 1980; Engelder and Gross 1993; Engelder and Lash 2008; Engelder et al. 2001; Engelder and Whitaker 2006; Fakundiny, 1986; Fakundiny et al., 1996; Isachsen et al., 1993; Isachsen and McKendree, 1977; Isachsen et al., 1991; Zho and Jacobi 1996). THP – Tug Hill Plateau.

The direct observation of basement faults that had control on the distribution and subsequent deformation of Paleozoic cover is limited to the exposure of Proterozoic rocks of the Adirondack dome (i.e. Isachsen, 1975; Isachsen, 1981; Roden-Tice et al., 2000; Jacobi and Mitchel, 2002). Roden-Tice et al. (2000) and Jacobi and Mitchel (2002) suggested that the Adirondack dome uplifted in segments that were accommodated by faults related to the Iaptian rift. Jacobi and Mitchel (2002) also discovered that the distribution of the Ordovician strata was controlled by fault bounded structural blocks, demonstrating fault activity during that period. Roden-Tice et al. (2000) and Roden-Tice and Tice (2009) demonstrated that the Adirondack region underwent differential uplift with the formation of the dome, and was controlled by change in bulk compression direction and reactivation of faults. Through apatite fission track dating, Roden-Tice et al. (2000) constrained the age of the dome uplift to mid-Jurassic through the Cretaceous. Wallach and Rheault (2010) suggested that the gentle southwestern incline of the Ordovician strata west of the Adirondacks, to be directly the result of basement faulting and uplift of the Adirondack dome. They further suggested that reactivation of the Carthage-Colton shear zone and movement on the Black River fault contributed to the formation of the Tug Hill plateau. Earlier, Isachsen (1981) reported on Adirondack fault zones that form small graben that contain remnants of the Paleozoic strata that covered the Adirondacks prior to uplift. Additionally, Isachsen (1975; 1981) proposed that neotectonic activity continues in the Adirondacks, and this is partially supported by seismic activity on the Saint Lawrence fault zone and in the Champlain Valley (Barosh, 1986; 1990; 1992; Faure et al., 1996; Mareschal and Zhu, 1989; Wallach, 2002). A statistical correlation between faults and seismic activity led Deneshfar and Ben (2002) to conclude that northwest striking faults in the Adirondacks are more likely to exhibit seismic activity.

It is apparent that faulting in the Proterozoic basement of the Adirondack dome exhibits a protracted history, ranging from late Proterozoic through the Cretaceous, and possibly even neotectonic in nature. Although it was demonstrated that basement faults penetrate the Paleozoic rocks adjacent to the Adirondack dome, the extent of joint and fracture zone development associated with rise of the dome has yet to be described. The intent of this investigation was to address this issue though a detailed study of joints and fracture zones in the Middle Ordovician rocks that directly overly the western flank of the Adirondack dome. This is the region of the Tug Hill plateau, where Wallach and Reault (2010) recently identified basement faults that extend into the Paleozoic strata (Figure 2), and where deformation associated with the Appalachian basin (i.e. Jacobi, 2002; Engelder, 2009) is most likely overlapped by deformation related to the uplift of the Adirondack dome.

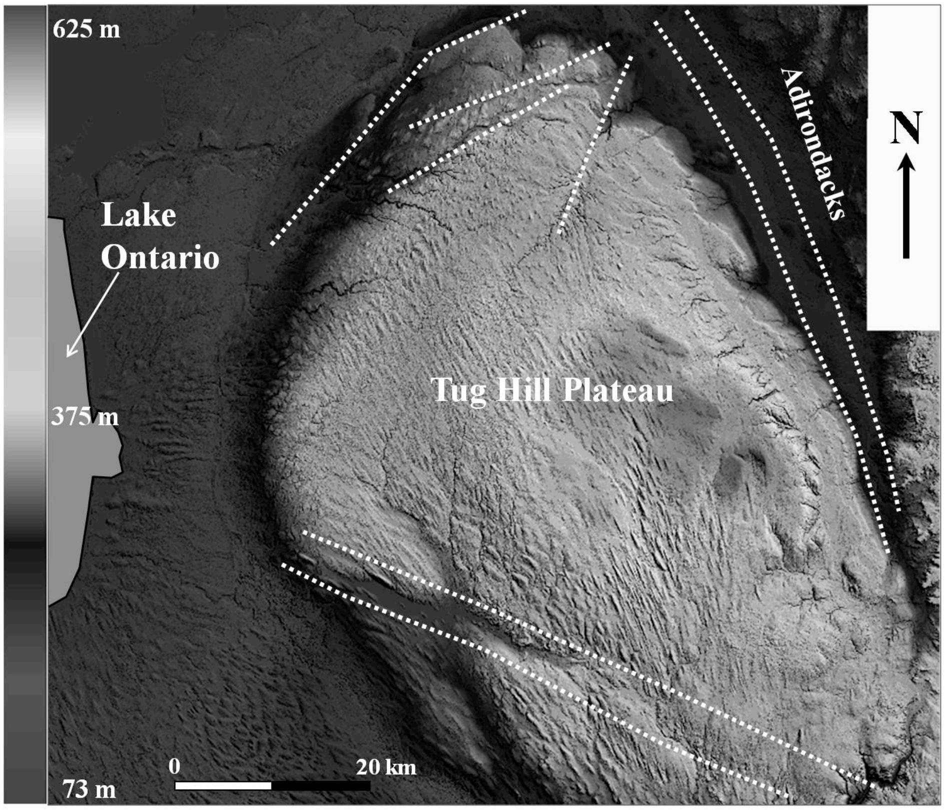


Figure 2 - Digital elevation model for the Tug Hill plateau. The dashed lines are faults of Wallach and Reault (2010).

THE TUG HILL PLATEAU

Situated between the Adirondack dome to the east and Lake Ontario to the west, the Tug Hill plateau rises about 500 meters from the lake shore to the highest point (Figure 2). The plateau covers an area approximately 3500 square kilometers and is underlain by Middle Ordovician strata that sit nonconformably on the Mesoproterozoic basement. These strata are inclined 2° to 5° toward the southwest, and Wallach and Reault (2010) propose that this incline to be the result of faulting in the basement during the uplift of the Adirondacks (Roden-Tice, 2000; Roden-Tice et al., 2009). Approximately 500 meters of strata are exposed along the rim of the plateau through river beds, gulfs, road outcrops, and quarries, but the center of the plateau is concealed by ground moraine, drumlins and wetlands. There are long linear escarpments and valleys on the southern, northern, and eastern flanks of the plateau, that were proposed to be the locations of faults (Wallach and Rheault, 2010). The southern flank of the plateau is bound by the Prospect Fault (Jacobi, 2002) and the northern flank of the plateau is coincident with the southwestern extension of the Carthage-Colton shear zone (Wallach and Rheault, 2010). Finally, the eastern flank of the plateau is the steepest slope, and the location of the proposed Black River fault (Wallach and Rheault, 2010). Superficially, it appears that the general geomorphology of the Tug Hill plateau was controlled by faults within the underlying basement that extend upward into the overlying sedimentary strata.

GENERAL STRATIGRAPHY

The Middle Ordovician strata of the Tug Hill plateau record the transition from marine to terrestrial deposition associated with carbonate sedimentation during the late stage of the Laurentian passive margin (Black River and Trenton groups), and siliciclastic deposition related to the onset of the Taconic orogeny (Lorraine Group) (Figure 3). The Tug Hill plateau sequence includes the Black River and Trenton Group carbonates that reside nonconformably on the crystalline basement, and the unconformity is exposed in the Black River valley. The thick limestone sequence that makes up the Trenton Group is directly overlain by the Utica black shale across an abrupt disconformity contact. With a progressive increase in the occurrence of siltstone beds, the Utica Formation transitions upward into the Whetstone Gulf Formation, and in turn the Whetstone Gulf Formation grades upward into the Pulaski Formation with the increase in the number of sandstone beds. Finally, with a marked decrease in the number of shale beds, the Pulaski Formation grades upward into the thick bedded sandstone of the Oswego Formation. Although earlier geologic maps (Isachsen and Fisher, 1970) show a few isolated occurrences of Queenston shale, the cap-rock for the Tug Hill plateau is primarily the Oswego sandstone.

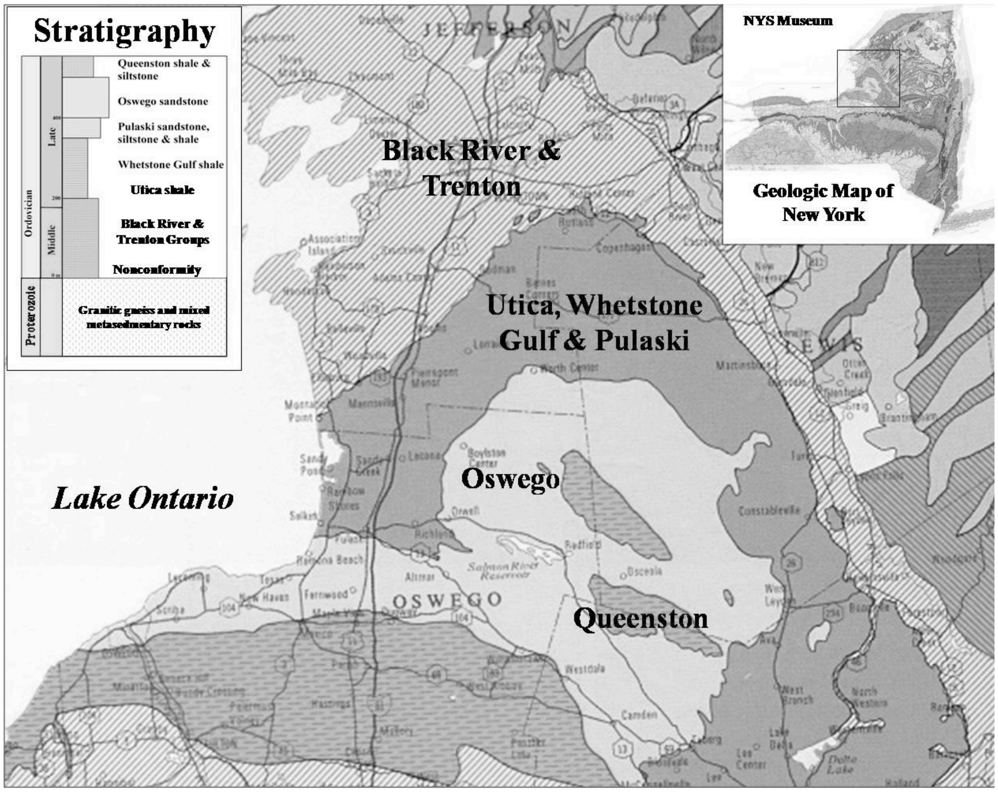


Figure 3 - Geologic map of the Tug Hill plateau (Isachsen and Fisher, 1970). The inset shows the general stratigraphy.

JOINTS, FRACTURE ZONES AND FAULTS

Three field seasons were spent documenting the orientation and density of various joint sets, fracture zones and faults in the strata of the Tug Hill – Eastern Lake Ontario region. Early in the investigation, it was noted that the orientation, spacing and even the occurrence of specific joint sets, vary from bed to bed within outcrops. In addition to mapping and characterizing joints and fracture zones throughout the region, it was prudent to document examples of the detailed variation to further understand the significance of lithology and bed thickness in the brittle deformation of each formation.

Outcrop scale analysis

Trenton Limestone.

Figure 3 is an example of a short stratigraphic column from the middle part of the Trenton limestone on the eastern flank of the plateau. In general, the Trenton limestone is made up of varied thickness inter-layered carbonate rocks. In this analysis, the orientation and density of various joint sets is shown. The structure symbols located to the right of each bed represent the averages of joint data collected for that bed. They are plotted as map symbols with north toward the top of Figure 3. As an example, within the bed of micrite located at 100 cm, there are three different joint sets that strike toward the northwest, east and south. The small numbers at the end of each symbol are the joint densities. In this case, the joint density was quantified for each joint set by counting the number of joints encountered over an average distance of 1 meter in the direction orthogonal to the joint surface. Therefore, the joint density is reported in joints per meter. Note the variability in joint attitude and density throughout this small example of Trenton limestone (Figure 4). To extrapolate these results in a regional analysis, the number of data collected is important due to the extreme variability. A minimum of several hundred joint orientations were measured at field sites to overcome the variability that occurs in some formations, as shown in this example.

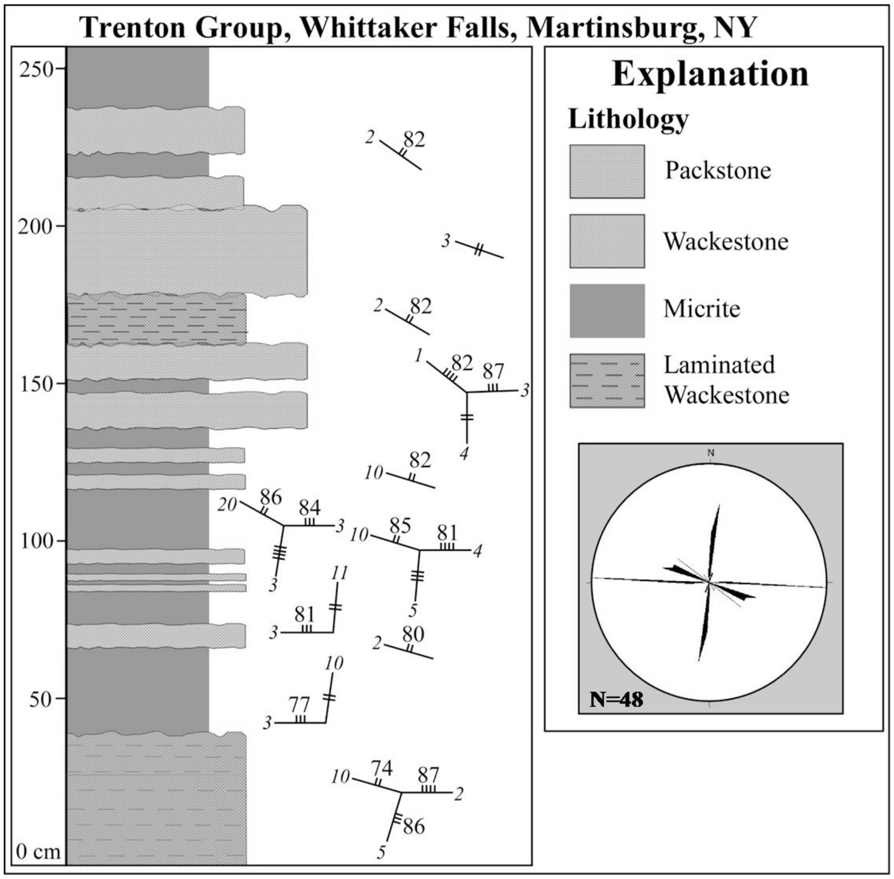


Figure 4 - Representative stratigraphic column for the middle section of the Trenton limestone showing the orientation of joints per sedimentary bed. The rose diagram represents all of the data for this section.

Within the Black River and Trenton limestone units, there are southeast striking normal faults with minor displacement on the order of several decimenters (Figure 5). Some of these faults exhibit no obvious throw, suggesting that they have a lateral displacement component. In some cases, the faults merge with bedding plane shear fractures (Figure 3), forming domino structure and mineralized slicken-sides. This class of fault was only observed in the limestone units.

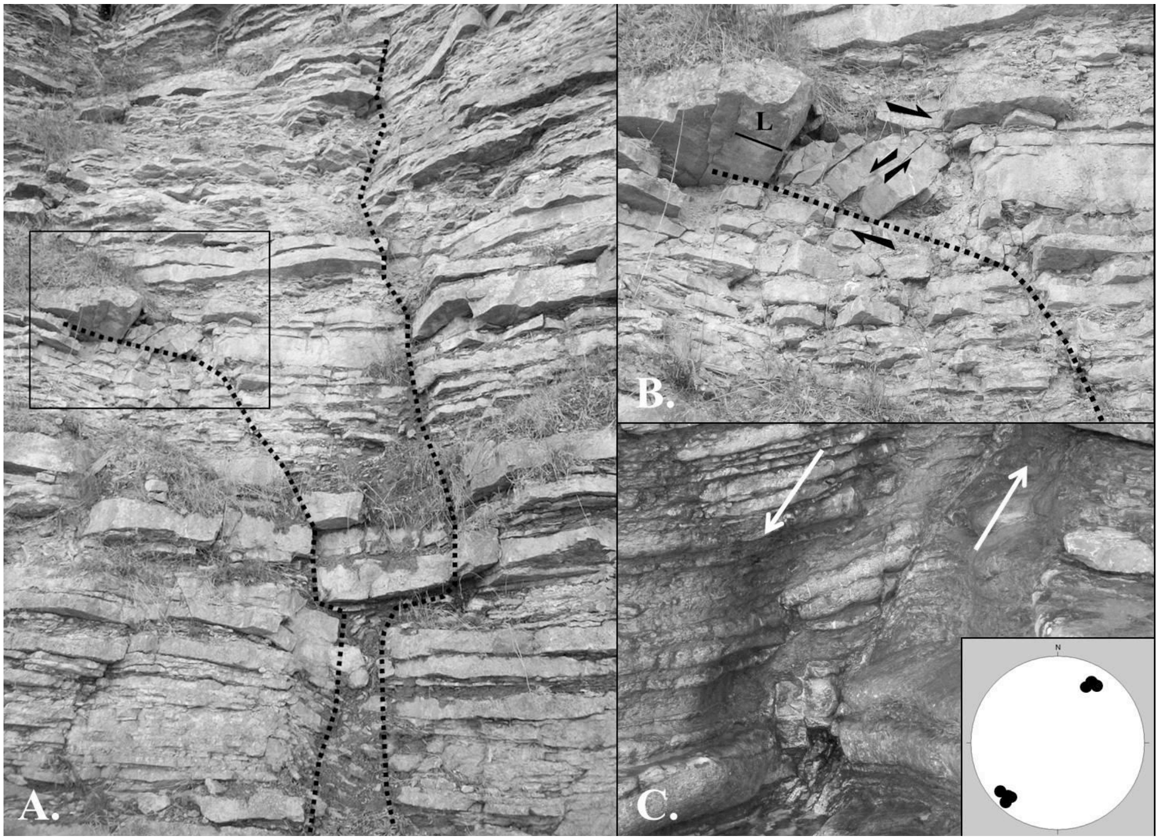


Figure 5 - Photographs of faults in Trenton Group limestone at Stony Point, New York. A. Complex fault with breccia zone, splays and no apparent dip-slip displacement. B. Close up of a domino structure where the fault merges with bedding. C. Example of a normal fault with several decimeters of displacement. The inset stereogram is the poles to fault planes.

*Utica Shale***.**

Continuing upward in the strata, the Utica black shale directly overlies the Trenton limestone. Four different fracture sets, with the highest density occur within the shale. Because shale is highly susceptible to weathering, there are few outcrops of the Utica shale away from the deep gulfs that occur in the flanks of the plateau. Within the gulfs, pavement style exposures provided an opportunity for detailed digital joint analysis. High resolution photo mosaics were collected and scaled using a computer program, so that total joint density could be quantified. In the example of Figure 6, once the different joint sets were identified (four in this case), the total joint length was summed for the outcrop surface. A plot of the results as vectors (total joint length/outcrop area verses azimuth), provides a visual representation for the joint set that is most abundant relative to geographic coordinates. The southeast striking joint set has a density that is three times greater than the east-northeast and northeast striking joint sets.

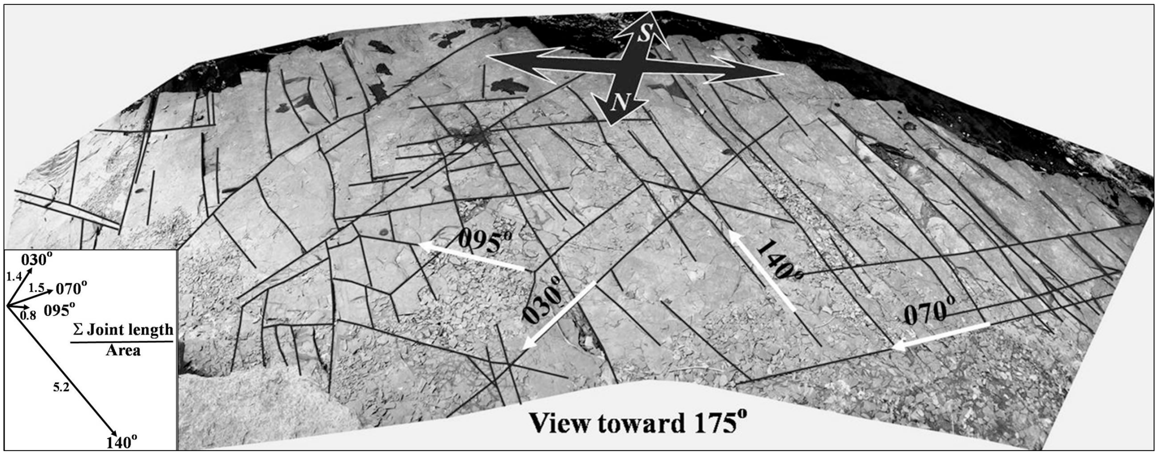


Figure 6 - Outcrop photograph of the Utica Formation with joint traces. The white arrows on the photograph show four joint sets with the average strike. The vector diagram in the lower left represents the total joint length per joint set divided by the outcrop area. See the text for details.

Whetstone Gulf Shale-Siltstone.

The Whetstone Gulf Formation consists of a sequence of black and gray shale that contains siltstone beds. The occurrence and thickness of siltstone beds increases stratigraphically upward. As well, there are thin sandstone layers that occur near the top of the formation. Northeast and southeast striking joint sets occur within the shale and siltstone beds, with rare occurrence of north-south striking joints in the sandstone beds (Figure 7). The regional consistency of the northeast and southeast striking joints can be seen in Figure 8, where the joints are shown to be consistent over an area of several square kilometers between Mooney Gulf and Totman Gulf on the western flank of the Tug Hill plateau. This is pattern of map-scale joint consistency is typical of the Whetstone Gulf Formation.

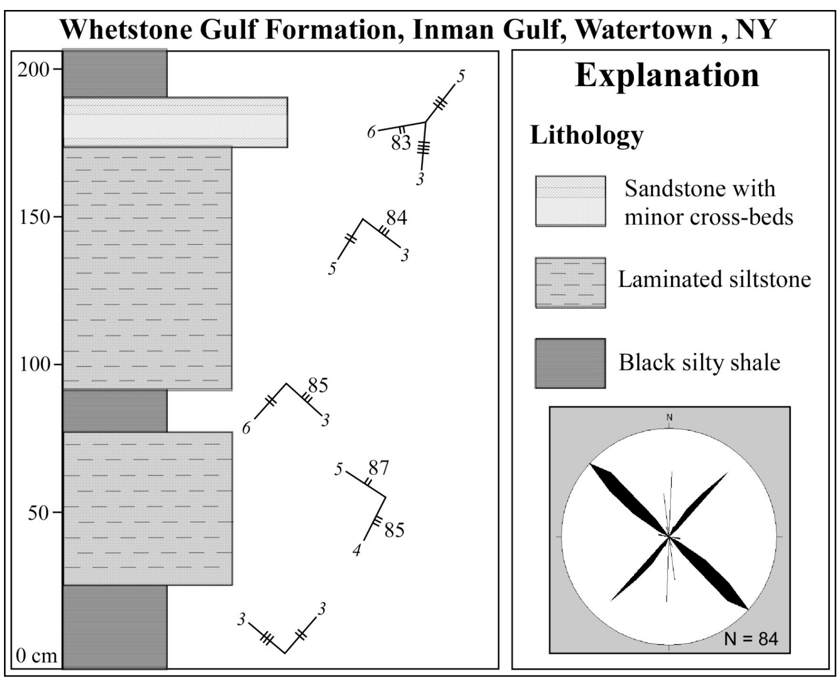


Figure 7 - Representative stratigraphic column for the Whetstone Gulf Formation at Inman Gulf. See text for details.

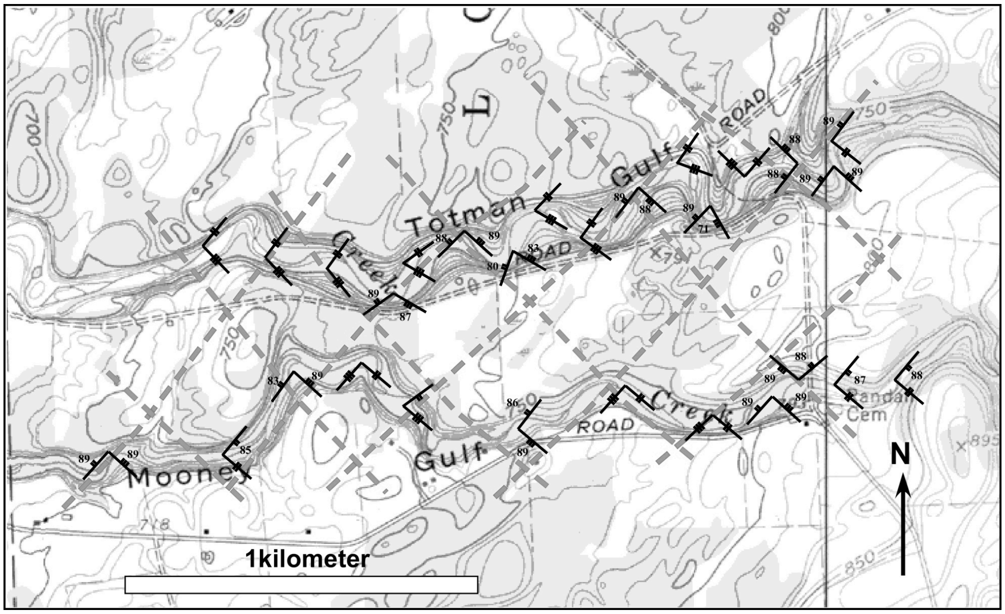


Figure 8 - Joint map for Totman and Mooney Gulfs, west flank of the Tug Hill plateau. The heavy dashed lines show the consistency in the strike of two joint sets in the Whetstone Gulf Formation over several kilometers.

Pulaski Shale-Siltstone-Sandstone.

Within the Pulaski Formation there are joints with strike variation on a bed-by-bed scale, where the shale, siltstone, and sandstone beds exhibit different joints sets and joint densities (Figure 9). Overall, there are northeast, east-northeast, southeast and south-southeast striking joint sets that appear to be tied to specific lithologies. This is the only formation in the Tug Hill plateau that exhibits both joint sets that appear to be associated with the Appalachian basin (J1 & J2) and with the crystalline basement. In addition, the shale units in the Pulaski Formation have extensive tightly spaced fracture cleavage ranging from 2 to 5 fractures per centimeter (Figure 10).

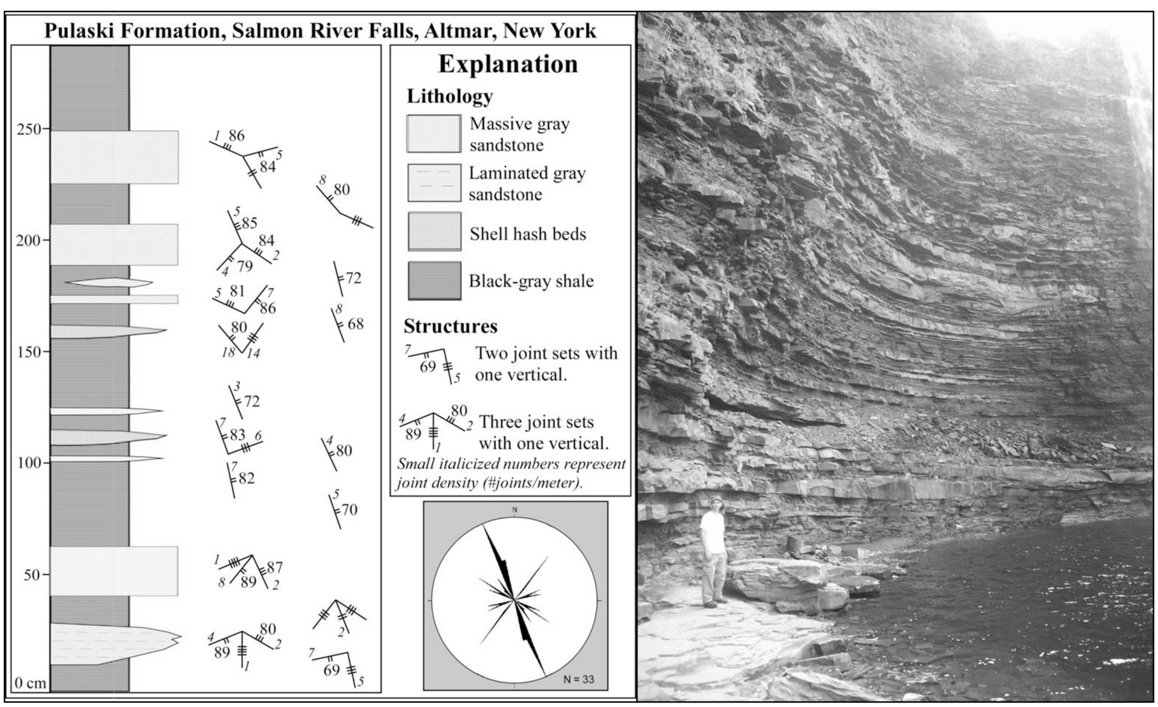
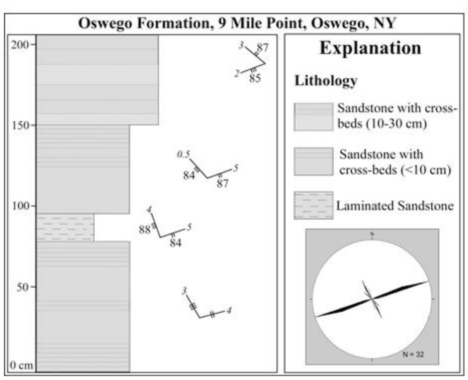


Figure 9 - Representative stratigraphic column for the Pulaski Formation at Salmon River Falls, Oswego County. The photograph shows the lithologic variability within the upper Pulaski Formation.

Figure 10 - Photograph of a shale bed that is sandwiched between two sandstone beds in the Pulaski Formation. Note that the shale contains fracture cleavage. The shale bed is about 10 cm thick in the photo.

Oswego Sandstone.

Finally, within the Oswego Formation, there are dominant east-northeast and southeast striking joint sets that persist from the region of Lake Ontario to the fringe of the Tug Hill plateau (Figures 11 and 12). These joint sets are consistent throughout the sandstone and only vary in strike where beds are either very thick (> 40 cm) or very thin (<2 cm). There is evidence for minor left lateral shear (2 – 20 cm) on the east-northeast striking joint set, where the southeast striking joints have been displaced horizontally. Small en-echelon fracture zones with left lateral geometry have a consistent strike with the individual fractures (Stilwell et al., 2005).



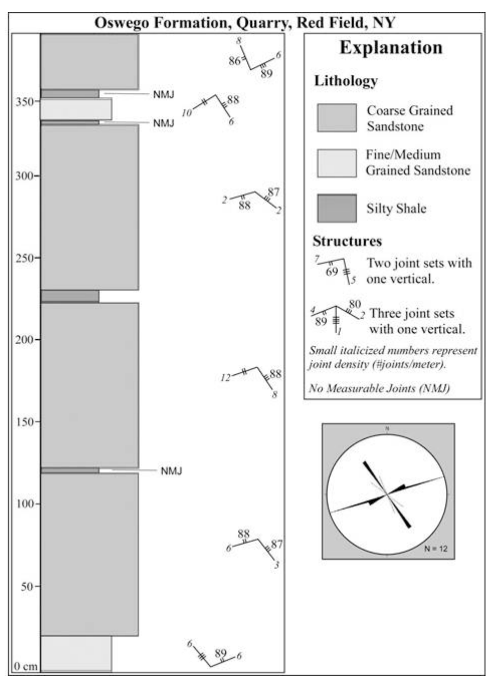


Figure 11 – Representative stratigraphic columns for the Oswego Formation at Nine Mile Point and Redfield, Oswego County, New York.

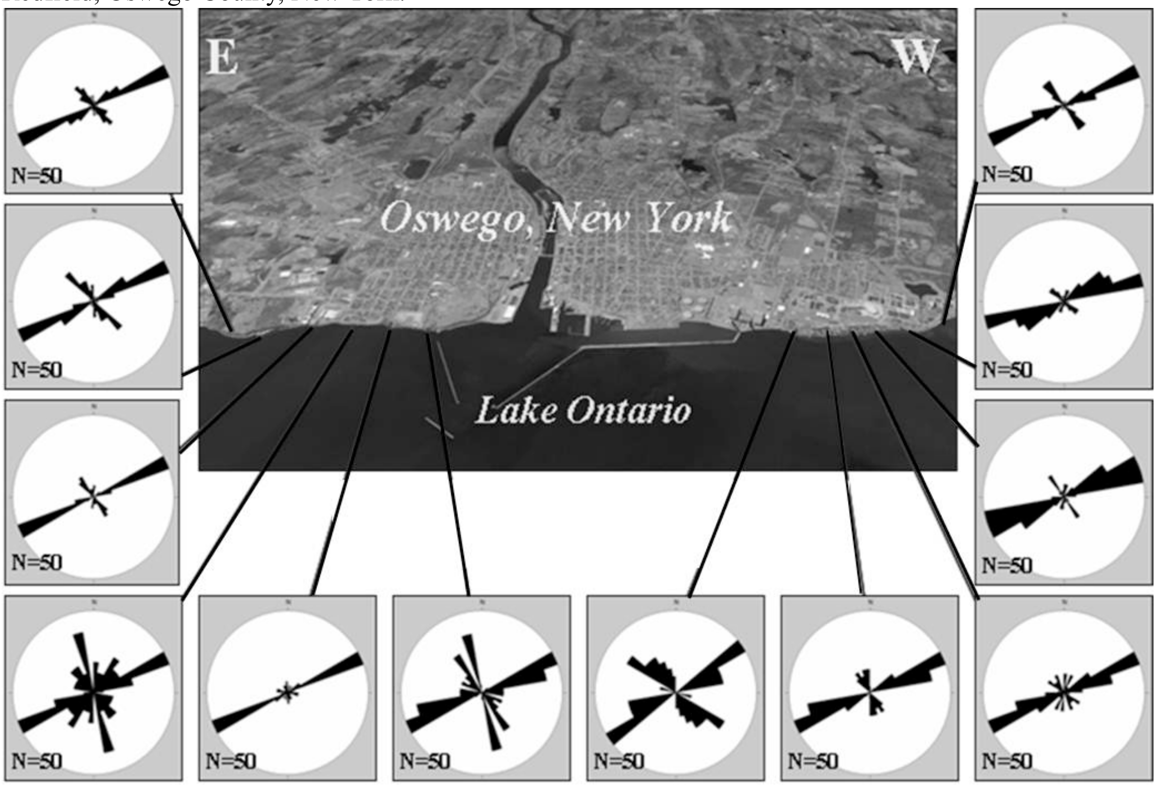


Figure 12 - Rose diagrams for the Oswego Formation at the type section in Oswego, New York. Each rose diagram is tied to the location along the Lake Ontario shore.

Regional fracture analysis

Joint density (joints/meter) is the inverse of joint spacing. For example, a joint density of 5 joints per meter is equivalent to a rock with average joint spacing of 20 cm. Joint spacing was plotted using box-and-whisker (Mcgill et al., 1978) diagrams (Figure 13) for comparison within rock formations across the region, and for comparison between the different formations. Two integrated box-and-whisker diagrams were plotted to portray variation in the strike direction of the joint sets against the variation in the joint spacing. This type of plot allowed for visual and quantitative comparison within and across rock formations.

Within the carbonate rocks of the Black River Group there are southeast, east-northeast and northeast striking joint sets that all have a mean joint spacing of about 30 cm. The limestone of the Trenton Group contains southeast, east-northeast, northeast and north-south striking joint sets with mean spacing of about 50 cm. Black shale of the Utica Formation has very tight joint spacing ranging from 5-10 cm for the east-northeast and northeast striking joint sets, however, east-northeast striking joints are spaced about 50 cm. Within the Whetstone Gulf Formation, there are southeast, east-northeast, northeast and north-south striking joint sets that exhibit tight 15 cm spacing in the eastern flank of the plateau, and wider 40 cm spacing in the western flank of the plateau. Mean joint spacing of 10 cm is consistent throughout much of the Pulaski Formation, where the southeast, east-northeast and northeast striking joint sets are dominant. Finally, within the sandstone of the Oswego Formation, there are only southeast and east-northeast striking joint sets with mean spacing of 25 cm. The Oswego Formation contains by far the most consistent joint-set occurrence, strike and spacing, probably reflecting the relatively homogenous lithology of the formation.

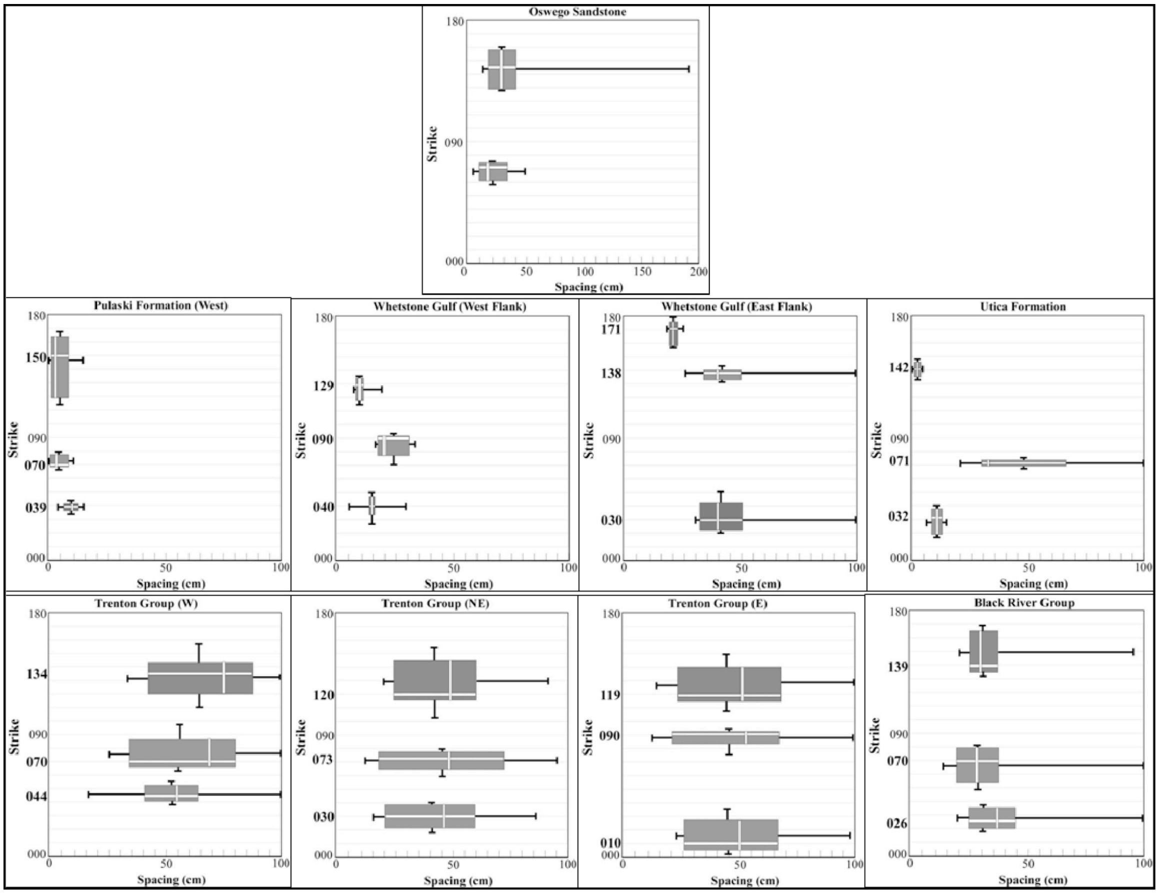


Figure 13 - Box and whisker plots of joint orientations and spacing for the Tug Hill plateau. The bottom row includes the carbonates of the Black River and Trenton Groups. The middle row includes the shale-bearing units of the Utica, Whetstone Gulf and Pulaski Formations. Note that the spacing scale for the Oswego sandstone is two times the other plots.

Adirondack region

Numerous northeast trending, long, narrow valleys and elongate lakes dominate the landscape of the Adirondack Mountains and define major topographic lineaments (Figure 2) that are interpreted as fault zone (Isachsen, 1975; Isachsen, 1981; Daneshfar and Ben, 2002). A few of these faults define the borders for small graben that contain Paleozoic carbonates (Isachsen, 1975; Isachsen, 1981). Although the fault lineaments are pronounced, the direct observation of the basement faults is obscured by glacial and lake cover, and mature forests, and the problem is compounded by intense weathering. In effect, there are few places where the faults can be directly studied and sampled without drilling. During this investigation, we examined joints and fracture zones at three locations within the Adirondack Mountains as a representative data set to compare with the joint data for the overlying Paleozoic rocks in the Tug Hill plateau. Joint orientation data was collected in the Moose River basin immediately east of the Tug Hill plateau. As well, joint data was collected in the regions of Piseco and Indian Lakes with the objective of demonstrating regional joint patterns in the basement. Figure 14 is a digital elevation model for the southern Adirondacks with pronounced topographic lineaments. Composite rose diagrams (several hundred joint strikes per diagram) of joint strikes for the three locations show a strong correlation between northeast trending lineaments and the dominant joint set. A second joint set generally strikes east-west to southeast.

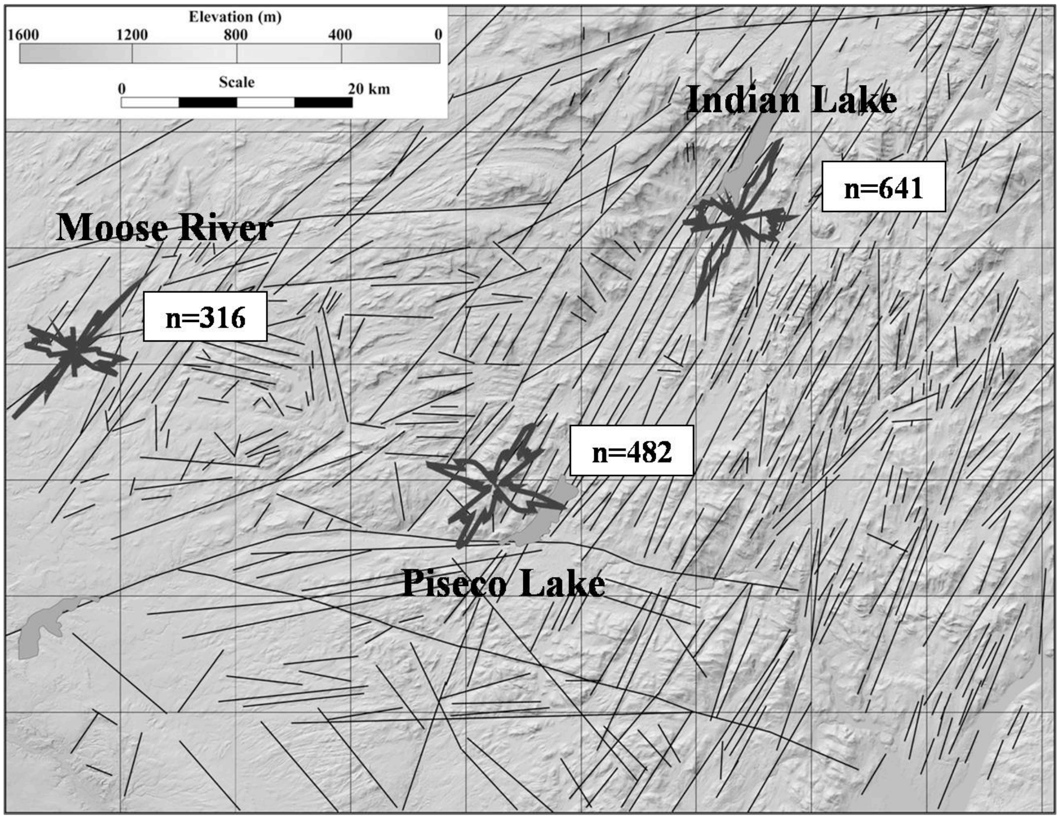


Figure 14 - Digital elevation model for the southern Adirondacks. The black lines highlight major topographic lineaments. The composite rose diagrams represent the distribution of the strikes for joints at three locations. See text for details.

DISCUSSION AND CONCLUSIONS

From the field observations during this study, it is apparent that joint density and strike are variable at the outcrop scale. However, at the regional scale, it is also apparent that joint variability is related to the proximity to the crystalline basement both laterally and stratigraphically, and related to the variation in rock type with mechanical characteristic control on the joint and fault formation (Figure 7). Starting at the base of the plateau, the limestone units have the most variability in joint strike with all sets that are present. The joint densities are relatively moderate with ~5 joints/meter in the rocks of the eastern flank of the plateau, and 2-3 joints/meter in the rocks of the western flank. In addition, the limestone units contain the highest number of southeast striking normal faults (Figure 15), and they are roughly parallel with the Black River fault (Figure 2) that was proposed by (Wallach and Rheault, 2010; Wallach, 2002). Within the shale-bearing rock formations, faults do not occur, but the fracture density is high in all of the sets that are present, with density values that exceed 20 joints/meter. There are dominant northeast, east-northeast and southeat striking joint sets with only a minor occurrence of the north-south striking cross fold joints. Finally, in the sandstone that caps the plateau, the joint sets have a generally low density of <1-4 joints/meter, and only the east-northeast and southeast striking fracture sets are present. The east-northeast striking joint set has evidence for left lateral reactivation (Stilwell et al., 2005), which is evident in the offset of southeast striking joints and discrete en-echelon fracture zones.

Figure 15 summarizes the occurrences of the four joint orientations and faults. The lower limestone groups contain the highest number of faults. These faults exhibit normal displacement, some terminate by merging with bedding plane fractures, and they strike northwest-southeast, a direction that is roughly parallel to the Black River Fault of Wallach and Reault (2010). The limestone groups also contain all four joints to some extent with the northeast, southeast and east-northeast striking sets being dominant. No faults were observed in the shale bearing formations of the Utica, Whetstone Gulf and Pulaski. However, these formations have the highest joint densities regardless of the joint set. All three formations contain well developed northeast, southeast and east-northeast striking joints, and only minor occurrences of N-S striking cross-fold joints. Overall the Oswego sandstone contains the least number of joints, with the east-northeast and southeast striking sets occurring in every outcrop. Minor occurrences of north-south striking joints were observed, and none of the northeast striking joints. The northeast striking joint set in the basement also occurs in the Paleozoic rocks, except that this set was not observed in the Oswego formation.

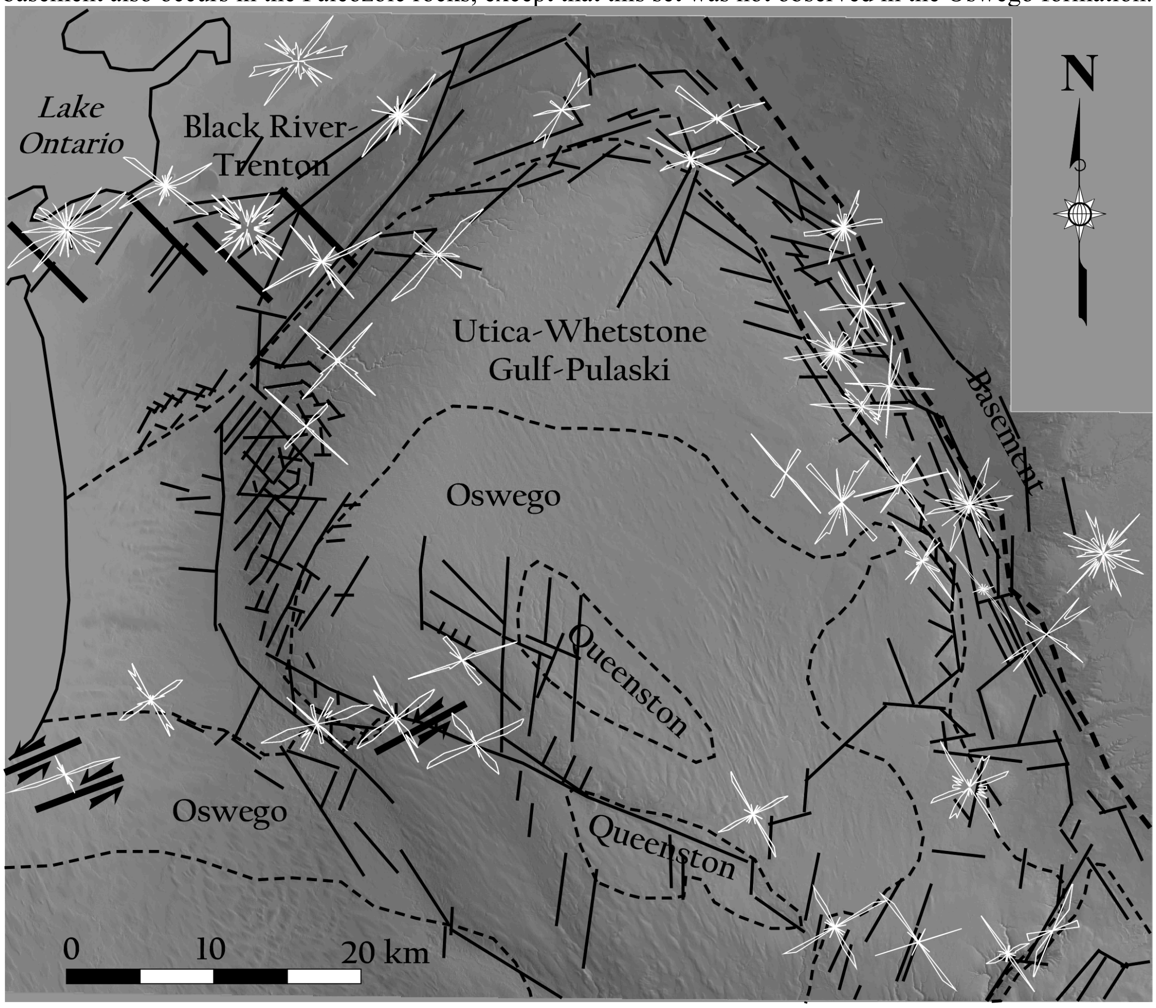


Figure 15 - Digital elevation model for the Tug Hill plateau with geologic formations, joint rose diagrams, and faults. The Black River fault is from Wallach and Rheault (2010). Minor faults are represented by short heavy lines represent observed faults. Those faults exhibit normal displacement in the northwest region and sinistral displacement in the Oswego Formation. The lighter weight black lines are interpreted topographic lineaments.

Four systematic joint sets were documented in the Tug Hill plateau. As before, Lash and Engelder (2009) demonstrated that the regional joint sets (J1 and J2) in the Appalachian basin Devonian strata are the result of hydrocarbon maturation fluid pressure during the Alleghanian orogeny. The east-northeast striking (J1) joints are well developed within all of the Ordovician rock units of the Tug Hill plateau (Figure - chart of joint sets), and there may also be some occurrences in the underlying basement. It is possible that maturation fluid pressure associated with the black shale of the Utica, Whetstone Gulf and Pulaski Formations was similar to the Devonian shale as explained by Lash and Engelder (2009). However, one should note that J1 is well developed in the underlying Trenton and Black River Group carbonates, suggesting that a mechanism other than maturation fluid pressure must also be responsible. Considering the persistence of these joints in all of the Ordovician strata, then perhaps they reflect the general east-west subhorizontal compression associated with the Alleghanian orogeny in the northern Appalachians, and the impact of that event on the underlying basement. On the contrary, the north-south striking cross-fold joints (J2) are only of minor occurrence, most likely due to the distal location relative to the Alleghanian thrust and fold belt.

The northeast and southeast striking joints (ADK 1 and ADK 2) that dominate the crystalline basement (Figure 16) are also developed in the overlying Ordovician strata of the Tug Hill plateau, but the occurrence is variable. The northeast striking joints have direct correlation with a suite of steeply dipping northeast striking normal faults in the southern Adirondacks (Roden-Tice, 2000; Valentino et al., 2011), however, no dip-slip deformation has been documented for the northeast striking joints in the Tug Hill plateau. Barosh (1990, 1992) proposed the northwest trace of oceanic fracture zones into eastern North America, showing several possible extensions into the Adirondack and Tug Hill regions. These proposed fracture zones are the likely cause of the southeast striking (ADK2) joint set, in addition to the southeast striking faults that occur within the Trenton Group. In the statistical analysis of Adirondack faults, Deneshfar and Ben (2002) also concluded that the northwest striking faults to be candidates for seismic activity. Wallach and Reault (2010) demonstrated that southeast striking faults account for the southwestern regional tilt of the Ordovician strata and accommodated the vertical rise of the crystalline basement during formation of the Adirondack dome. During this investigation, it was demonstrated that the east-northeast striking sinistral faults in the Oswego Formation cross-cut the southeast striking joints. This relative timing suggests that the sinistral deformation post-dates, or is synchronous with uplift of the Adirondack dome. This is most likely reactivation of the earlier developed J1 Appalachian joints as the result of differential uplift of the Adirondacks. These faults were only observed in the Oswego Formation, suggesting that the underlying shale units accommodated strain through partial plastic deformation. Note the development of cleavage in the Pulaski Formation and the lack of intense joints in some of the sections of the Whetstone Gulf formation.

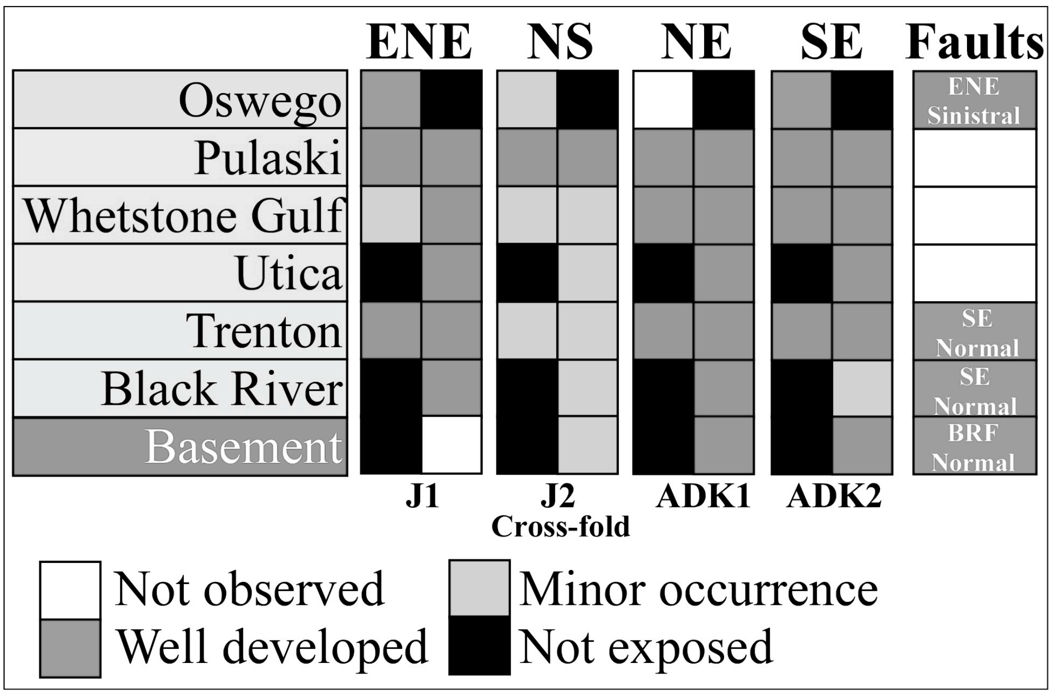


Figure 16 – Summary of joint occurrence and development in the Tug Hill plateau region. For each joint set there are two columns. The left and right columns are for the western and eastern flank of the plateau respectively.

As the Adirondack dome uplifted in the Late Jurassic to Cretaceous (Roden-Tice et al., 2000) the Paleozoic strata of the plateau underwent deformation (Figure 9), resulting in the regional 2^{o}-5^{o} southwestern tilt of the plateau strata, displacement on minor faults, and development of joints. This investigation demonstrates that there is a geometric and most likely genetic relationship between the southeast striking joint sets of the Adirondacks, the parallel Black River fault and the smaller faults found in the lower limestone units of the Tug Hill plateau. With the uplift of the Adirondack dome, the stratigraphic units of the plateau had different responses depending on structural competency and position, and thus had control on the development of joints and their intensity (Figure 17). The limestone layers responded to the uplift by developing southeast striking normal faults and high joint density closer to the basement. Within the shale bearing rock units, the response was development of high density joints in addition to local fracture cleavage consistent with a less competent rock body. Finally, the Oswego sandstone was not as severely affected by the uplift but instead there was reactivation of the east-northeast striking joints as minor left lateral faults. Considering the relative strength of a thick sandstone body, the Oswego sandstone should exhibit more variability in joints. However, the thick underlying shale formations probably absorbed much of the joint forming stress through development of fracture cleavage and low-temperature plastic flow. Even in the lithified state, the clay that makes up the shale formations would have experienced some degree of grain-boundary sliding in conjunction with fracture cleavage formation resulting in plastic deformation.

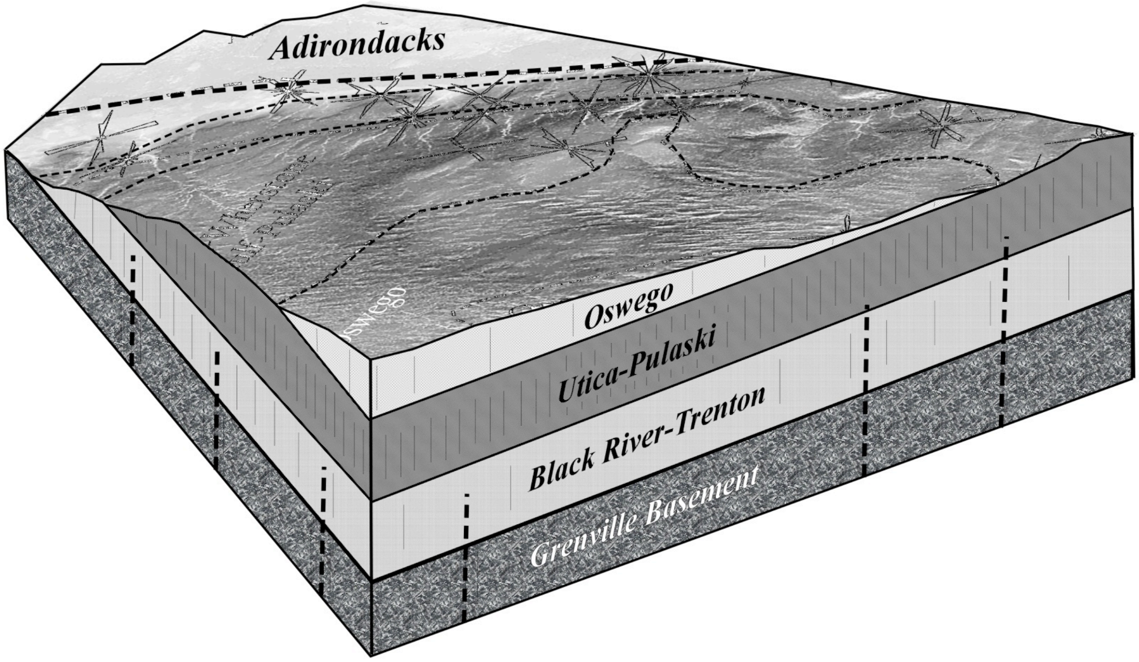


Figure 17 – Conceptual model for joint and fault distribution in the Tug Hill plateau.

Road log and field guide

This field trip circumnavigates the Tug Hill plateau starting in Oswego, New York, on Lake Ontario, traversing up the west side of the plateau at the Salmon River, Mooney Gulf and Stony Point, followed by a traverse down the eastern flank stopping at Whetstone Gulf and Wells Creek.

Stop 1: Lake Ontario at SUNY Oswego

The field trip begins at the parking-lot for Mary Walker Health Center on the campus of SUNY Oswego. From the parking-lot, follow the path along the lake shore toward the east to the first metal stair case. The outcrop of interest is at the bottom and to the west of the stair case.

A. Thick to medium bedded Oswego sandstone.
B. Two prominent joint sets (east-northeast and southeast striking).
C. Sinistral fault in the pavement outcrop.
D. Evidence for minor sinistral displacement on individual east-northeast striking joints.
E. East-northeast striking en-echelon fracture zones with sinistral geometry.

Road log:

At the entrance to the parking-lot, turn left on Rudolph Road.
0 – 0.5 miles to Sheldon Ave., turn right.
0.5-0.7 miles to Washington Blvd., turn left.
0.7-1.0 miles to Bridge Street, continue straight through traffic light.
1.0-2.5 miles to 7th Street, turn left.
2.5-2.7 miles to Schuyler Street, turn right.
2.7-3.0 miles to Helen Street, turn left.
3.0-3.4 miles to Stop 2.

Stop 2: Lake Ontario in East Oswego

From the parking-lot, walk east along the lake shore to the first large exposure of the Oswego sandstone.

A. Thick bedded Oswego sandstone.
B. Two prominent joint sets (east-northeast and southeast striking)
C. Fracture zone about 1 meter wide.

Road log:

3.4-3.6 miles Back-track on Helen Street to the intersection with Mitchel Street and turn left.
3.6-4.9 miles to Middle Road on Mitchel Street, proceed straight across the intersection.
4.9-10.9 miles on Middle Road to State Route 104, turn left.
10.9-26.6 miles to Co. Rt. 22, turn left.
26.6-31.0 miles to Co. Rt. 13, turn right.
31.0-31.2 miles to Cemetery Street (Co. Rt. 22 again), turn left.
31.2-35.7 miles to Falls road, turn right.
35.7-37.0 miles to Stop 3 (Salmon River Falls parking-lot)

Stop 3: Salmon River Falls

From the parking-lot, follow the trail northeast to the top of the falls, and then follow the trail back toward the parking-lot to the entrance to the gorge trail.

A. Oswego sandstone at the top of the falls.
B. Pulaski formation in the gorge.
C. Joints and fracture zones in the Oswego sandstone.
D. Fracture cleavage in the shale beds of the Pulaski formation.
E. Normal fault on the southeast side of the base of the falls.

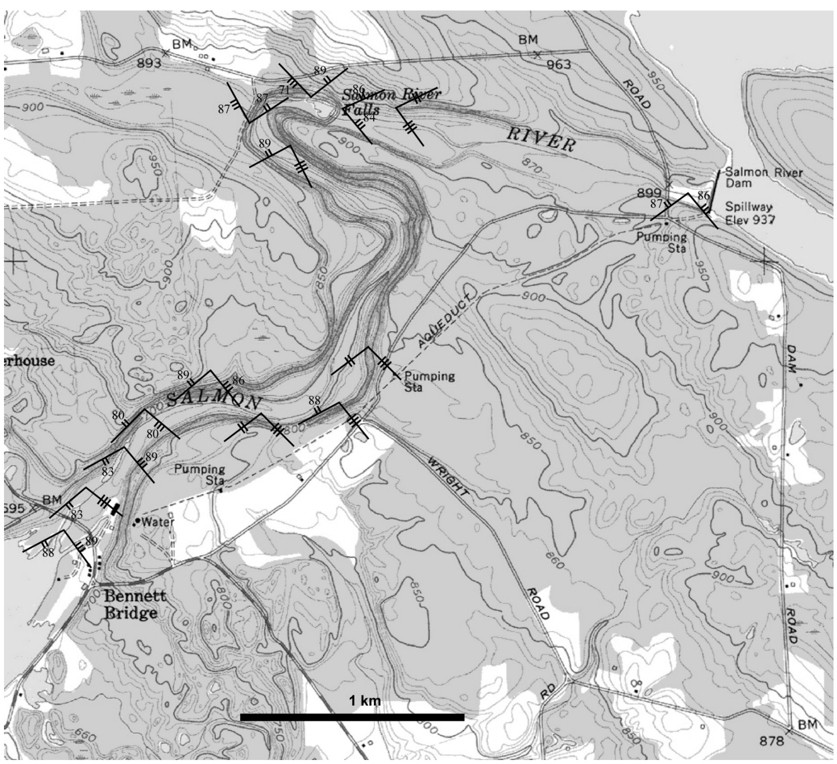


Figure 18 – Joint map for the Salmon River gorge, Oswego County, New York.

Road log:

37.0-38.3 miles on Falls Road to Co. Rt. 22, turn left.
38.3-42.8 miles on Co. Rt. 22 to Co. Rt. 13, turn right.

42.8-48.4 miles to Co. Rt. 2A, turn right.
48.4-50.5 miles to the entrance of I81 headed north.
50.5-62.7 miles on I81 north to Route 193, turn right.
62.7-63.4 miles to Route 11, turn left.
63.4-65.6 miles to Lemay Road, turn right.
65.6-67.7 miles on Lemay Road to a gravel pull-off on the left.

Stop 4: Mooney Gulf.

From the parking area, follow the gravel road into the gulf. There are excellent exposures on the north side of the gulf. See Figure 8 for a detailed joint map of Mooney and Totman Gulfs.

A. Whetstone Gulf formation.
B. East-northeast and southeast striking joints are present.
C. Some joints can be traced 10's of meters.

Road log:

67.7-69.8 miles on Lemay Road back to Route 11, turn right (north).
69.8-73.1 miles to Route 178, turn left.
73.1-85.9 miles on Route 178 to Military Road, proceed straight.
85.9-87.3 miles to North School House Road, turn right.
87.3-87.4 miles to Robert Wehle State Park entrance, turn left.
87.4-87.7 miles on the park road to the parking-lot.

Stop 5: Robert Wehle State Park at Stony Point.

From the parking-lot, walk northwest to the lake shore. There are several trails to the lake shore, but the closest is located behind the tennis court. Once on the lake shore trail, follow it toward the northeast and see the topographic map inset for reference. It will be necessary to traverse down the steep embankment to reach lake level. See Figure 19 for details.

A. Trenton Group limestone.
B. Minor normal faults with mineralized fractures.
C. Fault with no apparent displacement. The fault merges with bedding.
D. Large rock-fall that was controlled by two prominent joint sets.

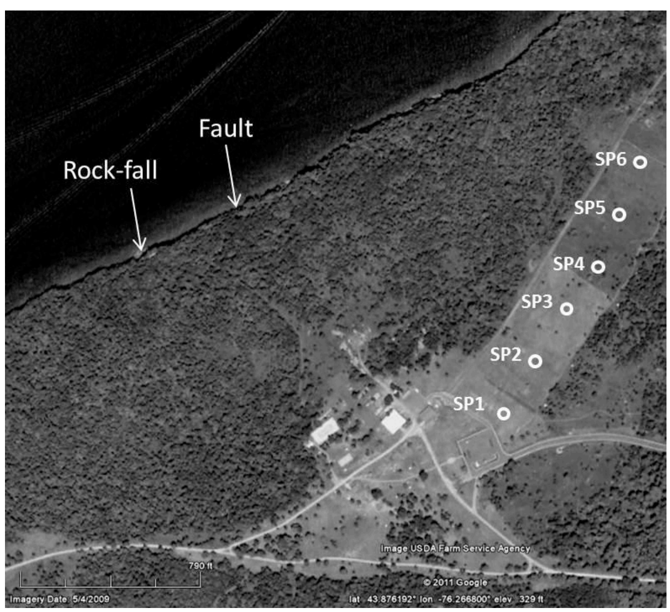


Figure 20 – Google Earth image for Stop 5 showing the locations of outcrops at the lake shore and the locations that azimuthal electrical resistivity surveys were run to reveal joint directions within the park.

Azimuthal Electrical Resistivity for Joint Analysis

Azimuthal ER (AER) involves a Wenner survey around a single point to look for directional variation, which can be impacted by the direction of groundwater-filled joints. The direction of the lowest ER values should correlate with the direction of the most prominent joints (Taylor and Flemming, 1988). At Stony Point, the limestone bedrock is very close to the surface with thin soil. In an attempt to locate joint patterns away from the lake shore, six azimuthal electrical resistivity (ARE) surveys were conducted (Figures 21 and 22). Each AER survey was completed with an automated 24 node system with the azimuthal increment of 30 degrees.

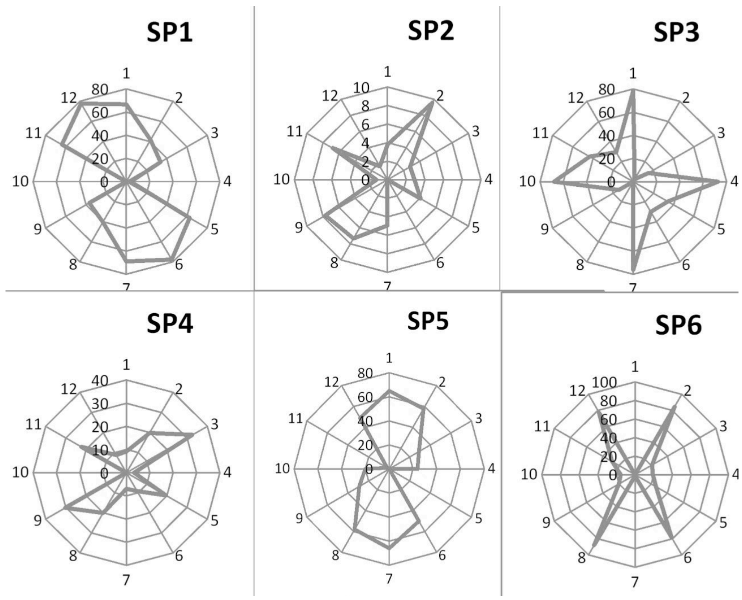


Figure 21 – Azimuthal electrical resistivity surveys that were collected at Stony Point. The ER scale is in ohm-meters. North is in the direction of #1 on each radial graph 30 degree increments for each ER reading.

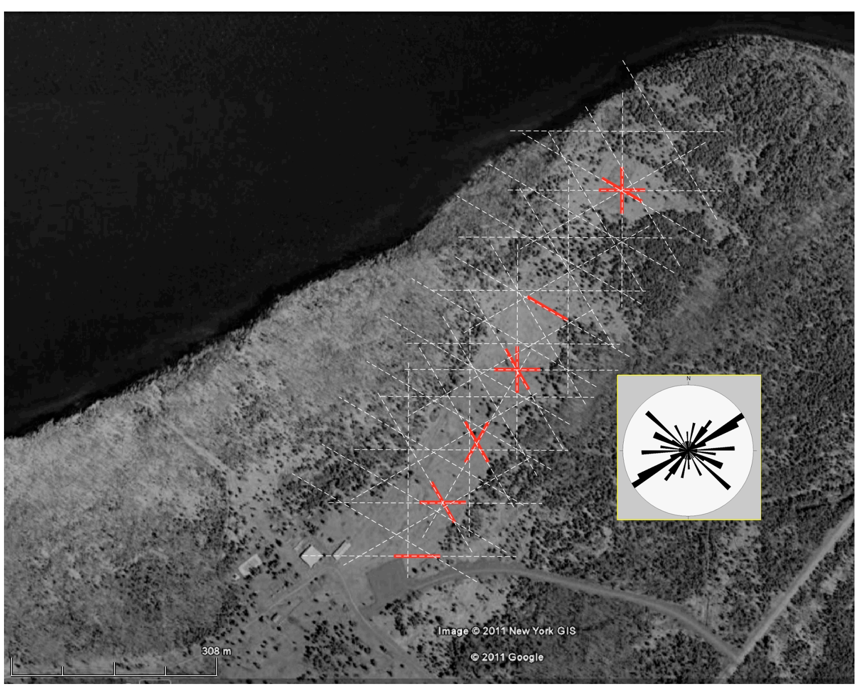


Figure 22 – Interpreted joint directions from the AER surveys at Stony Point. The inset rose diagram is for joint data that was collected at the outcrops on the lake sure (n = 244). The white dashed lines represent the inferred joint directions in the Trenton limestone from the AER results.

Road log:

87.7-88.0 miles Follow the park road back to North School House Road, turn right.
88.0-88.1 miles to Military Road, turn left.
88.1-89.5 miles to Route 178, proceed straight.
89.5-101.9 miles to the entrance of I81 on the left.
101.9-106.0 miles on I81 north to S. Harbor Road (one exit on high-way), turn right at end of exit ramp.
106.0-106.7 miles on S. Harbor Road to Route 177. Proceed straight through the intersection.
106.7-131.4 miles on Route 177 headed east across the Tug Hill plateau. Turn right on Co. Rt. 29.
131.4-139.1 miles on Co. Rt. 29 to the entrance of Whetstone Gulf State Park on the right.
139.1-139.5 miles on the park road to Stop 6.

Stop 6: Whetstone Gulf State Park.

Excellent exposures of Utica black shale occur in the bed of the creek that parallels the park road into the gulf. As well, the Whetstone Gulf formation is exposed in numerous high-walls that can be accessed along the road side farther up the gulf.

A. Utica black shale.
B. Whetstone gulf shale and siltstone beds.
C. Complex joint sets in the Utica formation (see text for details).
D. Follow the trail into the gulf to view the transition from Whetstone Gulf formation to Pulaski formation.
E. Follow the Rim Trail to the head of the gulf to view thick sandstone beds on the lower Pulaski formation.

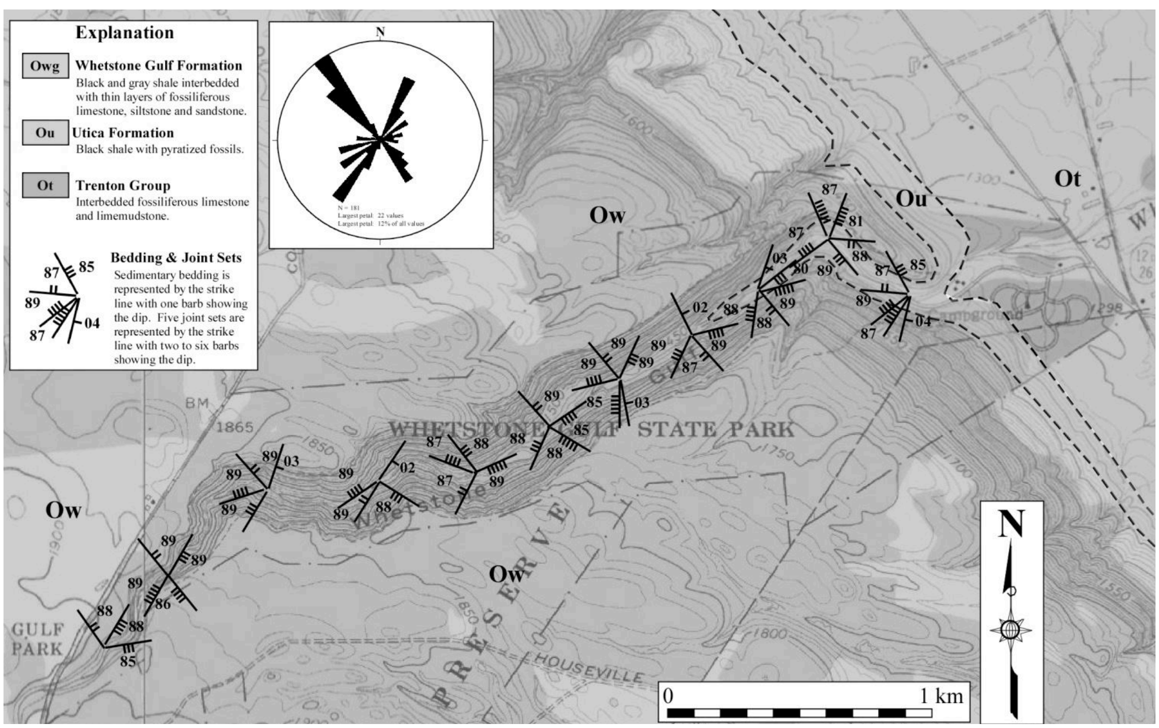


Figure 23. Geologic map of Whetstone Gulf, Lewis County, NY, with emphasis on joints. Mapping completed by A.P. O'Hara, 2009.

Road log:

139.5-139.9 miles on the park road back to the entrance. Turn right onto Co. Rt. 29.
139.9-140.0 miles on Co. Rt. 29 to Co. Rt. 26, turn right.
140.0-148.0 miles to Route 12D, proceed straight.
148.0-156.6 miles to Route 46 (Gorge Road), proceed straight.
156.6-171.2 miles on Route 46 to Wells Creek and Stop 7.

Stop 7: Wells Creek.

Walk along the road that parallels Wells Creek to view excellent exposures of the Utica black shale that directly overlies the Trenton limestone. The limestone is exposed in the bed of the creek.

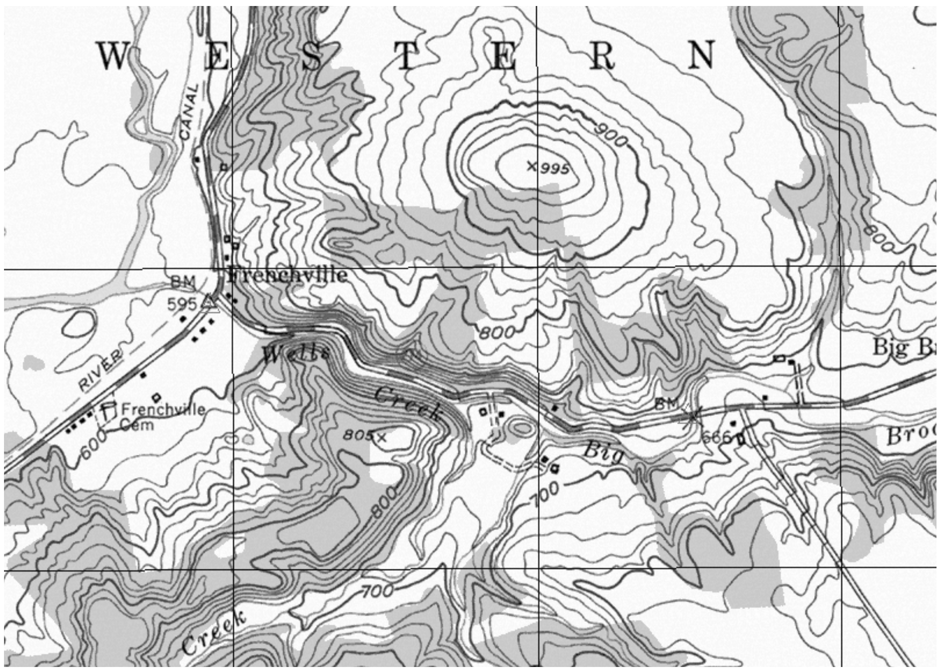


Figure 24 – Topographic map for the Wells Creek area, the location of Stop 7.

End of trip.

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Trip A-2

THE MARCELLUS SUBGROUP IN ITS TYPE AREA, FINGER LAKES AREA OF NEW YORK, AND BEYOND

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INTRODUCTION

The Marcellus "shale" was long one of the neglected "ugly ducklings" of New York stratigraphy that was known mainly for its black shale facies, the cephalopod-rich Cherry Valley Limestone, its potential as a source rock, and occasional interesting fossil finds like the famous *Devonaster* starfish trove found in upper Marcellus sandstones of the Hudson Valley. Generally, it was seldom the focus of much attention, however, during the last five years Marcellus strata have been the center of scientific and public attention, both as a regional symbol of economic opportunity and of controversy regarding exploitation of hydrocarbons.

The Marcellus in New York State is considered a subgroup of the Hamilton Group, which includes two different major lithofacies suites: the classic thin, black shale/dark gray mudstone succession, with minor carbonates developed in central to western New York; and the thick, coeval synorogenic siliciclastic succession of basinal to marine shoreface and terrestrial deposits in eastern New York. Beginning in underlying upper Onondaga strata, the succession represents two major ("third order") sea level cycles, which were apparently deposited over a 2-4 million year period. Total Marcellus thickness across the state ranges from less than 7.5 meters in the west near Buffalo to over 580 m in the Hudson Valley region (Rickard, 1989).

Herein we assemble a broad set of perspectives and knowledge on the strata of the Marcellus subgroup in the northern Appalachian Basin. Owing to the complexity of this topic the reader will find the paper divided into four chapters, written by various co-authors, as well the road log and brief stop descriptions at the end.

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CHAPTER 1. THE MARCELLUS SUBGROUP OF NEW YORK AND BEYOND:

STRATIGRAPHY, SEQUENCES, MUDROCKS, AND SEDIMENTOLOGY

(Chuck Ver Straeten, Gordon Baird, Carl Brett and Jeff Over)

Introduction

Geological Setting. The strata termed "Marcellus" (Hall, 1839) were deposited in New York and across much of what is termed the "Appalachian Basin" during the early Middle Devonian (late Eifelian to early Givetian stages; Figure 1-1). At that time, the Appalachian Basin was on the order of 30° south of the equator (van der Voo, 1983; Witzke, 1990; Scotese and McKerrow, 1990), and formed a retroarc foreland basin system adjacent to a tectonically-active mountain belt (Acadian Orogeny). The Marcellus was

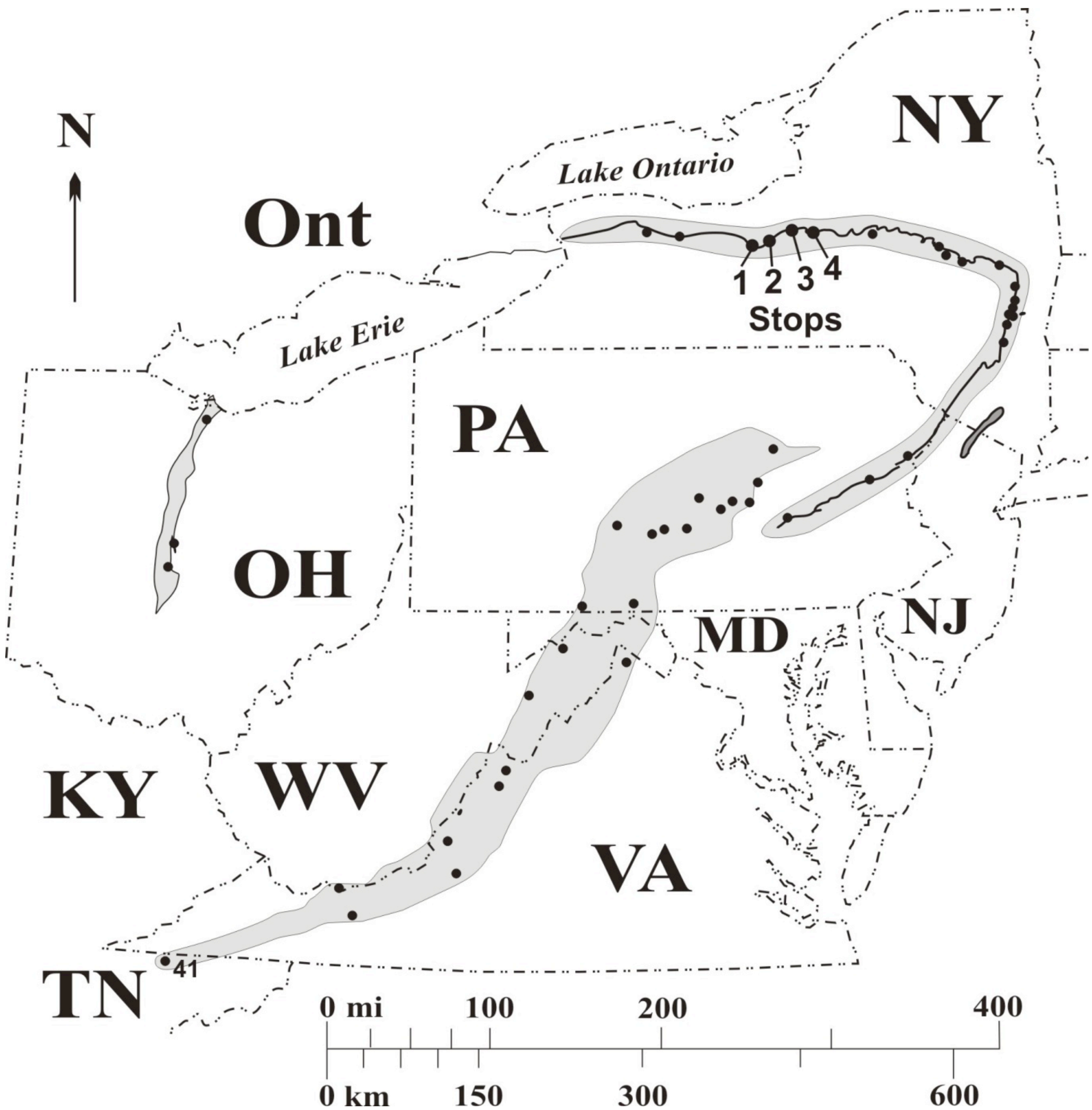

Figure 1-1. Outcrop map of the Marcellus subgroup of New York, and correlative strata in the Appalachian Basin, U.S. Marcellus rocks are found in the subsurface in between. Dots mark significant outcrops. In New York to eastern Pennsylvania, Marcellus outcrops occur in a narrow outcrop belt (dark line). In the Valley and Ridge (central PA, MD, VA, WV, TN), Marcellus strata outcrop in multiple belts within shaded area. In central Ohio, the shaded area represents the area of outcrop.

deposited over a period of what may comprise approximately two to four million years, longer than estimated by Kaufmann (2006), during the late stages of a Mid-Paleozoic greenhouse climatic event.

Synorogenic siliciclastics, organic-rich shales, and correlative carbonates were deposited during the bulk of two separate major ("third order") depositional sequences corresponding to two transgressive-regressive (T-R) cycles. Synorogenic sedimentation and flexure of the foreland basin system were the result of renewed Middle Devonian mountain building on the eastern margin of Laurentia (North America), during a third stage of the Acadian Orogeny. Paleontologically, lower and upper Marcellus strata feature two separate, very distinct faunas ("Stony Hollow" and "Hamilton" faunas), with very few species in common.

Marcellus-age strata in the Appalachian Basin can be divided into three major lithologic facies associations: the classic thin, organic carbon (OC)-rich shale and mudrock association; a thick, synorogenic clastic-dominated association; and a carbonate-dominated association largely on the cratonward margin of the foreland basin.

The term Marcellus is used in various ways in different areas of the Appalachian Basin, by different groups of geologists. In New York, it has long been considered a chronostratigraphic unit, bounded by time-specific units at its base and top. In other parts of the basin, it is more commonly treated as a lithostratigraphic unit, represented by black shales and dark gray mudstones, with little or no reference to age; or is locally considered part of another, more time-rich unit (e.g., Millboro Formation, in part, in the southern Appalachian Basin region).

Since the mid-1990s, New York Devonian researchers have utilized the term "Marcellus" as a subgroup of the Hamilton Group (Ver Straeten et al., 1994; Ver Straeten and Brett, 2006; and numerous other publications). The "Marcellus subgroup" comprises three formations (Union Springs Formation below, and coeval Oatka Creek and Mount Marion formations above), and several members (Figures 1-2, 1-3).

The early Middle Devonian Marcellus subgroup falls within the late Eifelian and early Givetian stages (*costatus* to *hemiansatus* conodont zones). Over et al. (Chapter 4 below) report that the Eifelian-Givetian stage boundary occurs in the lower part of the Oatka Creek-Mount Marion formations at or closely above the Cherry Valley Limestone. An altered volcanic airfall tephra bed a short distance below the Onondaga-Marcellus contact (Tioga B K-bentonite) has been dated at 390.5 +/- 0.5 Ma (Roden et al., 1990).

<u>The Acadian Orogeny.</u> The Late Silurian to Early Carboniferous Acadian Orogeny was the second of three Paleozoic-age mountain-building events in eastern North America, the result of continent-continent type collisional tectonics (Figure 1-4). The Acadian Orogeny (Rodgers, 1967; Ettensohn, 1985a; Osberg et al., 1989; Roy and Skehan,1993; Rast and Skehan, 1993; and Ver Straeten; 2009, 2010; alternatively, Acadian and Neo-Acadian orogenies of van Staal et al., 2009) is continuous through time with Silurian orogenesis from East Greenland to maritime Canada (Caledonian Orogeny), and is as a whole interpreted to have resulted from oblique collision of eastern North America (Laurentia) with one or more landmasses (e.g., Avalon, Rast and Skehan, 1993; Avalon, Meguma and Carolina terranes, van Staal et al., 2009; Hatcher, 2010; Hibbard et al., 2010).

Acadian tectonics resulted in the formation of an elongate mountain chain that extended from Newfoundland to Alabama. Recently dated igneous rocks, including altered airfall volcanic ashes ("tephras") within sedimentary rocks from Maine and adjacent areas indicate a Late Silurian beginning in New England (Bradley et al., 2000). The Acadian Orogeny is characterized by significant plutonic/volcanic activity, regional metamorphism, and large-scale deformation along the orogen.

Between the Late Silurian and earliest Late Devonian (ca. 40 m.y.), the Acadian deformation front migrated over 240 km cratonward (non-palinspastic distance) from coastal Maine into Quebec, accompanied by analogous migration of the adjacent, cratonward Acadian foreland basin (Bradley et al. 2000). A similar time-transgressive evolution of the paired Acadian hinterland- foreland basin system would have similarly advanced cratonward across central to southern New England and its extension into New York State and southern Ontario.

<u>Foreland Basins.</u> A foreland basin forms as an elongate trough, or "moat," on continental crust between an orogenic belt and the adjacent craton (Dickinson, 1974; Miall, 1995; DeCelles and Giles, 1996). Basins form and flex due to orogenic loading and the influence of other factors (e.g., Beaumont, 1981; Jordan , 1981; Quinlan and Beaumont, 1984; Flemings and Jordan, 1989, 1990; and references in Ver Straeten, 2010).

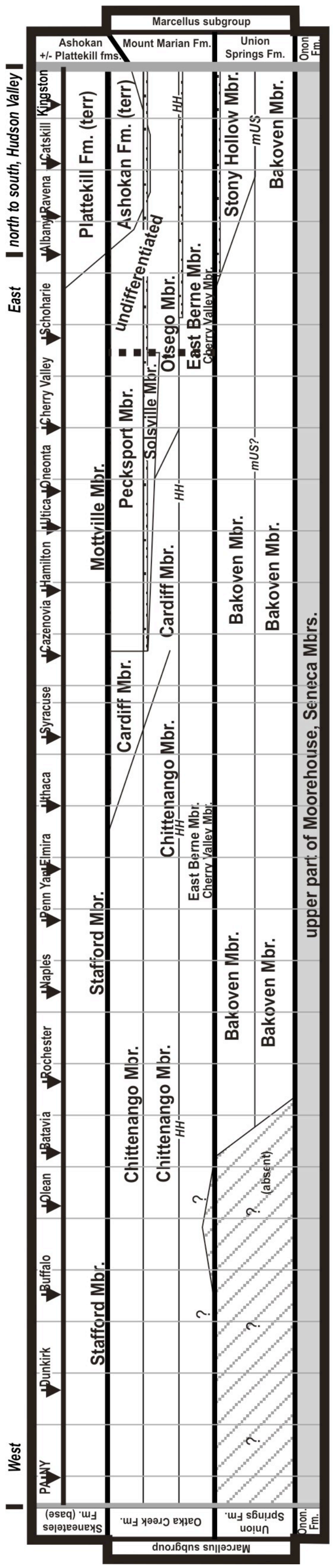

Figure 1-2

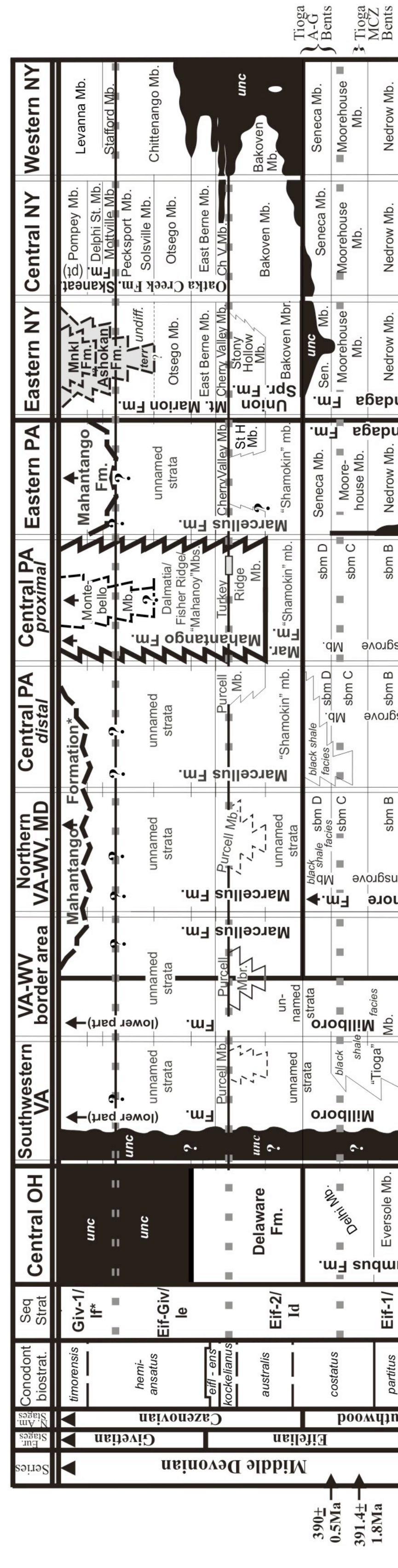

Figure 1-3

Figure 1-2. Marcellus subgroup stratigraphy in New York, west to east and through the Hudson Valley. Shaded area at bottom = upper Onondaga Formation, in lower part of Depositional Sequence Eif-2/Id of text. In the east, the Ashokan and Plattekill represent non-red and red terrestrial strata, respectively. Thick vertical dotted line west of latitude of Schoharie = boundary of coeval Oatka Creek and Mount Marion formations.

Figure 1-3. Basinwide outcrop stratigraphy of the Onondaga Formation, Marcellus subgroup and lower Skaneateles Formation and equivalents, Appalachian Basin. International conodont and goniatite zones shown, along with known range zones of key goniatites from the Appalachian Basin. Dashed lines bounding zones indicate position of biozone bases are unknown. Biostratigraphic data from House (1978, 1981), Becker & House (1994, 2000) and Klapper (1971, 1981). Black areas, unconformities. Geochronologic age dates from Roden et al. (1990) and Tucker et al. (1997). Abbreviations: bents., K-bentonites; *Cabrieroc*., *Cabrieroceras*; Fm., Formation; Mbr., Member; MCZ, Middle Coarse Zone; Mt Mar Fm, Mount Marion Formation; On., Onondaga Formation; sbm, submember. Modified after Ver Straeten (2007).

Foreland basin systems are comprised of four distinct depozones, from the orogenic front to the margin of the craton (DeCelles and Giles, 1996; Figure 1-5). These are termed, respectively, the wedge-top, foredeep, forebulge, and back-bulge basin, which as a set may migrate laterally over time as the basin evolves.

A *wedge-top* is situated over the front of the orogenic fold-and-thrust belt, where coarse-grained sediments are deposited, and commonly deformed. A *foredeep* is a subsiding trough cratonward of the wedge-top, characterized by a thick (ca. 2-8 km-thick) succession of dominantly synorogenic sediments that thin distally. This area, the focus of many foreland basin studies, is generally on the order of ~100–300 km wide. A *forebulge*, a zone of possible flexural uplift cratonward of the foredeep, may be on the order of 60–470 km wide. It is typically characterized by little to no siliciclastic deposition and/or carbonate deposition if flooded, as well as possible periods of erosion, and consequent stacked unconformities. A *back-bulge basin*, near the cratonic margin of the foreland basin system, is characterized by subsidence on a smaller scale than that in the foredeep. If flooded by marine waters, thin, tabular sediments may consist of in situ-formed carbonates, or sediments from the orogenic belt, craton, or forebulge (descriptions after DeCelles and Giles, 1996).

<u>The Appalachian Basin and the Acadian Foreland Basin.</u> The greater Appalachian Foreland Basin was an elongate trough that formed in the early Paleozoic, adjacent to the Appalachian -Taconic Orogen. The basin extended from Newfoundland to Alabama, on the cratonward side of the mountain belt. Its origin during the Middle Ordovician marks the transformation of a coupled passive margin-adjacent epicontinental sea to an active foreland basin system as a result of the collision of eastern Laurentia with a volcanic island arc (Taconic Orogeny). Two subsequent, major continent-continent collisions (Late Silurian to Early Carboniferous Acadian Orogeny, Figure 1-4; and Late Carboniferous to Permian Alleghenian Orogeny) resulted in reactivation and reorganization of the basin into an active foreland basin system. Between orogenies, erosional unloading of the mountain belt led to crustal relaxation, uplift and erosion of proximal portions of the basin, as well as cratonward seaway migration. As the Acadian Orogeny evolved through multiple phases of uplift to relative quiescence through time (e.g., Ettensohn, 1985a), the associated Acadian foreland basin system flexed, changed sedimentation regimes, and underwent overall cratonward migration (Ver Straeten, 2010).

The area commonly termed the “Appalachian Basin” (Figure 1-1) is an area of slightly deformed to undeformed sedimentary rocks preserved in parts of nine states (NY, NJ, PA, MD, VA, WV, TN, KY and OH) and southern Ontario. It is a subset of the greater Appalachian (or “Acadian”) foreland basin. Portions of the greater foreland basin fill from Newfoundland to Alabama (e.g., central to western New England) are missing due to syn- to post-orogenic weathering and erosion, or preserved in some areas as highly altered metasedimentary rocks in the orogenic belt.

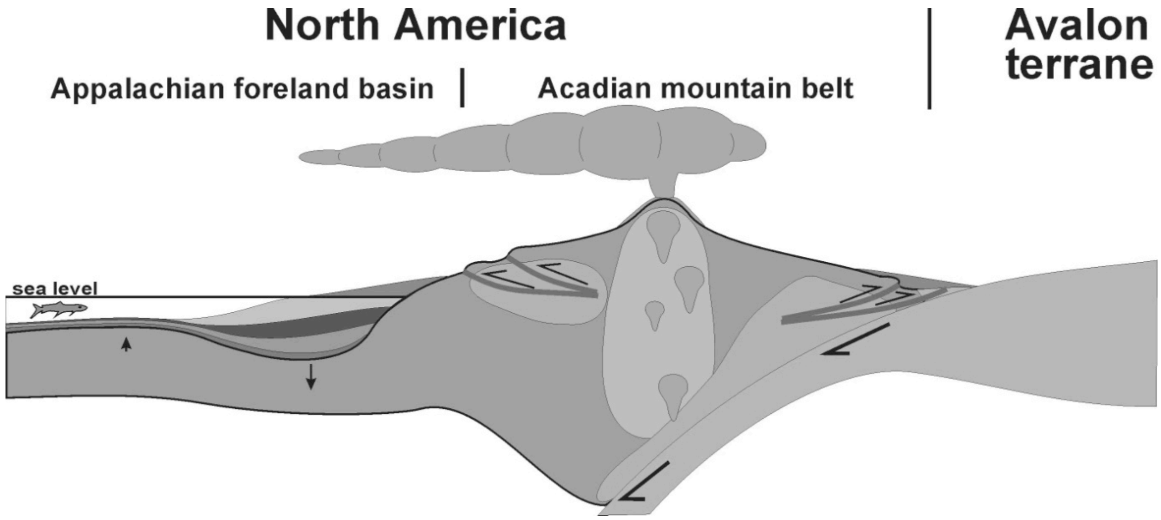

Figure 1-4. Idealized cross-section of Acadian orogen, and Acadian retroarc foreland basin system across central New England and New York into Ontario.

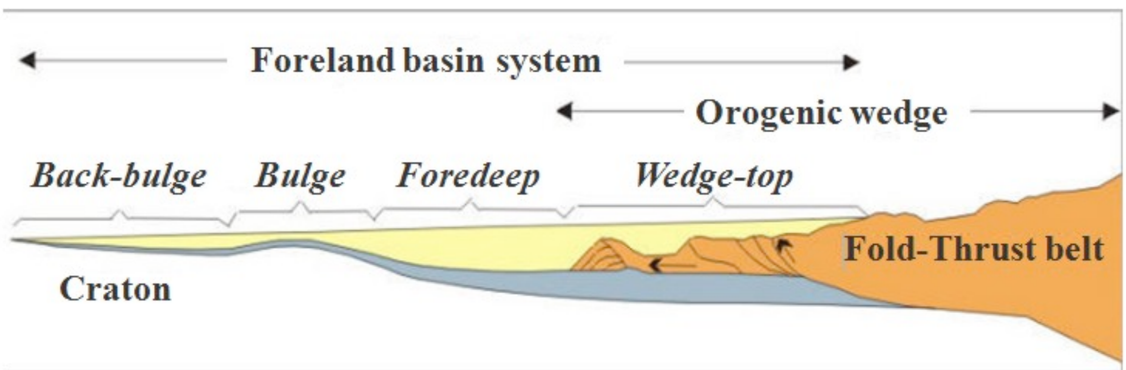

Figure 1-5. Idealized cross-section of an orogenic belt and adjacent foreland basin system. Modified after DeCelles and Giles (1996).

GEOLOGIC PERSPECTIVES

Sequence Stratigraphy

The concepts of sequence stratigraphy provide a powerful tool for the analysis of sedimentary rock successions (Wilgus et al. 1988; Van Wagoner et al. 1988; Emery and Meyers 1996). Sequence stratigraphy permits a chronostratigraphic subdivision of the rock record into cyclic, unconformity-bound, genetically related successions of strata at several scales (Van Wagoner et al. 1988), controlled by changes in relative sea level/base level. A "depositional sequence," the fundamental unit of sequence stratigraphy, is formed by a cyclic change in base level (relative sea-level) through the interaction of tectonics, eustatic sea-level change, and/or sedimentologic factors. Sequences can be subdivided into "systems tracts," composed of smaller-scale cycles, deposited during different stages of a transgressive-regressive cycle.

Two basic sequence stratigraphic models are currently in wide use, the "Exxon" (e.g., Mitchum et al. 1977; Van Wagoner et al. 1988; Catuneanu 2002) and "T-R Cycle" models (e.g., Embry, 1995, 2010). In Chapter 1 of this paper, Marcellus depositional sequences will be discussed in the context of the Exxon model, applied to outcrop and core studies. In Chapter 2, Lash and Blood will discuss Marcellus depositional sequences in the context of the T-R cycle model, as applied to subsurface data of the Marcellus.

Whichever model is applied, two elemental questions lie at the core of sequence stratigraphy. Where in the succession is the shallowest point? And where is the deepest point? No matter the concepts used to address the rest of a sequence/cycle (e.g., Exxon versus T-R cycle models), these are the cardinal positions to be clearly delineated.

In the Exxon model, four subdivisions are recognized within a sequence (lowstand, transgressive, highstand, and falling stage systems tracts; following Catuneanu, 2002). The base of a sequence is placed at the base of the lowstand systems tract, or at a subaerial unconformity and correlative conformity.

A "Lowstand Systems Tract" (LST) forms above a subaerial unconformity or a correlative marine conformity during an initial rise of base level and records sediment progradation (normal regression) during slow sea level rise; it may or may not be preserved at the base of a sequence (Catuneanu, 2002). A succeeding "Transgressive Systems Tract" (TST) is deposited during the middle stage of a relative sea-level rise, as marine waters flood over previously exposed lands. In siliciclastic-dominated systems, this leads to thin sediment-starved, deposits; in carbonate systems, in situ sediment production may continue, and form thick successions. A TST culminates in a "surface of maximum flooding," when sea level reaches its maximum.

A "Highstand Systems Tract" (HST) forms above the maximum flooding surface as the rate of sea level rise slows, depositing thin aggradational to progradational strata. A sequence concludes with a "Falling Stage Systems Tract" (FSST), deposited during a relative sea level fall. Progradation and deposition of offlapping stratal packages characterize the FSST in the basin, with formation of a subaerial unconformity above sea level.

Black shales, commonly thought to form during the late TST alone, actually form through the late TST to early HST. Detailed analyses of Devonian mudrock sequences (Ver Straeten et al., 2011) indicate that the maximum flooding surface occurs above basal TST deposits (e.g., limestones), within overlying black shales.

Muddy Sediments and Mudrocks: Recent Perspectives

Introduction. Since the publication of Potter, Maynard and Pryor (1980) "Sedimentology of Shale," studies of fine-grained siliciclastic sediments and rocks have revolutionized the understanding of the character and processes associated with these relatively neglected strata. Through higher resolution analyses and new analytical approaches to muds and mudrocks in modern and ancient settings, geologists now have a more actualistic perspective on physical transport and sedimentation, and the physical, biological and chemical processes operative during mud transport and deposition to final burial, and subsequent diagenesis.

Numerous important studies, directed to the voluminous Devonian mudrocks in New York and eastern North America, have played a part in these recent advances. Units examined include the classic New York Devonian black shales: Marcellus, Geneseo, Middlesex, Rhinestreet, Pipe Creek, and Dunkirk; and dark gray to gray mudstones such as various Hamilton Group strata, and many Upper Devonian units such as the Penn Yan, Cashaqua and Hanover, as well as mudstones, shales, and paleosols from Devonian terrestrial environments in the Catskills of eastern New York.

Mud deposition in marine shelf/epicontinental sea/foreland basin settings. The greater focus on mud and mudrock sedimentology in recent years indicates that mud erosion, transport and deposition in marine environments is more complex than previously thought, based on older, long-held assumptions (e.g., Hjulstrom's 1955 predictions of current velocities needed to erode cohesive muds; long distance transportation and settling out of suspended discrete clay grains). Many more factors than were formerly realized affect muds in marine environments, including clay mineral flocculation, water mass density (salinity, temperature), water content of muddy sediments, bioturbation, sea floor "armoring" (by biomats, sea grass, shell deposits, etc.), volume of clay sediments present, type of clay-rich sediment particles (discrete grains, floccules, fecal pellets, organic matter-clay agglomerates), formation of mixed clay-water slurries of various densities, low density suspended-clay nepheloid layers, and physical processes active in the environment. An overview of some of these factors is given in Schieber (1998a) and Macquaker and Bohacs (2007).

Macquaker and Bohacs (2007, p. 1735) state "…many of our preconceptions about fine-grained rocks are naïve," applies as well to muddy sediments. In contrast with commonly held assumptions, clays in marine environments are rarely transported as discrete particles, to settle out across broad regions. For example, flume studies by Schieber et al. (2007) found that flocculated clays are transported at the same velocities as sand grains, and may be deposited in ripples and low angle cross-sets that would not be visible after compaction.

Modern oceanographic studies indicate that most clay particles settle out close to their input center (e.g., river delta), due to flocculation of clays into larger aggregate clay grains on contact with saline waters (Syvitski, 1991; Hill et al. 2000). Floccules rapidly settle out into local shallow marine shelf/ramp/pro-delta environments (Milligan et al. 2007; Hill et al., 2007). Additional processes also may combine discrete clay grains into larger particles (e.g., fecal pellets, and organic matter-clay agglomerate or inorganic aggregate grains; Syvitski, 1991), increasing the speed of clay sedimentation.

Subsequent transport of muddy sediments occurs chiefly through resuspension and transport by storm events, bottom currents, and bottom-hugging density flows of mixed mud and water ("density underflows", "fluid mud flows", or "marine hyperpycnal flows"; Mulder et al., 2003; Hill et al., 2007; Sommerfield et al, 2007). In areas of steeper slopes (>7^o), these latter basinward-directed flows may be driven by gravity alone (Oggston and Sternberg, 1999; Oggston et al., 2000); on lesser slopes (<3-4^o), bottom-hugging mud flows need to be initiated and/or driven by other processes, such as waves or tides, combined-flow storm currents, sediment-laden fluvial flood events, or dilution of saline marine waters by fresh water during long duration flood events, etc. (Mulder et al., 2003; Hill et al., 2007; Sommerfield et al, 2007; Macquaker et al. 2011). Individual flows are deflected by the Coriolis effect, or by complexities of circulation on the shelf/ramp, and do not generally reach far basinward – therefore, muds may require multiple, discrete transport events to be distributed distally, and not reach more distant portions of a basin.

Another, albeit less significant mechanism of clay transport, is via low density suspended clay "nepheloid layers" (Biscaye and Eittreim, 1977; McCave, 1984, 2001; Ransom et al., 1998). These contribute smaller amounts of fine clay sediments to the sea floor, but transport them further basinward, and scatter light in the lower part of the water column, decreasing the depth of the photic zone.

To summarize, current understanding of mud transport and deposition indicates that distal, deeper portions of basins may remain largely starved of fine-grained siliciclastics for extensive periods of time, until episodes of forced regression during sea level fall enable sediment to prograde into such areas, far from siliciclastic sources. Most muds settle in proximal areas to their source, in offshore, lower energy settings. However, when rates of mud sedimentation are high enough, muds may settle and largely remain in shallow, higher energy environments, especially down-shore of major mud input sources (Rine and Ginsburg, 1985).

THE MARCELLUS SUBGROUP OF NEW YORK

What's in a Name? Marcellus Stratigraphy

Marcellus Stratigraphy. The Marcellus Shale was named by James Hall in 1839 for exposures around Slate Hill in the town of Marcellus, Onondaga County, NY (Stop 3). The name has its origins in classical Roman history, as do many in upstate New York; the town itself was named around 1794, after Marcus Claudius Marcellus, a Roman general. Since the work of Cooper (1930a, b, 1933, 1941) the term Marcellus in New York has been treated as a chronostratigraphic unit, comprising nearly all strata between the top of the Onondaga Limestone and base of the Skaneateles Formation. The exception to this has been uppermost Marcellus-age terrestrial strata in eastern New York.

Based on several factors, the Marcellus in New York was redefined in the mid-1990s, and assigned the status of "Marcellus subgroup" within the Middle Devonian Hamilton Group (Ver Straeten et al., 1994; Ver Straeten and Brett, 2006; Figure 1-2). Reasons for this change include: 1) the Marcellus comprises two major depositional sequences; each of the other three New York Hamilton Group formations (Skaneateles, Ludlowville, and Moscow) comprise a single sequence; 2) the two Marcellus divisions are markedly different in proximal facies of eastern New York, where they are both of great thickness (up to ca. 170 and 410 meters, respectively; Rickard, 1989); and 3) the presence of two completely distinct lower and upper faunas ("Stony Hollow" and "Hamilton" faunas; Brett and Baird, 1995; DeSantis et al., 2007). If previous researchers (e.g., G.A. Cooper in the 1930s) had begun lower Hamilton studies in eastern New York, not in the highly condensed facies of central to western New York, they would likely have concluded that the Marcellus comprised two separate formations. Note that the term "subgroup" is not capitalized; hence "Marcellus subgroup".

The Marcellus subgroup of New York State is subdivided into three formation-level units (Ver Straeten et al., 1994; Ver Straeten and Brett, 2006; Ver Straeten 2007). In this stratigraphic framework, lower Marcellus strata (of Depositional Sequence Id, also known as Sequence Eif-2) are assigned to the Union Springs Formation. The succeeding black to dark gray mudrock-dominated upper Marcellus succession in central to western New York is assigned to the Oatka Creek Formation (Sequence Ie/Eif-Giv); in eastern New York, thick, time-equivalent siliciclastic-dominated basinal to shoreface marine facies of the upper Marcellus are termed Mount Marion Formation. The boundary between the lower and upper formations of the Marcellus subgroup is placed at the base of the Hurley Member, a generally thin but correlatable unit throughout the Appalachian Basin.

Over the last 15 years, this stratigraphy has been widely recognized, it is utilized by most non-industry researchers working on the Marcellus time slice interval in New York, regionally, and internationally. This stratigraphy is outlined in Figure 1-2; its relationship to other Marcellus-age strata around the Appalachian Basin, based on study of more than 300 outcrops of Marcellus and Marcellus-equivalent rocks basinwide by Ver Straeten, is shown in Figures 1-3 and 1-6.

Only two members are recognized in the Union Springs Formation in New York (Figure 1-2): black to dark gray shale facies of the Bakoven Member, and calcareous, generally buff-colored calcareous mudstones, siltstones, and sandstones of the Stony Hollow Member. The Stony Hollow laterally replaces upper Bakoven black shale facies above a widespread marker bed ("mid-Union Springs K-bentonite"). It represents more proximal mid-ramp, neritic facies that occur in the Helderbergs-Hudson Valley to Port Jervis outcrop belt, and into eastern Pennsylvania.

There are a greater number of members in upper Marcellus strata across New York (coeval Oatka Creek-Mount Marion formations; Figure 1-2). This reflects a more complex, heterogenous lithologic suite, ranging from basinal black shales to shoreface sandstones, and even thin conglomerates. The upper Marcellus terrestrial facies which consist of channel sandstones, dark gray wetland mudstones and non-red paleosols in the Helderbergs and Hudson Valley have been assigned to the lower part of the Ashokan Formation.

Recently, in a paper focused on subsurface Marcellus strata, Lash and Engelder (2011) discussed Marcellus stratigraphy in the northern and central Appalachian basin (New York, Pennsylvania). They presented their understanding of the established stratigraphy of Marcellus strata as published by Ver Straeten and Brett (2006) and Ver Straeten (2007). In that section, they retained the term Marcellus as a formation-level unit; and proposed using a simplified tripartite division of Marcellus strata for subsurface analyses. However, Lash and Engelder's (2011) stratigraphy is an inaccurate portrayal of Marcellus strata in New York and Pennsylvania, as shown in herein (Figure 1-7).

A resulting problem in Lash and Engelder's (2011) paper is their proposed application of the term "Cherry Valley" to any and all calcareous facies in the middle of the Marcellus in the subsurface. In more distal parts of the basin, this represents the Hurley and Cherry Valley members at the base of the Oatka Creek Formation and equivalents. However, in increasingly proximal well logs, Lash and Engelder (2011) also apply the term Cherry Valley to sub-Hurley and Cherry Valley strata of the Stony Hollow Member. Applying the term "Cherry Valley" to any and all mid-Marcellus calcareous strata through the entire interval in this way reverts to an older pre-1930s definition; but, problematically, it creates a second, distinct definition of the Cherry Valley and hence, does not follow the North American and International stratigraphic codes (North American Commission on Stratigraphic Nomenclature, 1983; Salvador, 1994). In addition, it confuses the stratigraphy, and will confuse workers in the present and future.

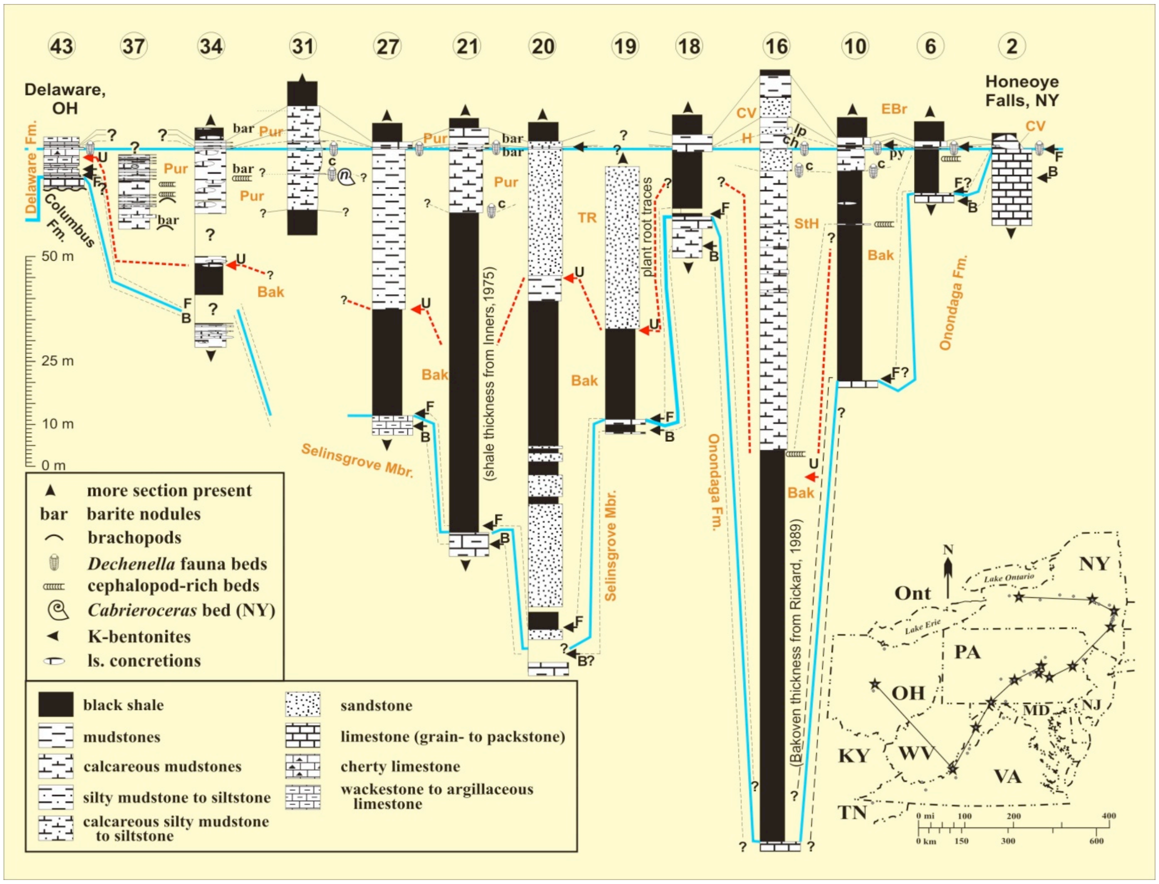

Figure 1-6. Basinwide correlations of the Marcellus subgroup and correlative strata, Appalachian Basin, New York to Virginia and Ohio. Dashed and solid lines show correlations. Bold solid lines separate New York formations and correlative horizons across basin. Non-bold solid lines separate member-level units of the Union Springs and Oatka Creek–Mount Marion Formations and correlative horizons. Datum = base of Hurley Member (H) of the Oatka Creek–Mount Marion Formations and equivalent position basinwide. Abbreviations: B, Tioga B K-bentonite, at base of Seneca Member (Onondaga Fm) in New York; Bak, Bakoven Member of the Union Springs Formation; c, base of fine sandstone cap of Stony Hollow Member of the Union Springs Formation and correlative position; ch, Chestnut Street submember of the Hurley Member; CV, Cherry Valley Member of the Oatka Creek-Mount Marion formations; EBr, East Berne Member of the Oatka Creek–Mount Marion Formations; F, Tioga F K-bentonite, just above base of Union Springs Formation in New York; H, Hurley Member of the Oatka Creek–Mount Marion formations; lp, Lincoln Park submember of the Hurley Member; StH, Stony Hollow Member; TR, Turkey Ridge Member of the Mahantango Formation; U, mid-Union Springs K-bentonite. Strata at Delaware, OH, are assigned to the Delaware Formation. Undifferentiated calcareous shales to limestones in VA and WV and central PA, are assigned to the Purcell Member of the 'Marcellus Formation' or Millboro Formation. Modified after Ver Straeten (2007).

In the central to southern parts of the Appalachian Basin, the term Purcell Member (Cate, 1963) has long been applied to undifferentiated calcareous mid-Marcellus strata. Ver Straeten (1996a, b, 2007; Figures 1-3, 1-6, 1-7) showed that in distal areas of the basin the Purcell is time-correlative with the Hurley and Cherry Valley members and, in intermediate areas, with the Hurley and Cherry Valley and part to all of the underlying Stony Hollow Member of New York. This is exactly what Lash and Engelder (2011) wanted – a term for any and all undifferentiated mid-Marcellus calcareous strata. Therefore, to avert the confusion of multiple definitions, to conform to the U.S. and International Stratigraphic Codes, and to use a familiar well established name for the same strata, it is proposed that the term "Purcell Member" also be applied to undifferentiated calcareous mid-Marcellus strata in the subsurface of the Appalachian Basin, as is done in outcrop.

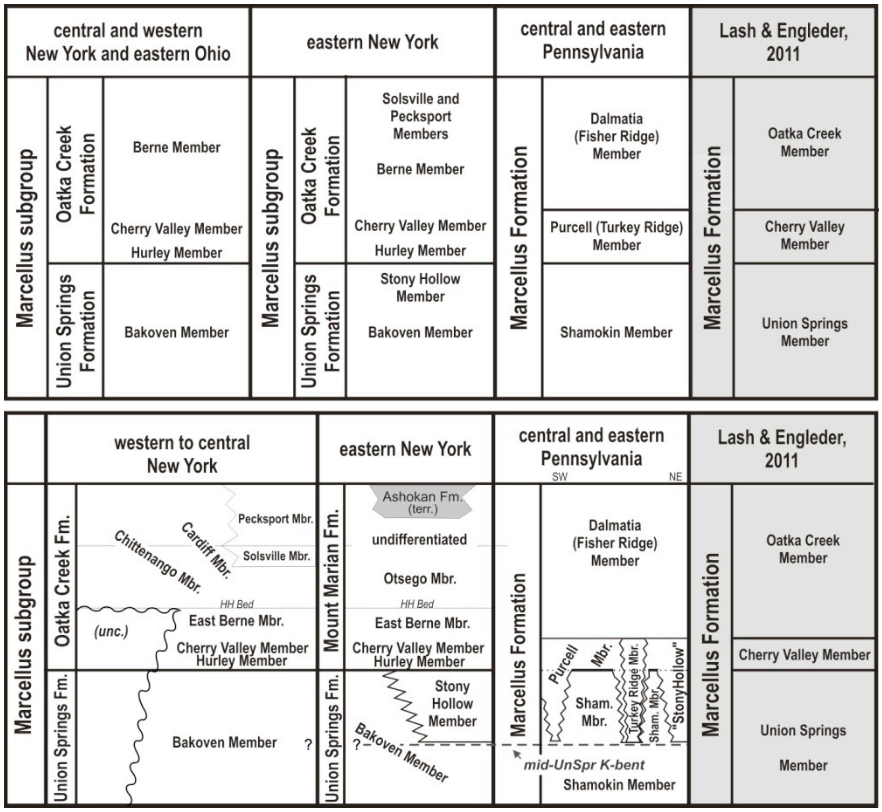

Figure 1-7. a: Comparison of Marcellus outcrop stratigraphy in New York and Pennsylvania as portrayed by Lash and Engelder (2011); b) Marcellus stratigraphy as outlined by Ver Straeten (2007), and for post-East Berne strata above. Note complex time-correlative relationships of strata, especially in central Pennsylvania.

The Character and Sedimentology of Marcellus Black Shale Facies

One of the key aspects of Marcellus strata, of great interest currently, is the organic-rich mudrock facies. The classic Marcellus black shales, developed in both the Union Springs and Oatka Creek formations, are characterized by high weight percent TOC (from 1 to >17%), dark color (generally N1-N2), well laminated sediment fabrics (typical bioturbation index of 1, on a scale of 1-7), dominantly fine-grained (clay-size) siliciclastic detritus (high weight % Al) with little silt to sand-size terrigenous (fluvial-derived) detritus (low Ti/Al ratios), moderate Si/Al ratios (related to eolian quartz silt), high Mo, and generally low although varying $CaCO_3$ content (sometimes none).

The concentration of TOC in marine strata reflects the varying interactions of multiple factors, including sediment condensation, circulation changes, anoxia, increased nutrient availability and primary production, and burial efficiency (Arthur and Sageman, 2005). Sea level rise, and its concomitant effects, is often an important factor in TOC enrichment on continental shelves and in epicontinental seas (Arthur and Sageman, 1994). Such potential effects include: 1) relative starvation of siliciclastic and carbonate sediment components, allowing for greater enrichment of TOC; 2) increased nutrient availability and productivity, including from recycling/remineralization; 3) increased anoxia +/- euxinic conditions, potentially affected by increased water depth and/or increased productivity, and affecting the preservation potential of TOC; and 4) weathering and erosion of orogenic highlands, may also lead to greater nutrient input, driving increased productivity.

Marcellus sedimentation was affected by relative sea level change at several scales, associated with both eustatic and tectonic-loading processes. During overall middle Lower to Upper Devonian sea level rise (Johnson et al., 1985), Marcellus strata were deposited during a time of more significant global eustatic sea level rise (above background trends) and of a pulse of increased subsidence of the foreland basin system (onset of a new tectonically active phase of the Acadian orogeny; Ettensohn, 1985a; Ver Straeten and Brett, 1995; Ver Straeten 2009, 2010). In addition, trends at the level of third to fourth order cyclicity (as well as finer scales) also affected the burial of TOC, with greater concentrations surrounding maximum flooding surfaces during late TST-early HST.

In distal, deeper areas of the foreland basin (e.g., west-central NY), much of the non-limestone portion of the Union Springs and Oatka Creek formations are comprised of organic-rich shales. In more proximal parts of the basin (e.g., east-central NY approaching the Helderbergs and Hudson Valley) or the far margin of the basin (e.g., central Ohio) only the deeper portions of the 4th order or finer-scale cycles preserve organic-rich facies. In the Helderbergs and Hudson Valley, black shale facies are only preserved in the deeper portion of the third order Id (Eif-2) and Ie (Eif-Giv) sequences; overlying parts of both sequences (Union Springs and Mount Marion formations) in those areas are characterized by thick, shallower progradational siliciclastic facies.

A series of papers published over roughly the last decade analyzed various sedimentologic, paleontologic and geochemical characters through 600 m of Middle to Upper Devonian mudrock-dominated facies from western New York. The succession, studied continuously through two overlapping cores, includes Eifelian- to Famennian-age strata from the Onondaga to Gowanda formations. Papers variously focused on specific organic-rich intervals (e.g., Geneseo Formation, Murphy et al., 2000a, b; Oatka Creek Formation, Werne et al., 2002); processes involved in OC enrichment and burial (Sageman et al., 2003); and sequence stratigraphic development and delineation in different mudrock facies (Ver Straeten et al., 2011).

The Oatka Creek Formation was analyzed in depth by Werne et al. (2002). Based on multiple lines of evidence (lithologic, paleontologic and geochemical), they concluded that different factors were involved in deposition of black Oatka Creek shales, but ultimately it was controlled by relative sea level rise (dominantly eustatic rise, with some component of tectonic-driven subsidence). Maximum OC concentrations in the succession coincided with a shift from anoxic (no oxygen in sediments) to anoxic and euxinic (sulfide-rich water column, with no oxygen) conditions. This correlated with the point of minimum fluvially-derived synorogenic siliciclastic input, and the highest concentration of eolian-transported quartz silt. Biogeochemical recycling of phosphate and nitrogen (P and N) from sediments led to enhanced levels of primary production in surface waters, which contributed to maintenance of euxinic conditions, even as terrigenous sediments prograded as far basinward as western New York in upper Oatka Creek strata.

Sageman et al. (2003) compared multi-proxy data sets through seven different Middle to Upper Devonian black shale units from the two western New York cores (Union Springs, Oatka Creek, Geneseo, Middlesex, Rhinestreet, Pipe Creek and Dunkirk formations). Based on various geochemical data, their findings included: 1) that organic carbon burial is interdependent on processes of sedimentation, primary production and microbial metabolism, in contrast to older arguments focused on "preservation versus production;" 2) that only the black shales of the Union Springs and Oatka Creek formations were largely deposited under both anoxic conditions and sulfidic (euxinic) water columns; and 3) that maximum organic richness in the Devonian black shales correlated with maximum siliciclastic and carbonate sediment starvation, during times associated with maximum transgression of the shoreline.

Sageman et al. (2003) concluded that three main controlling variables explained the origins of western New York Devonian black shales. These were: 1) Seasonal thermoclines, not long term stable pycnoclines as formerly interpreted – except in the case of the Marcellus subgroup black shales, which did have evidence for stable long-term anoxic and euxinic conditions; 2) nutrient supply, including remineralization of phosphate and nitrogen from sediments to the water column, which stimulated primary productivity; and 3) relative sea level, including both eustasy and tectonic-load related subsidence of the basin. They interpreted that water depths related to these two factors were at their maximum in western New York during Union Springs and Oatka Creek time. Because of greater depths, seasonal mixing only rarely penetrated bottom waters in the basin, and so anoxic and euxinic conditions were maintained over a long period of time.

Marcellus Paleontology and Paleoecology

The Marcellus subgroup was deposited during at least four conodont zones (Klapper, 1981; DeSantis et al., 2007; Over et al., Chapter 4 of this paper). The lower Union Springs (Bakoven Member) black shale probably represents the *australis* Zone, although it has not yielded diagnostic conodonts. The Hurley and Cherry Valley members fall within the *kockelianus* Zone, as indicated by assemblages from both the Chestnut Street Bed of the lower part of the Hurley Member and conodonts obtained at the top of the Cherry Valley Member (Klapper, 1981). The overlying East Berne Member is thought to have been deposited during the *ensensis* Zone near the end of the Eifelian, while the remainder of the Oatka Creek

Formation probably lies in the earliest Givetian *hemiansatus* Zone, although, again, conodonts are sparse and thus far the nominal species of this zone has not been found in eastern North America. Goniatites are also important in defining biostratigraphic zones during the Middle Devonian. The rich fauna of *Agoniatites* cf. *vanuxemi* in the Cherry Valley limestone indicates a late Eifelian age. Recent discovery of tornoceratids in beds just below the Dave Elliott Bed, toward the top of the East Berne Member in the Hudson Valley, indicates that the Eifelian-Givetian boundary lies near this level (Bartholomew et al., 2009; however, see Over et al., Chapter 4 of this paper).

Marcellus dark shale fossil assemblages (when present) are dominated by a pelagic fauna of small, conical styliolinids and dacryoconarids, straight and coiled nautiloid and goniatite cephalopods, and, on some bedding planes, low-oxygen adapted leiorhynchid brachiopods and small bivalves (Brett et al., 1991; Werne et al., 2002; Sageman et al., 2003; Boyer and Droser, 2007, 2009; Boyer et al., 2011; Ver Straeten et al., 2011). Shallower, more oxic biofacies of the Stony Hollow Member and the Mount Marion Formation in the Hudson Valley area are represented by much more diverse benthic assemblages, including diverse brachiopods, small corals, mollusks, echinoderms and trilobites (see Ver Straeten, 1994; Brett et al., 2007).

It should be noted that, in detail, the lower Marcellus fauna ("Stony Hollow Fauna;" from the upper Union Springs Formation and basal-most Oatka Creek-Mount Marion formations) is quite distinct from those of the upper Marcellus or the older Onondaga biofacies and is recognized as an Ecological Evolutionary Subunit (EESU) by Brett et al., (2009). A major bioevent, termed the Stony Hollow event (Brett et al., 2009; DeSantis and Brett, 2011) has been recognized within the Union Springs Formation and it appears to have caused local extermination of much of the diverse Onondaga fauna and its replacement by a low diversity, warm water associations including taxa derived from the tropical Old World Realm of present day Arctic Canada; (DeSantis and Brett, 2011). This unusual biota, typified by brachiopods such as *Variatrypa arctica* and the trilobite *Dechenella haldemanni* occurs widely over eastern North America during the late Eifelian time (Koch and Day, 1995).

Moreover, a major biotic change, also originally recognized as an EESU boundary by Brett and Baird (1995; corresponding to the global Kačák bioevents) appears to occur within the East Berne Member and close to the Eifelian-Givetian Stage boundary, as presently identified (Brett et al. 2009, Bartholomew et al., 2009; Over et al., Chapter 4 of this paper). This turnover led to a demise of the Stony Hollow fauna and the incursion of the long-lived Hamilton fauna, which appears high in the East Berne Member and close to the Eifelian-Givetian boundary.

These changes are associated with carbon isotopic evidence for major changes in the global carbon cycle; for example, a major positive excursion in $\delta^{13}C$ occurs in close proximity with each of the biotic overturns (Brett et al., 2009; DeSantis and Brett, 2011). Also, there are both biogeographic and some oxygen isotopic evidence for abrupt temperature changes at these times (Joachimski et al., 2004, van Geldern et al., 2006). Although these biotic changes are not the primary focus of the present paper, they do point to distinct environmental conditions during the deposition of the lower vs. upper Marcellus that require consideration. The Union Springs and Oatka Creek black shales are superficially similar, but they have unique geochemical and sedimentological features that may relate to these differences. For example, trace metal studies of Werne et al. (2002) indicate that the Union Springs Shale was deposited in lower dysoxic to anoxic conditions, while the Oatka Creek black shale records truly euxinic conditions. Hence, different environmental conditions, in terms of water circulation patterns, organic productivity, etc., must have pertained.

Marcellus Sequence Stratigraphy

As has been clearly shown in outcrop studies over the last 25+ years, strata of the Marcellus subgroup comprise two major, third order sequences (approximately equivalent to T-R Cycles Id and Ie as originally defined by Johnson et al., 1985, and subsequently refined by Brett and Ver Straeten, 1994; Ver Straeten and Brett, 1995; Brett and Baird, 1996; Ver Straeten, 2007; Brett et al., 2011; Ver Straeten et al., 2011). These same sequences have now also been delineated in the subsurface utilizing geophysical, sedimentologic, paleontologic, and geochemical data sets (Lash and Engelder, 2011; Ver Straeten et al., 2011).

For the purpose of retaining clarity between geologists across North America and globally, a naming scheme for Devonian sequences should be standardized, and used by all. Whether from academia or industry, , using a standardized, globally-utilized terminology for Devonian depositional sequences is important and useful. When interacting with peers from other basins on the same or other continents on the same-age strata, or if one's work shifts to another basin, a globally-standardized set of terms for sequences allow for immediate comparison, compared to unique names for sequences in from basin to basin. The use of local to regional names for the sequences (e.g., MSS1 and MSS2 of Lash and Engelder, 2011) will camouflage relationships and hinder communication about sequences beyond a small subset of individuals.

Currently, there are two international Devonian schemes for naming Devonian sequences. The first is the refined designations of Johnson et al. (1985; e.g., Marcellus Sequences Id and Ie); the second is based on which international Devonian stage a sequence occurs within (e.g., where Eif=Eifelian and Giv=Givetian, "Marcellus Sequences Eif-2 and Eif-Giv;" after Ver Straeten, 2007; and Brett et al., 2011). There are advantages to each; the first is already familiar and established. However, the second immediately tells you when in the Devonian a depositional sequence occurs. In this paper we will utilize both terms.

Numerous subdivisions (4th to 6th order cycles) are recognized within Sequences Id/Eif-2 and Ie/Eif-Giv in outcrop. Some of these, even at the smallest scale, can be correlated basinwide (Ver Straeten, 2007) and beyond. Three fourth order sequences are clearly delineated in the upper sequence (Ie/Eif-Giv); analogous fourth order sequences within the lower sequence (Id/Eif-2) are as yet indistinct.

As noted, Johnson et al. (1985) presented an initial model of Devonian T-R cycles/depositional sequences, synthesizing data from multiple basins on different continents, which continues to be refined. The broad geographic scope of this initial effort did not allow the authors to detail the precise positions of regressive and transgressive turnaround points. In most cases, such data were not even available at the time. Therefore, they placed the base of their T-R cycles at the most obvious changes in depth-related lithology, in many cases at maximum flooding surfaces rather than sequence boundaries. For example, in the Appalachian basin they chose the base of T-R Cycle Id/Eif-2 at the often starkly contrasting Onondaga-Marcellus contact. Here was a very obvious lithologic change, clearly related to transgression.

However, subsequent detailed study of the Onondaga-lower Marcellus succession in New York State indicated that the actual turnaround point of regression to transgression (at the sequence boundary) occurred in the upper middle Moorehouse Member of the Onondaga Limestone, well below the Onondaga-Marcellus contact (Brett and Ver Straeten, 1994; Brett and Baird, 1996; Ver Straeten, 1996a, 2007). Various sedimentologic and paleobiologic data (e.g., grain size, bedding thickness, sedimentary structures, fauna and faunal characteristics, etc.) mark the position of shallowest water conditions, below the Tioga A K-bentonite bed, with overall progressively deeper litho- and biofacies through overlying upper Onondaga strata into lower Union Springs black shales.

Subsequent basinwide correlation utilizing various marker beds (e.g., Tioga A-G K-bentonites and additional mid-Onondaga K-bentonites) by Ver Straeten (1996a, b, 2007) showed that this sequence boundary is synchronous basinwide, marked throughout the basin locally by the relatively shallowest water litho- and biofacies. It occurs at the same time-stratigraphic position in the Onondaga Limestone of New York and eastern Pennsylvania, the Selinsgrove Member (Needmore Formation) from central Pennsylvania southward through Maryland, Virginia and West Virginia, and in the Columbus Limestone in central Ohio (Ver Straeten, 2007).

The TST of Sequence Id/Eif-2 continues into the lower part of the black shales (Bakoven Member of the Union Springs Formation) above, to a surface of maximum flooding. Succeeding aggradational strata mark a HST in the lower part of the regressive hemicycle. The third order FSST, marked by progradation of siliciclastics is very distinct in eastern New York (Stony Hollow Member). Westward, however, much of Sequence Id/Eif-2 remains in black shale facies upward to the Id-Ie (Eif-2 and Eif-Giv) sequence boundary at the base of the Hurley Member of the Oatka Creek Formation. Upper Union Springs strata leading up to the Hurley Member may, however, show upward increasing carbonate content (e.g., nodular to bedded micritic/styliolinid limestones).

Shallowest water litho- and biofacies at the base of Sequence Ie/Eif-Giv occur at the base of the Hurley Member in New York. Litho- and biofacies indicate upward deepening through the Hurley, Cherry Valley and lower part of the East Berne members (TST), up to a surface of maximum flooding. Sedimentary aggradation follows during the HST. A distinct and diachronous, basinward-younging base of the progradational FSST is developed across New York in Sequence Ie/Eif-Giv. In the Hudson Valley, progradation of siliciclastics began well down within the East Berne Member. In eastern New York, this progradation culminated in overfilling of the proximal foredeep, locally occurring as low as the late FSST of the second of three 4th order Ie/Eif-Giv sequences; subsequent upper Marcellus in the Hudson Valley and Helderbergs are fluvial dominated terrestrial facies. In central New York, the progradational aspect of the FSST occurs later, perhaps marked by a diachronous transition from black shales (Chittenango Member) to dark gray mudstones (Cardiff Member). East of Syracuse these strata grade upward into a sandstone-shale package (Solsville and Pecksport members). In western New York, the Ie/Eif-Giv sequence remains largely in OC-rich black shales. Three fourth order sequences occur within the Ie/Eif-Giv Sequence (Oatka Creek-Mount Marion formations) , designated Eif-GivA to Eif-GivC by Brett et al. (2011). The base of each lies at, respectively, the base of the Hurley Member, the Halihan Hill Bed of Ver Straeten (1994), and topmost sandstones of the Solsville Member and correlative strata.

Multiproxy analyses (sedimentologic, paleontologic and geochemical data) were carried out on a core in western New York by Ver Straeten et al. (2011; Figure 1-8). For Sequence Ie/Eif-Giv (Oatka Creek Formation), the authors found that the sequence and component systems tracts were well delineated by redox-related proxies TOC and Mo. Through most of the sequence, a total to near total lack of burrowing to bioturbation was only of assistance determining sequence trends at the lower and upper margins of the sequence. The concentration of Al, a proxy for fine-grained siliciclastic input, helped distinguish the HST and FSST. An elevated Si/Al ratio within intervals of greater TOC and MO concentrations reflects non-fluvial sources of silica deposited around the maximum transgression, associated with eolian, volcanic and/or diagenetic processes.

DISCUSSION: MARCELLUS DEPOSITION

The Marcellus "Shale", When It Is: The Marcellus Depositional System, Appalachian Basin

Nearly everyone, even the public, talks about the Marcellus "shale" these days. However The Marcellus in New York has long included more than the classic black shale lithology, of great interest these days. The black shale is only a single facies of the broad, basinwide Marcellus depositional system, which is one reason that Ver Straeten et al. (1994) and Ver Straeten and Brett (2006) designated Marcellus a subgroup.

The Marcellus and time-correlative strata preserved in the Appalachian Basin can be subdivided into three major lithofacies associations: 1*) thin, condensed mudrock facies*; 2) *thick, synorogenic clastic facies association*; and 3) *carbonate-dominated facies*. Although these interfinger at medium to smaller scales, overall they comprise major facies belts, respectively, in the distal foredeep and forebulge, proximal foredeep, and back-bulge basin regions of the Acadian Foreland Basin system.

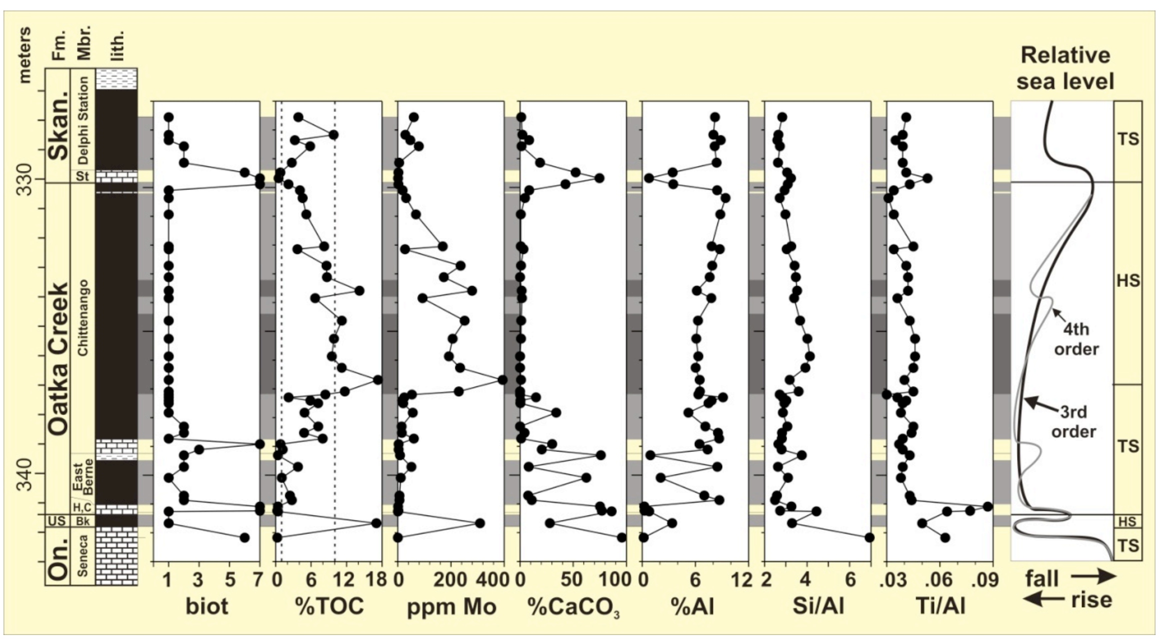

Figure 1-8. Development of Sequence Ie/Eif-Giv in black shale-dominated facies, as outlined by multi-proxy data analyses. Sedimentologic, paleobiologic and geochemical data of the Oatka Creek Formation from a subsurface core, Genesee Valley, western New York. %Al functions as a proxy for clays; Si/Al ratio as a proxy for siliciclastic silt to sand, including fluvial-derived terrigenous, eolian +/- volcanogenic quartz; Ti/Al as a proxy for coarse fluvial-derived clastics. Bioturbation ("biot") measured on rank scale, where 1 = fully laminated fabric, 7 = fully bioturbated fabric, and 0 =no fabric visible. LS= lowstand systems tract; TS= transgressive systems tract; HS= highstand systems tract. Abbreviations: Bk = Bakoven; Fm. =

Formation; H,C = Hurley and Cherry Valley members; lith. = lithology On = Onondaga; Skan = Skaneateles; St = Stafford; US = Union Springs. Light gray bands between data columns = TOC between1 and 10%; Dark gray bands = TOC≥10%. From Ver Straeten et al. (2011).

The *thin, condensed mudrock facies association* is characterized by relatively thin, distal deposits of black, organic carbon (OC)-rich shales and dark gray mudstones (Figure 1-9). Carbonates may or may not be present;

they may be diagenetically precipitated limestone concretion/concretionary limestone beds, styliolinid limestones, or fine-grained carbonate beds of calcisilt and/or small shells (Figure 1-9). Faunas in shales and carbonates represent pelagic forms, or benthic forms living under dysaerobic/poikiloaerobic (low/intermittent oxygen) conditions. In some intervals, silt- to sand-sized grains may also occur, associated with transport by distal hyperpycnal/tempestite/turbidite flows, eolian processes, or of diagenetic origin. Minor amounts of silt- to sand-size volcaniclastic grains may also be present, sourced from explosive felsic volcanism in the Acadian orogenic belt.

Black OC-rich shales are categorized as having greater than 1% total organic carbon; concentrations range up to 17% or greater in some Marcellus strata (Werne et al., 2002; Ver Straeten et al., 2011). They are typically laminated/fissile, with no burrow mottling. Fossils within the black shale facies are restricted to pelagic forms, which lived in the water column; these include small conical styliolinids and dacryoconarids (Lindemann, 2002), and straight and coiled nautiloid and goniatite cephalopods; specific genera and species vary through the Union Springs and Oatka Creek formations.

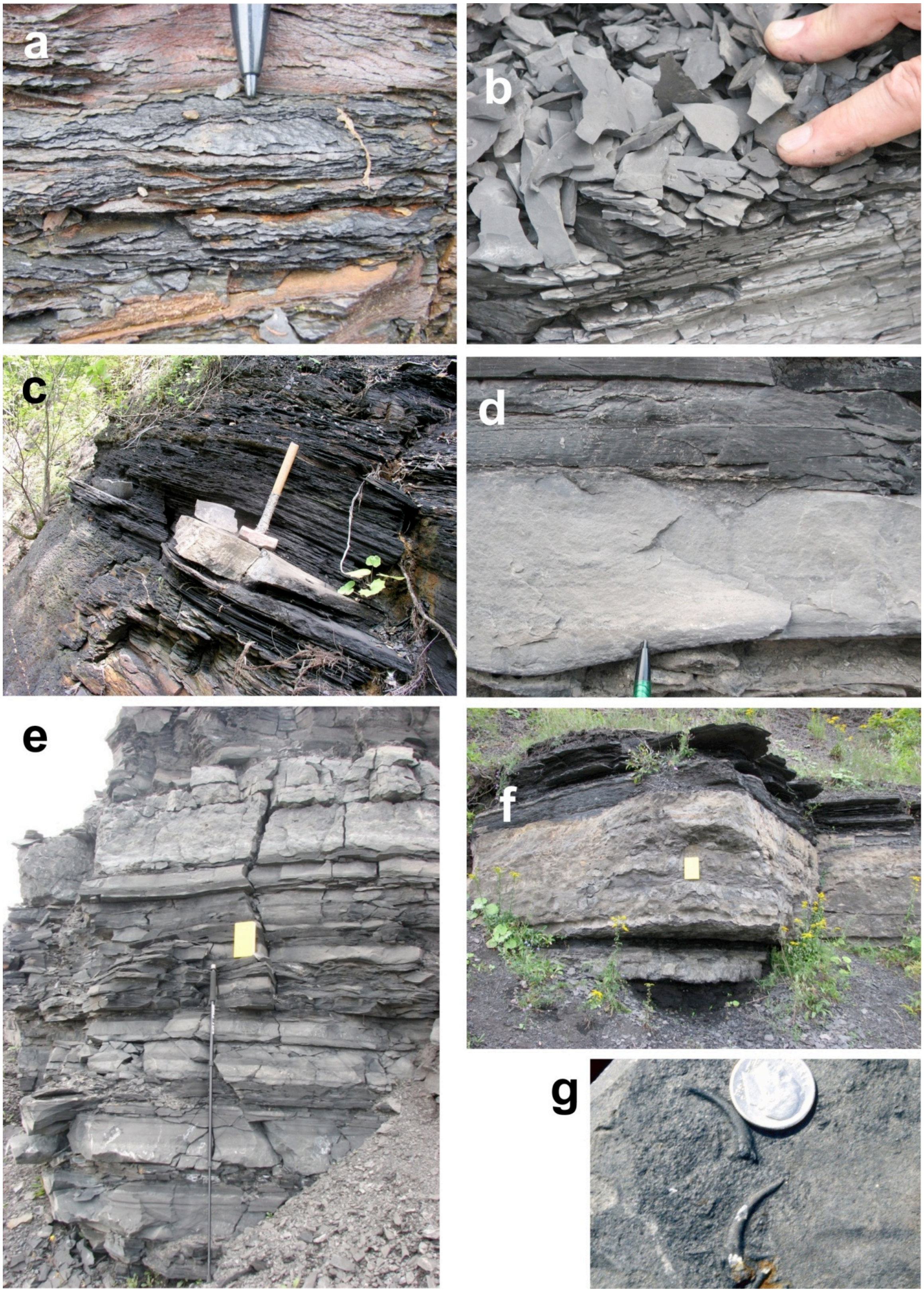
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Figure 1-9. Photos of the Marcellus *thin, condensed mudrock facies association*. a) black, fissile, pyritic shales (Chittenango Mbr., Marcellus, NY); b) medium dark gray mudstones (Cardiff Mbr., Half Acre, NY); black shale with concretionary limestone (Bakoven Mbr., near Catskill, NY); d) close-up of black shale and nodular limestone (Bakoven Mbr., Seneca Stone Quarry, Canoga, NY); e) black shales, and diagenetic to benthic/pelagic limestones (Bakoven through Chittenango mbrs., Seneca Stone Quarry, Canoga, NY); f) Hurley and Cherry Valley limestones. and bounding black shales of Bakoven and East Berne mbrs. (Cherry Valley, NY); g) Onychodid fish teeth in bone bed (basal Bakoven Mbr., Seneca Stone Quarry, Canoga, NY).

Dark gray mudstone facies are distinguished by a subtly lighter color, burrow-mottled non-laminated textures, with varying concentrations of silt- to very fine to fine sand-size siliciclastic grains. Faunas are characterized by both pelagic and generally small, low-oxygen tolerant benthic species, such as leiorhynchid brachiopods, “diminutive” brachiopod, and select mollusks (e.g., small *Leiopteria laevis*, medium to large *Panenka* bivalves, and small gastropods).

These two facies may interfinger in the *thin, condensed mudrock facies association*, even down to laminae-scale, indicative of fluctuating conditions at the sea floor. At medial scales, this may reflect small-scale (5th to 6th order) cyclicity. At the finest (mm) scales, they may reflect seasonal or other fluctuations (e.g., temporary oxygenation events associated with hyperpycnal, tempestite or other bottom-hugging flows reaching distal portions of the basin).

In the Union Springs Formation, black shale facies are assigned to the Bakoven Member. In the Oatka Creek Formation, the black shale facies is assigned to the East Berne or Chittenango members; the dark gray mudstones facies, which comprises an increasing amount of the upper Oatka Creek eastward to the Syracuse area and beyond, is termed the Cardiff Member.

The Marcellus *thick, synorogenic clastic facies association* is characterized by a broad range of organic-poor, terrigenous mudstones, siltstones, sandstones and even thin conglomerates (Figure 1-10). The sediments were sourced from rising Acadian highlands to the east/southeast, during the onset of a new, third phase of orogenic uplift (Ver Straeten, 2010; Tectophase II of Ettensohn, 1985a; onset of the Neoacadian orogeny of van Staal et al., 2009). The siliciclastic-dominated succession is thickest in eastern New York, where not-so-thin black *thin condensed mudrock facies* are overlain by an overall coarsening up succession with a total thickness estimated by Rickard (1989) as >580 meters.

The Marcellus *thick, synorogenic clastic facies association* is well developed in the Hudson Valley of eastern New York. The facies association includes marine strata deposited in relatively deeper water to delta toe to shallow shoreface marine environments of the foreland basin foredeep. Uppermost Marcellus-age strata in the Hudson Valley and Helderbergs were deposited in terrestrial coastal plain environments, as channel sandstones, wetland mudstones, and floodplain mudrocks and paleosols (lower Ashokan Formation).

Overlying typical black to dark gray Union Springs mudrocks in the Hudson Valley is a package of siliciclastic-dominated buff-weathering, calcareous mudrocks, siltstones and sandstones assigned to the Stony Hollow Member. These strata laterally replace typical upper Union Springs black to dark gray mudrocks of the Bakoven Member from the Helderbergs down the Hudson Valley. The facies change is gradational, from the top down, increasing southward. At Kingston the Stony Hollow comprises the upper half of the Union Springs succession.

The Marcellus *thick, synorogenic clastic facies association* in New York is best developed in the Oatka Creek-equivalent Mount Marion Formation (Rickard, 1975; Ver Straeten and Brett, 2006). This thick succession of dominantly silty to sandy mudstones and argillaceous to arenaceous sandstones and thin conglomerates, is locally over 400 m-thick (Rickard, 1989). Even the Hurley and Cherry Valley members, developed as carbonates from near Albany to western New York, change laterally into mudstones and

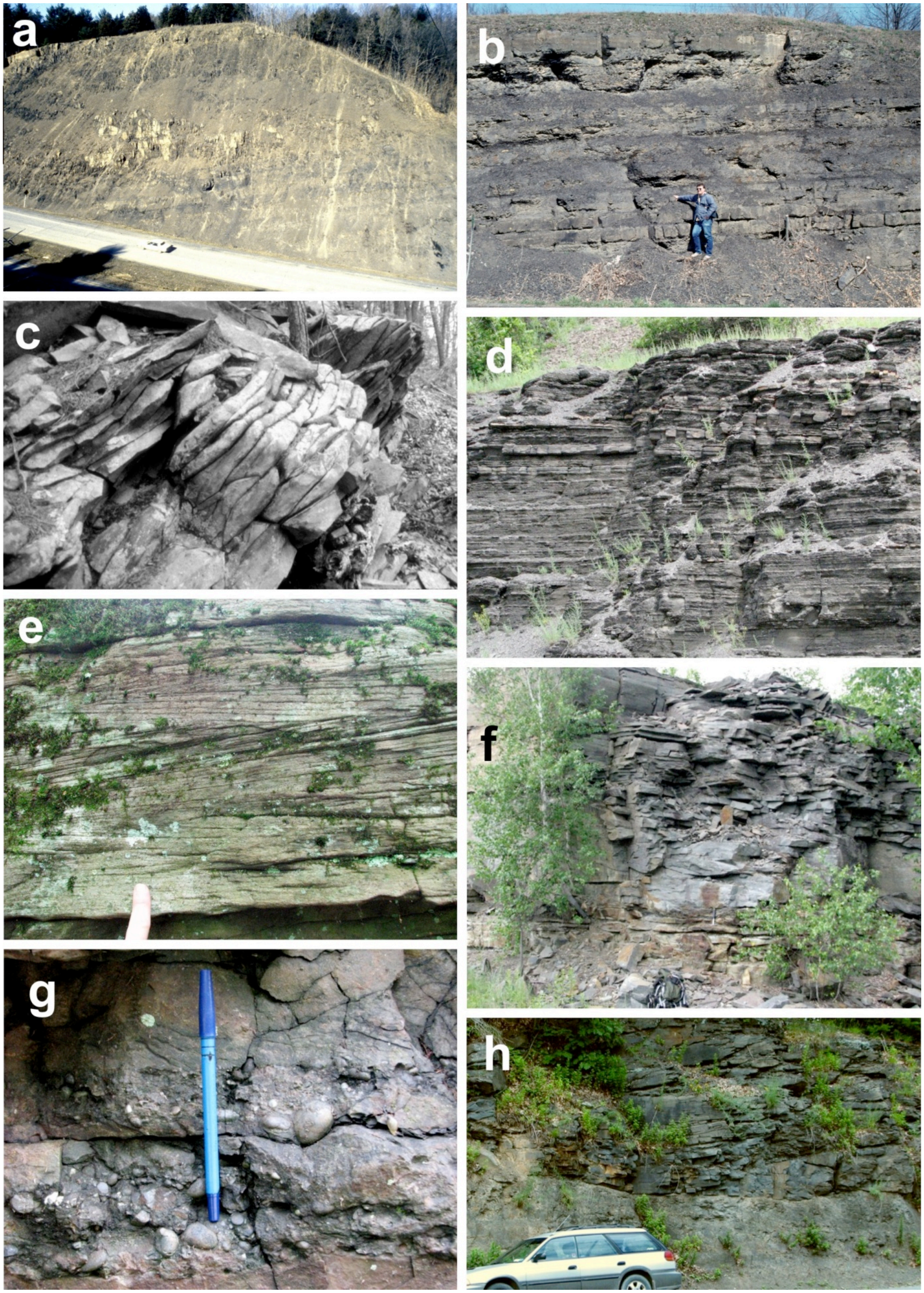
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Figure 1-10. Photos of the Marcellus *thick, synorogenic clastic facies association*. a) Buff-colored, calcareous shales to sandstones (Stony Hollow Member, Kingston, NY); b) thick siliciclastic-dominated upper part of Hurley and lower part of Cherry Valley mbrs. (Kingston, NY); c. Coarse upper Union Springs-equivalent Turkey Ridge Mbr. sandstones (Thompsontown, PA);d) interbedded sandy mudstones and thin sandstones (Mount Marion Fm., Cobleskill, NY); e) cross-bedded nearshore sandstones (upper Mount Marion Fm., East Berne, NY); f) mixed terrestrial (below) and nearshore marine (above) sandstones (upper Mount Marion, Hudson Valley, New York); g) polymict conglomerate with erosive base (upper Mount Marion Fm., Quarryville, NY); floodplain paleosols (below) and fluvial channel sandstones (above), with erosional contact (upper Marcellus-equivalent lower Ashokan Fm., Kingston, NY).

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sandstones southward along the Hudson Valley, and thicken from meter-scale to as much as 17 m-thick (Griffing and Ver Straeten, 1991; Ver Straeten et al., 1994, Figures 1-6, 1-10b).

Post-Union Springs strata in the easternmost outcrop belt of the Hudson Valley and Helderbergs undergo an initial fining-up from somewhat fossiliferous, bioturbated mudstones and pyritic, bioturbated sandstones (Hurley and Cherry Valley members) into thin black shales, followed by a thick overall coarsening upward succession from dark gray mudstones to sandstones in the East Berne and Otsego members, and undifferentiated upper Mount Marion strata. Progradation of a large volume of Acadian synorogenic siliciclastics overfilled the basin to above sea level during deposition of the Marcellus, apparently represented by upper Solsville Member-equivalent strata. At the same time, sands prograded nearly as far west as Syracuse (e.g., Solsville Sandstone; Figure 1-2); later, in the during deposition of at the base of the Skaneateles Formation (Mottville Sandstone), sands prograded even further west at least to the Cayuga Lake meridian (Figure 1-2).

In contrast with Union Springs strata, Mount Marion siliciclastic-dominated strata are rarely calcareous. As noted, even limestones (e.g., Hurley and Cherry Valley members) become argillaceous to arenaceous in more proximal facies. In central to western New York, some limestone concretions may be found; however, small and sometimes abundant concretions in medial to nearshore siliciclastics of the Mount Marion Formation are sideritic to possibly ankeritic in composition.

The Marcellus *thick, synorogenic clastic facies association* is found all along the proximal margin of the Appalachian Basin outcrop belt from eastern New York, through parts of New Jersey, Pennsylvania, Maryland, and Virginia, to eastern Tennessee (Figures 1-3, 1-6). Union Springs and lowest Oatka Creek equivalent sandstone-dominated facies are best developed in the vicinity of Harrisburg, PA, in the Turkey Ridge Sandstone. In one outcrop, Ver Straeten (2007) reports apparent terrestrial plant root structures at some unknown position within the Turkey Ridge Member. If true, this represents the lowest reported occurrence of terrestrial facies in Marcellus strata basinwide. Laterally, these Stony Hollow- to Cherry Valley-equivalent sandstones transition to offshore buff-weathering calcareous facies of the same character as seen in those units in the Hudson Valley. Further basinward they transition to the *thin, condensed mudrock facies association,* capped by thin carbonates laterally equivalent to the Hurley and Cherry Valley members. These offshore to basinal equivalents of New York's Stony Hollow, Hurley and Cherry Valley members are lumped together from central Pennsylvania southward as the Purcell Member.

The Marcellus *carbonate-dominated facies association* in the Appalachian Basin outcrop belt is characterized by limestone-rich strata (Figure 1-11). These are predominantly developed on the western margin of the basin as the Delaware Formation in central Ohio, on the western, cratonward margin of the back-bulge basin sector of the Acadian Foreland Basin system. However, Marcellus-age carbonate sediments occur scattered throughout the basin at different intervals, which are characterized by four different types: benthic shelly, pelagic styliolinid, calcisilt/micrite, and diagenetically precipitated carbonates that occur as bedded to concretionary limestones (Ver Straeten et al., 2011). In the central New York area of this fieldtrip, all of these carbonate types are found. The lower Hurley Member carbonates are a mix of benthic shelly-type carbonates with varying concentrations of the other three types; the benthic fauna becomes a minor component in the overlying Cherry Valley, with a greater concentration of the other types, accompanied by common nektonic cephalopods and styliolinids/dacryoconarids. Thin styliolinid limestones are not uncommon in the Union Springs Formation, as are concretionary beds and nodules, and likely some calcisilt/micrite sediment-type sediments, composed of transported, fine-grained carbonate grains.

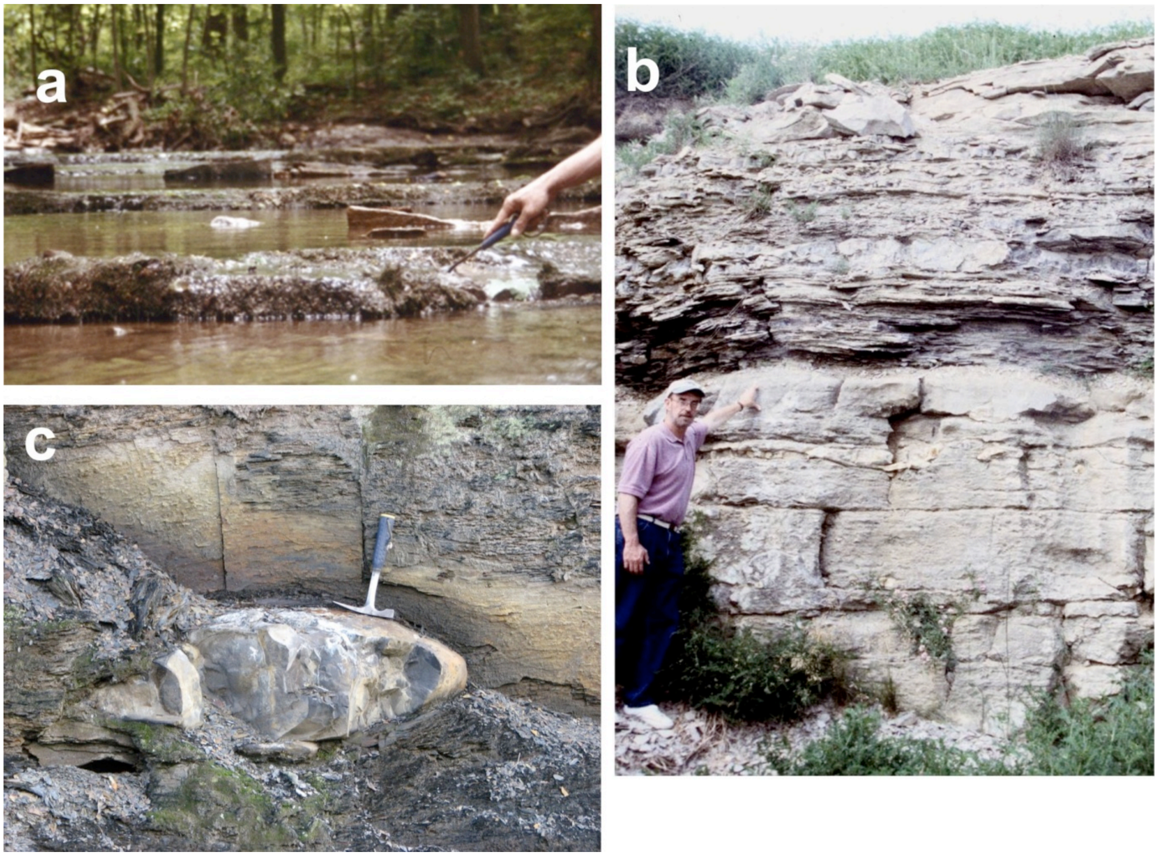

Figure 1-11. Photos of the Marcellus *carbonate-dominated facies association*. a) Hurley and Cherry Valley mbrs.-equivalent strata in the Delaware Limestone (Delaware, OH); b) shallow to deep ramp limestones and argillaceous limestones. The latter, in the center of the photo, represent the maximum transgression of Sequence Eif-2/Id in the back-bulge basin (Delaware Fm., Columbus, OH); c) large limestone concretion in black shale (Chittenango Mbr., Marcellus, NY. Additional carbonate facies photos can be seen in Figure 1-9.

These three major Marcellus facies associations generally dominate specific sectors of the Acadian Foreland Basin system. The *thick, synorogenic clastic facies association* occurs primarily in the proximal to medial foredeep of the foreland basin system. The *thin, condensed mudrock facies association* is developed primarily in the distal foredeep to submarine/flooded forebulge portions of the basin. Due to poor to no outcrop exposure it is difficult to tell what is found in the bulk of the back-bulge basin; however, on the distal cratonward margin normal carbonate ramp limestones of the *carbonate-dominated facies association* are found.

Interfingering and temporary shifts of the major facies occur across the basin at times, dependent on several factors. These may variously include sea level history, sediment input/dilution, climate shifts, shifts in fluvial points sources, degree of oxygenation, chemical conditions and other controls.

The Onondaga-Marcellus Contact in New York: Eustasy meets Crustal Flexure

The regressive- transgressive turnaround point at the base of Sequence Eif-2/Id in the upper Onondaga Limestone is synchronous in outcrops basinwide at a position in the upper middle Moorehouse Member, and correlatives, well below the Onondaga-Marcellus contact. Thus the significant limestone to black shale change at the contact, utilized by Johnson et al. (1985) as the base of their original T-R cycle Id, marks the cessation of carbonate production and transport. In central New York, a few remnant dark, fine-grained carbonate beds may be found locally above the Tioga F K-bentonite (which makes a convenient marker bed for a chronostratigraphic Onondaga-Marcellus contact in most areas, independent of facies). Essentially, Onondaga-type benthos-derived carbonate sediment was no longer being produced above the Onondaga-

Marcellus contact across the basin except in central Ohio, nor imported from surrounding areas into the main body of the Appalachian Basin. The generally few limestone beds found through the Union Springs Formation (locally common, as at Stop 1) are chiefly diagenetically precipitated or pelagic in origin (styliolinid limestones).

With the end of carbonate production, the only sediments arriving at the sea floor during Union Springs time were small amounts of distally transported siliciclastic muds, wind-blown eolian silts, organic matter, and perhaps some pelagic carbonate (e.g., styliolinid shells). Initially, it appears that little if any sediment was deposited across the main body of the basin. This resulted in formation of a sediment-starved unconformity, or "surface of maximum starvation", commonly marked by a fish bone bed. This is not a sequence-bounding unconformity; this type of unconformity occurs within a transgressive succession, associated with little to no sediment input ("sediment starvation"), during the time of the fastest rate of relative sea level rise. In this case, the distinct limestone-black shale contact (sometimes poetically called "The Edge of Night") is associated with the shutdown of carbonate production across much of the Appalachian basin

The position of the limestone-black shale contact does vary to some degree around the basin. In deeper portions of the central to southern part of the Appalachian Basin carbonate production largely shut down at an earlier time (e.g., Frankstown, PA; more southwesterly parts of the Virginia-West Virginia outcrop belt). In contrast, carbonate production continued until mid-Oatka Creek time on the distal margin of the foreland basin in central Ohio. Finally, the end of carbonate production/deposition was diachronous from central to eastern New York (Albany Co.), clearly shown by a progressive top-down absence of the Tioga B to F K-bentonites from the central Finger Lakes to the Helderbergs. Otherwise around the basin, the end of carbonate production and deposition was relatively synchronous.

Clearly a major transgression occurred through upper middle Moorehouse strata into lower Union Springs strata and equivalents basinwide, as seen in the gradation of sedimentologic and paleontologic data trends (e.g., grain size, bedding thickness, sedimentary structures, biofacies). What mechanism or mechanisms (e.g., global eustasy and/or tectonic subsidence) drove this relative sea level rise?

Johnson et al. (1985) and subsequent work across the U.S. and worldwide clearly show that both Marcellus cycles represent major global transgressions, more so than the 15 other previous and subsequent third order Helderberg- to Hamilton-age Devonian sequences (including Sequences Ia/Prag-1 through Ii/Giv-4; Johnson et al., 1985; Ver Straeten, 2007; and Brett et al., 2011). For example, Sequence Id/Eif-2 and more so Ie/Eif-Giv marine waters flood the cratonic Iowa basin, and Sequence Ie/Eif-Giv transgresses over the North American transcontinental arch, respectively for the first time during the Devonian (Johnson et al., 1985; Day et al., 1996), and subsequently fall through subsequent upper Hamilton Sequences If/Giv-1 to Ii/Giv-4 of Brett et al. (2011). There clearly is a global, eustatic component to the two Marcellus depositional sequences; and they represent major global sea level transgressions relative to older and younger sequences below and above them.

The diachronous end of carbonate deposition from proximal to distal areas of the basin (e.g., central PA, VA, WV) is well correlated with lateral changes in depth-related sedimentology and fossils assemblages; carbonate production and transport ended earlier in deeper areas, only slightly above the Id/Eif-2 cycle boundary (or even lower in places). On the distal margin of the basin in Ohio, overall shallower settings and little to no siliciclastic input permitted ongoing carbonate production. However, the end of carbonate deposition (Onondaga-Marcellus contact) across New York is time-transgressive; carbonate production and deposition ends earlier in the east (upper Moorehouse Member, Albany area), and progressively shuts down later and later into the central Finger Lakes area (e.g., Stops 4 versus 1). In contrast with the first case, it ended initially in areas of shallower water Onondaga facies of eastern New York, while continuing in deeper water areas (central Finger Lakes of central New York; Ver Straeten et al., 1994; Ver Straeten, 2007, 2010; Stop 1).

According to Dorobek (1995), controls on carbonate production/sedimentation in foreland basins include siliciclastic dilution, concentration of fine-grained siliciclastics in the water column (decreases light penetration, interferes with suspension-feeding organisms), transport efficiency, and depth relative to the euphotic zone (generally ca. 100 meters depth; Schlager, 1981).

Strong evidence indicates that OC-concentration and burial in distal deposits of the Union Springs and Oatka Creek formations is associated with extreme siliciclastic sediment starvation (Werne et al., 2002; Sageman et al., 2003; Ver Straeten et al., 2011) and eliminates siliciclastic dilution as a mechanism. Furthermore, siliciclastic sediment starvation would leave a rather clear water column through which sunlight could penetrate relatively deeply, and not disturb benthic communities; thus, the second mechanism is also not plausible. Transport efficiency decreases with greater distance from source, and/or with a decrease in hydrodynamic energy or gravity-related processes that drive transport. Across the contact, carbonate sources are obviously no longer local, and transport has largely shut down, which are perhaps best explained as depth related. Therefore, significant and non-synchronous deepening (sub-photic zone) appears to be the most parsimonious explanation for the end of carbonate deposition along the New York Onondaga-Marcellus contact. The diachronous timing of this transgression across eastern to central New York in late Onondaga time was hence controlled largely by tectonic subsidence, superposed over the significant Id/Eif-2 eustatic transgression.

At about this time, a new phase of mountain building and associated crustal loading began along the eastern Laurentian (North American) margin. Termed "Acadian Tectophase II" by Ettensohn (1985a), it actually represents a third pulse of orogenesis in New England during Late Silurian to Middle Devonian time (Ver Straeten, 2010). This phase of mountain building has recently been associated with the onset of van Staal et al.'s (2009) "Neoacadian Orogeny," the result of collision of the Meguma terrane (beginning at ca. 395 Ma).

The conjunction of data on a global sea level rise and non-synchronous cessation of carbonate production across New York outcrops indicate both eustatic and tectonic influences in the formation of the Onondaga-Marcellus contact, and subsequent deposition of Marcellus rocks in New York and the Appalachian Basin.

A Marcellus Forebulge in Western New York

Forebulges in foreland basin systems are often subtle features in the sedimentary record, and may or may not have any topographic expression (DeCelles and Giles, 1996). Ver Straeten et al. (1994) and Ver Straeten (2010) discussed outcrop evidence for the apparent uplift of a forebulge in western New York during upper Marcellus time.

Throughout the Appalachian Basin outcrop belt the Hurley Member features relatively shallower litho- and biofacies than the overlying upper Cherry Valley. However, in the Honeoye Falls Quarry south of Rochester in western New York, an anomalous local inversion of topography is indicated by distinctly shallower litho- and biofacies, coarser lithologies, and a major thickening of the Cherry Valley Member (Ver Straeten et al., 1994), clearly not related to eustatic sea level processes.

Recent subsurface analyses by Lash and Engelder (2011) indicate an anomalous interval of relatively thinner middle to upper Oatka Creek strata across western New York, relative to coeval strata lying to the east in the distal foredeep and the west in the back-bulge basin. They also relate this to the uplift and existence of a forebulge through Marcellus time, associated with uplift along older vertical faults.

However, the anomalous reversal of depth-related litho- and biofacies between the Hurley and Cherry Valley members at Honeoye Falls, in opposition to basinwide trends, and lateral grain-size and faunal trends in the Hurley Member along the east-west New York outcrop belt indicate that uplift of the forebulge did not occur south of Rochester until after deposition of the Hurley. Moreover, the thin Union Springs succession, which shows continuous thinning from central New York to a feather edge in western New York, was more likely related to distal sediment starvation (as well as erosional beveling at the basal Sequence Ie/Eif-Giv LST/initial TST).

Furthermore, deposition of six or more meters of mudrocks over the proposed forebulge during middle to upper Oatka Creek Formation above the Halihan Hill Bed (thickness from Rickard, 1989) suggests that for some reason, topographic expression of the bulge became minimal. As noted above, recent studies indicate that mud is distally transported by bottom hugging density flows of mud-laden waters (e.g., hyperpycnal flows), which would be unlikely to climb up the elevated topography of a forebulge. While the shale thickness is less over the forebulge than in the adjacent distal foredeep (east) and back bulge basin (west), deposition over the bulge seems to imply that progradation of thin synorogenic muds via bottom processes had infilled the closely adjacent distal foredeep, allowing deposition of muds over the bulge; or that the bulge underwent some degree of flexural subsidence for unknown reasons (e.g., migration, sediment loading). In either case, it appears that the topographic expression of the bulge was overall rather minimal through its existence during mid to upper Marcellus time.

How Shallow Is Shallow, How Deep Is Deep?

In Chapter 2 of this paper, Lash and Blood mention a shallow water hypothesis for the deposition of Marcellus organic-rich facies, briefly discussed in Lash and Engelder (2011). This touches on an old debate about the "deep" versus "shallow" origin of black shales (e.g., Conant and Swanson, 1961; Rich, 1951; Ettensohn, 1992). It is clear, even in the Late Silurian and Devonian strata of New York, that black to dark gray shales and mudstones can form in various settings, from the generally interpreted "deep" foreland basin black shales of the Union Springs and Oatka Creek formations and other units; to tidal facies of the Manlius Formation of the Hudson Valley; and terrestrial, fluvial-dominated facies of the Middle to Upper Devonian of the Catskills region. The shallow-water depositonal model for Devonian and Ordovician organic-rich mudrocks in New York includes: 1) deposition in relatively shallow waters, less than 30 meters deep; 2) deposition on the cratonward side of the basin; 3) shales overlie, onlap, and pinchout onto unconformities; 4) some unconformities are demonstrably subaerial in origin; 5) depositon occurs during times of orogenic load-induced subsidence; but 6) eustatic sea level was low during black shale deposition; and 7) black shale was deposited in the shallowest water in the basin.

There are several difficulties with this model, some of which can be debated, and several that cannot. Certainly central to western New York was not on the cratonward side of the foreland basin during the Middle Devonian; that lay in central Ohio and areas of Ontario to the north (Cincinnati-Findlay arch), on the order of 250 to 400 km cratonward of the central to western New York Marcellus. Unconformities do abound in highly condensed distal basin facies, far from siliciclastic input sources, as occurs in the central and especially western New York Marcellus. However, this is to be expected in distal, sediment-starved siliciclastic facies, especially in light of modern perspectives on clay transport.

No Marcellus-age unconformities in the foreland basin of "demonstrably subaerial in origin" are known to long-term New York Devonian researchers. In fact, this is a time when the mid-continent arches were flooded, as were the Michigan, Illinois and Iowa basins, and even the transcontinental arch, the latter two for the first time during the Devonian (Johnson et al., 1985; Day et al., 1996). It is very clear that eustatic sea level was demonstrably rather high globally during development of both Marcellus third order sequences, especially during Sequence Ie/Eif-Giv.

It is indeed also clear that the Marcellus was deposited in the greater Acadian Foreland Basin system during a time of strong load-induced subsidence at the onset of renewed mountain building in the Acadian Orogen, with subsidence superposed over two major eustatic deepenings (Sequences Id/Eif-2 and Ie/Eif-Giv).

Accumulation of 580 meters of sediment (compacted) of prior to shallowing to sea level/overfilling the basin in the foredeep in eastern New York (Rickard, 1989), indicates substantial accommodation space during Marcellus deposition, which might imply that depths at least into central NY could be rather significant. Foreland basin subsidence, which does decrease basinward, is generally understood to be controlled largely by orogenic loading; subsidence related to sediment loading has a lesser effect. So, there should be "significant" depths cratonward of the toe of the siliciclastic delta – until progradation infills both eustatic- and subsidence-related accommodation space.

Four lower Hamilton tongues of sands prograded basinward during the late FSSTs of fourth order sequences through the uppermost Union Springs and Oatka Creek into basal Skaneateles formations. Each of these tongues extended progressively further basinward, from the Hudson Valley (top Stony Hollow Member) to western Albany County (upper East Berne Member) to western Madison County east of Syracuse (Solsville Member), and to the central Finger Lakes (Mottville Member, basal Skaneateles Formation). Each of the sand bodies thin to a feather edge at the distal ends, rather than thickening as might be expected if they encountered rising sea floor topography at their distal terminus. The progradation of these sand bodies far cratonward across New York would seem to imply extension of the submarine delta downslope and basinward into deeper water far into central New York during Marcellus time. Yet much accommodation space remained following each progradational event.

Lash and Engelder (2011) based their hypothesis of a shallow marine setting for western New York at least in part on the thin Marcellus succession observed in well logs, which they interpreted to reflect erosional removal of muds over the Acadian forebulge. However, with an understanding of mud transport and deposition, it is perhaps more likely that only minor amounts of mud was deposited over the bulge until basin topography became relatively leveled out in the vicinity of the bulge. With little to no rainout of suspended clays in marine settings, clay transport to distal portions of the basin would have been largely due to relatively dense "hyperpycnal" mud flows, and related processes. Such denser flows would not run upslope onto a prominent topographic high, which implies that there may not have been much topographic expression of the western forebulge at that during mid to upper Oatka Creek time.

Thin successions, with multiple marine unconformities and erosion would be expected to be common in condensed distal basin facies, associated with the failure of distal mud transport. Furthermore, with low sedimentation rates, processes related to bottom and contour currents, possible impingement of internal waves along density boundaries/pycnoclines on the sea floor, combined with distal combined-flow storm currents and mud-laden hyperpycnal currents, would lead to reworking of sediments on a starved basin floor or elevated topographic feature like a forebulge, without having to invoke a shallow marine setting/processes. As noted by Rine and Ginsburg (1985) muds may be deposited and remain in shallow water environments – where mud sedimentation rates are extremely high, not where they are very condensed.

Carbonate platforms or ramps generally form on the distal cratonward margin of foreland basin and over a forebulge, or if the latter is exposed, on its flanks (Dorobek, 1995). If western New York was elevated into shallow marine environments during Union Springs and at least lower Oatka Creek time it would be expected for carbonates to have developed, as were coevally deposited on the shallow margin of the foreland basin in central Ohio and Ontario. The western New York area was quite starved of siliciclastic sediments, related to the problems of distal clay transport. Another component is that very little clay would have been suspended in the water column so far basinward, allowing for good penetration of light, encouraging carbonate production and deposition. Clearly, in the time-correlative Delaware Limestone, such carbonates developed in even relatively offshore settings below normal wave base and possibly, during the late TST to HST of Sequences Id/Eif-2 and Ie/Eif-Giv, deeper than storm wave base: In some areas of central Ohio, black/dark gray shales (Dublin Member) were deposited during the maximum flooding interval of Sequence Id/Eif-2, in the dark gray to black Dublin Shale Member, (lower part of the Delaware Formation). However, in other areas, carbonates were deposited at that time.

Carbonate production and deposition was certainly possible in the condensed, black shale region of central to western New York during times of relatively shallow water conditions during basal TSTs +/- LSTs, as with the Hurley and Cherry Valley members. Except in interpreted deepest basinal areas or proximal siliciclastic-dominated facies, the Hurley Member is composed of fine-grained, moderately fossiliferous limestones, with deeper water benthic- to pelagic-dominated faunas in the overlying Cherry Valley. Yet, in between the shallowest water settings in the upper Onondaga, Hurley-Cherry Valley Limestones and basal Skaneateles Formation, no such carbonates formed in the hypothesized shallow water settings of western New York.

A commonly cited model in the debate about shallow- versus deep-water origin of black shales is the Upper Devonian Chattanooga Shale in Tennessee and adjacent areas. It has been variously interpreted to provide evidence for "deep" (e.g., Potter et al., 1982; Ettensohn, 1985b) and "shallow" (e.g., Conant and Swanson, 1961; Schieber, 1994, 1998b) water deposition of black shales.

The Chattanooga Shale was deposited through the upper Givetian to Famennian stages of the late Middle to late Late Devonian, in an epicontinental sea cratonward of the Appalachian Foreland Basin (Over, 2007), over a period of time approximating 24 million years (using the time scale of Kaufmann, 2006). Generally on the order of less than 10 meters in thickness (Schieber, 1994), it is a time-rich, relatively condensed mudrock-dominated unit.

In an interesting new study, Witzke (2011) reports preliminary minimal estimates of eustatic sea level changes for upper Middle to Upper Devonian sequences from the tectonically stable Iowa Basin. Based on depths of incised valleys, he documented sea level shifts during sequences that at minimum ranged from 15 to 90-140 meters for Devonian T-R Cycles IIa-2 to IIf (>15m, >35 m, > 30m, > 90-140m, >70m, and "onlaps 35 m at shelf margin," respectively). These sequences comprise the same interval as the Chattanooga Shale in Tennessee. If little flexure was acting on the Chattanooga depositional basin, cratonward of the foreland basin system, then obviously the Chattanooga Shale was deposited in waters of sometimes greatly varying depths through time.

The obvious point to this is that any given site of Devonian black shale deposition was relatively shallow during lowstands, and relatively deep during highstands. Depth varied cyclically, with varying magnitudes of change, and as shown by Witzke (2011) what was sometimes "shallow" was also sometimes "deep".

Translated this to Marcellus strata. Relative sea level would have been relatively shallow – at times (e.g., the relatively deep water Hurley and Cherry Valley members), and relatively deep at others – deeper than the Hurley-Cherry Valley limestones. No quantitative measurements of Marcellus-age eustatic change is documented, nor is the amount of flexural change. But, except for local upward flexure (e.g., forebulge, which may or may not have had much topographic expression), the relatively deep water Hurley-Cherry Valley depositional environments would have been the shallowest point within the Marcellus succession until progradation began to fill the Hudson Valley outcrop belt to near sea level. Considering that Sequences Id/Eif-2 and especially Ie/Eif-Giv were associated with major global eustatic flooding events, even transgressing over the transcontinental arch during the highstand of Sequence Ie/Eif-Giv, the surface of maximum flooding in both sequences would have been rather deep throughout most of the basin, including much of the area where the black Marcellus shales were deposited. This would include the sea floor over a forebulge that may have had only relatively subtle topographic expression.

Organic-rich black shales can form in many marine settings, at various depths. Overall, however, a broad set of geological and paleontological evidence and modern perspectives argue against a shallow water setting for the Marcellus subgroup in New York, even over the forebulge in western New York.

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CHAPTER 2: PETROPHYSICAL PROPERTIES OF THE MARCELLUS FORMATION AS REFLECTED IN SEQUENCE STRATIGRAPHY

Gary G. Lash and Randy Blood

Introduction

The rapid evolution of the Marcellus Formation gas shale play into development mode is shifting exploration focus from assessment of fairway dimensions to production optimization. Crucial to the latter aspect of the play is the proper stratigraphic positioning of horizontal wellbores or laterals within the Marcellus. Optimal lateral placement necessitates consideration of a number of reservoir properties, including (1) location of the greatest concentration of hydrocarbon and (2) post-stimulation deliverability of the formation. Source rock quality and petrophysical reservoir rock properties of the Marcellus Formation can be linked to the abundance of organic matter, quartz, and clay, as well as the diagenetic history of the rock. Systematic variations of these and other properties reflect changes in depositional environment controlled by base level fluctuations (Blood, 2011). This chapter considers several of the more important rock properties of the Marcellus Formation in terms of the sequence stratigraphic paradigm (refer to Lash and Engelder (2011) for details of the Marcellus subsurface sequence stratigraphy). This work, preliminary as it is, reflects a multi-faceted approach that encompasses well log analysis, examination of outcrop and core material, and geochemical analysis of well cuttings.

Stratigraphic Framework

In a series of papers spanning nearly 15 years, Ver Straeten et al. (1994), Ver Straeten and Brett (1995, 2006) and Ver Straeten (2007) proposed a Marcellus stratigraphy that seeks to reduce the accumulated, sometimes confusing, stratigraphic verbiage of more than 150 years of study. The revised stratigraphy links the generally fine-grained Marcellus succession of the more distal, western region of the basin with that of the proximal eastern basin where the Marcellus Formation is part of a generally shallowing-upward trend from basinal black shale to nearshore sandstone and fluvial deposits. In this paper, however, we adopt a lithostratigraphy more in line with that employed by Rickard (1984) and one that lends itself to subsurface correlation of wireline log signatures (Fig. 1). Our basal unit of the Marcellus Formation is the Union Springs Member, an organic-rich unit that passes upward into the Cherry Valley Member (Fig. 1) and the partially correlative Purcell Limestone in Pennsylvania. The Cherry Valley Member, which comprises variable amounts of interlayered carbonate, shale, and sandstone, is overlain by the Oatka Creek Member, a succession of black and gray shale and lesser siltstone and limestone which underlies the Stafford and Mottville members of the Skaneateles Formation (Fig. 2-1).

Sequence Stratigraphic Framework Of The Marcellus Formation – The T-R Sequence Paradigm

The application of the sequence stratigraphic approach to source rock and reservoir analysis enables one to subdivide basin fill into a framework of systems tracts and internal and bounding surfaces. The resulting stratigraphic architecture can mitigate risk in frontier regions of the basin or areas of poor data control (e.g., Partington et al., 1993; Emery and Myers, 1996; Singh et al., 2008). Lash and Engelder (2011) adopted the transgressive-regressive (T-R) sequence stratigraphy described by Embry and Johannessen (1992) and further refined by Embry (2002, 2010) in their recent stratigraphic investigation of the Marcellus Formation. Indeed, we find the T-R sequence stratigraphic paradigm to be especially well suited to the log-based analysis of siliciclastic successions. In essence, T-R sequences are similar to Types 1 and 2 depositional sequences in marginal regions of basins where sequence boundaries comprise subaerial unconformities or unconformable shoreface ravinements (Embry, 2002, 2010). In the basinal or conformable succession, however, the maximum regressive surface (MRS) serves as an objectively recognizable sequence boundary that correlates with the unconformable shoreline ravinement (Embry, 2002). A single T-R sequence comprises a transgressive systems tract (TST), a succession that records rising base level, overlain by regressive systems tract (RST) deposits that accumulated during the subsequent base level fall (Embry and Johannessen, 1992; Embry, 2002). Transgressive

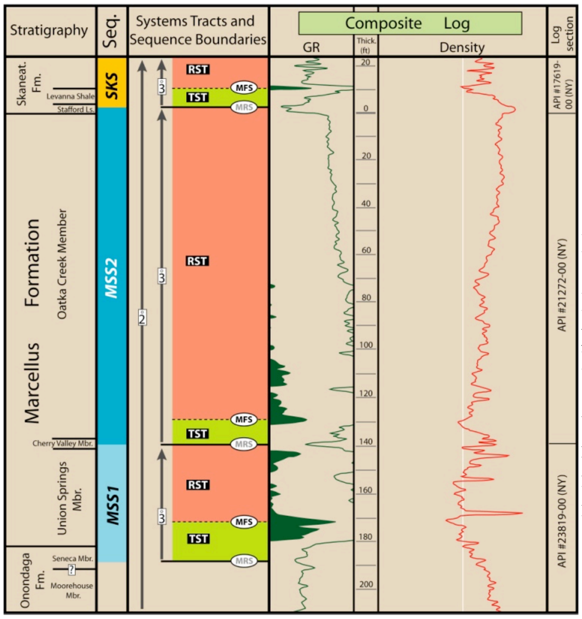

Figure 2-1: Sequence stratigraphic "type section" of the Marcellus Formation that encompasses the upper part of the underlying Onondaga Formation and the lower interval of the Skaneateles Formation; TST = transgressive systems tract; RST = regressive systems tract; MFS = maximum flooding surface; MRS = maximum regressive surface. See Lash and Engelder (2011) for details.

systems tract and overlying RST deposits are separated by a maximum flooding surface (MFS), which marks a change in water-depth trend from deepening to shallowing (i.e., the time at which transgression ends and regression begins; Embry, 2010). The MFS, arguably the most readily recognized sequence surface in well logs, approximates the horizon of deepest water in nearshore environments (Embry, 2010) and may correspond with, or pass distally into, a condensed interval (Emery and Myers, 1996; Partington et al., 1993).

The Marcellus Formation encompasses the bulk of two T-R sequences herein referred to as MSS1 and MSS2, in ascending order (Fig. 2-1). These sequences, approximate equivalents of Johnson et al.'s (1985) T-R Cycles Id/Eif-2 and Ie/Eif-Giv and Ver Straeten's (2007) Eif-2 and Eif-3 sequences, span ~1.8 MY, extending from the upper *costatus* conodont Zone through the *hemiansatus* Zone (Kaufmann, 2006; Ver Straeten, 2007). The relatively short duration of MSS1 and MSS2 is consistent with third-order base level cycles (Mitchum and Van Wagoner, 1991). Moreover, the generally basinwide extent of Marcellus T-R sequence boundaries, manifestly transgressive deposits overlying sequence boundaries, and the presence of unconformities (unconformable shoreline ravinement) only on basin flanks are consistent with third order T-R sequences (Embry, 1995).

Marcellus Shale Sequence Stratigraphy Reflected In Petrophysical Properties

The balance of this paper considers a number of parameters critical to the placement of laterals within the Marcellus reservoir in terms of sequence stratigraphic controls. We recommend Blood's (2011) recent contribution for further discussion.

Gas-in-Place and Organic Carbon

Gas-in-place (GIP) comprises a four-part system being the sum of gas retained in matrix porosity, gas adsorbed onto organic particles, gas held within fractures, and gas dissolved in liquid hydrocarbons. Given the high *in situ* confining stresses carried by the Marcellus Formation at drilling depths of 1,525 – 2,438 m (5,000 – 8,000 ft), natural fractures are likely closed (if not mineralized), retaining minimal porosity. Further, the presence of much of the Marcellus play within the dry gas window diminishes the amount of gas likely to reside in liquid hydrocarbons. Gas-in-place of the Marcellus Formation, then, reduces to the sum of the free (~ 60-70% of the total GIP) and adsorbed gas. Core calibrated log-derived GIP estimates of the Marcellus Formation (Fig. 2-2) are locally specific to account for changes in stratigraphy over short geographic distances. Regionally, the greatest gas saturations recorded in the Marcellus (locally >1 Bcf/sqmi/ft) are associated with MFS/condensed intervals where detrital clay volume is lowest and quartz content is highest. Perhaps more important, though, is the strong correlation of GIP and total organic carbon (TOC; Fig. 2-3). This relationship, too, is reflected in sequence stratigraphy.

Authigenic uranium (U) and molybdenum (Mo) concentrations suggest that the Marcellus Formation accumulated under episodically anoxic or even euxinic conditions (Fig. 2-4), especially conducive to the preservation of organic matter. Further, enhanced primary productivity of surface waters, perhaps triggered by an infusion of nutrient-rich waters into the foreland, likely contributed to a high organic carbon flux to the sea floor. The relatively shallow water depth hypothesized for the Marcellus basin (Smith, 2010; Lash and Engelder, 2011) and consequent short residence time of organic particles in any oxygenated portion of the water column in tandem with an absence of organic consuming benthos would have favored the preservation of organic carbon.

The complex interrelationship of sedimentation rate and organic carbon preservation (i.e., Ibach, 1982) is nicely illustrated by the Marcellus Formation. The general reduction of TOC up-section through the Marcellus reflected in gamma-ray and density log behavior is suggestive of dilution by detrital sediment, principally clay. The generally basin-wide increase in aluminum (Al), a proxy for detrital clay, upward through the Marcellus (Fig. 2-5) appears to confirm a close relationship of TOC and detrital flux. Indeed, calculated original TOC is lowest in those regions of the basin where detrital clay lobes recognized in well logs, are thickest (Fig. 2-6). On the other hand, the relatively high calculated (based on compacted thickness) sedimentation rate of the Marcellus Formation [9-70 m/MY (30-230 ft/MY)] would have quickly removed accumulated organic matter from zones of chemical and biotic degradation near the sediment-water interface thereby enhancing its preservation. Still, the siliciclastic sediment (clay) flux during accumulation of TST and condensed interval deposits of the Marcellus (Fig. 5; also see Fig. 10) appears to have been low enough to have precluded dilution of organic matter but high enough to have averted any significant organic matter degradation.

The robust co-variance of TOC and gas saturation of the Marcellus Formation reflects two critical aspects of the contained organic matter; 1) organic particles are the sites of adsorbed gas and 2) amorphous organic matter and bitumen represent the dominant sites of intraparticle porosity development (Fig. 2-7). The latter point is best illustrated by porosity trends documented from across much of the basin. In essence, total porosity co-varies linearly with TOC to values as high as 12% porosity (Fig. 2-8). Recent investigations (e.g., Loucks et al., 2009) have demonstrated the relationship of TOC and porosity to be principally a function of thermal maturity. At relatively low thermal maturity, perhaps to a maximum level of $\%R_o \approx 1.0$, organic grains behave in a ductile manner that, during burial-related compression, results in the occlusion of pore throats and consequent diminished porosity and permeability of organic-rich intervals (Lash and Engelder, 2005). Indeed, heavily bioturbated organic-lean gray shale is generally more porous (Fig. 2-9) and, judging from data obtained by mercury injection capillary pressure analysis, markedly more permeable than associated organic-rich shale (e.g., black shale permeability = 0.00028 md; gray shale permeability = 0.00528 md). At higher levels of thermal stress ($\%Ro > \approx 1.1$), however, organic particles host development of nanoporosity thereby increasing the gas storage potential of the most organic-rich intervals of the succession (e.g., Loucks et al., 2009). Thus, assuming that thermal maturity has attained a threshold level ($\%R_o > \approx 1.1$, depending on organic matter kinetics), TST/condensed interval deposits, in addition to being especially organic-rich, may have a markedly enhanced GIP storage potential.

Silicon (quartz) Enrichment

The presence and distribution of quartz and clay within the reservoir are integral to the structural integrity of the rock and its ability to initiate and maintain conductivity across hydraulically fractured intervals. The abundance of quartz in the Marcellus Formation and other Devonian black shale units appears to reflect base-level fluctuations as expressed in sequence stratigraphy. X-ray diffraction analysis of sidewall core samples recovered from TST deposits of the MSS1 T-R sequence in northern Pennsylvania, for example, reveals the

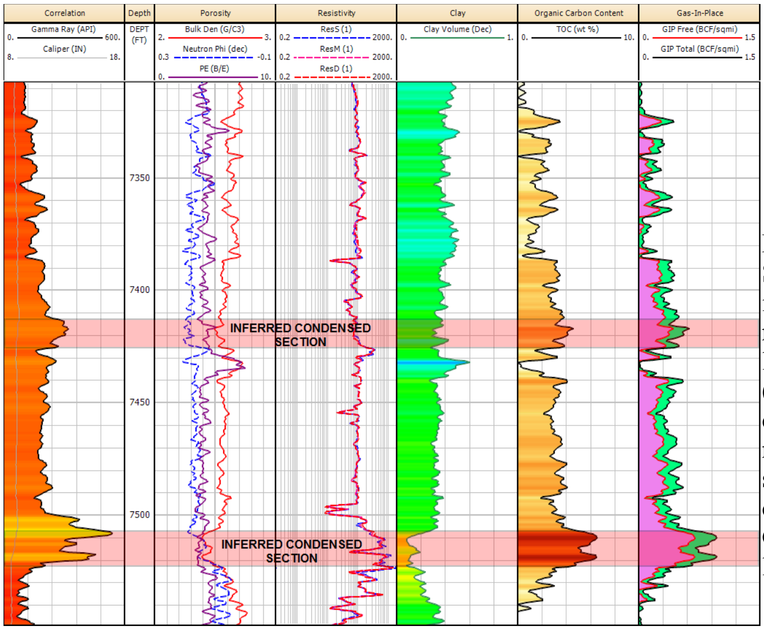

Figure 2-2: Marcellus Shale log suite, including bulk density, neutron, and resistivity logs. Clay volume (third from right) is calculated from the neutron-density log separation. Core-calibrated, log-derived GIP is illustrated in the log on the right.

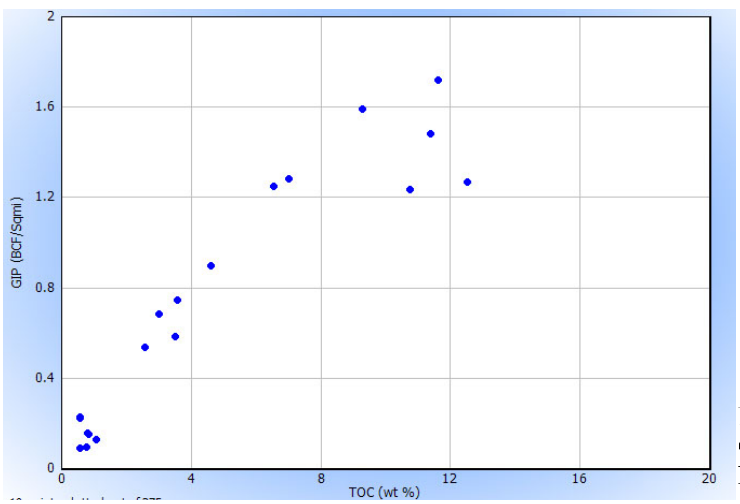

Figure 2-3: TOC versus GIP; data from a Marcellus well.

presence of quartz well in excess of that observed in immediately overlying RST deposits (Fig. 2-10). It is noteworthy shift of marine environments at this time. Thin section and scanning electron microscopy reveals that much of the quartz in the TST /condensed section is microcrystalline and likely derived from the dissolution of opaline silica tests. Occasional angular quartz and feldspar grains may be windblown detritus. It is worth noting that calcite is as much as three-times as abundant in TST deposits as in overlying RST (Fig. 2-10). Much of the calcite occurs as single crystals or patches of microspar and microcrystalline aggregates that originated from styliolinid fragments. Finally, peaks in pyrite and TOC are roughly coincident with the inferred MFS (Fig. 2-10).

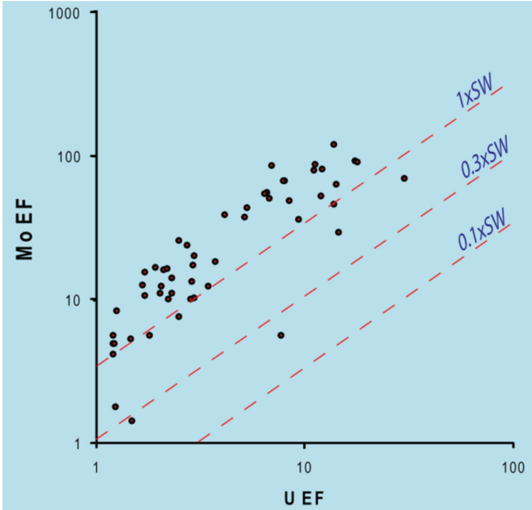

Figure 2-4: Mo_{EF} versus U_{EF} for the Marcellus Formation based on XRF data obtained from a Marcellus core. EF (enrichment factor) enables one to compare sample elemental abundances normalized to aluminum relative to elemental abundances of an "average" shale. Dashed red lines show Mo/U molar ratios equal to seawater (1xSW) and fractions of seawater; the dataset reveals a trend parallel to that of seawater, but enriched by a factor of 2-3X that of normal seawater. Such Mo-U co-variation is most consistent with operation of a Mn-Fe particulate shuttle system as a means of accelerating Mo transfer from intermittently sulfidic bottom waters to the sea floor (Algeo and Tribovillard, 2009).

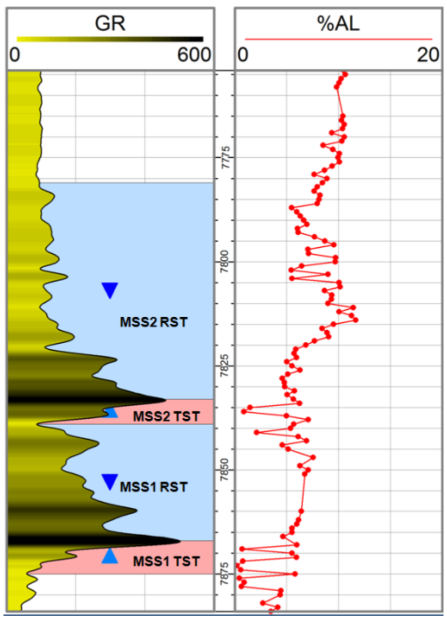

Figure 2-5: Gamma-ray (left) and %aluminum (right) logs through the Marcellus Formation (including the T-R sequences). Note the (1) general increase in Al through the unit and (2) markedly reduced Al in TST deposits.

Abundances of Al, potassium (K), and silicon (Si) suggest that total quartz in the Marcellus exceeds that amount expected for derivation from a siliciclastic source alone. The strong co-variance of K and Al, both proxies for detrital clay, illustrates an unmistakable siliciclastic signal (Fig. 2-11A). However, the Si-Al cross-plot (Fig. 2-11B) reveals Si in excess of that expected of a detrital trend (i.e., strong co-variance with Al). The bulk of the "excess" Si resides in TST deposits (see Fig. 2-13) most likely as authigenic quartz. Indeed, thin section analysis of Marcellus samples reveals that detrital quartz grains, defined by angular edges, are far more common to RST deposits than TST and condensed section deposits.

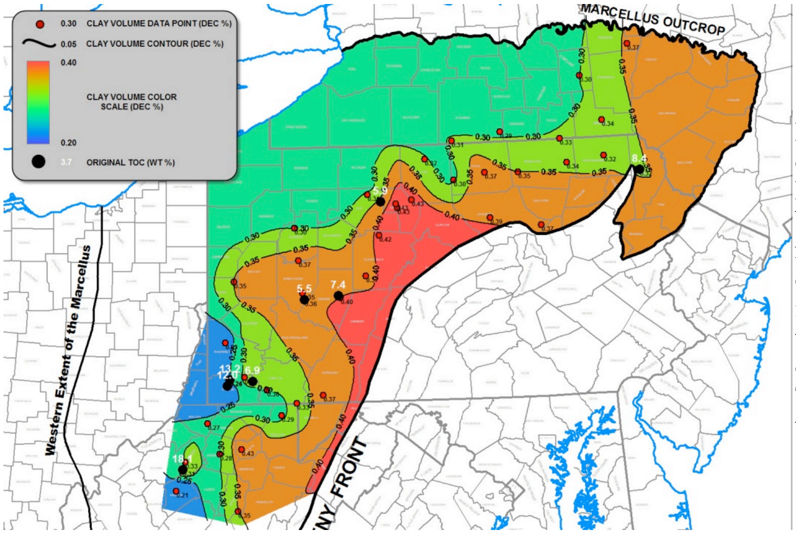

Figure 2-6: Percent clay map for the Marcellus Shale based on use of a neutron-density double clay indicator calibrated to resistivity. Note calculated original TOC values in white.

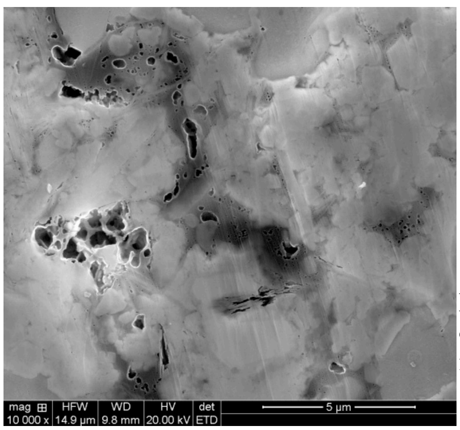

Figure 2-7: Intraparticle porosity development in organic grains (dark patches). Sample is argon ion milled.

Some have suggested that much of the quartz present in the Marcellus Formation is of eolian origin. The titanium (Ti) to Al ratio has been used as an indicator of wind-transported sediment (Boyle, 1983), the premise being that lighter clay grains are winnowed from sediment leaving a relatively coarse-grained deposit enriched in quartz and denser Ti-bearing minerals. Ti/Al ratios over much of the Marcellus succession are less than 0.05 (Fig. 2-12). Further, notably elevated Ti/Al values observed in TST deposits (Fig. 2-12), as much as an order of magnitude higher than Ti/Al values documented from oceanic sites of known eolian influx (e.g., Boyle, 1983; Bertrand et al., 1996), reflect the lack of clay (Al) and perhaps the effects of winnowing, both expected during rising base level, rather than abnormally high abundances of Ti. High Ti/Al values documented from within MSS2 RST deposits in several studied Marcellus cores from across the basin (Fig. 2-12) likely record a pulse of coarse-grained sediment delivered to the basin at this time of lowering base level.

In summary, we suggest that the most likely source of silicon (quartz) enrichment in the Marcellus Formation is the early dissolution and recrystallization of opaline tests (radiolarian). Such an interpretation is further supported by a generally strong co-variance of TOC and quartz (Fig. 13). However, the debate over the source and distribution of quartz in the Marcellus is more than an academic issue. Biogenic quartz, by virtue of its mode of formation, permeates the clay grain fabric thereby enhancing the structural rigidity of the rock. Detrital or eolian quartz grains are certainly high modulus particles, yet their more randomly disseminated distribution throughout the clay matrix or, alternatively, concentration in discrete laminae may not yield a uniformly high modulus rock. The more pervasive nature of authigenic quartz would be expected to increase the rigidity of the rock beyond that associated with detrital quartz. As such, TST/condensed section deposits are likely to be more favorable to fracture stimulation than are adjacent RST strata. Indeed, analysis of well behavior histories appears to confirm the proposed relationship between well stimulation and resultant production and sequence stratigraphy (i.e., TST/condensed intervals).

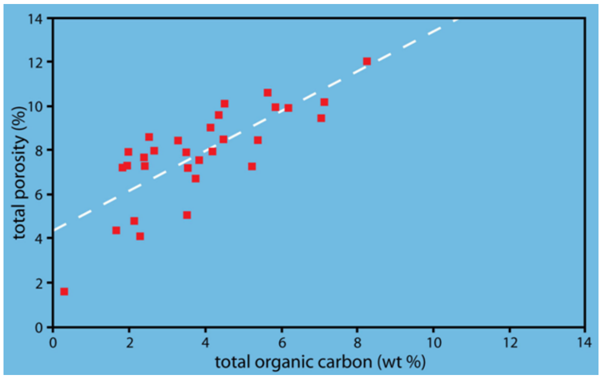

Figure 2-8: TOC – porosity cross-plot, Marcellus Formation; %Ro >1.4%.

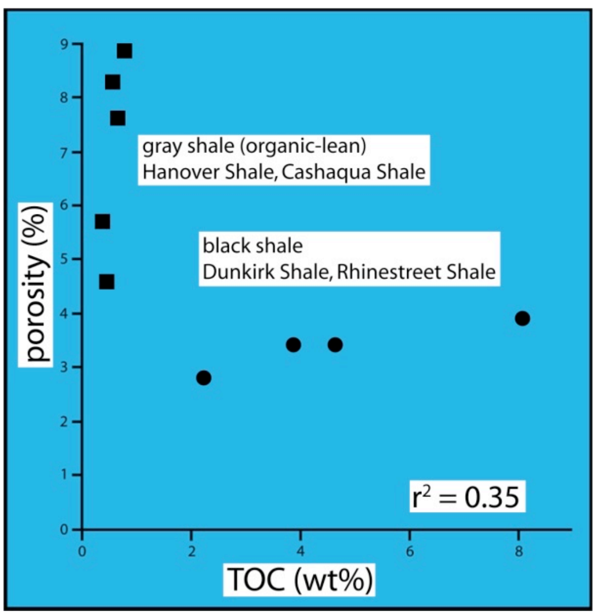

Figure 2-9: TOC – porosity cross-plot, Upper Devonian outcrop samples, western New York; %Ro = 0.6-0.74%.

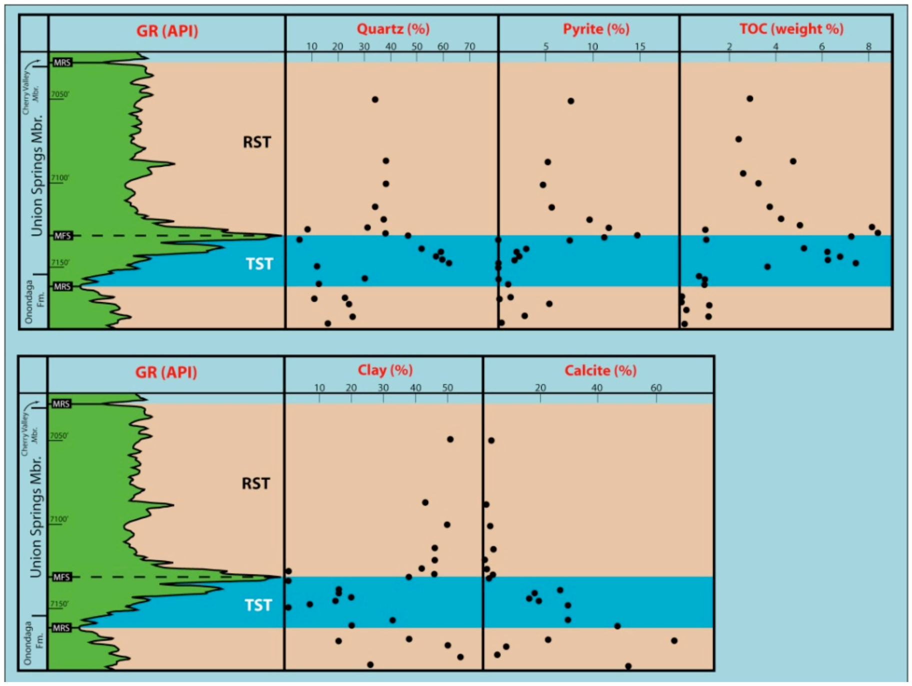

Figure 2-10: Mineralogy and TOC trends through the MSS1 depositional sequence. Data based on X-ray diffraction analysis of a suite of sidewall core samples recovered from a Marcellus well.

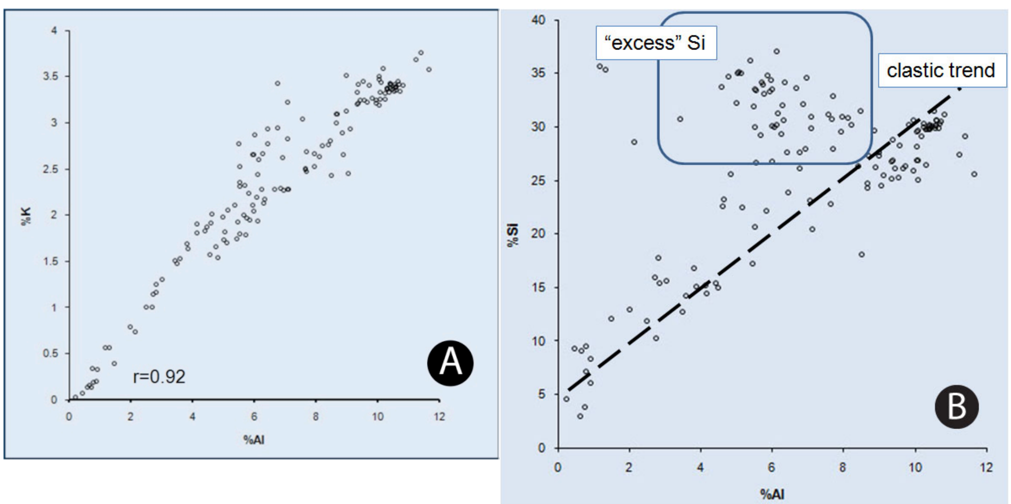

Figure 2-11: %K-%Al (A) and %Si-%Al (B) cross-plots; data obtained by handheld XRF analysis of a Marcellus Formation core.

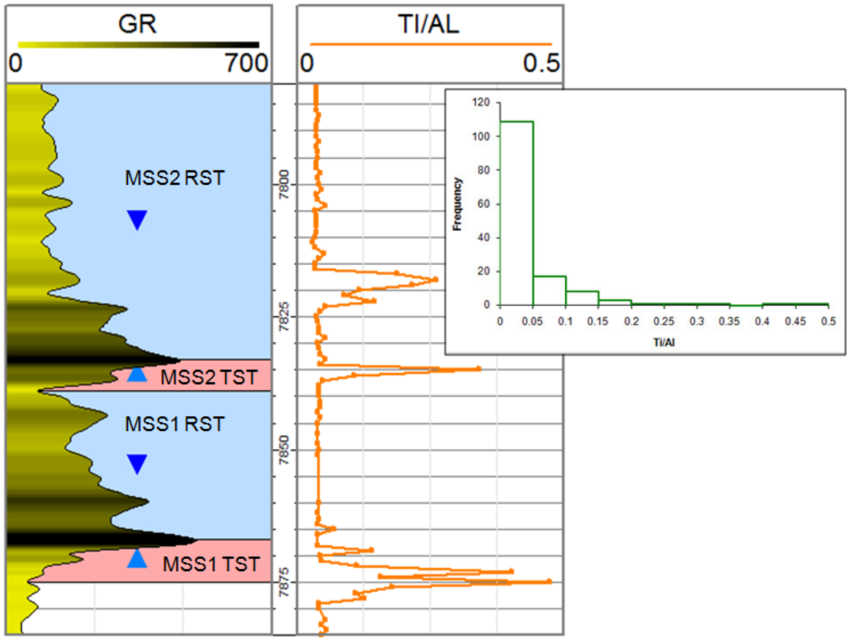

Figure 2-12: Gamma-ray log and corresponding plot of Ti/Al for a Marcellus well and core; inset shows a frequency plot of Ti/Al values (n=142). Note the great number of Ti/Al values less than or equal to 0.05. Note also the Ti/Al excursion in MSS2 RST between 7,816 and 7,822 feet.

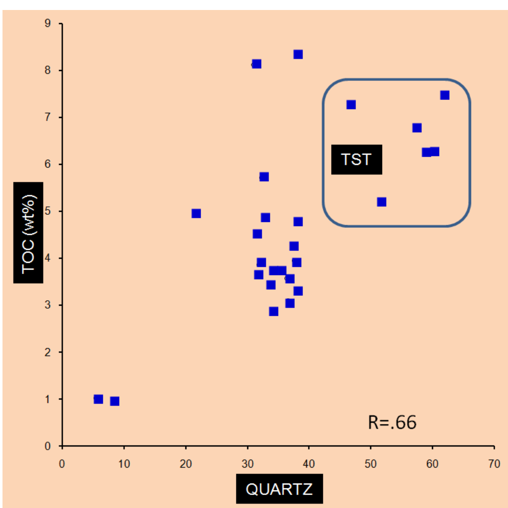

Figure 2-13: TOC – quartz cross-plot; data from a suite Marcellus core samples. The box labeled "TST" denotes those samples recovered from transgressive systems tract deposits

<u>The distribution of clay and its bearing on mechanical rock properties</u>

The clay fraction of the Marcellus Formation comprises illite and lesser amounts of chlorite and smectite/mixed layer clays (Fig. 2-14). Clay volume, reflected in the abundance of Al, increases upward through the Marcellus into low-TOC gray shale of the Hamilton Group (Fig. 2-5), likely a reflection of protracted progradation of the Catskill Delta complex across the basin. For a given Marcellus T-R sequence, clay is generally lowest in TST deposits, increasing through the overlying RST succession (Fig. 2-10). Increasing clay is accompanied by a general reduction of reservoir quality, principally a consequence of the dilutive effects of clay on organic matter, the principal gas reservoir, and reduced biogenic quartz. Moreover, increased clay is accompanied by elevated volumes of bound water.

There is little doubt that certain mechanical rock properties are reflective of clay volume. Indeed, Poisson's ratio of a clay-rich rock generally exceeds that of a rock comprised of lesser clay. Further, the tendency of a rock to expand laterally under an applied load, the Poisson effect, may be enhanced by a laminated texture and/or anisotropic platy-grain fabric, especially in clay-rich successions (Fig. 2-15). The strongly oriented clay grain microfabric of some Marcellus samples, especially those recovered from unbioturbated clay-rich intervals within RST successions, yields an *in situ* stress anisotropy that increases rather dramatically from negligible values (< 2% in TST/condensed section deposits) to as high as 20% in clay-rich horizons (Fig. 2-16). Further, the Poisson effect, being a rough measure of ductility, is important to considerations of the degree of proppant embedment one might expect during stimulation (e.g., Kinley et al., 2008). That is, those intervals defined by elevated Poisson's ratio may experience greater degrees of proppant embedment and consequent loss of conductivity of induced hydraulic fractures. Moreover, continued production of the reservoir and consequent reduction of reservoir pressure would yield an increase in the effective normal stress on the rock thereby increasing the Poisson effect (i.e., *in situ* confining stress) in clay-rich intervals.

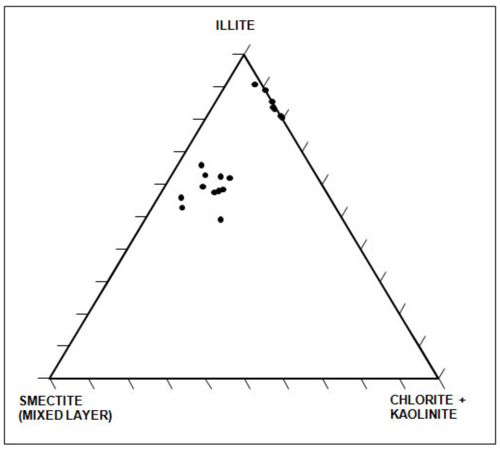

Figure 2-14: Clay mineralogy plot for data from a Marcellus Formation well, northeastern Pennsylvania.

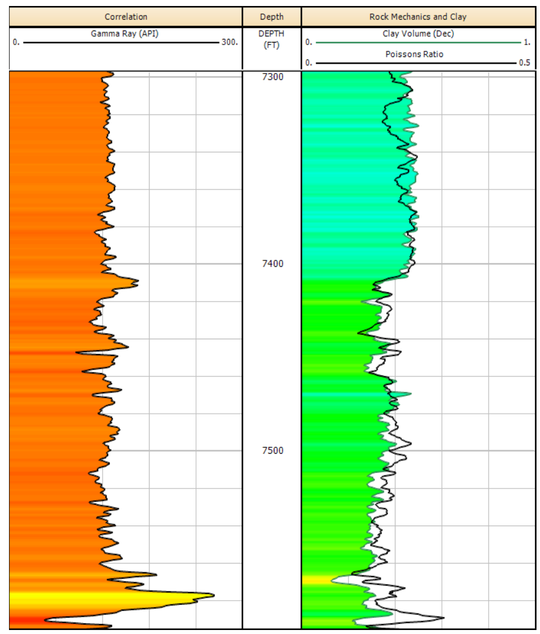

Figure 2-15: Gamma-ray log (left) and combined rock mechanics (Poisson's Ratio) and clay volume logs (right).

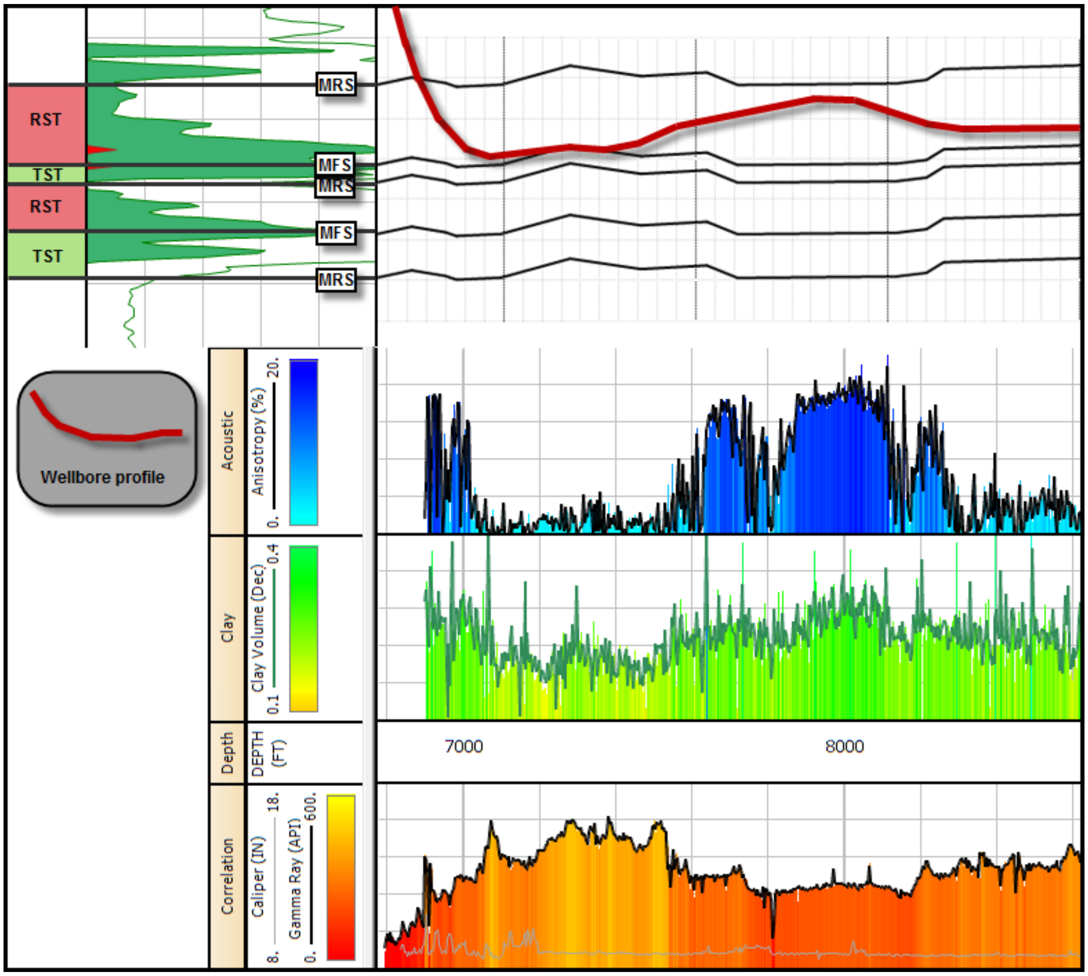

Figure 2-16: Correlation of acoustic shear wave anisotropy (top horizontal curve) and clay volume (middle curve) calculated from a suite of geophysical well logs run along a horizontal wellbore. Note the increase in clay volume and shear wave anisotropy as the lateral drifted upward away from the MSS2 MFS farther into clay-rich RST deposits. Note also the variation in gamma-ray response (lower horizontal curve) relative to the MSS2 MFS.

Conclusions, Chapter 2

The Marcellus Formation T-R sequence stratigraphy offers a predictive framework for source rock and reservoir assessment that can be extrapolated into areas of poor data control (Lash and Engelder, 2011) and can also be very useful to stimulation strategies as well as decisions related to the targeting of laterals (Blood, 2011). Compositional attributes that influence such critical reservoir properties as porosity and brittleness, including the abundance and type of quartz and clay volume, vary predictably as a consequence of base level oscillations. Consideration of a number of petrophysical attributes of the Marcellus Formation in terms of the T-R sequence stratigraphic framework demonstrates that TST/ condensed interval and lower RST deposits contain the greatest abundance of organic matter. Further, in those regions of the basin that have been thermally matured to the dry gas window, the development of intraparticle porosity has enhanced the gas storage potential of the most organic-rich deposits. Moreover, relatively porous organic-rich TST/ condensed interval deposits contain the greatest abundance of biogenic quartz and minimal clay yielding a rock that is more susceptible to fracture stimulation. The heterogeneous nature of petrophysical properties of the Marcellus Formation, reflected in the unit's base level history and consequent sequence stratigraphy, are certainly important considerations to the successful exploitation of the unconventional Marcellus Formation gas play.

Chapter 2 Acknowledgements

Lash and Blood thank EQT and Chief Oil and Gas for permitting us to publish proprietary data used in this study. The support of Chesapeake Energy, EQT, Shell, Seneca Resources, and ThermoFisher, especially Chris Smith, formerly of ThermoFisher, has enabled Lash and Blood to begin a regional study of trace element and metals in the Devonian shale succession.

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CHAPTER 3.
SEDIMENTOLOGY OF UNION SPRINGS FORMATION

Ceren Karaca and Teresa Jordan

Introduction

The Devonian successions of the Appalachian Basin have been discussed widely and a stratigraphic framework for the entire basin has been established by many authors, including Cooper (1930a, 1930b), Rickard (1975), Sevon and Woodrow (1985), Brett and Baird (1996), Ver Straeten (2007), Ver Straeten et al. (2011), and Lash and Engelder (2011). One of part of Devonian successions, the Eifelian-aged Marcellus subgroup (Ver Straeten and Brett, 2006) has recently been the center of interest for its significance as an unconventional shale gas reservoir. The Marcellus subgroup has two organic rich black shale units: 1. Union Springs 2. Oatka Creek. Ver Straeten and Brett (2006) proposed that the ranking for both members be raised to formation level as the Union Springs Formation and the Oatka Creek Formation, respectively. Here, we present the results of a detailed sedimentological study of the Bakoven Member, which comprises the entire Union Springs Formation in central New York State. This study is a part of an integrated black shale research project that utilizes sedimentology, paleontology, geochemistry, and stratigraphy in order to understand the paleoenvironmental conditions during Unions Springs deposition.

Union Springs Formation at Seneca Stone Quarry

A complete succession of Bakoven Member (Union Springs Formation) was investigated in Marcellus shale exposures at the Seneca Stone Quarry. There, the Bakoven Member overlies the Tioga F K-Bentonite Bed (Brett and Ver Straeten, 1994) and underlies the Cherry Valley Limestone (Ver Straeten et al., 1994; Figure 3-1). It is an organic rich calcareous black shale facies that has a total thickness of 3.80 meters.

The recently excavated northern wall of the quarry features fresh exposures of black calcareous shale. The rocks are mostly hard, faintly weathered and show a low degree of fissility. At the outcrop scale the approximately 4-meter thick package of rock shows subtle vertical variations. The most prominent features are the early diagenetic calcitic concretions that are traceable throughout the quarry. Bedding planes are visible in cross section. However a careful examination of samples iwas s necessary to retrieve further data. The study of fresh rock slabs and ultra-thin sections reveal significant small-scale variations that are not visible at the outcrop scale. Also, oiling and polishing the slabs enhance the visibility of small-scale structures within the shales.

Detailed Sedimentology of Union Springs Formation

Features of note include the grain composition and size, type of matrix support, composition, fossil content, sedimentary structures including lamination, bioturbation and soft sediment deformation, and diagenetic structures including concretions. Based on those features, the Union Springs Formation is subdivided into three lithofacies that have distinct characteristics (Figure 3-1).

Packstone Lithofaices. The lowest lithofacies starts with a thin 10-15 cm thick black organic-rich laminated carbonate mudstone layer and, with increased abundance of bioclasts upward, grades from wackestone into packstone (Dunham, 1962) towards the top (Figure 3-2). Overall, this facies is dominated by calcite composition, constituting a clay-rich limestone, with a packstone texture, so it is referred to as the “packstone facies.” It is composed of articulated and disarticulated ostracod shells, bivalve debris, echinoid spicules, and conical fossil grains, supported by a matrix of silt-sized calcite grains and clay cement. Laminations, although disturbed by

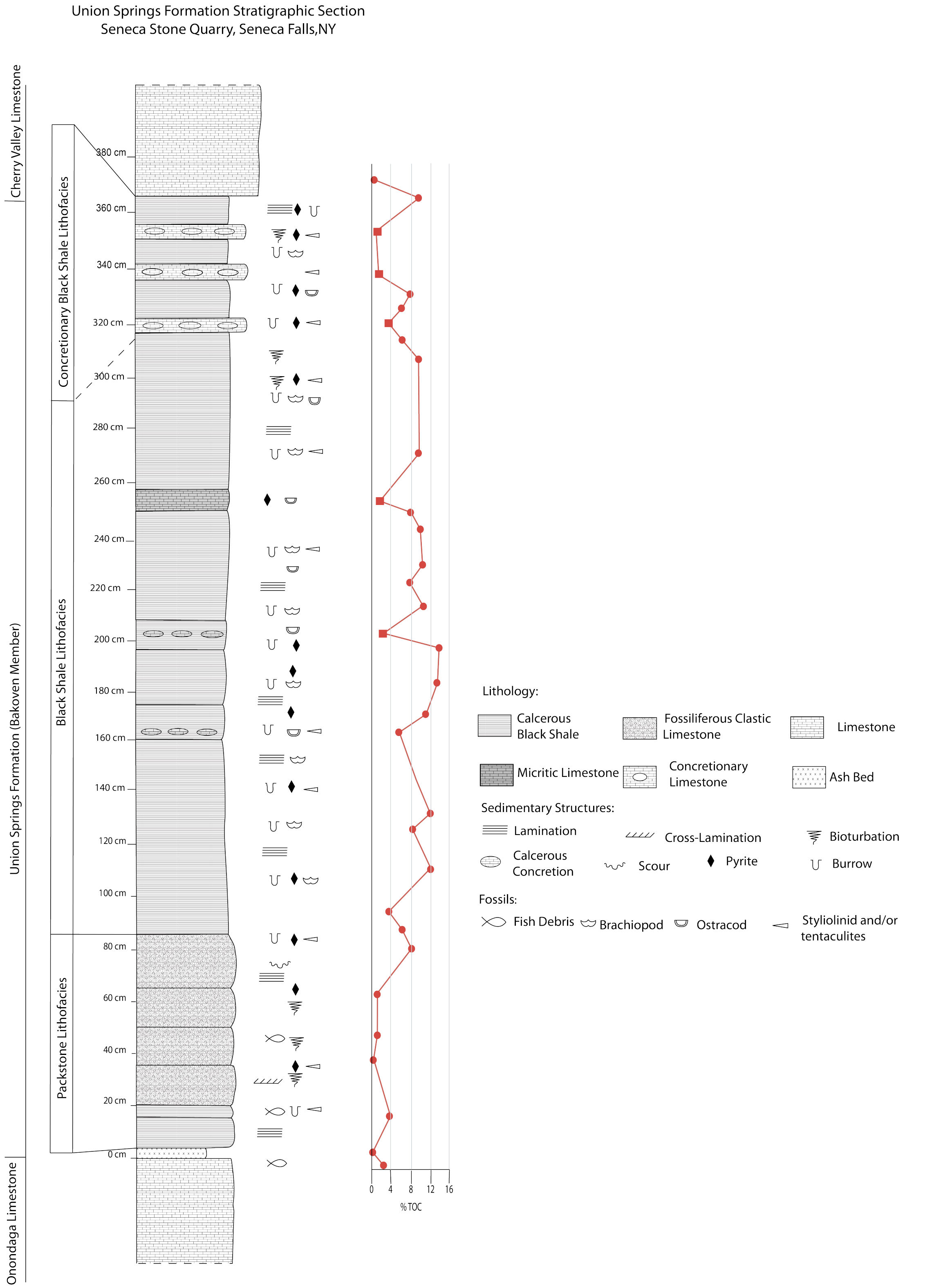
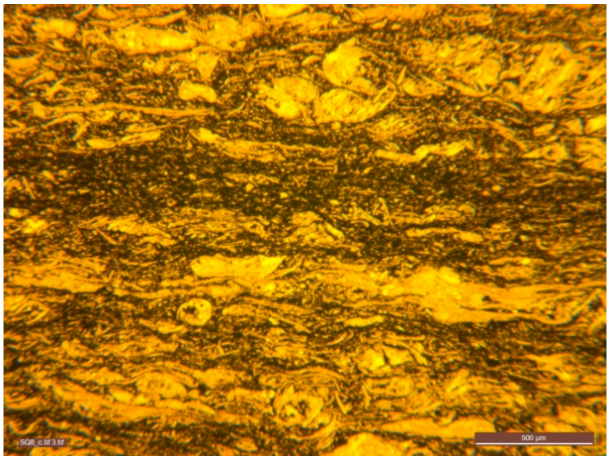


Figure 3-1: Stratigraphic section of Union Springs Formation with associated TOC values. Filled circles represent samples from shale whereas filled boxes represent samples from concretions.

bioturbation and stylolites, are still visible. At the lowest level of the succession, cross laminations are also observed (Figure 3-3). At silt rich layers, we recognized micro-erosional surfaces and micro-graded bedding. While original laminations are disturbed by bioturbation, pseudo black laminations that formed by pressure solution (e.g., stylolaminated fabric, Flugel, 2010) are common in this lithofacies (Figure 3-2). Vertical and horizontal burrows are mostly compressed and original calcite burrow fills are replaced by pyrite at some sections. In the thin section of a sample collected 40-50 cm above the Tioga F bed, four conodont or fish fragments (Jeff Over, personal communication, 2011) are recognized (Figure 3-4). We correlate this bone-bearing interval to the Bakoven bone bed of Ver Straeten et al. (1994). In addition to the conodonts, densely packed shell beds are the other features of this layer. Towards the top, the packstone facies grades into grainstones; the bioclasts increase in abundance and they are supported by sparry calcite. At this level, sphalerite associated with pyrite is present as ostracod fillings. The total organic carbon (TOC) is generally low; around 1% for this facies. However, at the sphalerite-rich zone, TOC peaks to 8%.

Figure 3-2: Transmitted photomicrograph of stylolaminated packstone. Dark layers are pressure solutions that are formed due to loading or tectonism. Scale bar is 500 µm.



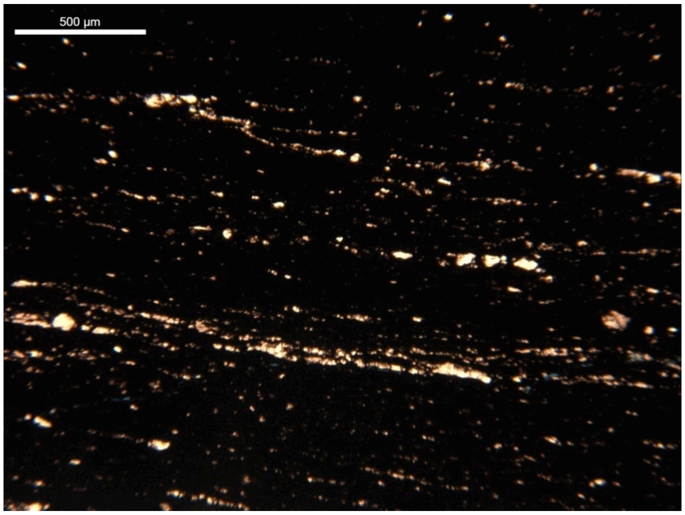
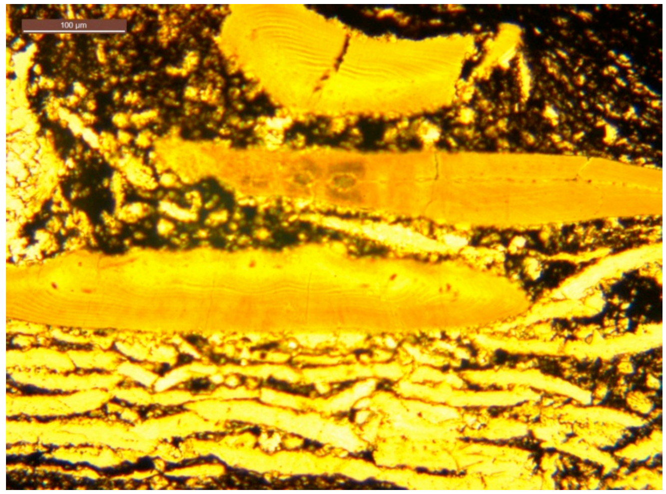


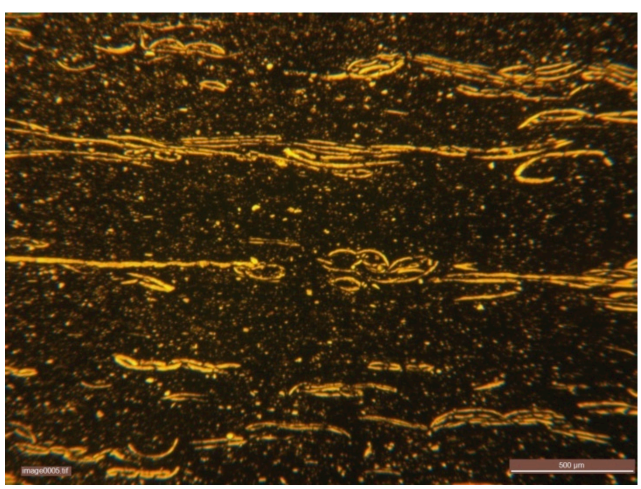
Figure 3-3: Transmitted photomicrograph of organic rich layer showing cross laminations. The sample is from the lowest level of the Bakoven member. The uppermost lamination, which parallels the top edge of the thin section, is the horizontal reference. Scale bar is 500 µm.

Figure 3-4: Transmitted photomicrograph showing conodont elements (large) and compressed and fragmented ostracods (lower half). Scale bar is 100 µm.



Black Shale Facies (Bakoven Member). Overlying the packstone facies is a 2-meter thick, black organic-rich shale succession within the Union Springs Formation. The fauna of the black shale facies is limited to styliolinids, ostracods, and the brachiopod *Leiorhynchus*. Fine silt-sized calcite grains are present in an organic matter and clay rich matrix. This facies is characterized by parallel laminations that are fully preserved. In thin section, the contacts between alternating laminations are faint, whereas in hand samples they are identifiable by subtle color changes in tones of dark grey and black. Layers rich in disarticulated shells are concordant with laminations (Figure 3-5). Load structures under the shell fragments were formed due to the differential compaction of the underlying muds (Figure 3-6). There are horizontal burrows present and these are usually filled with microscopic pyrite crystals (Figure 3-7). These highly compacted burrows can be misinterpreted as discontinuous laminations. In addition to burrow fillings, pyrite is present as replacement crystals of shell material and as pyrite framboids that are recognizable at thin section scale. Sometimes whole laminae are composed of pyrite crystals and these are visible at the outcrop scale as shiny fine gold streaks. The highest TOC of the entire succession is measured in this black shale lithofacies. The average TOC is 10% and the highest measurement is 14%.

Figure 3-5: Transmitted light photomicrograph of black shale lithofacies. Ostracod and brachiopod shell fragments are concordant with laminations and they are concentrated at specific levels. Scale bar is 500 µm.



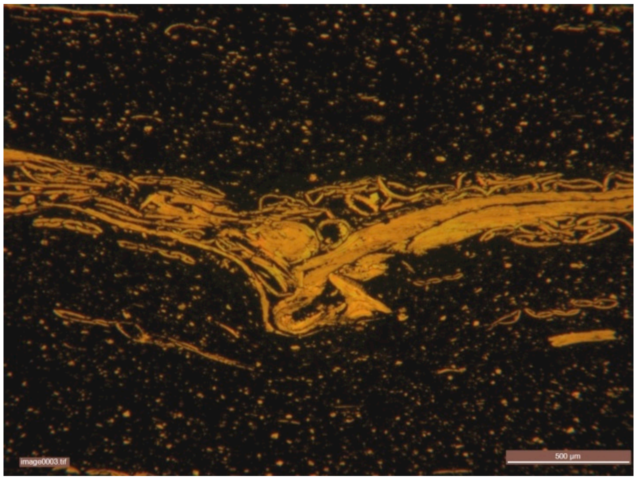
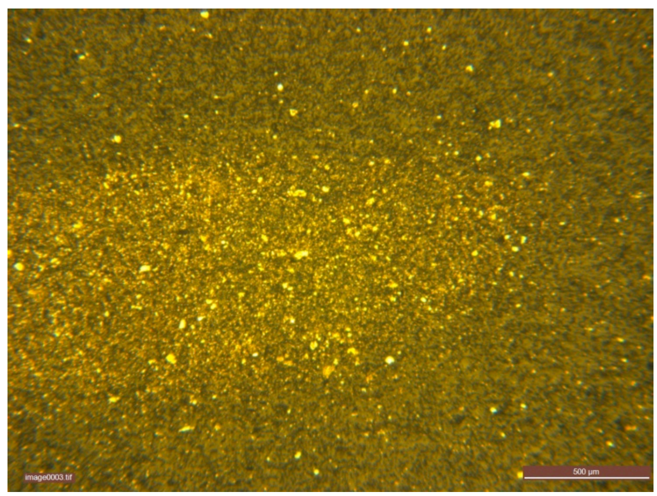


Figure 3-6: Transmitted light photomicrograph showing the load structure under the shell load. Scale bar is 500 µm.

Figure 3-7: Reflected light photomicrograph showing a pyritized burrow. Probably the organic material in the burrow is replaced by pyrite. Scale bar is 500 µm.



Concretionary Black Shale Facies. The top 50 cm section of Bakoven Member has significant calcitic concretions (Figure 3-8) and at some levels these concretions become laterally continuous defining this interval as the concretionary black shale facies because black shale layers alternate with concretion-rich zones. The transition from underlying organic-rich black shale facies to concretionary black shale facies is not distinct. The grain size increases slightly from fine to medium silt size. The grains are mostly made up of calcite with some quartz in accessory amounts. The black shale layers have lower TOC values (around 7%) than the underlying black shale facies. The grains are composed of styliolinids and ostracods in a clay and organic matter-rich matrix; in this facies we no longer see *Leiorhynchus*. Silt- and pyrite-filled burrows are present and mostly compacted, however in the concretions we can see uncompacted silt-filled burrows. If not close to a concretion, black shale laminations are horizontal and parallel. Around the concretions, the laminations fold and follow the relief of the concretion. The TOC of the black shale layers is around 8%; however, the concretionary layers are significantly lower (around 2%).

Summary and Discussion

In the deposiiton of lithofacies with different characteristics, important roles were played by base level variations, clastic input, paleo-circulation and paleo-oxygen levels, biologic productivity (Bohacs et al., 2005, Demaison and Moore, 1980, Pedersen and Calvert, 1990, Werne et al., 2002).

Figure 3-8: An example of a calcitic concretion in the concretionary black shale lithofacies.

The degree of bioturbation and distinctive fossil assemblage of several species indicate that during the deposition of the wackestone-packstone facies paleo-oxygen levels were high enough to maintain biological activity. The low TOC values also support the interpretation of oxygenated bottom waters, because, if not, organic matter would have been preserved through sulphate reduction when there was no oxygen present for aerobic decay (Demaison and Moore, 1980, Pedersen and Calvert, 1990, Bohacs et al., 2005). The cross laminations at the beginning of the sequence and the erosional features show that current activity played a role in the deposition of this fine grained lithofacies.

The conditions under which the shales of the black shale and concretionary lithofacies were deposited were different than the previous conditions. The oxygen level declined drastically, permitting limited species to live, such as *Leiorhynchus* which is a brachiopod that is restricted to gray and black shale facies and tolerates dysoxic (low oxygen) conditions (Thompson and Newton, 1987). The abundance of various types of pyrite indicates that the low oxygen conditions favored pyrite formation via several processes such as pyritization of organic matter in the burrow fills, pyrite framboids, replacement of shell material, and pyritization of entire lamina (Fisher and Hudson, 1985, Schieber, 2002a, 2002b, Wilkin et al., 1996).

The genesis of the concretions of the third lithofacies is not fully understood. The behavior of the black shale lamina around the concretions (folding and differential compaction) indicates an early diagenetic origin. If these are early diagenetic concretions, formed during the final stages of Union Springs deposition, probably the clastic input was low (Taylor et al., 1995).

In order to better understand the paleoenvironmental conditions under which the Union Springs was deposited, a detailed integrated study of sedimentology, paleontology, geochemistry, and stratigraphy is mandatory. Here we have presented the results of a high resolution sedimentological study of the Union Springs Formation. Future work will include incorporation of chemical analysis for whole rock bulk elemental composition and clay mineralogy. Those chemical data will enable a better understanding of the nature and origin of this organic-rich fine grained facies.

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CHAPTER 4.

CORRELATION OF MARCELLUS SUBGROUP AND RECOGNITION OF THE EIFELIAN-GIVETIAN BOUNDARY UTILIZING MAGNETIC SUSCEPTIBILITY IN WESTERN NEW YORK

D. Jeffrey Over, Shannon Rabideau, Matthew Travis, Brynne Grady, and Charles Ver Straeten

Distinctive lithologic units and recognition of sequence stratigraphic packages enable correlation of strata within the Marcellus subgroup across the Appalachian Basin (Ver Straeten, 2007) as well as to global event horizons (Brett et al., 2011). Recognition of the Eifelian-Givetian (E-G) boundary has proven more problematic in eastern North America due to the dearth of conodonts which does not allow recognition of the conodont zones found in Europe and North Africa, and endemic nature of dacryoconarids that are used to recognize and define the stage boundary. At the Global Stratotype Section and Point located at Jebel Mech Irdane in Morocco the E-G boundary (Walliser et al., 1995) is also marked by a magnetic susceptibility low during the Kačák-*otomari* Interval (House, 1985; Walliser, 1996; Crick et al., 2000) and a shift to higher MS values corresponds to the Eifelian-Givetian Boundary. Magnetic susceptibility of sedimentary rocks is essentially a determination of the concentration of detrital iron-bearing minerals, the result of weathering, and relates to climate change and sea-level fluctuation (Ellwood et al., 2000, 2011) that has great potential for high resolution correlation. In the Appalachian Basin the placement of the E-G Boundary has been difficult due to poor conodont control and the numerous black shale intervals that have been questionably associated with the Kačák-*otomari* Interval.

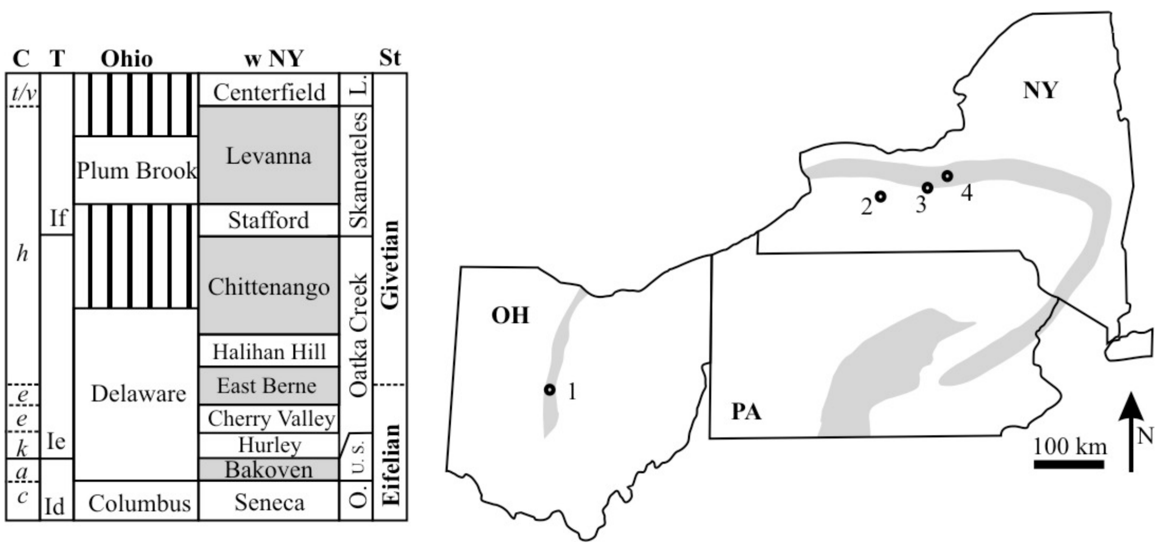

Figure 4-1 – A. General stratigraphy, conodont zonation, and major T-R cycles of the Marcellus subgroup (Unions Springs and Oatka Creek formations) and equivalents. O = Onondaga Formation; US = Union Springs; L = Ludlowville Formation; *c = costatus; a = australis; k = kockelianus; e = eiflius; e = ensensis; h = hemiansatus; t/v = timorensis/varcus*. B. Map of Marcellus outcrop distribution and location of measured sections/cores illustrated in Figures 4-2 and 4-3. 1 = Lazarus Run; 2 = East Groveland, NY; 3 = Seneca Stone Quarry; 4 = West Limestone Creek.

Magnetic susceptibility in the E-G boundary interval was determined from four closely spaced cores in the Genesee River Valley near East Groveland – in the relatively thin distal black shale portion of the Marcellus subgroup – as well as sections measured in outcrop from Ohio and New York (Figures 4-1 to 4-3). This demonstrates depositional packages of varying thickness, even in close proximity, within the distal Appalachian Basin, and tentative placement of the E-G Boundary in the lower part of the East Berne Member of the Oatka Creek Formation (Figure 4-3).

Marcellus subgroup Depositional Packages in Western New York

Four drill cores that penetrated the Marcellus Subgroup near East Groveland, New York (Figure 4-1B) were analyzed for magnetic susceptibility at 5 or 20 cm intervals from the top of the Seneca Member of the Onondaga Formation, through the entire Marcellus, to the Stafford Limestone Member of the Skaneateles Formation. All MS values were determined using an Agico MFK1-A Kappabridge and internal/inhouse standards. The magnetic susceptibility data are framed within distinct lithostratigraphic units, notably the Hurley and Cherry Valley members of the Oatka Creek Formation, the Halihan Hill Bed at the base of the Chittenango Member, and the Stafford Limestone Member (Figure 4-2). The Bakoven and Hurley-Cherry Valley have a relatively low magnetic susceptibility value, overlain by strata of the East Berne and Chittenango members that have relatively higher values. Within the Oatka Creek shale-rich strata there are five discrete packages characterized by shifting trends in the MS values (Figure 4-2). The nature of these trends are not clear in regard to lithologic changes, but they persist within the Oatka Creek Member which thins from 12 to 10 meters across 5 km (Figure 4-2). Within the Oatka Creek the East Berne thickens from 2 to 3.2 m. Similar changes in thickness over short distances were described by Ver Straeten et al. (1994) in the Bakoven from the Honeoye Falls Quarry further to the north.

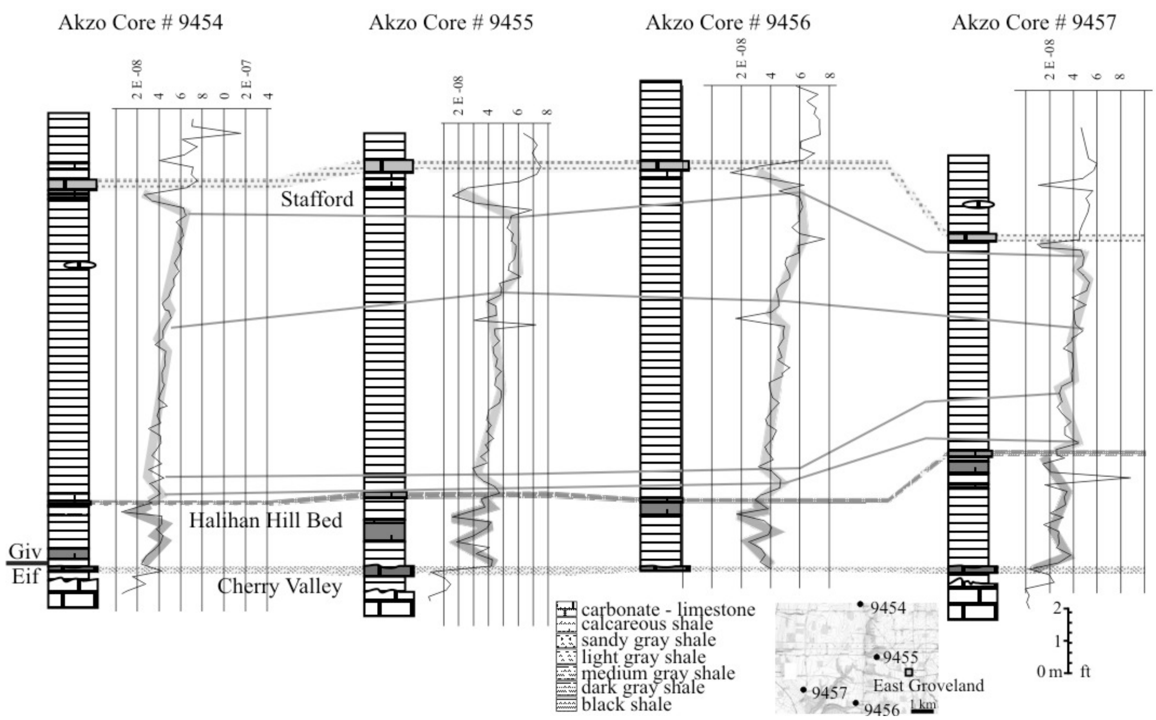

Figure 4-2 – Lithology and magnetic-susceptibility (M3/kg) of four drill cores in the Genesee River Valley near East Groveland. Key beds are correlated by patterned intervals – datum = top of the Cherry Valley Member. MS values are shaded by splined trends which are correlated by light gray lines.

Eifelian-Givetian Boundary

The Eifelian Givetian Boundary is a time of significant global change in the Middle Devonian, marked by the end of the Kačák-*otomari* events and defined by the first occurrence of the conodont *Polygnathus hemiansatus.* Based on conodonts, the Hurley and Cherry Valley Limestone members are in the *kockelianus* Zone (Klapper, 1981). The next higher zone defining conodont, *Polygnathus timorensis*, has been recovered in the Centerfield Limestone at the base of the Ludlowville Formation marking the *varcus* Zone and the end of the lowest Givetian *hemiansatus* Zone (Klapper 1981). Givetian macrofossils such as *Mediospirifer audaculus*, *Athyris cf. A. cora*, and *Tornoceras mesopleuron* first occur in the Dave Elliot Bed and the Halihan Hill Bed, as well as an un-named bed below the Dave Elliot Bed in eastern New York (Bartholomew et al., 2009; Brett et al., 2011). Based on these fossils and a distinct positive shift in magnetic susceptibility the E-G boundary in central and western New York is tentatively placed at the top of the Cherry Valley Limestone, or slightly above in the East Berne Member, where higher MS values become stable (Figure 4-3). This shift is evident in the Delaware Limestone in central Ohio, the platform carbonate equivalent to the Marcellus subgroup, above a distinct carbonate bed that is equivalent to the Hurley and Cherry Valley limestone members based on bio- and sequence stratigraphic interpretation

(DeSantis et al., 2007). Delineation of the E-G Boundary in the lower East Berne indicates that the lower Oatka Creek Formation is equivalent to the Kačák-*otomari* interval (see Brett et al., 2011).

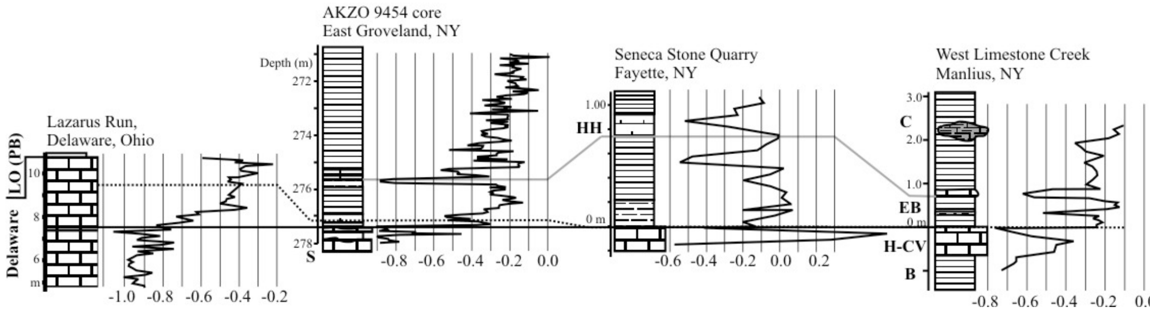

Figure 4-3 – Lithology and magnetic susceptibility (δMS) of Marcellus subgroup strata and equivalents in Ohio, western New York, and central New York – note different scales of measured sections. Datum = top of the Cherry Valley Limestone and equivalent; the Eifelian-Givetian Boundary is marked by a dotted line. LO (PB) = Lower Olentangy (Plum Brook); S = Seneca; B = Bakoven; H-CV = Hurley - Cherry Valley; EB = East Berne; HH = Halihan Hill Bed; C = Chittenango.

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ROAD LOG FOR MARCELLUS FIELD TRIP

CUMULATIVE MILEAGE	MILES FROM LAST POINT	ROUTE DESCRIPTION
000	00	**Start of Road Log — Heroy Hall, Syracuse University.**
000.5	00.5	North on Crouse Drive – Crouse Avenue to Harrison Street
000.8	00.3	Turn west (left) on Harrison Street to signal at Almond Street
001.0	00.2	Signal, turn north (right) onto Almond Street - get in left lane
001.6	00.6	From Almond Street, ramp to I-81 North to I-690 toward Baldwinsville
010.4	08.8	Exit onto I-690 West toward Baldwinsville
025.5	15.1	Take Exit 1 to I-90 West – toll road - toward Buffalo
026.2	00.7	Exit 40 – NY 34 to Weedsport/Auburn ($ 0.70)
026.5	00.3	Turn south (right) onto NY 34 South
034.9	08.4	Signal, turn east (right) onto NY 31 West
039.3	04.4	Turn south (left) onto NY 90 South
044.5	05.2	Signal, turn west (right) onto NY 5 / US 20 West to Seneca Falls
048.6	04.1	Signal, continue straight onto NY 414 South
049.7	01.1	Turn east (left) onto Canoga Road –Co Road 121. Turn into Seneca Stone Quarry, proceed through the gate, and park on the left.

STOP 1. Marcellus subgroup at Seneca Stone Quarry (parking area on left, inside the gate)

This large active quarry exposes Devonian strata of the Eifelian and lower Givetian, as well as minor folding and a thrust fault distinctly visible on the southern wall; the Onondaga Limestone is the primary production unit. In the lowest level the Lower Devonian Manlius and Oriskany formations floor the quarry. Four members of the Onondaga are recognized in respective upward succession: Edgecliff, Neadrow, Moorehouse, and Seneca. The Seneca Member is easily recognized on the quarry faces as it is bound by two distinct ash beds that weather and stain the walls orange. The lower ash is the Tioga B – Onondaga Indian Nations Ash, which marks the base of the Seneca; the Tioga F is used to recognize the base of the Marcellus subgroup (Figure RL-1A). The entire Bakoven member of the Union Springs Formation (Figure RL-1B) and the succeeding lower Oatka Creek Formation, consisting of the upward succeeding Hurley, Cherry Valley, East Berne, and Chittenango members (Figure RL-1C) can be clearly seen in the quarry benches and walls.

The Bakoven Member, approximately 4 m thick here, overlies dark gray biowackestones of the Seneca Member and represent deposition in the HST and FSST of the 1d/Eif-2 sequence. The Tioga F Ash, approximately 10 cm thick, defines the base of the Marcellus and Union Springs Formation. The lowest part of the Bakoven is dark gray muddy biowackestones and packstones that contain ostracodes, bivalves, pelmatozoa, and dacryoconarids. Seventy centimeters above the base of the Bakoven is a disconformity that preserves abundant fish material, including acanthodian spines, crassopterygian teeth, and placoderm plates, preserved in thin grainstones that are developed in decimeter-scale wide shallow channels. These channels may be analogs of younger, linear erosional furrows associated with a channeled disconformity surface within the lower part of the dark Levenna Shale Member in Erie County, New York (see:Baird et al., 1999; stop 6). Above this disconformity are finely laminate black shales and thin calcareous beds, where the carbonate material is

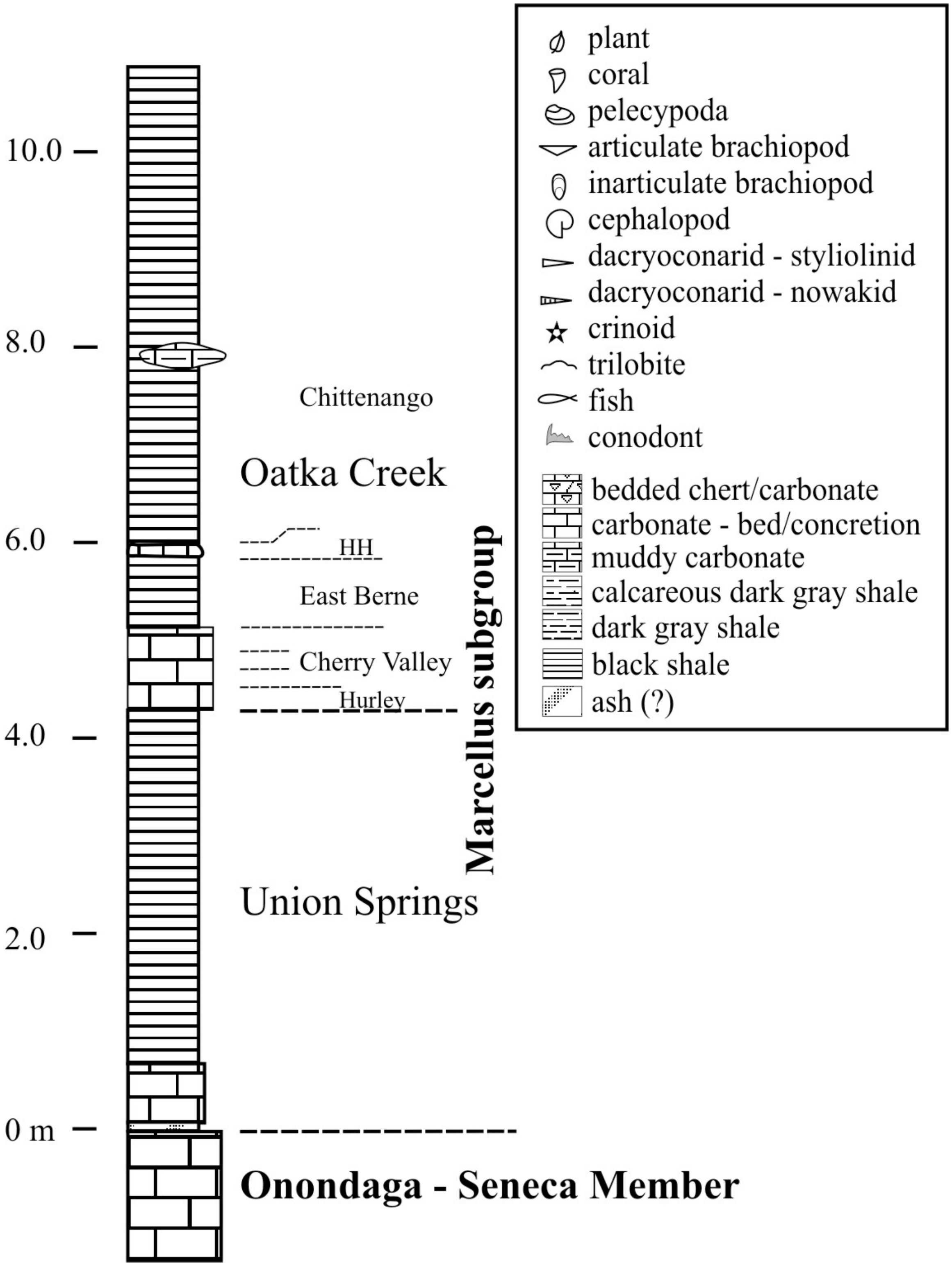

Figure RL-1A. General stratigraphy of Marcellus subgroup strata at Seneca Stone Quarryand key to symbols; HH = Halihan Hill Bed.

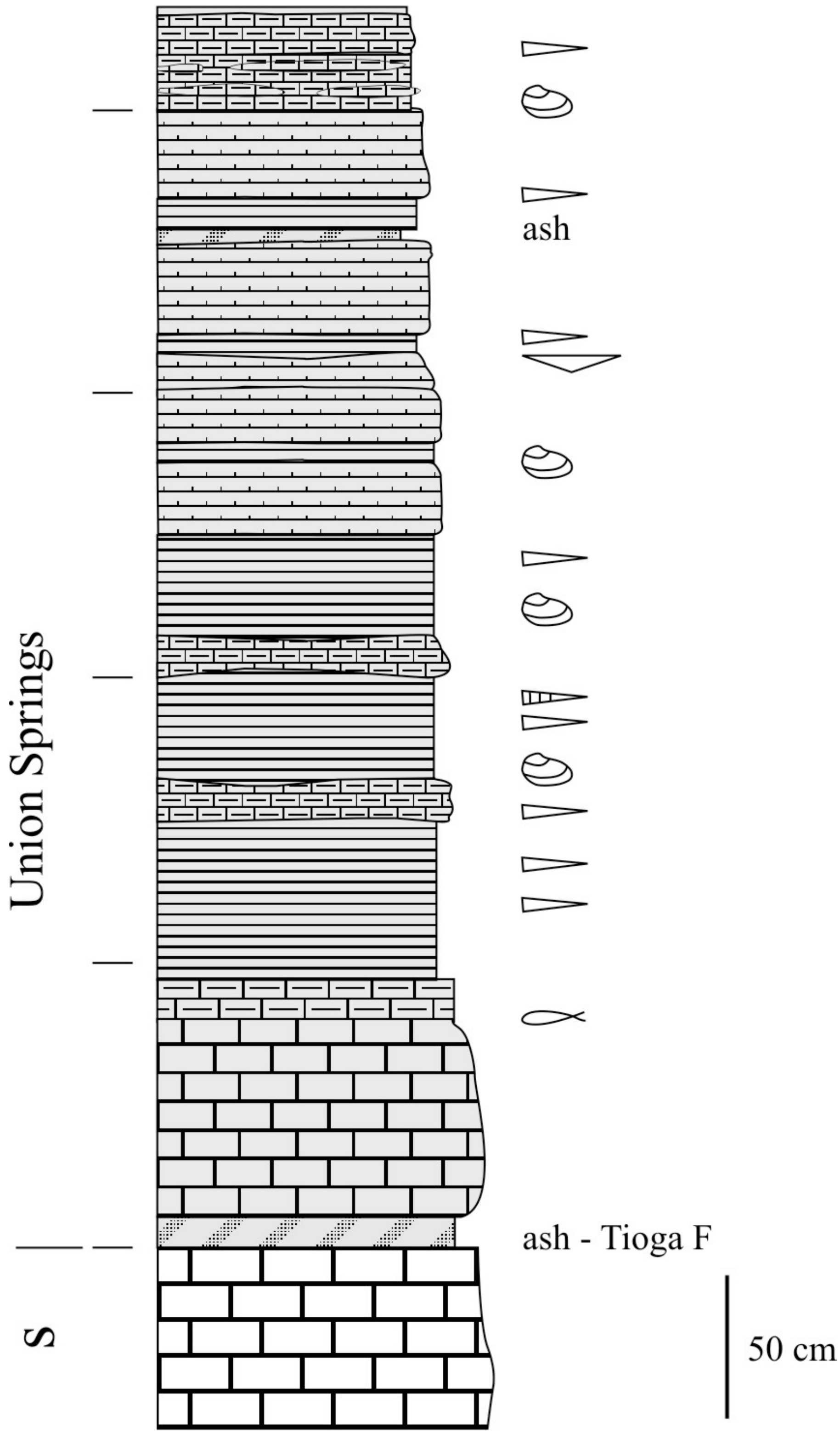

Figure RL-1B. General stratigraphy of the Union Springs Formation above the uppermost Seneca Member of the Onondaga Limestone at Seneca Stone Quarry.

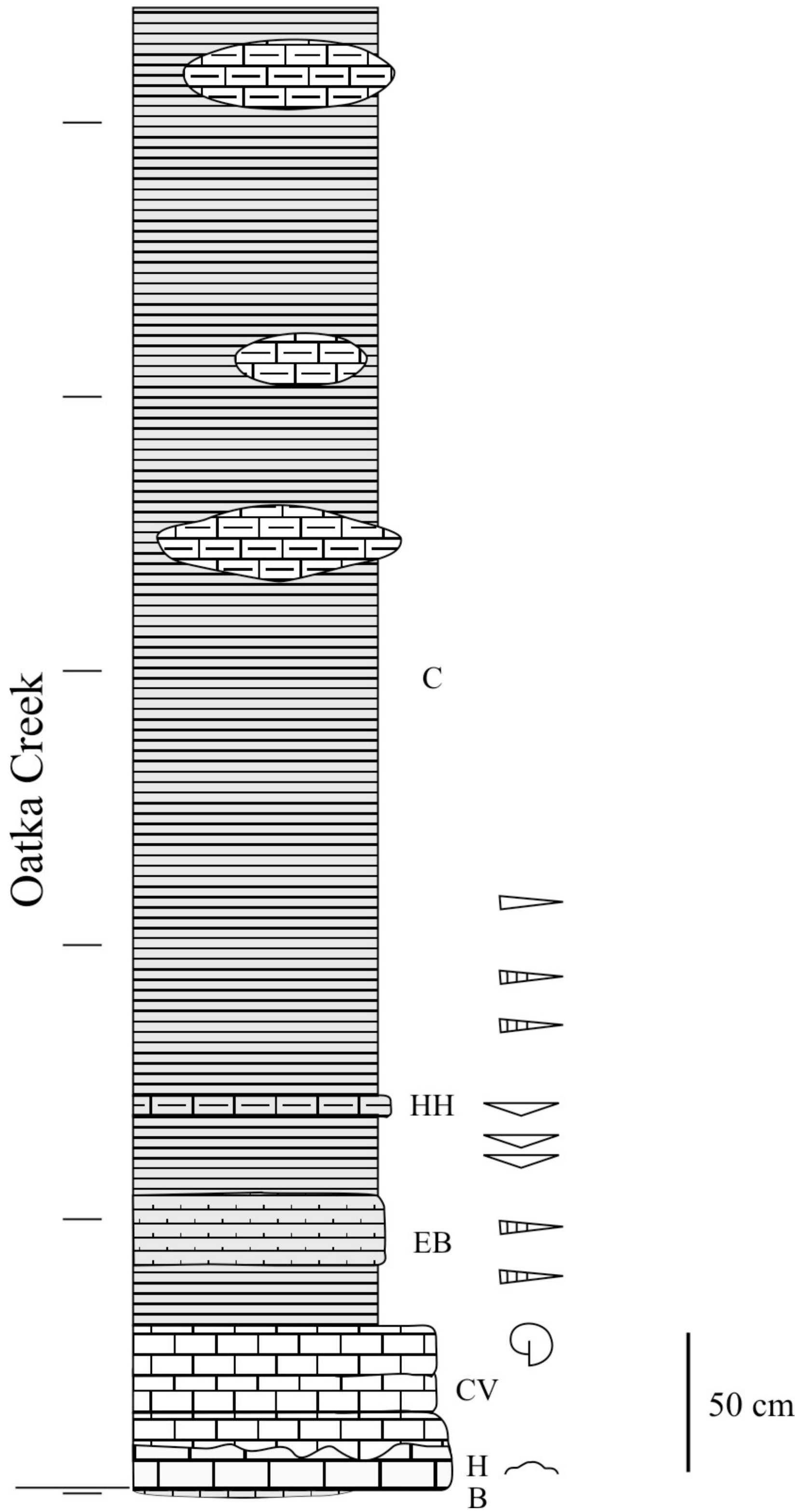

Figure RL-1C. General stratigraphy of the Oatka Creek Formation at Seneca Stone Quarry. B = Bakoven; H = Hurley; CV = Cherry Valley; EB = East Berne; HH = Halihan Hill; C = Chittenango.

dominated by *Styliolina*, other dacryoconarids, and thin brachiopods that sometimes form thin grainstones and packstones. The upper Bakoven contains nodular carbonates and laminated black shales.

The Oatka Creek Formation is comprised of four ascending members: Hurley, Cherry Valley, East Berne, and Chittenango. The base of the 1e/Eif-Giv sequence is the Hurley Member which consists of light-weathering very fossiliferous biopackstone-grainstone, characterized by the proetid trilobite *Dechenella*, which is amalgamated onto the base of the Cherry Valley Member to form a single ledge-forming carbonate approximately 50 cm thick. The Hurley – Cherry Valley contact is an irregular hardground surface. The Cherry Valley is brown weathering, fossiliferous biopackstone-grainstone, and is remarkable for numerous large, truncated orthicone nautilods and *Agoniatites vanuxemi* exposed on a corrosion surface, that based on sedimentological evidence, must have been exposed on the Devonian sea floor. These strata record desposition during the TST.

Black styliolinid-rich shales of the East Berne Member overlie the Cherry Valley Member. Several brachiopod-rich horizons occur in the lower 50 cm, which represent the first appearance of the Hamilton fauna, characterized by diverse corals and brachiopods. The Halihan Hill Bed is a 20 cm thick gray calcareous mudstone that marks the appearance of a more diverse shelly fauna, including *Mediospririfer, Ambocoelia, Athyris, Pseudoatrypa*, and others, and delineates the base of the Chittenango Member. These strata fall within the late TST of the 1e/Eif-Giv sequence. Based on magnetic susceptibility, the Eifelian-Givetian boundary is placed at the Cherry Valley – East Berne contact. The lower Chittentango is composed of organic-rich black shale containing variably rich styliolinid- and brachiopod-bearing horizons. Three horizons of large – upto 1 m in diameter - calcareous concretions are developed between 3 and 5 m above the base. These concretions are apparently also developed in the lower Chittenango Member at the Oatka Creek type section in LeRoy, NY as well as in strata near Marcellus (Stop 3). The Chittenango represents deposition during the HST and FSST; the base of sequence If/Giv-1 will be examined at Stop 2.

050.5	00.8	Exit parking lot, turn east (left) on Canoga Road-Co. Road 121
050.7	00.2	Turn north (left) onto Seybolt Road-Co Road 121
051.7	01.0	Turn east (right) onto Canoga Street-Co Road 121
054.7	03.0	Turn north (left) onto NY 89 North
054.8	00.1	Turn east (right) onto Willows Hill Road toward Cayuga Lake
054.9	00.1	Turn north (left) onto Lower Lake Road-Co Road 116 to Cayuga Lake State Park

LUNCH STOP: Cayuga Lake State Park

055.0	00.1	Turn south on Lower Lake Road-Co Road 116
055.1	00.1	Turn west (right) on Willows Hill Road
059.0	03.9	Turn north (right) on NY 89 North
067.5	08.5	Signal, turn east (right) onto NY 5 / US 20 East
068.0	00.5	At blinking Signal turn south (right) onto Half Acre Road
069.6	01.6	Stop sign and junction with NY 326 – continue on Half Acre Road
069.7	00.1	Turn west (right) onto farm lane – park near calf shed

STOP 2. Lockwood Family dairy farm quarry, Half Acre/Oakwood

This small quarry exposes strata of the uppermost Oatka Creek Formation (Cardiff Member) and lowermost Skaneateles Formation (Mottville Sandstone Member; Figure RL-2).

Medium dark gray, burrowed silty mudstones of the Cardiff Member (Oatka Creek Formation) comprise the lower approximately 2.3 m of the section here. The Cardiff Member here is slightly calcareous, and poorly fossiliferous. Coiled and straight nautiloid cephalopods are part of the sparse fauna; the “hitch-hiking” bryozoan *Reptaria stolonifera* was found on one of these.

The dark gray mudstones of the Cardiff are overlain by buff-colored, calcareous very fine sandstones of the Mottville Member. In this area, basal Skaneateles strata are in a transition from sandstone- to limestone-dominated facies (coeval Mottville and Stafford members, respectively). Baird et al. (1999) used the term

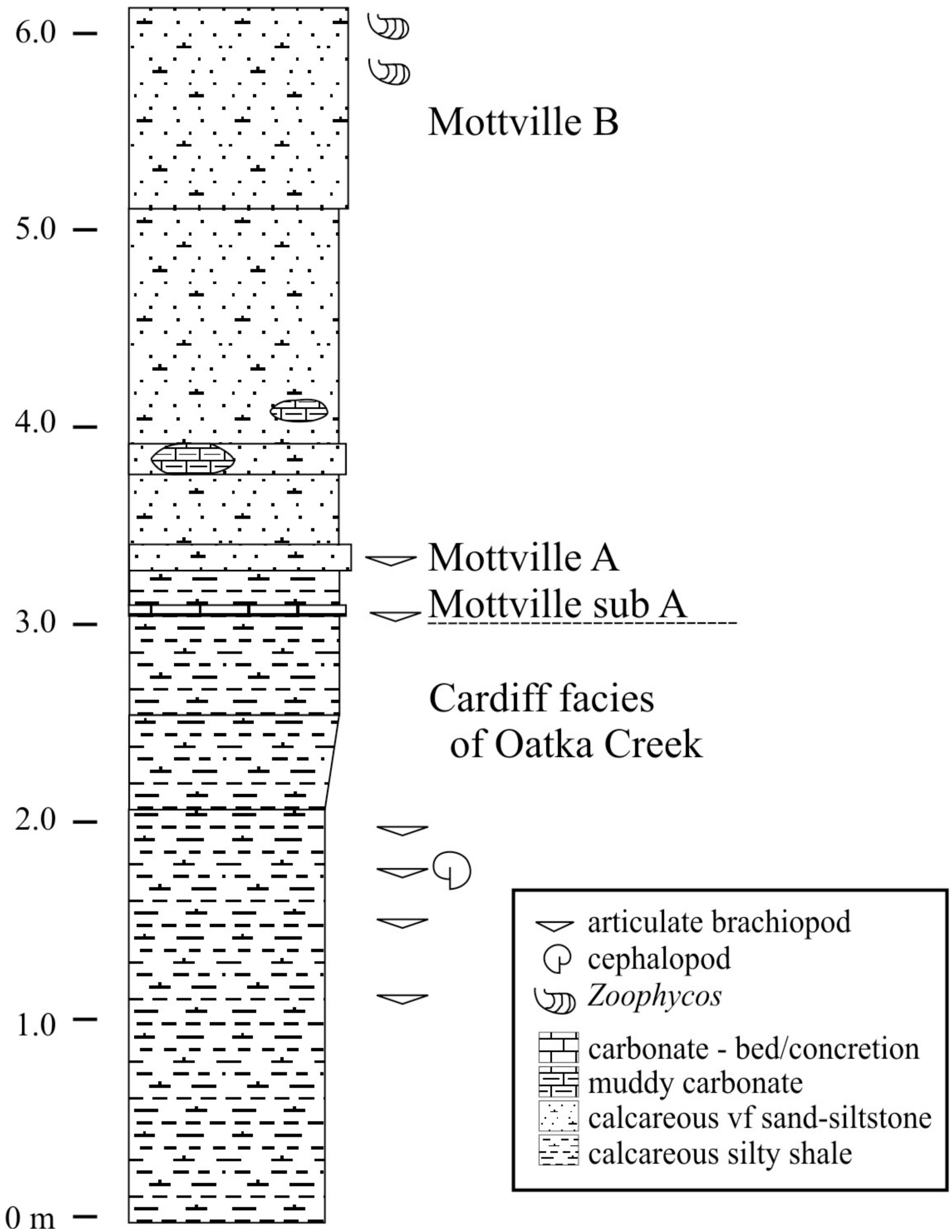

Figure RL-2. General stratigraphy of the upper Oatka Creek Formation and base of Skaneateles Formation – Mottville Member at Lockwood Farm.

Stafford Member here. Closer examination in the lab indicate an apparent greater percent sand to calcite at this locality, so the term Mottville Member is better applied.

Relatively low in the Mottville Member here a thin, fossiliferous highly calcareous sandstone bed 3.2 m above the base of the section, termed the Mottville A bed (Figure RL-2), is characterized by very fine to lower fine sand-size grains, and benthic fossils including the brachiopods *Cupularostrum*, *Emanuella* and *Devonochonetes*, auloporid corals, and other fauna. The upper part of the Mottville Member here, including the

Mottville B bed, is less fossiliferous; however, at least some parts of it are heavily bioturbated by the trace fossil *Zoophycos* which may, alternatively mark even shallower water conditions than the Mottville A bed below. The greater degree of *Zoophycos* churning of the Mottville B interval may actually indicate the shallowest water conditions, and mark the Ie/Eif-Giv - If /Giv-1 sequence boundary. Although all of the buff-colored, calcareous sand-rich strata have been assigned to the Skaneateles Member, technically all of the strata below the shallowest point in the Mottville are progradational strata that are a part of the Sequence Ie/Eif-Giv FSST and belong to the Oatka Creek Formation. See Baird et al. (1999) for more details of Mottville Member facies and correlations.

069.8	00.1	Return to Half Acre Road – NY 326
071.4	01.6	Turn north (left) onto NY 326
081.0	09.6	Turn east (right) onto Genesee Street – NY 326. Continue through Auburn
085.1	04.1	US 20 joins Genesee Street – continue east through Skaneateles
089.5	04.4	Turn north (left) onto NY 175 toward Marcellus. Roadcuts south of the NY 175-174 intersection with Slate Hill Road at the SE edge of Marcellus.

STOP 3. Marcellus subgroup along NY 175/NY 174 near Slate Hill Road, Marcellus, NY

This roadcut along the east side of NY Routes 174 and 175 at the northwestern base of Slate Hill exposes the lower black shales of the Oatka Creek Formation (Chittenango Member). Additional nearby sections, however, expose most of, or all, of the Marcellus in its type area, around Slate Hill (Figure RL-3). Hall (1839) did not name a specific type section, but it likely would have been placed at Slate Hill.

Cherty fossiliferous limestones of upper part of the Seneca Member (Onondaga Limestone) are visible in roadcuts on either side of the highway to the north. Intervening strata of the Union Springs Formation and lowest Oatka Creek Formation (including the Hurley, Cherry Valley and East Berne members and the Halihan Hill Bed) are covered southward along the highway to Slate Hill Road.

However, the top Seneca, entire Union Springs through basal East Berne succession, and during some years the Halihan Hill Bed, is exposed in a nearby creek bed, approximately 500 m (0.3 miles) east of the intersection of Slate Hill Road and NY 174-175. The Onondaga-Marcellus contact is sharp here, with no K-bentonite bed at the contact. Topmost beds of the Seneca Member are overlain by 3.1 meters of Bakoven Member black shales with styliolinid to bedded/concretionary diagenetic limestones; the shale/limestone ratio approximates 3:1. These Union Springs strata are capped by a 70 cm-thick ledge of the Hurley and Cherry Valley limestones. Only one 12 cm-thick bed of the Hurley Member is recognized along the stream. Several centimeters of the overlying East Berne black shales were recently observed above the Cherry Valley ledge, and in earlier years the Halihan Hill Bed was also visible along the creek.

Beginning in the ditch near the intersection with Slate Hill Road is an extensive exposure of black, rusty-weathering pyritic fissile shale with multiple levels of small-to-large discoidal septarian concretions. The base of this succession in the lower Chittenango Member is likely close to the top of the Cherry Valley Member and the overlying Halihan Hill Bed. The cut continues to exhibit the same lithology up through the section, exposed southward along NY 174-175 for several 100 meters.

Much of the rest of the Oatka Creek Formation can be studied along a small gully that ascends the northwest side of Slate Hill, beginning at a culvert toward the south end of the roadcut. The section, most of which is exposed in and along the creek bed, rises approximately 40 meters (130 feet) above the road level. The upper part of the succession transitions into dark gray mudstones (Cardiff Member) in the vicinity of a large bend in the creek roughly 23 m in elevation above the road, but the top-Cardiff contact with the overlying Mottville Member is not reached in this locality. The Mottville Member forms the cap of a steep cascade near the top of a deep gorge – "Jacknife Ravine" - approximately 1 km (0.7 miles) to the southeast of the roadcut. This ravine has yet to be fully explored and has potential to yield the most complete Marcellus section in the area.

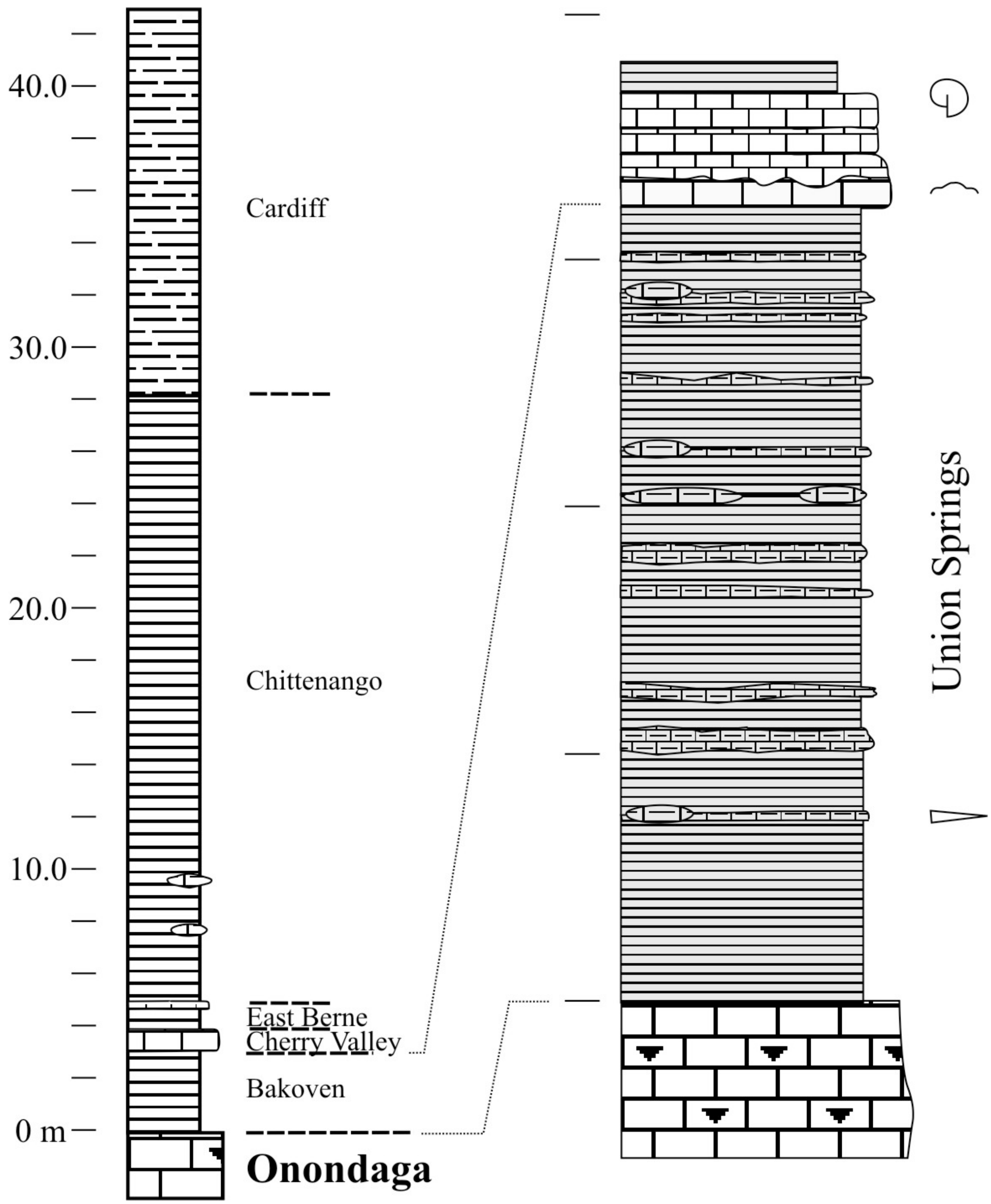

Figure RL-3. General stratigraphy of the Marcellus subgroup in the type area near Slate Hill Road, Marcellus, NY. See Figure RL-1A for key to symbols.

089.8	00.3	Slate Hill Road and Marcellus Fire Department
090.3	00.3	Continue north on NY 175
098.5	08.2	Turn east (right) onto Seneca Turnpike-NY 175
101.1	02.6	Continue onto NY 173

104.3	03.2	Turn south (right) continuing on NY 173
106.5	02.2	Signal and junction with US 11 – continue straight on NY 173
107.2	00.7	Continue on NY 173
110.3	03.1	Signal, turn south (right) on NY 173 toward Jamesville
110.9	00.6	Signal, turn north (left) onto North Street (Jamesville Hardware at corner)
111.1	00.2	Turn east (right) onto Solvay Road and cross Butternut Creek to quarry entrance. Proceed to far southeast end of quarry.

STOP 4: Upper Onondaga Formation and lower Marcellus subgroup at Hanson-Jamesville Quarry

Outcrops at the far southeast end of the Hanson Aggregates quarry at Jamesville expose the Onondaga Limestone and the lower part of the Marcellus subgroup. Key features at this site include a thinner Seneca Member at the top of the Onondaga Formation, a prominent bone bed just above the Onondaga Marcellus contact, and black shales of the Bakoven Member (Union Springs Formation).

Between the Seneca Stone and Jamesville quarries (Stops 1 and 4), the Seneca Member thins from 7.1 m to 5.4 m. Regional correlation of the Tioga A-G K-bentonites indicate that while the Tioga F bed lies at the Onondaga-Marcellus contact at Stop 1, both the Tioga E and F beds are missing from the section. A 5-8 cm-thick bed below the topmost Onondaga Limestone bed is the Tioga D bed. The eastward thinning and top-down absence of Tioga K-bentonites and upper Onondaga strata continues to the Albany area (Seneca Member = 3.2 m at Oriskany Falls; 2.0 m at Cherry Valley; and zero in the Helderbergs). This is indicative of diachronous east to west tectonic-load induced subsidence of proximal areas of the foreland basin during the renewed onset of orogenesis in the Acadian mountain belt.

Given the absence of the Tioga F Ash at Jamesville, the Onondaga-Marcellus contact is placed at the highest Onondaga-like limestone (32 cm thick, above the Tioga D Ash bed). The contact is overlain by a 0-5 cm-thick lag bed rich in phosphatic debris with numerous fish bone fragments and teeth. This bed marks a prominent sediment-starved surface, associated with the end of carbonate production and transport in this region, and the concentration of resistant phosphatic lag material on the sea floor.

The sharp upper surface of the bone bed is overlain by black shales of the Bakoven Member of the Union Springs Formation. Only the lower part of the Bakoven is exposed much here. At times in the past, the top of the Cherry Valley Member could be found in the weeds at the south end of the black shale exposure, 6.5 m above the top of the Onondaga Limestone.

In a nearby creek exposure on private land (south of NY 173 and east of Sweet Road), the Onondaga-Marcellus contact, and approximately 1.3 m of lower Bakoven Member black shales, are succeeded by shales interbedded with thin packages of limestone, followed, in turn, by a covered interval. Upper Bakoven strata are again visible in the stream still farther upsection, where black shales and concretionary to thin-bedded limestones appear to be difficult to correlate between separate stream forks.

In the upper part of this creek section, the prominent Hurley-Cherry Valley limestones ledge (10 and 73 cm-thick, respectively) is capped by black shales of the East Berne Member. The Halihan Hill Bed is found at the top of the creek exposure, 73 cm above the top of the Cherry Valley. Here it is comprised of two distinct layers, a lower 13 cm-thick dacryoconarid-rich bed, and an upper 16 cm-thick layer with brachiopods and other normal marine benthic fauna, indicative of well-oxygenated conditions on the sea floor at that time. The base of the Hurley Member and the base of the brachiopod-rich upper Halihan Hill Bed mark the base of two of three fourth order sequences within the Oatka Creek Formation (Sequence Ie/ Eif-Giv).

A near by abandoned quarry, also on private land, exposes a thick succession of mid (?) to upper Oatka Creek dark gray mudrock facies, mostly corresponding to the Cardiff Member interval. The Mottville Sandstone, the basal member of the Skaneateles Formation, is not seen in the quarry.

111.3	00.2	Take Solvay Road downhill to North Street – Jamesville Road
112.0	00.7	Turn north (right) onto North Street - Jamesville Road
115.3	03.3	Turn west (left) onto I-481 South toward Syracuse
117.6	02.3	Right exit onto I-81 North to Syracuse
117.8	00.2	Exit 18 – Adams Street (to Carrier Dome)
118.0	00.2	Signal, turn east (right) onto Adams Street

118.1	00.1	Turn south (right) onto Irving Avenue
118.2	00.1	Signal, turn east (left) onto Waverly Avenue
118.5	00.3	Turn right onto Crouse Avenue and proceed to Heroy Hall.

End of trip.

Trip A-3

THE GREEN VEDDER MEMBER – A HIGHSTAND SYSTEMS TRACT IN THE "PERITIDAL" MANLIUS FORMATION

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INTRODUCTION

Since the stratigraphic synthesis of Rickard (1962) and the sedimentologic interpretations of Laporte (1967, 1969), the Manlius Formation of the Helderberg Group has attained iconic status representing peritidal environments in the overall transgression of the Helderberg Sea. Although Laporte (1969) described Manlius facies that recorded subtidal conditions (e.g., stromatoporoid biostromes), most workers think of the Manlius as dominated by facies that represent supratidal and intertidal environments. The Manlius actually displays much greater variability of facies than is commonly thought. On this field trip, we will explore the Green Vedder Member of the Manlius Formation (Ebert and Matteson 2003a, b), a distinctive and traceable stratigraphic unit that records highstand (subtidal) conditions within the Manlius Formation. This Manlius highstand occurred well before the New Scotland highstand that is regarded as the transgressive maximum in the lower Helderberg sequence. Recognition of the Green Vedder highstand has also provided important biostratigraphic information with respect to the location of the Silurian/Devonian boundary (Matteson and Ebert 2011) in the Appalachian Standard Succession of New York State (Johnson and Murphy 1969).

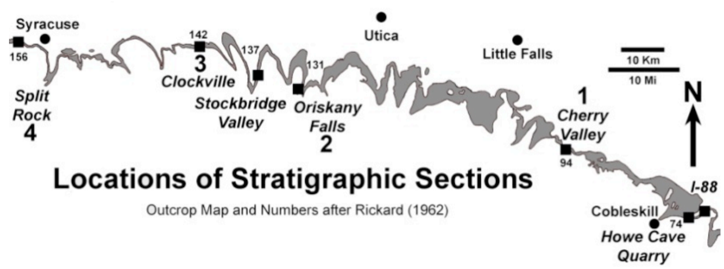

Fig. 1 Portion of the outcrop belt of the Helderberg Group in central New York. Small numbers are outcrop numbers of Rickard (1962). Large numbers 1-4 mark stops on this field trip.

STRATIGRAPHY OF THE GREEN VEDDER MEMBER OF THE MANLIUS FORMATION

The Green Vedder Member occurs in outcrops from the Syracuse area (e.g. Split Rock Quarry) eastward to Schoharie (Fig. 1), disappears in the region near Albany, and reappears as shallower facies in the Hudson Valley. This trip will focus on the member between Cherry Valley and Syracuse.

The Green Vedder Member of the Manlius Formation overlies the Thacher Member and is overlain by the Olney Member of the Manlius Formation in western outcrops and the Dayville Member of the Coeymans Formation to the east (Fig. 2).

The Clockville Unconformity at the Base of the Green Vedder Member

Throughout its extent, the Green Vedder Member is separated from the underlying Thacher Member of the Manlius Formation by the Clockville Unconformity (Ebert and Matteson 2003a, b). The Clockville Unconformity is always sharp and locally erosional. Where evidence of erosion is lacking, sedimentologic features consistent with sediment starvation are apparent. At Clockville, the unconformity is marked by several centimeters of relief (Fig 3). Topographic lows on this surface are filled with coarse, skeletal grainstones to packstones. These coarse lithologies include echinoderm debris, fragments of thick-shelled brachiopods, rare branching bryozoans, and the calcified green alga, *Garwoodia* (Laporte 1963, 1967). Erosional relief of several centimeters, lithoclasts of underlying Thacher lithology, isolated hardgrounds, encrusting stromatoporoids and pyrite impregnation highlight the Clockville Unconformity at Schoharie (I-88).

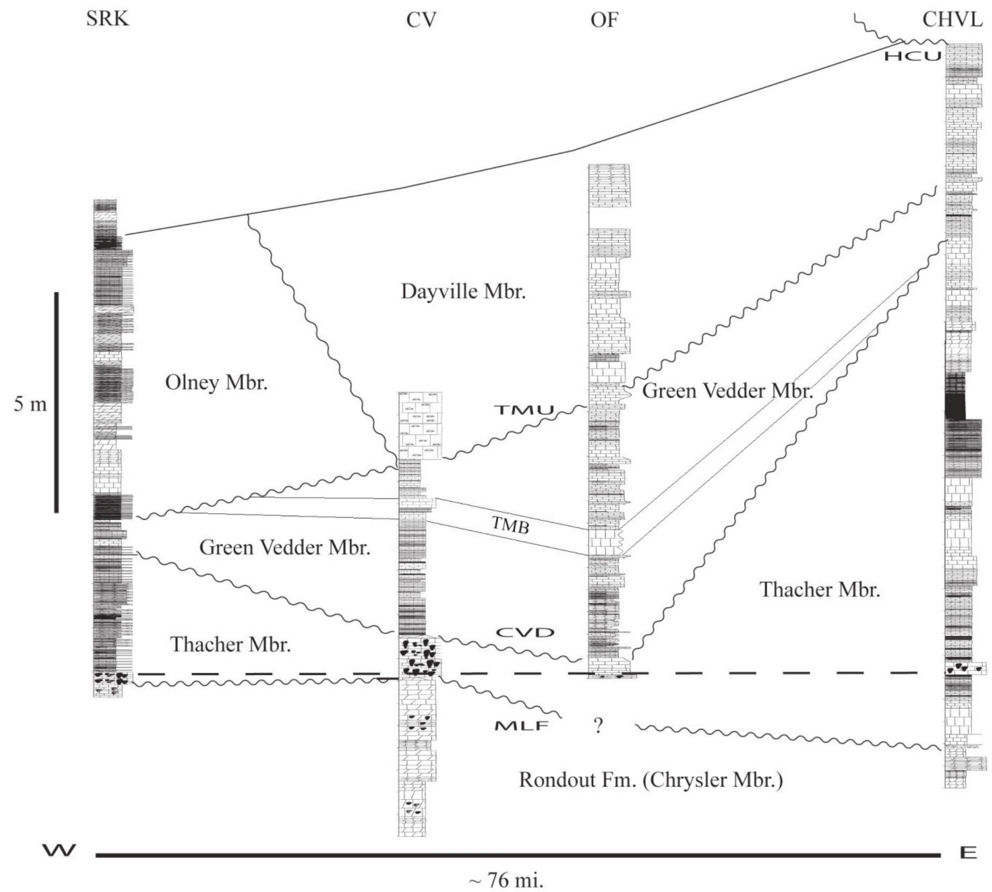

Fig. 2 Correlation and variation in thickness of the Green Vedder Member of the Manlius Formation across the study area. Datum for correlation is a distinctive zone of thrombolitic mounds in the Thacher Member of the Manlius Formation. Outcrop designations are SRK = Split Rock Quarry; CV = Clockville; OF = Oriskany Falls (Type section of the Green Vedder Member on Green Vedder Road); CHVL = Cherry Valley. MLF = Mine Lot Falls Unconformity; CVD = Clockville Discontinuity (unconformity); TMU = Terrace Mountain Unconformity; TMB = thick middle bed in Green Vedder Member.

At locations such as Cherry Valley, the Clockville Unconformity appears as an unremarkable bedding plane that separates drastically differing lithofacies. The Green Vedder Member has been recognized only recently in the Hudson Valley (Van Leuven Lake and Jefferson Heights outcrops). Here, the Green Vedder Member displays different lithologies (described below) compared to all other outcrops. However, the Clockville Unconformity remains recognizable. At Van Leuven Lake, the Green Vedder abruptly overlies mud-cracked laminites of the Thacher Member with extremely rare, bored, phosphatic nodules indicating sediment starvation.

We interpret the Clockville Unconformity as a flooding/transgressive surface because it bears evidence of sediment starvation and/or bypass and separates markedly deeper water facies (described below) of the Green Vedder Member from the underlying peritidal facies of the Thacher Member. Preliminary correlations (Ebert and Matteson 2005) suggest that the Clockville deepening event may be at least basin-wide in scale. The presence of a transgressive event recorded by the Clockville Unconformity and the overlying highstand systems tract of the Green Vedder Member indicate a more complex history of sea level change during Helderberg deposition than the simple transgression interpreted by Rickard (1962) and Laporte (1969).

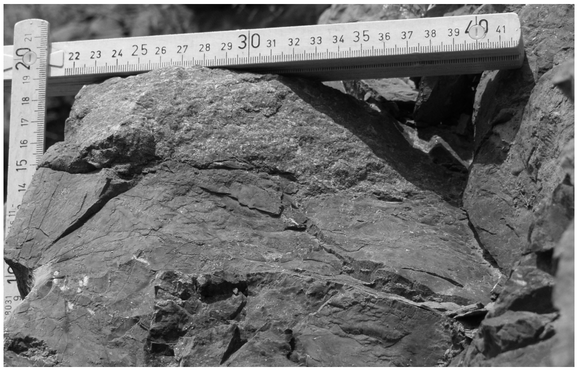

Fig. 3 Erosional relief and coarse skeletal fill of scoured pocket on the Clockville Unconformity. Clockville outcrop (**STOP 3**).

Correlation and Variations in Thickness of the Green Vedder Member

Wilson (2010) correlated the Green Vedder Member from the western limits of exposure near Syracuse through the Schoharie Valley (Fig. 2). The Green Vedder varies in thickness from of 0 to ~40 cm at Nedrow, 0 to ~30 cm at Split Rock and Clark Reservation State Park, to a maximum thickness of ~5.6 m at Oriskany Falls (type section of the Green Vedder Member). East of the type section, the Green Vedder thins to 70 cm at Cherry Valley, where only the upper beds of the member are present. It thickens to nearly 4 m in the Schoharie and Cobleskill valleys and is apparently absent east of Schoharie (Ebert and Matteson 2003a, b). However, recent fieldwork in the Hudson Valley has led to recognition of a more proximal, hardground-dominated facies of the Green Vedder Member (1.9 meters) within strata previously assigned to the Thacher Member (Matteson and Ebert 2011). Stratigraphic relationships among the Green Vedder Member, the subjacent Thacher Member and overlying strata previously referred to the Thacher Member in the Hudson Valley and Capital District are currently under investigation.

These variations in thickness and in the expression of the muddier interbeds led Rickard (1962) to assign strata above the quarry floor as Olney at Split Rock. Earlier and nearly contemporaneous interpretations conflicted with this assignment (see Sanders 1956; Chute and Brower 1964). Sanders (1956) noted a thin horizon similar with the upper Thacher (here referred to as Green Vedder) of alternating limestone and shales at Split Rock, which he interpreted as interfingering with the Olney Member. Chute and Brower (1964) reported 9 feet (~2.75 m) of Thacher Limestone above the Rondout Fm. at Split Rock Quarry and noted similar lithologic characteristics within the lowest 5 feet (~1.5 m) of the Thacher to the east. Indeed, the lowermost “ribbon limestone” from Split Rock to Jamesville shares similarities with counterparts of the Thacher from Cherry Valley eastward. The overlying Green Vedder Member is exposed at Clark Reservation State Park, Nedrow, and Split Rock, although stratigraphically thinned (~0 to 40 cm). Mudstones and wackestones are slightly more dolomitic than in eastward sections; shales display the characteristic carbonized biota of the Green Vedder.

Terrace Mountain Unconformity above the Green Vedder Member

In central New York, the Green Vedder Member is overlain unconformably by the Olney Member of the Manlius Formation from Perryville westward and by the Dayville Member (currently assigned to the Coeymans Formation, but see Ebert and Matteson 2003a, b) to the east of Perryville (Fig. 2). This surface is the Terrace Mountain Unconformity of Ebert and Matteson (2003a, b).

The Green Vedder Member holds biostratigraphic data pertinent to the location of the Silurian – Devonian boundary in the Appalachian Standard Succession (New York stratigraphy; Johnson and Murphy 1969). These data are discussed below, following the section on the sedimentology of the Green Vedder Member.

SEDIMENTOLOGY OF THE GREEN VEDDER MEMBER

The Green Vedder Member is a distinctive, heterolithic unit that bears a diagnostic fauna and unique taphonomy (Matteson, Natel and Ebert 1996; Ebert and Matteson 2003a, b). The Green Vedder Member encompasses skeletal and peloidal packstones, wackestones and mudstones (1-18 cm beds; average 5-7 cm) with interbeds of dark, carbonaceous shale (0-6cm).

(Fig. 4). An unusually thick set of burrow-amalgamated beds (up to 80 cm thick) occurs near the middle of thicker exposures of the Green Vedder. We refer to this marker informally as the thick middle bed (TMB; Fig. 4).Carbonate beds of the Green Vedder display sharp bases and planar to wavy laminated undulatory tops. Topmost laminae are dolomitic and slightly pyritized (Matteson et al. 1996). Many beds are discontinuous laterally because they comprise large-scale hummocks and swales. Individual hummocks are typically between 20 and 50 cm in wavelength with amplitudes of up to 12 cm (e.g., Stockbridge Quarry exposure). Hummocks alternate with much broader swales (~1 to 2 meters). To the East, at Schoharie, hummocks display much shorter wavelengths and beds appear nodular. These short wavelength hummocks appear to be a variant of large ripples or dune forms.

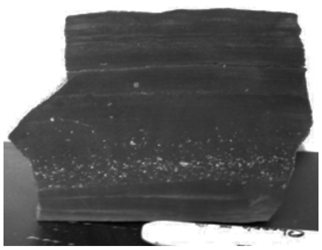

Fig. 5 Graded tempestite from the Green Vedder Member. Wall Street outcrop of Wilson (2010). Marker for scale.

Fig. 6 Gutter cast on the base of a thick tempestite bed in the lower Green Vedder Member at Clockville (STOP 3). Hammer for scale.

Though laterally discontinuous beds are common, other beds in the Green Vedder are more or less planar and persist across individual outcrops. Wavy to cross-bedded strata within the upper wave-rippled portions of the tempestites occur with winnowed skeletal debris.Both continuous and discontinuous beds in the Green Vedder exhibit the distinctive features of tempestite deposition (e.g., Aigner 1985) and many beds are clearly amalgamated from several storm events. Beds display sharp erosive bases that are overlain by graded wackestones to packstones with randomly oriented skeletal debris (Fig. 5). Gutter casts with up to 5 cm of erosional relief (Fig. 6) record initial scouring of consolidated sediment. Basal scouring is generally more pronounced at eastern outcrops of the Green Vedder (e.g., I-88 near Schoharie). Though remnants of grading are visible, extensive bioturbation obscures other internal structures. Infiltration fabrics are common in the coarser packstones. These features and the shallow-water aspect of the fauna suggest these beds were deposited as tempestites on an open shelf, below fair-weather wave base but above storm wave base.

Thin, skeletal-rich firmground/hardground grainstone horizons (sub mm to ~2 cm) and/or shell pavements are common in Green Vedder strata. Skeletal material within these condensed horizons is slightly pyritic. These grainstones commonly bifurcate into thin, discrete beds. Stacked hardgrounds comprise most of the member in the Hudson Valley.

Encrusting bryozoans adorn hardgrounds and shell pavements. Small *in situ* clusters of small rugose corals are less common. Thin, horizontal, simple, and singly branching burrows commonly mark the surfaces of these once semi-consolidated substrates. Hardgrounds also display linear to curvilinear fractures that taper downward from the lithified upper surfaces (Fig. 7). Though spar-filled, these fractures are clearly synsedimentary because thin veneers of sediment have been observed overlying the spar. We attribute these fractures to subaqueous slumping or liquefaction of sediments on a paleoslope beneath cohesive (hardground) surfaces.Black to gray, organic-rich shales are interbedded with the coarse, hummocky beds and hardgrounds. Some shales grade laterally into mudstones that are capped by thin laminae of sucrosic dolomite. Shales in the Green Vedder represent long intervals of sediment starvation during which many organisms were excluded, but a soft-tissue biota including bacterial mats and soft-bodied organisms were present (see below).

Medusaegraptus, a non-calcified, aspondyl, dasycladacean alga, preserved in these shales may have been opportunistic colonizers of an otherwise inhospitable seafloor or they were specialized for living in low oxygen environments (Brett and Baird 1986).

Fig. 7 Synsedimentary fracture in a mm-thick hardground from the Green Vedder Member at the Howe Cave Quarry, near the Schoharie. Such fractures cross-cut clasts and skeletal grains, taper downward and are filled with spar. Spar fill in fractures is draped by micritic or peloidal sediment in the upper portions of the fractures.

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The lithologic, sedimentologic, and paleontologic characteristics of the Green Vedder Member record episodic events of storm deposition that alternated with longer periods of relative sediment starvation. These features document the existence of a relatively deep, open shelf environment during early Helderberg time. This interpretation stands in stark contrast to the intertidal to supratidal environments that most workers (e.g., Laporte 1969; Goodwin and Anderson 1985; Kradyna 1987; 1991; Demicco and Smith 2009) associate with the Manlius Formation (See also Discussion of Demicco and Smith by Ebert, Matteson and Wilson 2010). The relatively deep, subtidal environment (below average wave base) recorded by the Green Vedder Member was brought about by transgression and subsequent highstand conditions following the non-depositional hiatus (flooding surface) marked by the Clockville Unconformity.

PALEONTOLOGY OF THE GREEN VEDDER MEMBER

Brachiopods dominate the macrofauna in the carbonate beds. *Mesodouvillina varistriata* and *Howellella vanuxemi* are most common. Fewer bryozoans and high-spired gastropods occur. Shell pavements comprising valves of the ostracod *Hermannina alta* are common. Echinoderm debris occurs throughout the member, much of which is attributable to the pelagic scyphocrinitid, *Camarocrinus* (see Biostratigraphy below). Rare, articulated specimens of *Lasiocrinus scoparius* and fragments of trilobites occur throughout the unit. Tentaculitids are rare in comparison to the underlying Thacher Member. Large orthocone cephalopods (~ 3 cm wide, over 20 cm long) occur in the upper beds at Schoharie. Bryozoan-encrusted spirorbid worm tubes, first recognized within the Thacher by Laporte (1967), are present within basal mudstone-wackestone beds at Clockville (Fig. 8). Skeletal fragments in the Green Vedder are typically angular and slightly abraded. Facies with *Howellella*, *Mesodouvillina*, and *Hermannina* are assignable to the BA 2 paleocommunity (Brett and Baird 1999), although hummocky cross-strata and the ichnofabrics described below suggest a deeper setting.

Carbonaceous shale interbeds hold a very distinctive carbonized biota (Fig. 9) that includes scolecodonts, poorly preserved annelid soft tissues, *Conularia* sp. and *Medusaegraptus* (Matteson, Natel and Ebert 1996).

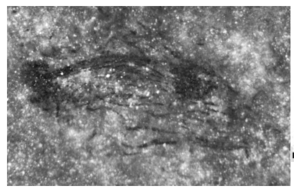


Fig. 9a *Medusaegraptus* from Green Vedder carbonized biota. Length = 1 cm.

Fig. 9b Scolecodont from Green Vedder carbonized biota. Scale bar = 1mm.

ICHNOLOGY OF THE GREEN VEDDER MEMBER

Green Vedder ichnofauna belongs predominantly to the *Cruziana* ichnofacies. Observed ichnogenera include *Planolites*, *Thalassinoides*, *Chondrites, and Helminthoides* (?). *Planolites* are displayed as simple, meandering horizontal burrows (<1 cm wide, up to 10 cm long) which are prevalent within post-event horizons. These likely represent opportunistic colonization of new substrates. *Thalassinoides* burrows are larger (~1-3 cm wide, up to 10+ cm long) and are lobate branching systems that occur at the base of the TMB and in some overlying beds of the upper Green Vedder. Burrows are slightly enlarged at bifurcations between branches and filled with fine, and dolomitized skeletal debris. The presence of *Thalassinoides* demonstrates the existence of cohesive burrowing paleosubstrates (Ekdale et al. 1984) which are commonly associated with knobby, irregular hyporeliefs ~ 2-4 cm in diameter (possible resting traces?). Uniformly small, thin, bifurcating networks of *Chondrites* occur most commonly in shales and mudstones/ wackestones in western (presumably deeper) sections of the Green Vedder (e.g., Clockville, Oriskany Falls) and are less common eastward. Exposures of the base of the TMB display remarkable networks of *Thalassinoides* and *Planolites* (Fig. 10). These traces occur with disarticulated ossicles of the pelagic crinoid *Scyphocrinites*. Small vertical borings (*Trypanites*; ~1 mm in diameter) in brachiopod shells occur along the base of deeply eroded horizons. *Helminthoides* (?) are present within few beds within the upper Green Vedder strata at Cherry Valley and Wall St. (Cedarville). *Helminthoides* appear as paired structures with pyritized cores that are surrounded by diagenetically leached dolomitic halos. Singular vertical burrows occur (up to 5 cm long and 1-4 mm wide) within the Green Vedder, although they are less common than horizontal and branching burrow systems.

Fig. 10 Networks of *Thalassinoides* and *Planolites* on the base of the "thick middle bed" from the outcrop at Oriskany Falls (**STOP 2**).

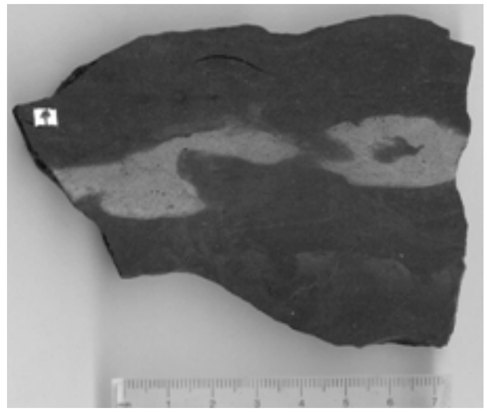


Fig. 11 "Tubular tempestites" – coarse skeletal fill of *Thalassinoides* burrows. Sample from Oriskany Falls. Scale is in centimeters.

Polished slabs from the uppermost strata of the Green Vedder Member at Oriskany Falls and Munnsville show pervasively dolomitized and chert-replaced thalassinoid burrow systems. These networks are filled passively with coarse grained and reworked dolomitic skeletal packstone (Fig. 11; i.e., "tubular tempestites" of Wanless et al., 1988). These uppermost Green Vedder beds are not present east of Oriskany Falls.

Thalassinoid burrows likely represent feeding or dwelling burrows of arthropods, similar to structures built by decapod crustaceans such as thalassinid shrimp in modern intertidal and subtidal environments (Myrow 1995). Aigner (1985) interpreted casts of *Thalassinoides* as evidence of deep exhumation (> 10cm) by storm scour. Across the study area, casts of *Thalassinoides* occur consistently on the base of the TMB (Fig. 10) and among upper beds of the Green Vedder. The producers of *Chondrites* may represent the last infaunal element in an increasingly dysaerobic environment that was inhospitable to other benthic organisms (Savrda and Bottjer 1986). The abundance of burrow systems within the Green Vedder contradicts Laporte's (1969) account of the rarity or absence of burrows within the Manlius, which he used to imply the presence of environmental stresses imposed by very shallow water and intermittent exposure. Rhoades (1967) indicated that in highly stressed environments, vertical burrows predominate over horizontal burrows and feeding traces, which is clearly not the case within the Green Vedder.Distinctive horizons of carbonaceous shale between the wavy bedded strata of the Green Vedder occur as partings between beds at many horizons. These shales vary in character across the outcrop belt, appearing as planar to wavy laminae that are black to dark grey in color west of Cherry Valley. Eastward, crinkly and less fissile shales to mudstones are found as partings at Schoharie, exhibiting a lighter grey, to a very slight reddish hue. A previously recognized distinctive biota from within the dark shales includes *Medusaegraptus,* poorly preserved annelid tissues, and scolecodonts (Matteson, Natel and Ebert 1996) (Fig. 9). Specimens of *Medusaegraptus* are common as individual ramifications (see Valet 1968) typically preserved as carbonized films and display a Y-shaped, singly branching morphology. Individual specimens of *Medusaegraptus* are typically ~2 mm wide and up to ~10 cm long. The best faunal preservation occurs within the upper and basal surfaces of the shales and among upper bedding surfaces within mudstone horizons, associated with few ostracods and small gastropods. Abundant ramifications of *Medusaegraptus* are found at Clockville as the sole faunal element in two distinctive beds of very dark, organic rich shales. At Schoharie, *Medusaegraptus* specimens are associated with vascular plant material, echinoderm debris, and rare tentaculites. Shales at Oriskany Falls have also yielded a previously undocumented microbial mat (Fig. 12). At Schoharie and Oriskany Falls (**STOP 2**), shales that are 0.5 to 1.0 m above the TMB have yielded loboliths of the pelagic crinoid *Camarocrinus* (Matteson and Ebert 2011; see below).

Tempestitic sequences in the Green Vedder are more condensed near Schoharie than in correlatives to the west. Beds at Schoharie contain lenses 2-5 cm thick consisting of winnowed strophomenid steinkerns in hydrodynamically stable positions. Farther east, in the Hudson Valley, the Green Vedder is almost exclusively comprised of hardgrounds. Crinoidal debris (individual ossicles and fragmental stems) are larger east of Cedarville as the spar/mud ratio increases eastward in the upper beds of the Green Vedder.

BIOSTRATIGRAPHY AND CARBON ISOTOPE CHEMOSTRATIGRAPHY OF THE GREEN VEDDER MEMBER AND THE SILURIAN – DEVONIAN BOUNDARY

Characteristics of the Silurian – Devonian Boundary

By international agreement, the base of the Devonian is defined by the first appearance of the graptolite *Monograptus uniformis uniformis* in the International Boundary Stratotype (GSSP) at Klonk in the Czech Republic (Chlupáč, Jaeger and Zikmundová 1972). Although the index conodont *Icriodus woschmidti woschmidti* has been used as a proxy for the boundary, there are distressing problems with the conodont taxonomy and biostratigraphy in this interval (e.g., Kleffner et al. 2009; Carls, Slavic and Valenzuela-Rios 2007). Biostratigraphically significant graptolites are conspicuously absent from the Helderberg Group in New York and the central Appalachian Basin. As a result, definitively locating the S/D boundary in New York has been problematic. Some authors have placed the boundary as low as the Rondout Formation (Rickard 1962) and potentially as high as the Kalkberg Formation (Kleffner et al. 2009).

Fig. 12 Pyritized microbial mat from shaly interbed in the Green Vedder Member at Oriskany Falls (**STOP 2**).

From numerous studies across the globe, other proxies have been developed to aid in locating the S/D boundary. There is growing recognition that the boundary is associated with transgressive/highstand conditions, a positive $\delta^{13}C$ excursion, and an epibole of pelagic scyphocrinitids with their distinctive floats (loboliths). In the Barrandian Basin (Czech Republic), scyphocrinitids first appear in the *ultimus* zone (Přídolí, Late Silurian), reach peak abundance at or just above the S/D boundary and disappear within the lower portion of the basal Lochkovian *uniformis* zone (Chlupáč et al. 1972). The large, ornate, pelagic scyphocrinitid crinoids *Scyphocrinites, Carolicrinus, Marhoumacrinus,* and *Camarocrinus* have been reported from strata close to the Silurian/ Devonian boundary from all over the world. These crinoids play a pivotal role in locating the S/D boundary in New York.

Scyphocrinitid Crinoids and the Silurian – Devonian Boundary One of the most distinctive features of scyphocrinitid crinoids is their float, or lobolith (Fig. 13). Loboliths consist of two main types: the plate-type, consisting of a large hollow chambered bulb with a double outer wall formed from numerous polygonal plates; and a presumably more primitive cirrus-type consisting of a bulb bearing more numerous, irregular chambers comprised of walls formed from numerous tightly packed branching rootlets (Haude 1972).

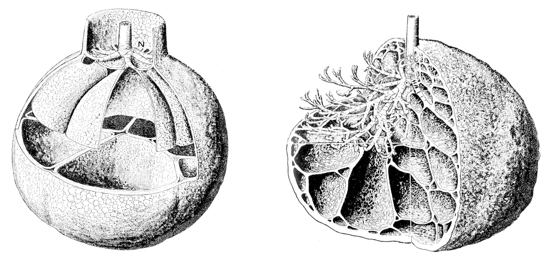


Fig. 13 Line drawings of scyphocrinitid floats (loboliths). Plate-type (left) and cirrus-type (right) are represented. Adapted from Haude (1972).

To date, not a single specimen exists of a fully articulated scyphocrinitid, with crown, stem, and lobolith. As a result, we can only infer which lobolith belonged to which crown. Curiously, plate type loboliths and cirrus type loboliths have not been found together on the same bedding plane (Haude 1972; 1989; Prokop and Petr 1994; 2001). In addition to loboliths, debris from these intriguing crinoids forms a considerable fraction the sediments surrounding the S/D boundary in the Czech Republic (Kriz et al. 1986; Chlupáč et al. 1972). Therefore, scyphocrinitid debris can also serve as a proxy for identification of the boundary interval in areas where biostratigraphic control is sparse.

Plate-type loboliths were first reported in James Hall's 1875 "Notice of Some Remarkable Crinoid Forms from the Lower Helderberg Group," in which Hall described material collected by John Gebhard Jr. at Schoharie, NY. The genus *Camarocrinus,* the first scyphocrinitid genus ever described, consists entirely of plate-type loboliths. Gebhard's original material, still archived at the New York State Museum, consists of several whole, well preserved loboliths that are filled nearly completely with dark, micritic matrix. Hall described these specimens and christened them as *Camarocrinus stellatus*. However, the question has remained – where in the Helderberg Group did Gebhard collect the first specimens of *C. stellatus* nearly two hundred years ago? Hall gives the occurrence as the "Tentaculite limestone" (an obsolete term for the Manlius Formation) but until recently the exact position and range of scyphocrinitid crinoids in the Helderberg of New York has been unknown.

At Schoharie, the Manlius Formation is represented by the Thacher Member, which is overlain by approximately four meters of the Green Vedder Member. The Terrace Mountain Unconformity (Ebert and Matteson 2003a, b) separates the Green Vedder from the Dayville Member of the Coeymans Formation. The Dayville, in turn, is truncated by the Howe Cave Unconformity (Ebert and Matteson 2003a, b) and is overlain by the Ravena Member of the Coeymans Formation. Ossicles bearing the characteristic stellate lumen (Fig. 14) as well as whole and partial plate-type loboliths (Fig. 15) have been recovered from the Green Vedder Member at Schoharie and Oriskany Falls. To date, loboliths and scyphocrinitid debris have been found only in the Green Vedder Member in central NY and the Keyser Member in the Central Appalachian Basin (Ebert and Matteson 2005; Matteson and Ebert 2010). Southeast of Schoharie, near Catskill, scyphocrinitid debris becomes increasingly abundant in the Green Vedder Member and includes partial to complete loboliths that range from 5 cm to more than 10 cm in diameter, stem ossicles, and discrete clusters of plates from the outer walls of plate-type loboliths. These occur with abundant current aligned orthocone cephalopods, bored stromatoporoids, and high-spired gastropods. This assemblage also includes the first examples cirrus-type loboliths from the Appalachian Basin and east of the Mississippi (Matteson and Ebert 2011).

Although plate-type loboliths can be quite large (in excess of 20 cm in diameter), the plates forming the interior and exterior walls are mere millimeters in thickness, and the entire double wall (including any separation between the inner and outer wall plate layers) is less than 5 mm (Haude 1992). As a result, specimens of loboliths can be difficult to find in the field, even when complete bulbs are present. Loboliths in cross section can be camouflaged, particularly when they have been completely filled by the surrounding matrix (including skeletal elements from other fauna) or the specimen consists of only a fraction of the outer wall of the float. Occurrences of whole and nearly complete loboliths (informally termed "cricket ball cemeteries," J. Hladil, written communication 2008), such as the spectacular deposits of *Camarocrinus ulrichi* from Oklahoma, bearing thousands of complete and three-dimensional loboliths (Springer 1917; Ray 1986), are likely more the exception than the typical occurrence. Plate-type loboliths in many parts of the Barrandian are represented by clusters of a few tens of plates or less, or even isolated individual plates (Hladil, J., written communication 2008). As a result, the true global geographic distribution of these crinoids may not be completely recognized.

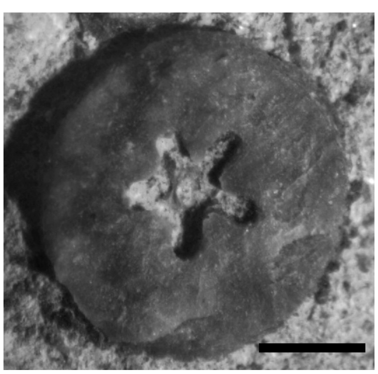


Fig. 14 Stellate lumen in scyphocrinitid columnal from outcrop at Van Leuven Lake (Hudson Valley). Scale = 1 mm.

Fig. 15 Cross section of a lobolith from the upper Green Vedder Member at Howe Cave Quarry near Schoharie. Scale is in cm.

Species of scyphocrinitid crinoids bear yet another distinctive feature – their long stems (which may have reached a meter or more in length) are constructed from numerous ossicles that bear a distinctly stellate central lumen, that is wider and more pentameral proximate to the calyx, and more narrow and cinqelobate near the junction with the lobolith (Springer 1917). *Scyphocrinites, Marhoumacrinus,* and *Carolicrinus* share this distinctive feature (Prokop and Petr 1987). Crinoid stems disarticulate very rapidly post mortem, leaving hundreds of readily available macroscopic elements that can be found in the field more easily than fragments and whole loboliths. Crinoid ossicles bearing stellate lumens conforming to scyphocrinitid crinoids are more abundant in eastern exposures of the Green Vedder Member and decrease in abundance westward from Schoharie to Clockville. Partial plate loboliths (Fig. 16) are present in outcrop at the Green Vedder type locality at Oriskany Falls (**STOP 2**).

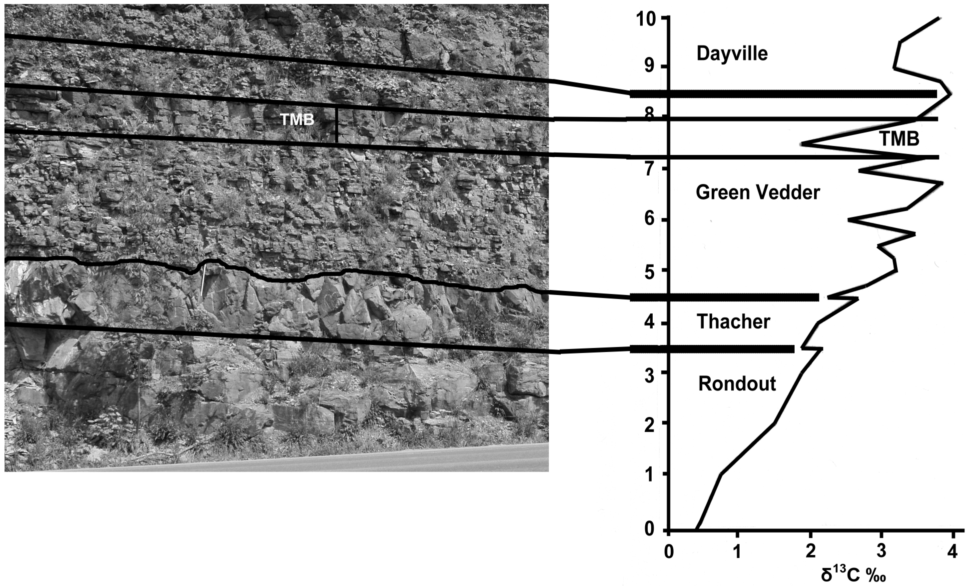
Fig. 16 Partial plate loboliths in the upper Green Vedder Member at the type section near Oriskany Falls (**STOP 2**).

Thus far, no traces of scyphocrinitid crinoids have been found in stratigraphically younger or older units in the Helderberg Group, including units that display more diverse faunal assemblages, such as the obrution deposits and coarse encrinites of the Dayville and Ravena members of the Coeymans Formation. The unique occurrence of debris of scyphocrinitid crinoids within the Green Vedder Member indicates close proximity to the S/D boundary and constitutes the best biostratigraphic control on this boundary yet determined in the Appalachian Standard Succession.

Carbon Isotope Chemostratigraphy and the Silurian – Devonian Boundary in the Green Vedder Member

Hladikova et al. (1997) reported a positive $\delta^{13}C$ excursion occurring across the Silurian/Devonian boundary at the Global Stratotype Section and Point (GSSP) at Klonk, Czech Republic, beginning in the Přídolí and ending in the early Lochkovian (*M. transgrediens*- early *M. uniformis* graptolite biozones). Subsequently, this positive carbon isotope excursion has been recognized globally (See Malkowski et al. 2009 and references therein). In the central Appalachian Basin, Saltzman (2002) reported the excursion from the uppermost Keyser Formation. The rediscovery of *Camarocrinus sp.* in the Big Mountain Shale/ upper Keyser at the Keyser type section (Ebert and Matteson 2005) supports the interpretation that this is the S/D boundary excursion.

In New York, Williams (2004) and Williams and Saltzman (2004) reported a positive $\delta^{13}C$ excursion at Cherry Valley (**STOP 1**) from strata reported as Manlius Formation. Kleffner et al. (2006, 2009) sampled a more complete section at Cherry Valley (**STOP 1**) and confirmed that the positive $\delta^{13}C$ excursion (peak at +3.65 ‰) at Cherry Valley and that it occurs within the Green Vedder Member. Based on the excursion at Cherry Valley and ambiguous conodont data (e.g., Carls et al. 2007), Kleffner et al. (2009) determined seven possible locations for the S/D boundary at Cherry Valley. However, the exact position of *Camarocrinus* in the succession was not known when Kleffner et al. (2009) was published.



Wilson (2010) included an examination of $\delta^{13}C$ chemostratigraphy from the more complete (~4 m of section) Green Vedder exposure at Clockville (**STOP 3**). Stable carbon isotopes vary through the lower Helderberg Group at Clockville by 3.53 ‰ (Wilson 2010; Fig 18). Values rise gradually through the Rondout Formation, from 0.88 ‰ to a maximum at 2.3 ‰ at the Mine Lot Falls Unconformity and then fall slightly at this horizon. Within the Thacher Member of the Manlius Formation, values increase slightly to a maximum of 2.79 ‰ at the Clockville Unconformity, followed by a slight decrease in the lower Green Vedder Member. Within the Green Vedder, $\delta^{13}C$ values fluctuate somewhat but show an overall positive trend. The most significant decrease in $\delta^{13}C$ (-1.74 ‰) occurs within the thick middle bed (TMB), a packstone-grainstone package associated with abundant pyritized pelmatozoan debris. A marked positive $\delta^{13}C$ excursion occurs within the upper Green Vedder and corresponds to the interval containing the most organic rich, black shales in the Clockville outcrop. Values of $\delta^{13}C$ between 3.58 ‰ and 4.12 ‰ occur within the upper 3.25 m of the member and peak in the uppermost Green Vedder bed. $\delta^{13}C$ values fall across the Green Vedder – Dayville contact (Terrace Mountain Unconformity), before climbing to a second peak (3.96 ‰) in the Dayville Member.

The $\delta^{13}C$ curve from Clockville (Fig. 17) compares well with the isotope curve for Cherry Valley from Kleffner et al. (2009). The carbon isotope data support the correlation of the dramatically thinned Green Vedder at Cherry Valley with the upper Green Vedder at Clockville (Wilson 2010).

The occurrence of loboliths and scyphocrinitid debris in the Green Vedder Member, along with a significant, positive $\delta^{13}C$ excursion, strongly suggest that the excursion is, in fact, the S/D boundary event. These data indicate that the S/D boundary occurs within the upper portions of the Green Vedder Member or in the erosional vacuity of the Terrace Mountain Unconformity that separates the Green Vedder from the overlying Dayville or Olney members. Chitinozoan studies (e.g., Bevington, Ebert and Dufka 2010) are currently in progress and may shed further light on the position of the boundary. Until such time, the combination of the $\delta^{13}C$ excursion and scyphocrinitids represents the most precise placement to date of the Silurian/Devonian boundary in the Appalachian Standard Succession.

SUMMARY

The Green Vedder Member records highstand deposition in an open-shelf setting in which the bottom waters or surface sediments were dysaerobic to anaerobic. As such, the member represents significantly deeper, subtidal conditions than the peritidal environments that are typically associated with the Manlius Formation. The initial pre-Green Vedder deepening is recorded by the Clockville Unconformity, which represents conditions of localized erosion and sediment starvation and/or bypass. We interpret the Clockville surface and immediately associated lithologies as a transgressive surface, perhaps closely linked with maximum flooding.

The Green Vedder open shelf was characterized by generally dysaerobic conditions recorded by interbeds of dark shale and carbonate mudstone to wackestone. This relatively quiet setting was interrupted episodically by storms that delivered skeletal and peloidal debris from shallower parts of the shelf to the zone below average wave base. Shell pavements that developed during storm winnowing and early-lithified hardgrounds hosted opportunistic, encrusting, and boring fauna that persisted as long as oxygenation was sufficient. Deposition of fine-grained, likely condensed beds resumed with the re-establishment of dysoxia.

Relatively thin accumulations of the Green Vedder Member near Syracuse represent onlap onto the western margin of the Appalachian Basin (Ebert 2008). East of Syracuse, the Green Vedder thickens owing to localized subsidence and then thins again at Cherry Valley, via onlap onto a local paleotopographic high that experienced little or no subsidence. East of Cherry Valley, subsidence facilitated a second area of thickened Green Vedder deposition that reached a maximum near Schoharie. The more proximal nature of tempestites in the Schoharie area indicates that the eastern sub-basin was shallower than the contemporaneous sub-basin in the west (Wilson and Ebert 2008).

The Green Vedder highstand established partial connection with the global ocean, which enabled loboliths of scyphocrinitids to drift into the otherwise restricted Appalachian Basin during the Silurian – Devonian transition. Loboliths and scyphocrinitid debris in the Green Vedder Member occur in the same strata as a strong, positive $\delta^{13}C$ excursion. Combined, these biostratigraphic and chemostratigraphic data strongly suggest that the Silurian – Devonian boundary in the Appalachian Standard Succession occurs in the Green Vedder Member of the Manlius Formation.

ACKNOWLEDGMENTS

We thank Jim Barrick and Mark Kleffner for the carbon isotope analyses of samples collected by RW for his thesis. We also thank Leigh Fall and Les Hasbargen for helpful suggestions on an earlier draft of this contribution.

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ROAD LOG FOR THE GREEN VEDDER MEMBER – A HIGHSTAND SYSTEMS TRACT IN THE "PERITIDAL" MANLIUS FORMATION

Road Log begins at the end of the ramp for Exit 30 (Herkimer/Mohawk) on I-90, the New York State Thruway, approximately 65 miles east of Syracuse (Fig. 18).

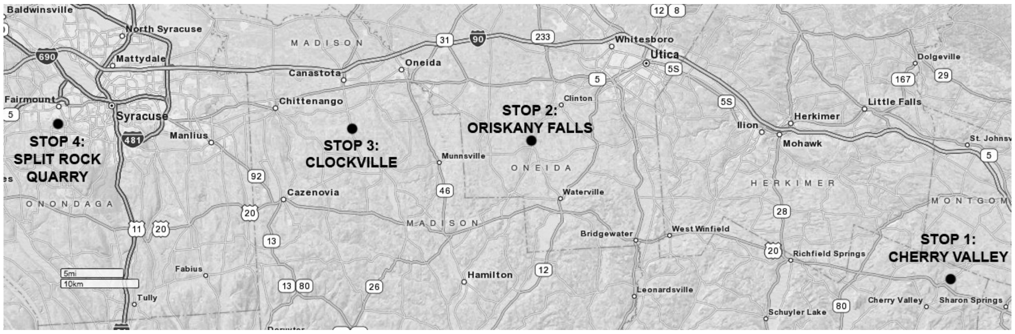

Fig. 18 Road map of field trip area with stop locations.

Note: This road log was compiled in Google Earth. Actual odometer-based mileage may differ somewhat.

CUMULATIVE MILEAGE	MILES FROM LAST POINT	ROUTE DESCRIPTION
0.0	0.0	Bottom of ramp for I-90 Exit 30, Turn left onto Rt. 28, Cross the Mohawk River
0.2	0.2	Turn right to continue on Rt. 28, concurrent with Rt. 5S
0.6	0.4	Turn left to continue on Rt. 28 south through village of Mohawk
1.3	0.9	Junction with Rt. 168, continue on Rt. 28
4.3	3.0	Denison Corners, intersection with County Rt. 46, continue on Rt. 28
11.8	7.5	Junction with U.S. Rt. 20, turn left (east) onto Rt. 20
18.6	6.8	Junction with Rt. 80, continue on Rt. 20
25.7	7.1	Exit Rt. 20 at exit for Rt. 166, Cherry Valley
25.8	0.1	End of ramp, turn left (north) onto Rt. 166
25.9	0.1	Cross under Rt. 20, road continues north as County Rt. 32, Sprout Brook Road
26.5	0.6	Continue northeast on Sprout Brook Road to **STOP 1.**

STOP 1: ROAD CUT ON SPROUT BROOK ROAD

The following stop description is modified from the description of Stop 1 of Ebert and Matteson (2003b) and Stop 1a by Ebert and Matteson in Ver Straeten et al. (2005). **STOP 1** is the stratigraphically lowest of a series of outcrops on Sprout Brook Road (Cty. Rt. 32), Rt. 166 and U.S. Rt. 20 that expose the upper Silurian (Ebert and Matteson 2003a, b) and much of the Lower Devonian section that is present in New York State (See Brett and Ver Straeten 1997). This outcrop corresponds approximately to Rickard's (1962) section 94, which was measured in nearby Judd's Falls.

The section begins at the north end of the outcrop with approximately one meter of the Rondout Formation (Přídolían). The Rondout is abruptly overlain (Mine Lot Falls Unconformity of Ebert 2008) by the Thacher Member of the Manlius Formation (11.2 m thick). A distinctive zone of thrombolitic mounds occurs two meters above the Mine Lot Falls Unconformity. Regional tracing of the thrombolitic zone shows it descending westward relative to the Mine Lot Falls Unconformity (westward onlap of the Thacher Member) and rising stratigraphically eastward (Ebert 2008). This thrombolitic marker zone is also present at **STOPS 3** and **4** on this trip and is important in recognizing the Thacher Member where it is thinned by truncation and onlap.

Less than one meter below the top of the Manlius Formation, the Clockville Unconformity (Ebert and Matteson 2003a, b) marks an abrupt change in the style of bedding to thinner (decimeter scale) limestone beds with interbedded dark gray to black shale (Fig. 19). These beds (total thickness = 0.82 m) comprise a thinned portion of the Green Vedder Member. The dark shale interbeds of the Green Vedder Member contain a distinctive carbonized biota comprising scolecodonts, annelid bodies, and non-calcified green algae (*Medusaegraptus*) (Matteson, Natel and Ebert 1996). The faunal content and general lithology of this thin remnant of the Green Vedder Member are strikingly similar to upper portions of the member from thicker sections to the west (e.g., Oriskany Falls) and to the east (I-88, Schoharie Valley). This similarity suggests onlap of the Green Vedder Member onto a paleotopographic high in the Cherry Valley area and that the Clockville Unconformity is more pronounced here than at sections that record greater subsidence to the west and east.

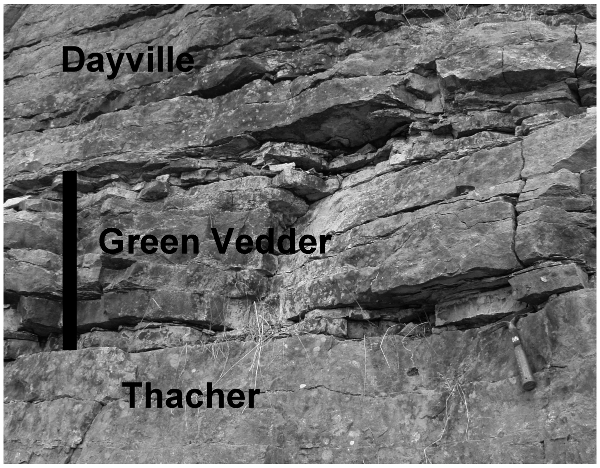

Fig. 19 Clockville Unconformity separates the Thacher Member of the Manlius Formation from the Green Vedder Member. The Green Vedder (only 0.82 m thick here) is cut by the Terrace Mountain Unconformity at the base of the Dayville Member. Hammer for scale.

The Green Vedder beds are cut by a sharp surface which is the Terrace Mountain Unconformity (13.27 m from base of section) (Ebert, Matteson and Natel 2001; Ebert and Matteson 2003a, b). The Dayville Member of the Coeymans Formation rests on this unconformity. The Dayville is comprised of coarse echinoderm grainstones and packstones, interbedded with mudstones and skeletal wackestones. Total thickness of the Dayville here is 4.03 meters (corrected from Ebert and Matteson 2003). The overlying Ravena Member of the Coeymans Formation is separated from the Dayville Member by the Howe Cave Unconformity (Ebert, Matteson and Natel 2001; Ebert and Matteson 2003a, b).

Combined, the Green Vedder beds and the Dayville Member appear to constitute a gradational transition (Grabau's [1906] "transition beds") between the Manlius and Coeymans formations, a key factor in Rickard's (1962) stratigraphic reconstruction. However, such a gradation does not exist owing to the presence of the Clockville Unconformity and the Howe Cave Unconformity, which bound the interval and the Terrace Mountain Unconformity, which occurs within it. The Green Vedder Member and Dayville Member are traceable across the northern portion of the outcrop belt of the Helderberg Group. These members are distinctive and do not comprise an ambiguous zone of transition.

CUMULATIVE MILEAGE	MILES FROM LAST POINT	ROUTE DESCRIPTION
26.7	0.2	Continue northeast on Sprout Brook Road to area with wide shoulder on the left. Turn around and return to Rt. 20
27.4	0.7	Junction with Rt. 20, get on Rt. 20 heading west
34.7	7.3	Junction with Rt. 80, continue on Rt. 20
41.6	6.9	Junction with Rt. 28, continue on Rt. 20
49.0	7.4	Junction with Rt. 51, continue on Rt. 20
51.9	2.9	Main intersection in West Winfield, continue on Rt. 20
55.0	3.1	Junction with Rt. 8 in Bridgewater, continue on Rt. 20
62.6	7.6	Junction with Rt. 12 in Sangerfield, continue on Rt. 20
68.0	5.4	Junction with Rt. 26/12B, turn right, north onto Rt. 26/12B
70.5	2.5	Intersection with Broad Street in Oriskany Falls
70.53	0.03	Turn Right onto Main Street
71.3	0.8	Pass quarry and turn left onto Green Vedder Road to **STOP 2**.

STOP 2: GREEN VEDDER ROAD – TYPE SECTION OF THE GREEN VEDDER MEMBER, MANLIUS FORMATION

STOP 2 (associated with R-131 of Rickard 1962), which is located on Green Vedder Road on the edge of the large active quarry near Oriskany Falls, comprises the type section of the Green Vedder Member of the Manlius Formation. The member reaches its greatest thickness (5.6 m) here. The lower contact (Clockville Unconformity) with the underlying Thacher Member is exposed near the base of the outcrop at the southern end of the exposure. The Terrace Mountain Unconformity marks the top of the Green Vedder Member and separates it from the overlying skeletal-rich, stromatoporoid-bearing packstones to grainstones of the Dayville Member (previously mapped here as Olney Member by Rickard [1962]).

The Green Vedder Member comprises decimeter-scale beds of skeletal/peloidal packstones and wackestones interbedded with gray, calcareous shales. The carbonate beds in the Green Vedder display broad hummocks (>1m; Fig. 20) that are much greater in wavelength than equivalent strata to the east (e.g., I-88 near Schoharie).

Fig. 20 Hummocky bedforms in the Green Vedder Member at Munnsville

We interpret the carbonate beds of the Green Vedder as tempestites that record episodic storms in the Helderberg Sea. Greater thicknesses of shaly interbeds and longer hummock wavelengths suggest that the Green Vedder at its type section represents deeper conditions than those that existed to the east.Some carbonate beds display *Thalassinoides* and *Planolites* burrows. These are particularly well developed on the base of the thick middle bed (TMB). Abundant *Garwoodia* (codiacean alga) occur approximately 20 centimeters below the top of the member. Skeletal debris of pelagic scyphocrinitid crinoids are abundant in the carbonate beds and a partial lobolith (*Camarocrinus*) is exposed on the bedding plane (Fig. 16) across the road from the main outcrop.

Shaly interbeds in the Green Vedder type section exhibit the carbonized biota that is so distinctive of this member (Matteson, Natel and Ebert 1996). *Chondrites* burrows are common in the shaly beds and a pyritized microbial mat (Fig. 12) was collected here from one of the shaly interbeds.

CUMULATIVE MILEAGE	MILES FROM LAST POINT	ROUTE DESCRIPTION
71.3	0.0	Return to Main Street/Rt. 26/12B heading south back through Oriskany Falls to Rt. 20
74.7	3.4	Junction with Rt. 20, turn right (west) onto Rt. 20
76.3	1.6	Village of Madison
77.6	1.3	Split between Rt. 20 and 12B (south). Remain on Rt. 20
78.5	0.9	Village of Bouckville
83.4	4.9	Village of Morrisville
83.7	0.3	Turn Right (north) on Cedar Street, which becomes County Rt. 101, which becomes Old County Rt. S
88.4	4.7	Bear right onto Pleasant Valley Road
89.2	0.8	Village of Peterboro
89.35	0.15	Bear left onto Oxbow Road (County Rt. 25) just north of Peterboro
94.9	5.5	Large outcrop on Oxbow Road **STOP 3.** Park in parking area on west side of road

STOP 3: CLOCKVILLE ROAD CUT Approximately four meters of the Chrysler Member of the Rondout Formation are exposed in this large road cut (Section 142 of Rickard 1962). The contact with the overlying Thacher Member of the Manlius Formation is the Mine Lot Falls Unconformity, a sharp surface upon which thrombolitic mounds have nucleated. We regard this massive, thrombolitic bed (0.9-2.0 m thick) as the total thickness of the Thacher Member at this location.

Fig. 21 Lowest Green Vedder beds lapping onto irregular topography of the Clockville Unconformity at the top of the Thacher Mbr. at Clockville (**STOP 3**).

The top of the Thacher Member displays up to 20 cm of erosional relief along the Clockville Unconformity (Fig. 3). This surface is overlain by a coarse, cross-stratified skeletal grainstone that is semi-continuous across the outcrop. Lithoclasts of Thacher lithologies are incorporated in this grainstone. The lithology and texture of this bed is atypical for both the Thacher and Green Vedder members of the Manlius Formation. Where the coarse bed is discontinuous, it is preserved locally in scoured pockets along Clockville Unconformity.

Approximately four meters of the Green Vedder Member rest on the Clockville Unconformity. The cm- to dm-scale beds of the Green Vedder contrast sharply with the massive Thacher below (Fig. 17). In several places along the outcrop, the lowest beds of the Green Vedder lap onto local high spots along the Clockville Unconformity (Fig. 21). Shaly interbeds in the Green Vedder contain the member's signature carbonized biota. The TMB is well-displayed in the Clockville road cut. Carbonate beds in the upper portions of the Green Vedder are noticeably thicker than in the lower half of the member. These beds also tend to be richer in skeletal debris than beds in the lower half. Thinner beds reappear in the uppermost 1.5 m of the member.

The Terrace Mountain Unconformity separates the Green Vedder Member from the superjacent Olney Member of the Manlius Formation. Here, the lowest Olney is a massive stromatoporoid biostrome.

CUMULATIVE MILEAGE	MILES FROM LAST POINT	ROUTE DESCRIPTION
94.9	0.0	Return to Oxbow Road and continue north
97.8	2.9	Intersection with Rt. 5, continue north on S. Peterboro Street through village of Canastota to I-90 on-ramp.
98.7	0.9	Merge onto I-90 westbound
113.6	14.9	Exit I-90 at Exit 34A for I-481 South, toward Syracuse
114.5	0.9	Merge onto I-481
117.0	2.5	Merge onto I-690 West
123.2	6.2	Exit I-690 at Exit 10, North Geddes Street
123.4	0.2	Turn left (south) onto North Geddes Street
123.6	0.2	Turn right onto West Genesee Street (Rt. 5)
125.2	1.6	Turn left onto Fay Road
127.0	1.8	Turn right onto Onondaga Boulevard
127.5	0.5	Intersection with Rt. 173, continue straight through intersection on Onondaga Boulevard

128.7	1.2	Park at gated extension of Onondaga Boulevard and walk to **STOP 4**, the inactive Split Rock Quarry.

STOP 4: SPLIT ROCK QUARRY

Rickard (R-156; 1962, p. 54 and 149) described the section exposed at Split Rock Quarry as "33 feet of fine-grained, even bedded limestones" of the Olney Member, overlain by 3 feet of Elmwood waterlimes. Rickard did not recognize any Thacher Member at this location. Rickard's "Olney" rests upon a sharp contact (Smith 1929; Logie 1933) with the underlying Rondout Formation, which is exposed in the floor of the quarry (Fig. 22). This sharp contact is the Mine Lot Falls Unconformity of Ebert (2008). In the northeastern section of the quarry, an additional 1.3 m of the Rondout, comprising dolomitic mudstones with abundant domal stromatoporoids are exposed below the Mine Lot Falls Unconformity. In this area, approximately 30 cm of the thrombolitic marker bed within the Thacher Member rests on the unconformity (Fig. 23). Above the thrombolitic horizon, an additional ~2.9 m of the Thacher Member are exposed. Here, the Thacher displays mudstones and skeletal wackestones that alternate with very fine grained dolomitic grainstones. Planar to wavy or lenticular beds, with laterally discontinuous and rippled laminae are common in this section of the Thacher.

Fig. 22 Thrombolites of the Thacher Member colonized the irregular topography of the Mine Lot Falls Unconformity at the top of the Rondout Formation. Split Rock Quarry (**STOP 4**).

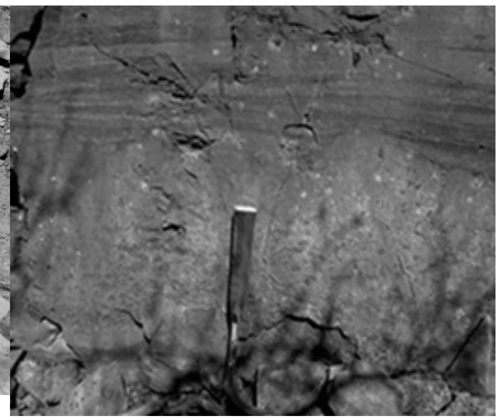

Fig. 23 Thicker development of the thrombolitic marker horizon in the Thacher Member. Split Rock Quarry (**STOP 4**).

The Clockville Unconformity separates the Thacher Member from approximately 30 cm of the Green Vedder Member, which is exposed in the western section of the quarry. The Green Vedder comprises thin to medium-bedded skeletal wackestones to packstones with dark shale interbeds containing *Medusaegraptus*, a representative of the distinctive biota that occurs in the Green Vedder throughout central New York. Fauna in the limestones includes *H. vanuxemi*, *M. varistriata*, ostracods, crinoid and trilobite debris, and rare pelecypods.

The Terrace Mountain Unconformity separates the Green Vedder from ~6.4 m of the Olney Member of the Manlius Formation. The Olney is a hard, grey, dolomitic skeletal limestone, which ranges from laminated to massively bedded. The sedimentology and ichnology of the Olney are described by Wilson (2010). Numerous sedimentologic and ichnologic features in the Olney indicate deposition under arid to semi-arid, evaporative supratidal (sabkha) to high intertidal conditions.

End of road log. After **STOP 4**, return to Syracuse University.

Trip A-4

QUARTZ-SILLIMANITE VEINS AND NODULES IN LYON MOUNTAIN GRANITE, AND THE OCCURRENCE OF PRISMATINE IN METAPELITIC GNEISS AND QUARTZITE, MOOSE RIVER, WESTERN ADIRONDACKS

by

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INTRODUCTION

The western Adirondacks contain a wide variety of rock types, and this trip will visit representatives of the major lithologies. Stops include units of Lyon Mountain granite recently dated by U-Pb geochronology including multi- and single-grain TIMS and SHRIMP II analyses as well as in-situ dating of monazites.

The Adirondack Mountains represent a southeastern extension of the Grenville Province via the Thousand Islands – Frontenac Arch across the St. Lawrence River (Fig. 1). The region is topographically divided into the Adirondack Highlands and Lowlands separated by the Carthage-Colton Mylonite Zone (CCMZ, Fig. 2). The former is underlain largely by orthogneiss metamorphosed to granulite facies and the latter by upper-amphibolite facies metasediments, notably marbles and calcsilicate. Both sectors have experienced multiple deformations resulting in refolded major isoclines. The intrusion history is summarized by a tripartite division, as summarized below and in Fig. 3. Broadly, the plutonic lithologies fall into the following five groups: 1350-1300 Ma tonalites rifted from Laurentia at ca. 1300 Ma.; 1250 Ma granodiorites; 1160-1150 Ma anorthosite-mangerite-charnockite-granites (AMCG suite); 1100-1090 Ma Hawkeye granite; and 1055-1040 Ma Lyon Mountain Granite (LMG). The four major orogenic events associated with these are: an Andean-type arc along southeast Laurentia, the Elzevirian (ca1250-1220), the Shawinigan (1210-1140 Ma), the Ottawan (1090-1030 Ma), and the Rigolet (1000-980 Ma) orogenies (Fig. 3). The Elzevirian involved a back-arc rift basin with protracted outboard arc magmatism and accretion and was followed by the Shawinigan Orogeny due to the collision between Laurentia and the Adirondack Highlands – Green Mountain Terrane. The Ottawan was a Himalayan-type collision of Laurentia with Amazonia (?). Most of the metamorphic and structural effects present in the Adirondacks are the result of the Ottawan Orogeny, but Shawinigan and Elzevirian features can be recognized where minimally overprinted by the Ottawan. Both the AMCG suite and the LMG are thought to be late- to post-tectonic manifestations of delamination of over-thickened orogens undergoing terminal extensional collapse.

Structurally, the central and western Adirondack Highlands are dominated by the same large, recumbent fold-nappe structures (F_2) as found in the southern Adirondacks (Fig. 2), and with fold axes oriented dominantly ~NE-SW and ~E-W parallel to stretching lineation. As in the southern Adirondacks, these Ottawan fold-nappes are thought to possess sheared-out lower limbs, but this has yet to be demonstrated at map scale. F1 folds of Shawinigan age occur as minor folds within the much larger F2 recumbent isoclines. At least two distinguishable upright fold events are superimposed on the nappes: F3 with shallow plunging ~E-W axes and F4 with shallow-plunging NNE axes. Both of these can be shown to be of late Ottawan (ca, 1050-1040 Ma) origin. All of three fold sets affect Hawkeye and older units, and thus must be of Ottawan age. This is also the case with the strongly penetrative rock fabric, including strong ribbon lineations that are present in these rocks and are largely associated with the large fold-nappes. Intense fabric and nappe structure are largely absent from the ca. 1050 Ma LMG; however, this unit is folded by F3 and F4, and this is interpreted to reflect its intrusion in late, post-nappe stages of the Ottawan orogenic phase of the Grenvillian orogeny. In the northern portion of the eastern Adirondacks the NNE, F4, folds become quite tight and have a strong lineation associated with them. This may be the result of rock sequences being squeezed between large, domical prongs of anorthosite during terminal Ottawan exhumation.

Most of the region likely experienced peak Ottawan temperatures of ~750-850^{o} C and pressures of ~5-8 Kbar (Kitchen and Valley, 1995, Florence et al, 1995, Spear and Markussen, 1997; Storm and Spear, 2005 and Darling et al., 2004). Evidence exists that similar P, T conditions were attained in the Shawinigan orogeny (Heumann et al., 2006). Based upon extensive oxygen isotopic work by John Valley and his students, it appears most likely that Ottawan metamorphism proceeded under fluid-absent conditions

(Valley and O'Neil, 1982, Valley *et al*., 1990). Note, however, that this does not exclude the presence of late, post-peak fluids associated with the emplacement of Lyon Mountain granite.

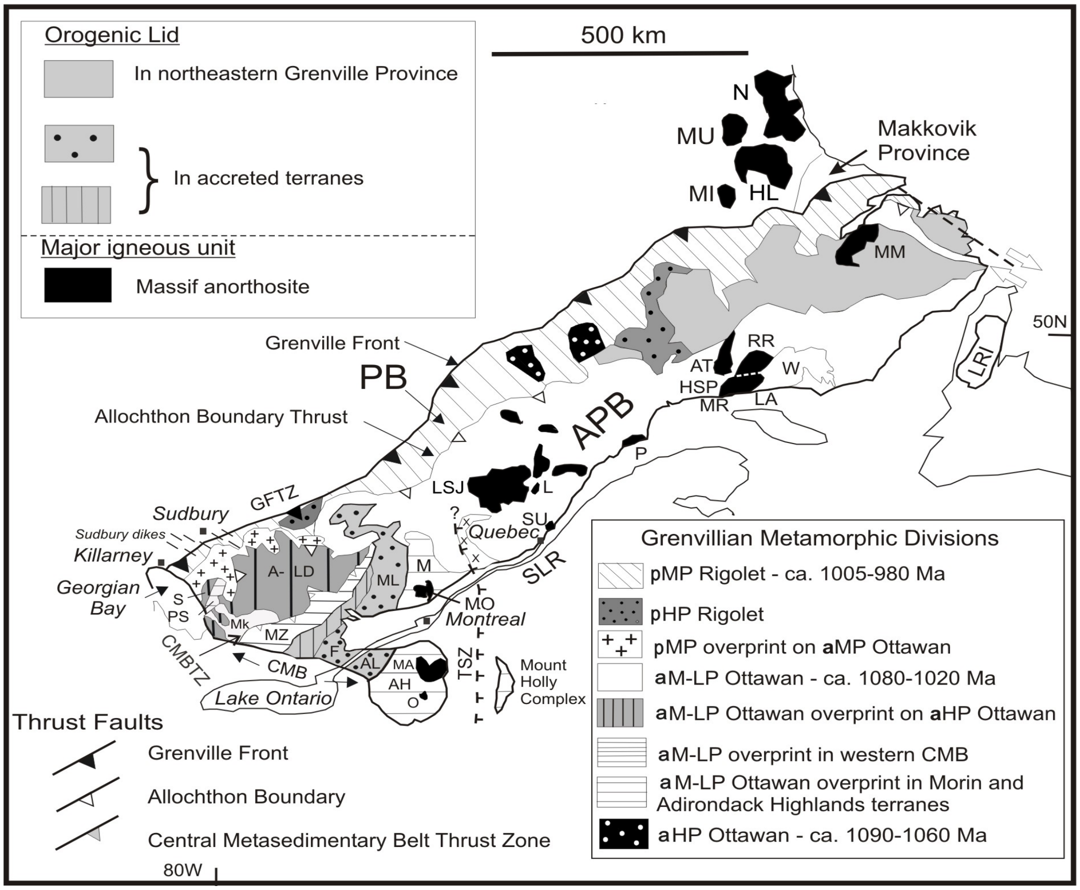

Fig. 1 Generalized map shows the Grenville Province whose three major tectonic divisions (Rivers, 1997) are indicated. The Orogenic Lid of Rivers (2008) is shown in light gray. The accreted ca. 1.3-1.4 Ga Montauban-La Bostonnais arc is shown by X- pattern. **Abbreviations**: A-LD- Algonquin- Lac Dumoine domain; AL-Adirondack Lowlands; AH- Adirondack Highlands; APB- Allochthonous Polycyclic Belt; CMB- Central Metasedimentary Belt; CMBTZ, Central Metasedimentary Belt thrust zone; F, Frontenac terrane; GFTZ- Grenville Front Tectonic Zone; LRI- Long Range inlier; M- Morin terrane; MK- Muskoka domain; ML- Mont Laurier domain; MZ- Mazinaw terrane; O- Oregon dome; PS- Parry Sound domain; RR- Romaine River, S- Shawanaga domain; SLR- St. Lawrence River, TSZ – Tawachiche Shear Zone with its southern projection, W- Wakeham terrane. **Abbreviations for metamorphic divisions**: p-MP- parautochthonous medium pressure belt; aM-LP- allochthonous medium to low-pressure belt; aHP- allochthonous high-pressure belt; pHP- parautochthonous high-pressure belt. **Major anorthosite massifs with ages**: AT- Atikonak (ca 1130 Ma); HL- Harp Lake (ca. 1450 Ma; HSP- Havre-St-Pierre, (ca. 1126 Ma, dashed white line is the Abbe-Huard lineament; L- Labrieville (1060 Ma); LA- Lac Allard HL) (ca. 1060 Ma); LSJ- Lac-St.-Jean (ca. 1155 Ma); MA- Marcy Massif (ca. 1155 Ma), MO- Morin (ca. 1153 Ma), MI- Mistastin (ca. 1420 Ma); MU- Michikamau (ca. 1460 Ma); MR- Magpie River (ca. 1060 Ma); N- Nain-(ca 1383- 1269 Ma); P- Pentecôte (ca 1350 Ma); S- St. Urbain (ca. 1060 Ma). *Modified after Rivers (2008 and McLelland et al. (2010a).*

The most recognizable fold patterns in the Adirondack Highlands are those of the ~E-W F3 and ~NNE F4 folds that span the entire region, e.g., Fig. 2, P (Piseco dome). These upright, relatively open folds exhibit E-W stretching lineations coaxial with the F3 folds. Because both these fold sets affect the ca. 1050 Lyon Mt. Granite and are crosscut by it, they are most likely synchronous with the terminal, extensional phase of the Ottawan orogeny. We interpret them as "a" and "b" domical folds that have been identified in large scale extensional orogens such as the Aegean core complex (Jolivet et al., 2004) and are due to the extension itself. Axes of the "a" folds form parallel to the principle extension due to constriction, whereas the "b" folds form approximately perpendicular to extension and slightly postdate the "a" folds. The axes of F2 folds are approximately parallel to extension and may have formed in this orientation and/or were rotated into it by extensional strain.

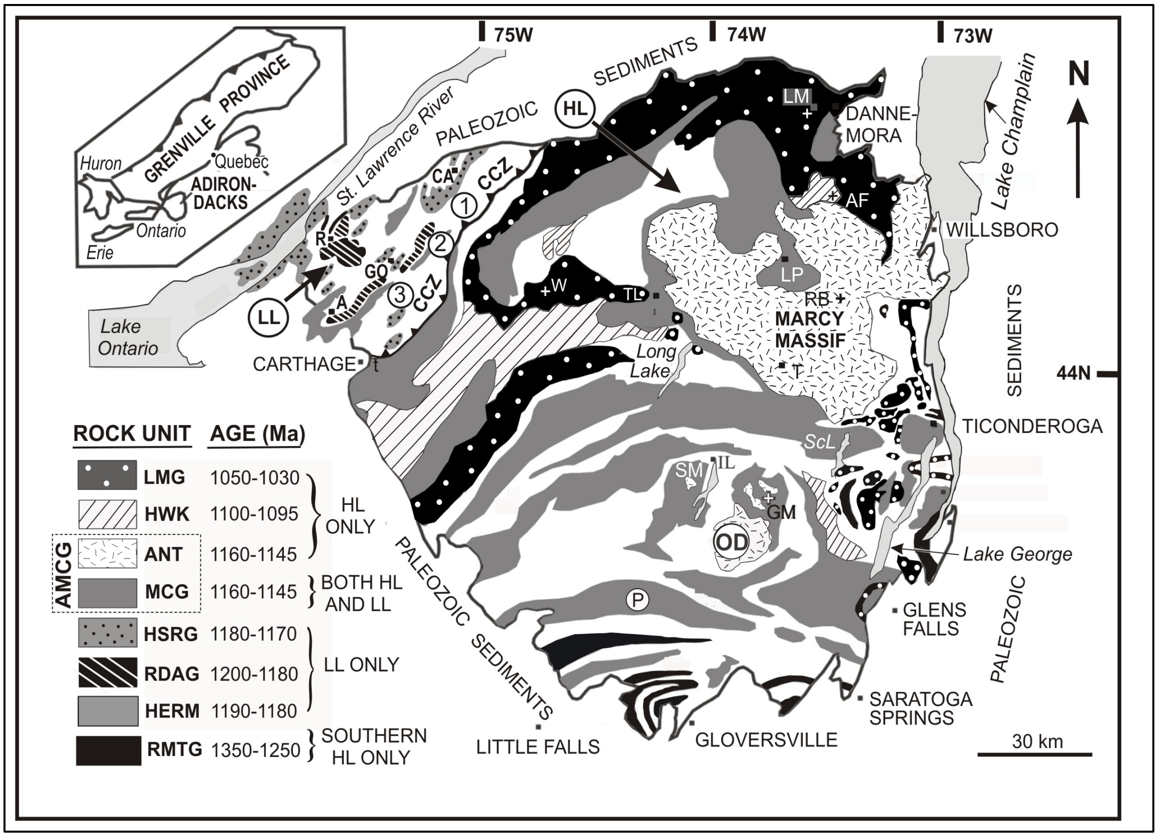

Fig. 2. Map showing generalized geology and geochronology of the Adirondacks. Units designated by patterns and initials consist of igneous rocks dated by U-Pb zircon geochronology with ages indicated in the legend. Units present only in the Highlands (HL) are: RMTG – Royal Mountain tonalite and granodiorite (southern and eastern HL only), HWK -Hawkeye granite, LMG – Lyon Mountain granite, and ANT – anorthosite. Units present in the Lowlands (LL) only are HSRG – Hyde School and Rockport granites (Hyde School also contains tonalite), RDAG – Rossie diorite and Antwerp granodiorite, RMTG – Royal Mountain tonalite and granodiorite (southern and eastern HL only), HWK -Hawkeye granite, LMG – Lyon Mountain granite, and ANT – anorthosite. Units present in the Lowlands (LL) only are HSRG – Hyde School and Rockport granites (Hyde School also contains tonalite), RDAG – Rossie diorite and Antwerp granodiorite. Granitoid members of the anorthosite-mangerite-charnockite-granite (AMCG suite) are present in both the Highlands and Lowlands. Unpatterned areas consist of metasediments, glacial cover, or undivided units. Other abbreviations: CCZ- Carthage-Colton Shear Zone; CL-Cranberry Lake, LL-Lowlands, OD- Oregon dome anorthosite massif, SM – Snowy Mountain anorthosite, IL-Indian Lake, L-Lyon Mountain. *Modified after McLelland et al., 2010a.*

Recent investigations in the Grenville Province demonstrate that the Ottawan orogen (Rivers, 2008) and its Adirondack outlier (McLelland et al., 2010a) underwent terminal extensional collapse that, in the latter case,
was accompanied by emplacement of LMG. The most complete local study of this sort is that of Selleck et al. (2005) that utilized zircon geochronology to document that down-to-the-west displacement along the northwest dipping Carthage-Colton Shear Zone (CCZ, Fig. 2) was coeval with intrusion of Lyon Mt. Granite (Fig. 2) into the fault complex at ca 1050-1045 Ma. It is thought that, at peak Ottawan contraction, the over-thickened lithosphere experienced delamination by foundering, convective thermal erosion, or both. Following delamination of the dense lithospheric keel, the orogen rebounded and hot asthenosphere moved to the crust-mantle interface where it underwent depressurization melting to yield aluminous gabbro

that differentiated into anorthositic crystal mushes that ascended, along with crustal granitic melts, to yield the ca. 1050 Ma AMCG suite that is exposed in Canada (Fig. 2) as the Lac Allard massif and the Labrieville to St. Urbain CRUML belt (Owens et al., 1994, Morrisett et al., 2010, McLelland, 2010b). The asthenospheric diapir likely underwent degassing, and caused fertilization and melting of deep crust to yield A-type LMG. As the granite ascended into the crust, it both lubricated and enhanced low-angle normal faults formed in response to increased topographic elevation, fluids, and anatexis; all leading to orogen collapse. In such situations, it is not uncommon to find that collapse is quasi-symmetrical around the orogen core thus giving rise to a mega-gneiss dome or double-sided core complex such as the Shuswap complex (McLelland 2010a, Wong et al, 2011).

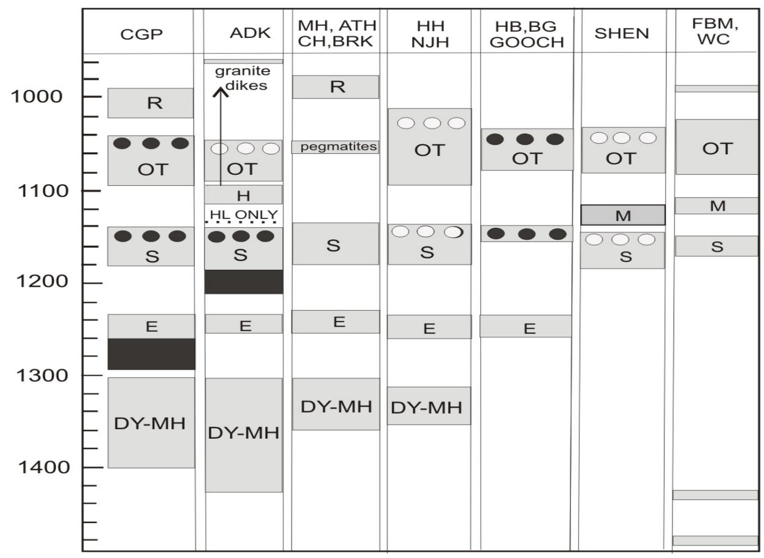

Fig.3 Chart summarizing Mesoproterozoic orogenic events in eastern North America during the interval 1.5-0.9 Ga. The gray blocks labeled R and OT represent the ca 1.09-1.03 Ga Ottawan orogenic phase and the ca. 1000-980 Ma Rigolet phase of the Grenvillian orogeny,respectively (Rivers, 2008). The black circles represent late- to post-tectonic AMCG suites, whereas the smaller, white circles represent MCG magmatism only. The black blocks in the Elzeverian (E) and Shawinigan (S) orogenies represent pre-orogenic arc magmatism. The events labeled DY-MH in the lower gray blocks represent ca.1.45-1.3 Ga continental arc magmatism along the southeastern margin of Laurentia. Portions of the arc rifted during the formation of the Central Metasedimentary Belt (ca. 1.3-1.22 Ga) and are now situated in the Adirondacks and the Mesoproterozoic inliers of the northeast Appalachians where they are known as the Dysart-Mt. Holly suite. Geographic abbreviations from left to right: CGP – Canadian Grenville Province, ADK – Adirondack Mountains, MH – Mount Holly, ATH – Athens dome, CH – Chester dome, BRK – Berkshire Mountains, HH – Hudson Highlands, NJH - New Jersey Highlands, HB – Honey Brook uplands, BG – Baltimore Gneiss domes, GOOCH – Goochland Terrane, WC– Wolf Creek, SHEN– Shenandoah Blue Ridge, FBM– French Broad Massif. *Modified after McLelland, 2010a.*

LYON MOUNTAIN GRANITE

The mineralogy and chemistry of Lyon Mountain granite has been thoroughly reported by Postel (1952), Whitney and Olmsted (1988, 1993), Whitney, 1992, and Valley et al. (2011) who showed it to be a ferroan, A-type leucogranite. McLelland et al. (2001a, 2001b, 2002) addressed the zircon geochronology of LMG and its close relationship with hydrothermal sodic alteration and Kiruna-type magnetite ores. Unequivocal evidence demonstrates that LMG is an igneous intrusive rock emplaced at ca. 1050 Ma (McLelland et al. 2001a,b; Selleck et al. (2005), Valley et al., 2011). The regional outcrop pattern of LMG is significant as it tends to occur around the perimeters of the Adirondack Highlands (Fig. 2) leading McLelland et al. (2010) and Selleck et al. (2005) to propose that it was intruded along, and helped to lubricate and promote, detachment faults formed during ca. 1050 Ma extensional collapse of the Ottawan orogen. The purpose of the first part of this trip is to examine some outstanding river-washed exposures of LMG and some of the interesting features within it including remarkable quartz-sillimanite veins and nodules. To this end further discussion will be presented in the Road Log and Field Stops.

PRISMATINE-BEARING METAPELITES AND QUARTZITES ALONG THE MOOSE RIVER

Prismatine, the boron-rich end-member of the kornerupine solid solution (Grew et al., 1996; ideally $[(\square,Mg,Fe)(Al,Mg,Fe)_9(Si,B,Al)_5O_{21}(OH,F)]$) occurs in metapelitic and quartzitic rocks about 12 km east of Ager's Falls (Stop 2) on the Moose River. Kornerupine-group minerals are generally rare, having been described from nine localities in the Grenville Province (Grew, 1996; Darling et al., 2004; Korhonen and Stout, 2005) including two in the Adirondacks (Farrar and Babcock, 1993; Farrar, 1995; Darling et al., 2004).

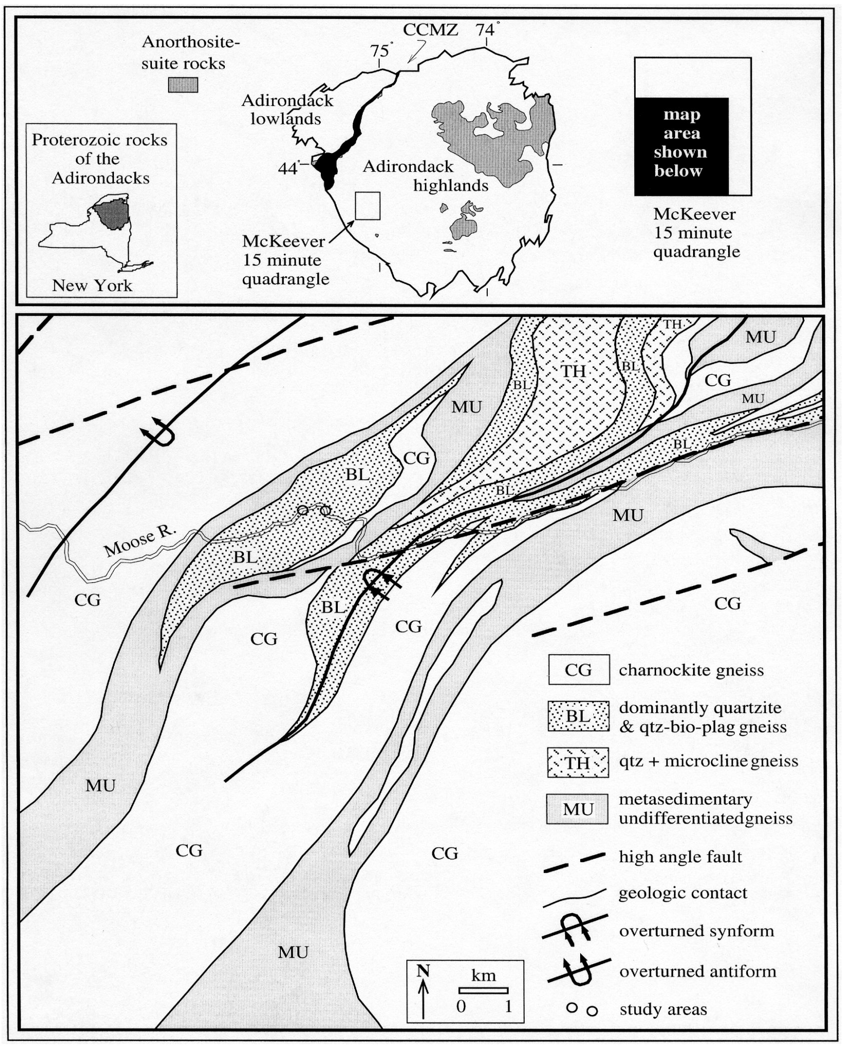

Fig. 4. - Map showing location of prismatine-bearing metapelites and quartzites (open circles on Moose River) and surrounding bedrock geology. Stop 6 is easternmost open circle. Geologic map units, structures, and relations illustrated are from Whitney et al. (2002). *Taken from Darling et al. (2004).* Bio—biotite; CCMZ—Carthage-Colton mylonite zone; pl—plagioclase; qtz—quartz.

Along the Moose River, prismantine occurs at two locations (separated by about 400 meters, Figure 4) within a unit of heterogeneous metasedimentary rocks (Figure 4, unit BL) mapped by Whitney et al. (2002). This unit comprises mostly quartzite and biotite-quartz-plagioclase gneiss with lesser amounts of calcsilicate rocks, and minor amphibolite, quartzofeldspathic gneiss, and calcite marble (Whitney et al., 2002). These rocks are interlayered with: 1) other metasedimentary undifferentiated rocks (Figure 4, unit

MU) that contain relatively more calcsilicate and less quartzite than BL, 2) thick tabular bodies of granitic to locally charnockitic gneiss (Figure 4, unit CG), and 3) quartz-microcline gneiss with quartzite and calcsilicate granulite layers (Figure 4, unit TH). These units occur in a complex, southeast-verging overturned synform bordered on the northwest by a tabular, northwest-dipping body of CG several kilometers thick, and on the southeast by a domical body of batholithic proportions consisting of relatively leucocratic CG. Although the granitic and charnockitic rocks have not been dated, they are lithologically and geochemically similar to felsic rocks of the ca. 1150 Ma anorthosite-mangerite-charnockite-granite (AMCG) suite found throughout much of the Adirondack Highlands (McLelland et al., 2001; Whitney et al., 2002).

In addition to prismantine-bearing assemblages, the surrounding rocks contain metapelitic assemblages of a) cordierite + spinel + sillimanite + garnet + plagioclase + quartz + ilmenite + rutile +/- biotite, b) cordierite + orthopyroxene + biotite + K-feldspar + quartz, and c) orthopyroxene + plagioclase + K-feldspar + quartz +/- biotite, +/- garnet (Darling et al., 2004). Additional descriptions and interpretations are included in the Road Log and Field Stop section.

ROAD LOG

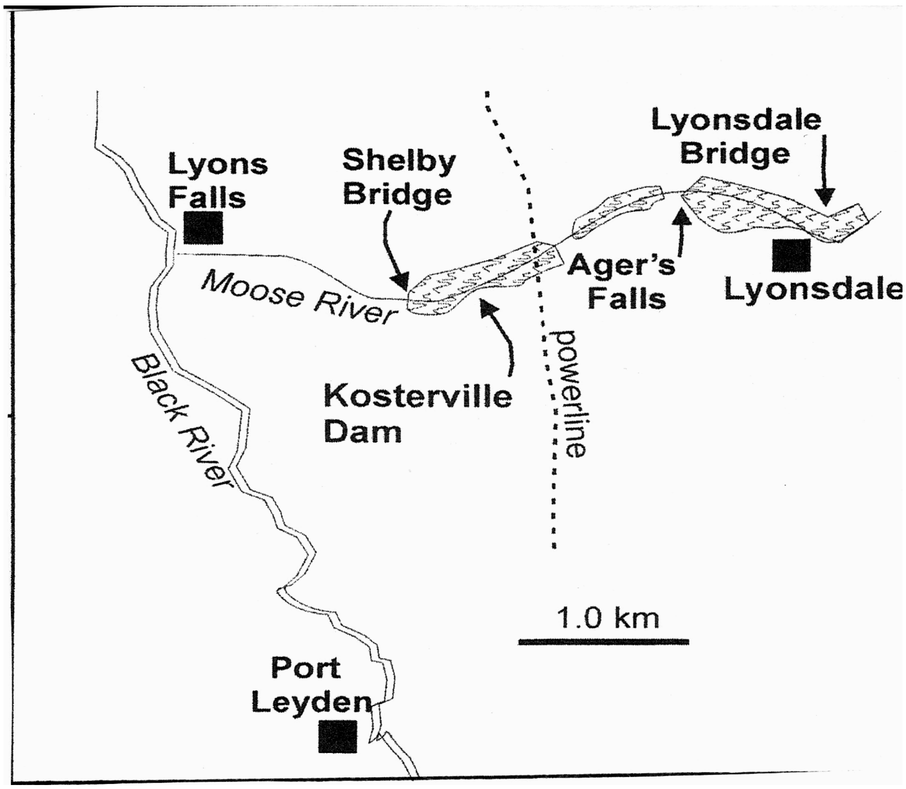

Fig. 5. Location map for Lyonsdale stops

CUMUL-ATIVE MILES	MILES FROM LAST STOP	ROUTE DESCRIPTION
0.0	0.0	Traffic light at intersection Rt. 12 and Main St., Port Leyden. Turn east on Main St.
0.5	0.5	Intersection with River Road. Turn left, Continue for 0.8 miles.
1.3	0.8	River Road forks left; continue straight ahead on Marmon Road.
2.7	1.9	Pass Penny Settlement Road on right; continue straight ahead.
4.5	2.6	Lyonsdale; turn right on Lowdale Road and cross three short one-lane bridges over the Moose River. Just before the third bridge park in parking area on left side of road. If no further room, continue over third bridge and park along shoulder. Meet at bridge.

STOP 1. LYONSDALE BRIDGE (45 MINUTES)

River-washed exposures begin at the bridge over the Moose River that adjoins the Burroughs Paper plant at Lyonsdale. A coarse, steeply east-dipping pegmatite can be seen descending down a steep cliff face and crossing the river onto the exposures immediately beneath the bridge. The pegmatite is zoned with a quartz core containing sillimanite and magnetite. McLelland et al., (2001) report a .single grain TIMS upper intercept age of

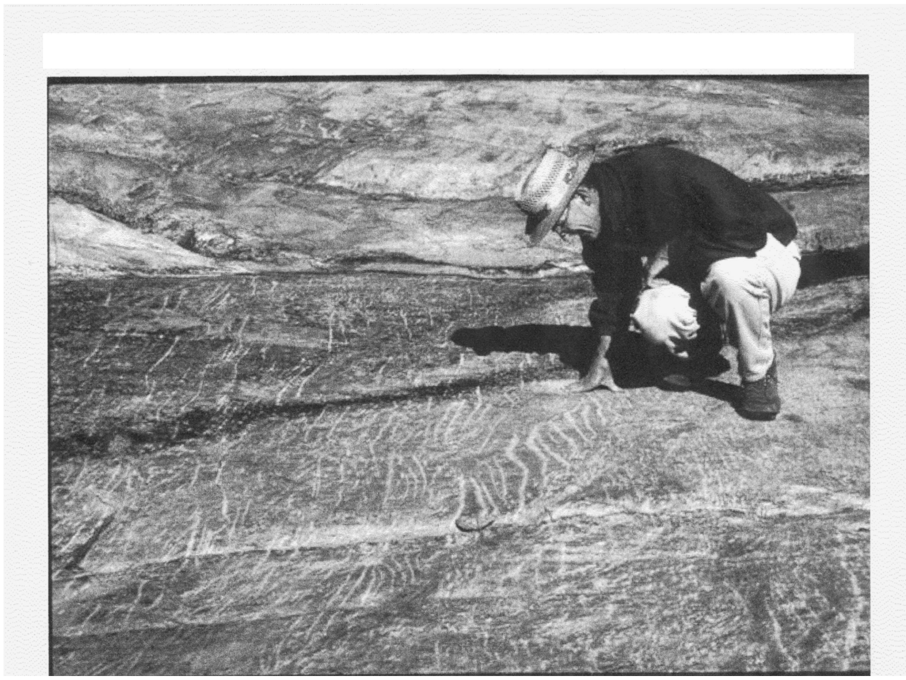

Fig. 6. Dugald Carmichael puzzles over the network of quartz-sillimanite veins and nodules below Lyonsdale Bridge.

1035 ± 10 Ma for the pegmatite (McLelland et al., 2002). A nearby sample of LMG yielded a SHRIMP and single grain TIMS age of 1046 ± 4.4 Ma (McLelland et al., 2001). The pegmatite age appears to be ~10 Ma too young and redating by SHRIMP would be very desirable. Nonetheless, it overlaps with the average LMG (n = 11) age of 1049 ± 10 Ma (McLelland and Selleck, 2011).

LMG consists of pink equigranular quartz and feldspar with small amounts of magnetite and occasional biotite. It shows virtually no penetrative grain shape fabric (Fig. 7b) and, despite examples of tight folding in the outcrop, appears to be undeformed by the nappe-like folding that affects all rocks older than 1050 Ma and imparts to them pronounced ribbon lineation and grain shape fabric. In contrast, LMG displays a good igneous texture resembling that of hypersolvus granite. Elsewhere it has been shown that late, upright folds trending E-W and NNE-SSW fold some LMG and were formed during the ca. 1050 Ma extensional collapse of the Ottawan orogen (McLelland et al., 2010). We are confident that the tight, disrupted, isoclines defined by quartz and granitic veins in the riverbed are the result of flow folding due to LMG magma emplacement (McLelland et al., 2002a,b).

As shown in Fig. 6, LMG has been penetrated by myriad white veins that consist of sillimanite and quartz with subordinate magnetite (Fig. 7a). There are several distinct generations of these, and it is clear that some groups are more deformed than others (Fig. 8a). The dominant, and perhaps youngest, trend is N50E. Careful examination of quartz nodules will show that they commonly form linear trends that are best explained by disruption or boudinage.

Where LMG intruded country rock, the former contains garnet. The original country rock was a calcsilicate but has been contaminated by the granite and its fluids. It is common to find aligned boudins of this rock in the stream bed. They tend to be angular and separated by granite and most likely represent blocks disrupted by the magma. Other important features to be seen are pegmatites that transition from two-feldspar to albite rich and even to quartz veins with magnetite. At one place a sillimanite-magnetite veinlet can be recognized when sunlight hits it in a certain direction. The same is true with several occurrences of very coarse (~6”) sillimanite.

A **B**

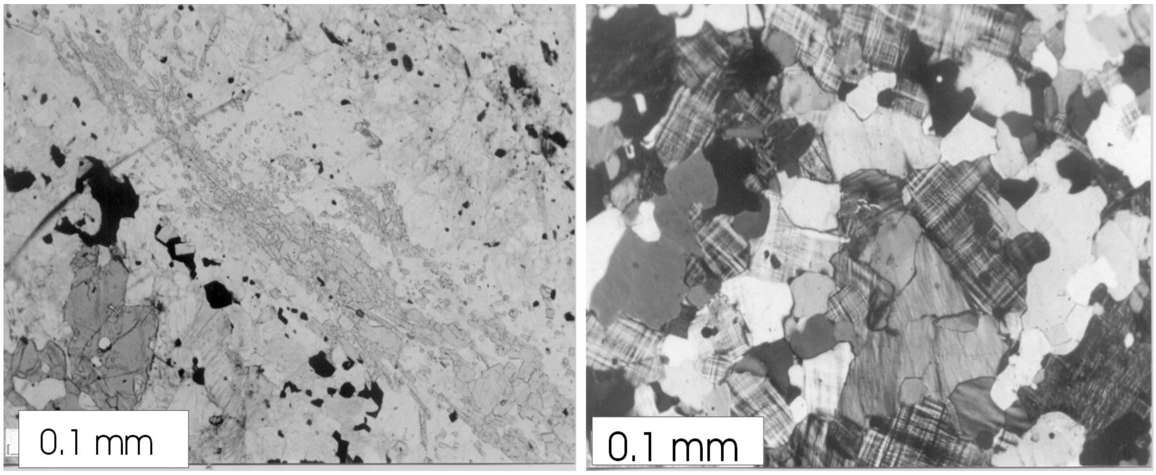

Fig.7. These images show photomicrographs of LMG collected at Lyonsdale Bridge. In 7a a quartz-sillimanite vein is shown with sillimanite needles at its core. Quartz (clear) occurs on either side of the sillimanite, and minor magnetite is sporadically present along the edges. Host LMG can be seen in the lower left hand corner.
7B shows a thin section of LMG under cross-polars. Microcline perthite and quartz dominate the field of view,
and there is no sign of deformation in the rock. *Modified after McLelland et al., 2002a.*

Mileage
4.7 0.2 mi. Return over bridges to Lyonsdale and turn right (west) on Marmon Road
5.1 0.4 mi. Turn right past gate at entrance to Ager's Falls power station. Take an immediate left and continue downhill to parking area.

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STOP 2. AGERS FALLS (45 MINUTES)

This stop exposes riverbed outcrops of LMG immediately southwest of a small hydroelectric facility on the Moose River. Fig. 8 presents some of the important relationships at Ager's Falls. Fig. 8A shows the same multiple, cross-cutting vein sets as seen at Lyonsdale Bridge. Note how the older set has been disrupted into nodules. In Fig 8B trains of nodules are shown crosscutting dark, xenolithic fragments of country rock calcsilicate. A similar scenario is shown in Fig. 8F that also contains late pegmatite veins. The crosscutting of the calcsilicate xenoliths is a critical relationship, because it establishes the post-intrusion origin of the quartz-sillimanite veins and nodules, i.e., they are not due to incorporation of "aluminous quartzite "into the magma. In Fig 8C several vein sets can be seen with the one pointed to by the pen clearly undergoing ductile boudinage consistent with extension in a fairly viscous magma. A very late, thin vein wanders across the field parallel to the pen. Figs. 8D and Fig 8G display a crucially important exposure in which a pegmatite-cored granite dike crosscuts, and ductiley deforms, the quartz-sillimanite vein-nodule system in a sinistral shear zone. The granite in the dike is good LMG identical to that in the country rock, but it contains no veins or nodules. This informs us that LMG magma was still present when the veins and nodules were being formed as a viscous magma was being fractured by rapid strain rate forces associated with emplacement into a fault controlled magma chamber. The granite then evolved a pegmatitic core as the magmatic episode drew to a close. It is likely that these processes were taking place near the top of the pluton, perhaps in a cupola. A self-consistent sequence of events is shown in Fig. 9 where a sheet of magma in a N50E shear zone has plucked off xenoliths parallel to its margins. As fluid pressure built in the crystallizing magma, movement along the shear zone triggered extensional fracturing parallel to sigma-1 and filled with hot, acidic magmatic fluids giving rise to quartz-sillimanite veins via hydrolysis. Sinistral shearing continued into frame 9C and resulted in rotation and boudinage of the earlier veins. By frame 9D, fluid pressure had built up again and a new set of extension fractures were produced along with subsequent rotation, boudinage, etc. In this way, the crosscutting vein system evolved with earlier veins getting rotated and boudinaged into nodules (McLelland et al., 2002 a, b).

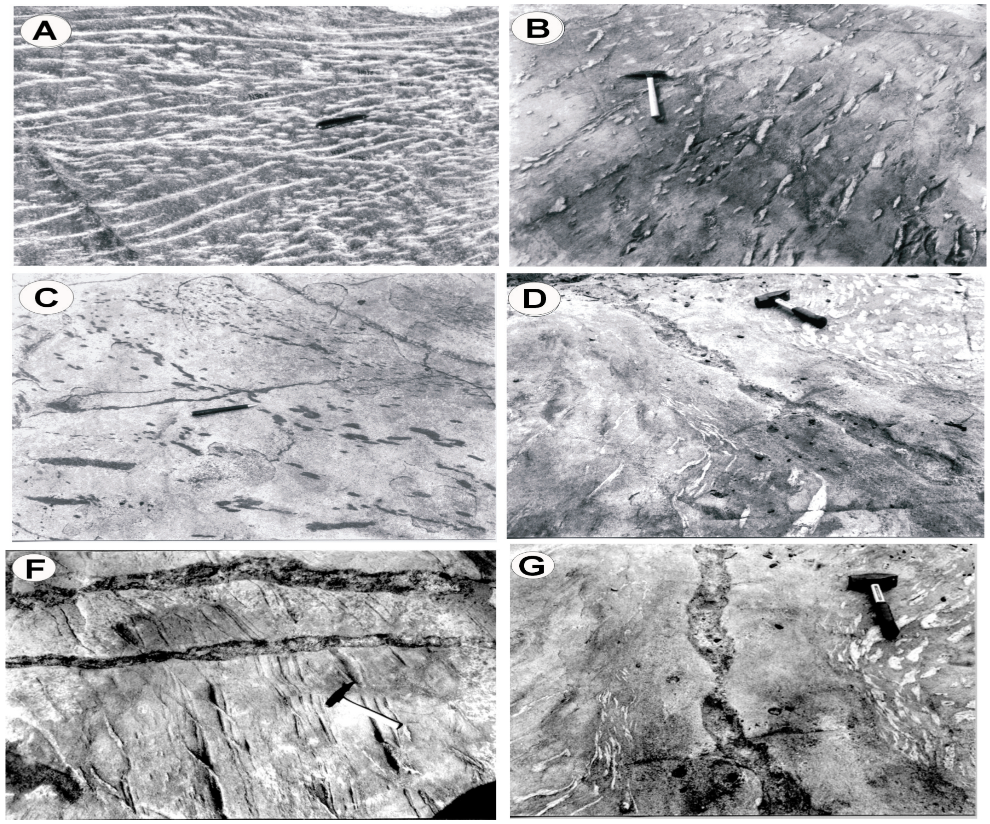

Fig. 8. Some granite-dike relationships at Ager's Falls, See text for explanation.
After McLelland et al. 2002a.

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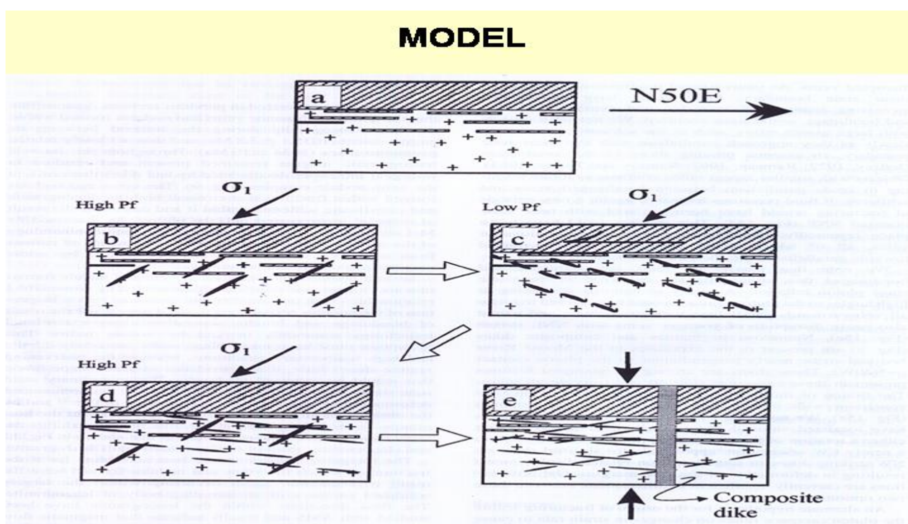

Fig. 9. The model shows the evolution of the vein-nodule complex.
See text for details. *After McLelland et al., 2002b.*

There are places in the Ager's Falls exposures where tight isoclinal folds defined by quartz veins are present, However these are disharmonic and the veins are discontinuous. The best explanation for these folds is that they are the result of flow-rate variations in a magma.

Fig. 10. Closeup showing veins crosscutting calcsilicate xenoliths.

As the Moose River is followed farther downstream, the quantity of quartz veins increases until the rocks pass into a thick, sillimanite-bearing megaquartz vein complex that we will visit at Stops 3, 4, and 5. The quartzite is clearly of hydrothermal origin, and we interpret it to represent the release of late magmatic fluids through the carapace of a large LMG pluton.

Mileage
5.75 0.65 mi. Proceed west along Marmon/Lyonsdale Road. Park in small parking area for trail access.

STOP 3. QUARTZ VEIN COMPLEX AT FISHING ACCESS SITE (15 MINUTES).

These stream exposures contain quartz veins, pegmatite veins, and veins of granite. The green and red coloration of portions of the rocks is the result of hydrothermal fluids acting on iron oxides and iron-bearing silicates. The wide expanse of quartz is due to widening of the vein system visited at the end of stop 2. The quartzite is interpreted as the end product of leaching of alkalis from the original granite leaving a residue rich in silica and alumina. Magnetite, hematite, and later chlorite result in the pink to green coloration in the rock.

Mileage
6.0 0.25 mi. Proceed west along Marmon Road. Park along east shoulder, walk down dirt road.

STOP 4. QUARTZ-SILLIMANITE ROCK & ILLITE-DIASPORE, KOSTERVILLE DAM (20 MINUTES).

Exposures at the base of the dam display ~100 feet wide coarse grained, massive, sillimanite-bearing quartzite discussed by Selleck et al., 2004). Broken-off fragments show randomly oriented sillimanite crystals up to 6” long. Scattered across the outcrops of quartzite are large (up to 20 cm.), porcellanous, and irregularly shaped salmon- and green-colored platelets consisting of diaspore and illite. It is common for these minerals to psuedomorph sillimanite. These features are interpreted as low T (<200C) alteration features related to later hydrothermal alteration that dissolved and redistributed silica and alumina as fluids permeated the fracture network. Fluid inclusion work indicates oxidizing and highly saline (>20% NaCl) fluids. Interestingly, the illite-diaspore platelets contain high concentrations of altered zircon that we interpret to be the consequence of leaching of labile constituents (e.g., alkalis) from the original granite leaving a residue rich in quartz, alumina, and zircon. A few viable zircons yield ages of ca. 1030-1050 Ma. Although the quartzite is poorly layered, there does appear to be a gently dipping tectonic foliation defined, in part, by the illite-diaspore platelets.

Mileage
6.9 0.9 mi. Continue down Lyonsdale Road to intersection with Shibley Road at Gould’s Mills. Turn right.
7.1 0.2 mi. Turn left at Kosterville Road. Park in open lot on right.

STOP 5. SHIBLEY BRIDGE COMPLEX (30 MINUTES).

These interesting exposures are a continuation of the mega-quartz vein occurrence seen at the last three stops. The border with the granite is visible near the west end of the waterfall. Beneath Shibley Bridge there occurs a rock consisting of what looks like a quartz boulder conglomerate. However, close inspection will prove it to be the megaquartzite that has been deformed in such a manner to produce boulder-size rounded boudins of coarsely crystalline quartz in finer, grain-size reduced quartz that superficially resembles a sedimentary conglomerate.

At the waterfall, staining of the rock with sodium cobaltinitrite reveals the presence of a fine grained, potassic phase that takes the yellow stain and clear quartz grains that do not. Thin sections reveal that the potassic phases consist of muscovite and K-feldspar that were deposited by fluids in fractures that have rendered the quartzite a mass of sand grain-size fragments. Directly above the waterfall, a dark, red-stained band heads in the direction of the road, Careful sampling reveals that it consists of very coarse sillimanite intergrown with magnetite. This assemblage was a principal iron ore at Benson Mines and, given its composition, must be hydrothermal.

STOP 6. PRISMATINE-BEARING GNEISSES (60 MINUTES)

CUMUL-ATIVE MILEAGE	MILES FROM LAST POINT.	ROUTE DESCRIPTION
0.0 mi.	0.0 mi.	Turn around and head back southwest on Shibley Rd.
0.6 mi.	0.6 mi.	Turn left onto River Rd.
1.9 mi.	1.3 mi.	Turn right onto Pearl St. (River Road; Co. Rt. 39) and continue south.
2.6 mi.	0.7 mi.	Make slight left and continue on Pearl St. for 0.1 miles.
2.7 mi.	0.1 mi.	Turn left onto Moose River Rd. and head east.
5.3 mi.	2.6 mi.	North-South Rd. on left.
7.7 mi.	2.4 mi.	Boonville / Moose River Rd. on right.
12.3 mi.	4.4 mi.	1876 Red School House on right.
12.4 mi.	0.1 mi.	Turn left into dirt parking area by Moose River and park.

Follow the all-terrain vehicle path downstream for about 200 meters. The path passes through the stone foundations of the former Moose River tannery and then follows rapids as the Moose River heads southwest. Here, the river cuts through northwest-dipping, calc-silicate gneisses and quartzites. Stay high on the river bank until the rapids disappear. The path will descend and cross a small, wet, muddy creek bed. Afterwards, the Moose River pools and turns north and the first outcrops on the west side of the river are the prismatine-bearing rocks. Please exercise caution while walking among the river boulders and talus at the base of the outcrops. Also, please DO NOT USE HAMMERS at this stop and refrain from collecting prismatine specimens unless you're planning to study them scientifically; a future geologist will be grateful someday.

The feature of geologic interest at STOP 6 is the exceptionally well-developed prismatine crystals in coarse-grained, feldspathic lenses. Here, prismatine crystals form dark greenish-black, euhedral, elongated grains (up to 10 cm in length) that commonly display radiating patterns in feldspathic lenses one to three cm thick (Figure 11A). The prismatine crystals *appear* to have grown only within the plane of the foliation. However, upon closer examination, the prismatine grains are seen to be arranged randomly, but the longest and best-developed crystals formed parallel to the foliation plane. Because of this, Darling et al. (2004) inferred that non-deviatoric pressure conditions prevailed locally during prismatine formation. It should also be noted that a number of prismatine-bearing feldspathic lenses are located adjacent to fine-grained tourmaline + plagioclase + biotite–rich zones near the north end of the exposed rocks. In these locations, the prismatine-bearing feldspathic lenses texturally embay, cross-cut earlier foliation, and appear to form at the expense of the tourmaline-bearing zones (Figure 11B). The embayed country rocks, coarser grain size, and the random arrangement of the prismatine crystals led Darling et al. (2004) to interpret the feldspathic lenses and the prismatine found in them to be of anatectic origin.

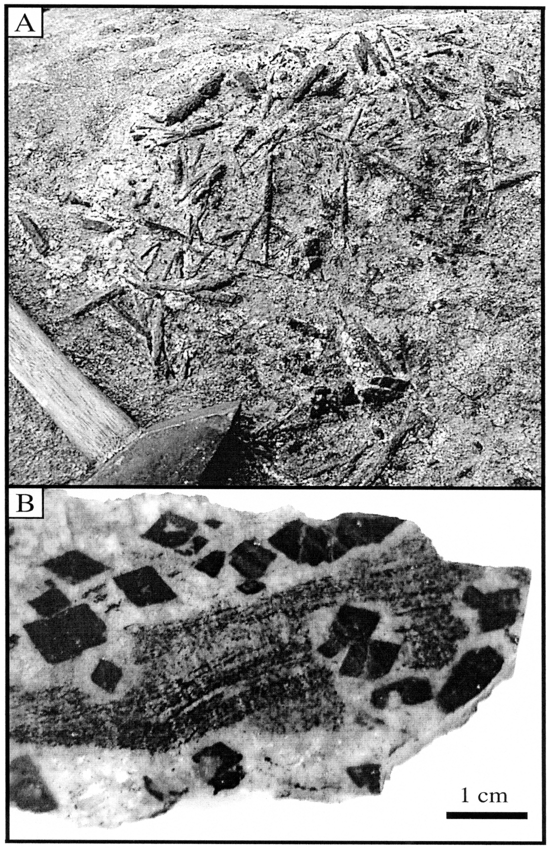

Fig. 11. - (A) Prismatine crystals (black) in coarse-grained feldspathic lens, taken parallel to plane of lens. Hammer for scale. (B) Euhedral prismatine (black) in coarse-grained feldspathic lens embaying fine-grained, foliated tourmaline + biotite + plagioclase–rich zones. *Taken from: Darling et al. (2004).*

Plagioclase, K-feldspar, minor quartz and rutile are the most common phases associated with prismatine, but biotite, cordierite, garnet and rarely sillimanite occur locally as well. The prismatine contains 0.73 to 0.79 formula units of B (out of 1.0) and has Mg / Mg + Fe between 0.70 and 0.73 (Table 3 of Darling et al., 2004). After determining the associated mineral compositions, Darling et al. (2004) proposed the following prismatine-forming reaction:

Tourmaline + sillimanite + biotite + cordierite → prismatine + rutile + melt (1).

This reaction is similar to a number of proposed prismatine-forming reactions from other granulite terranes (Grew, 1996), including those in sapphirine-free rocks found in the Reading Prong, New Jersey (Young, 1995), and in Waldheim, Germany (Grew, 1989). In those cases, garnet rather than cordierite was a proposed reactant.

Metamorphic temperatures and pressures are difficult to estimate from prismatine-bearing mineral assemblages as little is known about the stability of boron-rich kornerupine at pressures less than 10 kb (Schreyer and Werding, 1997). However, the prismatine occurs in proximity to low-variance metapelitic assemblages in the surrounding rocks. Specifically, thermobarometry calculations from net transfer and exchange equilibria record temperatures and pressures of 850° ± 20°C and 6.6 ± 0.6 kilobars for orthopyroxene + garnet assemblages and 675° ± 50°C and 5.0 ± 0.6 kilobars for cordierite + garnet + sillimanite + quartz assemblages (Darling et al., 2004). The former assemblage is interpreted to have formed during partial melting whereas the latter assemblage is interpreted to have formed on the early retrograde metamorphic path (Darling et al., 2004). The ~850°C temperatures derived from the orthopyroxene + garnet assemblage are reasonable for partial melting conditions. Although the cordierite + garnet + sillimanite + quartz assemblage occurs at STOP 6, it and the orthopyroxene + garnet assemblage are better developed further downstream at the second prismatine location (the westernmost open circle in Figure 4). These exposures can be reached by following the footpath on the south bank of the Moose River for a distance of about 400 meters.

Because many of the prismatine-bearing feldspathic lenses are saturated in both quartz and rutile, Storm and Spear (2009) intensely studied the prismatine-bearing lenses as part of a natural test of the titanium-in-quartz geothermometer of Wark and Watson (2006). Storm and Spear (2009) determined a wide range of metamorphic temperatures, specifically from 630 + 63 / -86 to 879 ± 8 °C, but most determinations fell between 700°C and 880°C (see Figure 8a of Storm and Spear, 2009). This is in good agreement with metamorphic temperatures determined by the aforementioned methods (Darling et al., 2004). Storm and Spear (2009) also provide convincing textural evidence that prismatine was locally replaced by leucosomatic quartz, most likely during melting of prismatine. Interestingly, it was the leucosomatic quartz that yielded the highest Ti-in-quartz temperatures (800-880°C; Figure 8a of Storm and Spear, 2009).

The timing of partial melting is unknown but is likely associated with either intrusion of the AMCG suite at ~ 1160-1145 Ma, or burial associated with the Ottawan phase of the Grenville Orogenic cycle and the associated intrusion of Lyon Mtn. granite at ~ 1050-1030 Ma (McLelland et al., 2010a).

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Trip A-5

SEQUENCE STRATIGRAPHY OF PLATFORM CARBONATES: DEVONIAN LIMESTONES OF JOHN BOYD THACHER STATE PARK, SOUTHWEST OF ALBANY, NY

In Honor of Dr. Gerald M. Friedman, a distinguished Geologist, educator, and a mentor.

The Text of this trip is adapted and revised from a paper written by **Dr. Gerald M. Friedman** which was published in the Field trip Guide for the 67th Annual Meeting of New York State Geological Association (October 13-15, 1995) and edited by John I.Garver and Jacqueline A. Smith.

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Figure *1. A plaque erected in 1933 in memoriam of those pioneer geologists whose researches in the Helderbergs in the nineteenth century*

ABSTRACT

The Lower Devonian strata that crop out on the Helderberg Escarpment illustrate the characteristics of marine platform parasequences in carbonate rocks. In the hierarchy of stratigraphic units, parasequences are a relatively conformable succession of genetically related strata bounded by surfaces of erosion. The parasequences exposed in this escarpment, which include stromatoporoid reefs, stromatolites, and lithified lime mud (micrite) show regressive fades separated by unconformities representing transgressive episodes. The lowermost part of the section exhibits classical karst-generated solution collapse breccia of the kind that hosts oils and gas in the subsurface. The Middle Devonian Onondaga fades at this site are full of coral-reef debris. In the subsurface, Onondaga reefs form gas reservoirs and now serve as reservoirs for gas storage. Sir Charles Lyell visited this classic site, part of the Helderberg Mountains, in 1841 and a monument commemorates his visit and that of Sir William Logan, James Hall, Amos Eaton, and others of the heroic age of Geology.

INTRODUCTION

In the hierarchy of stratigraphic units, parasequences are a relatively conformable succession of genetically related strata bounded by surfaces of erosion. The Lower Devonian parasequences exposed in this escarpment, which include stromatoporoid reefs, stromatolites, and lithified lime mud (micrite) show regressive fades separated by unconformities representing transgressive episodes. The Middle Devonian Onondaga fades Stratigraphically above the rocks at this site is full of coral-reef debris.

HISTORY

This classic site is on hallowed ground. A plaque erected in 1933 in memoriam of those pioneer geologists whose researches in the Helderbergs in the nineteenth century (Fig.1) made this region classic ground includes not only almost all American pioneer geologists, but in addition, lists pioneers of British and Canadian geology. Among those listed are Amos Eaton (1776-1842), the John Gebhards Sr. and Jr. (life-cycle dates not available), James Hall (1811-1898), William W. Mather (1804-1859), Lardner Vanuxem (1792-1848), James Eights (1797- 1882), Sir Charles Lyell (1797-1875), Benjamin Silliman (1779-1864), Edouard de Verneuil (1805-1873), James D. Dana (1813- 1895), Henry D. Rogers (1808-1866), William B. Rogers (1804- 1882), Ferdinand Roemer (1818-1891),
Louis Agassiz (1807-1873), and Sir William E. Logan (1798-1875). Sir Charles Lyell visited the "Helderberg Mountains", as he called them, in September 1841 and although he rejoiced, noting that "the precipitous cliffs of limestone, render this region more picturesque than is usual where the strata are undisturbed" (Lyell, 1845, p. 67), he was

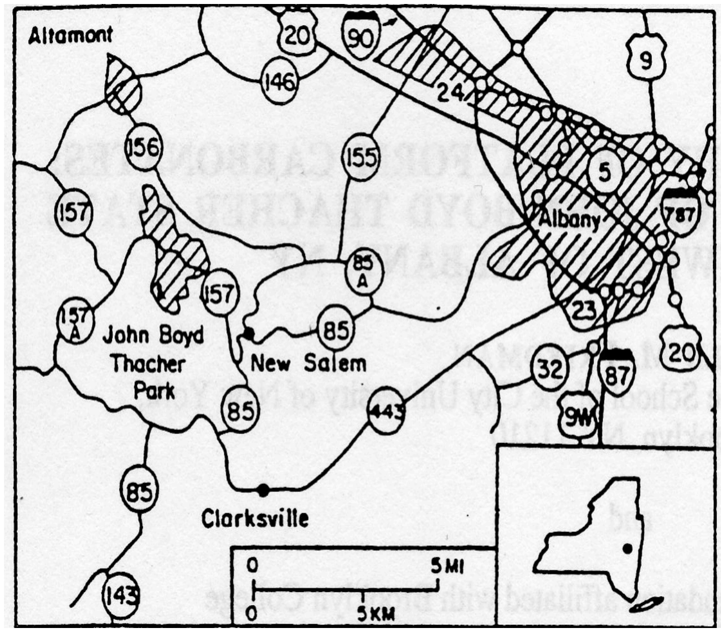
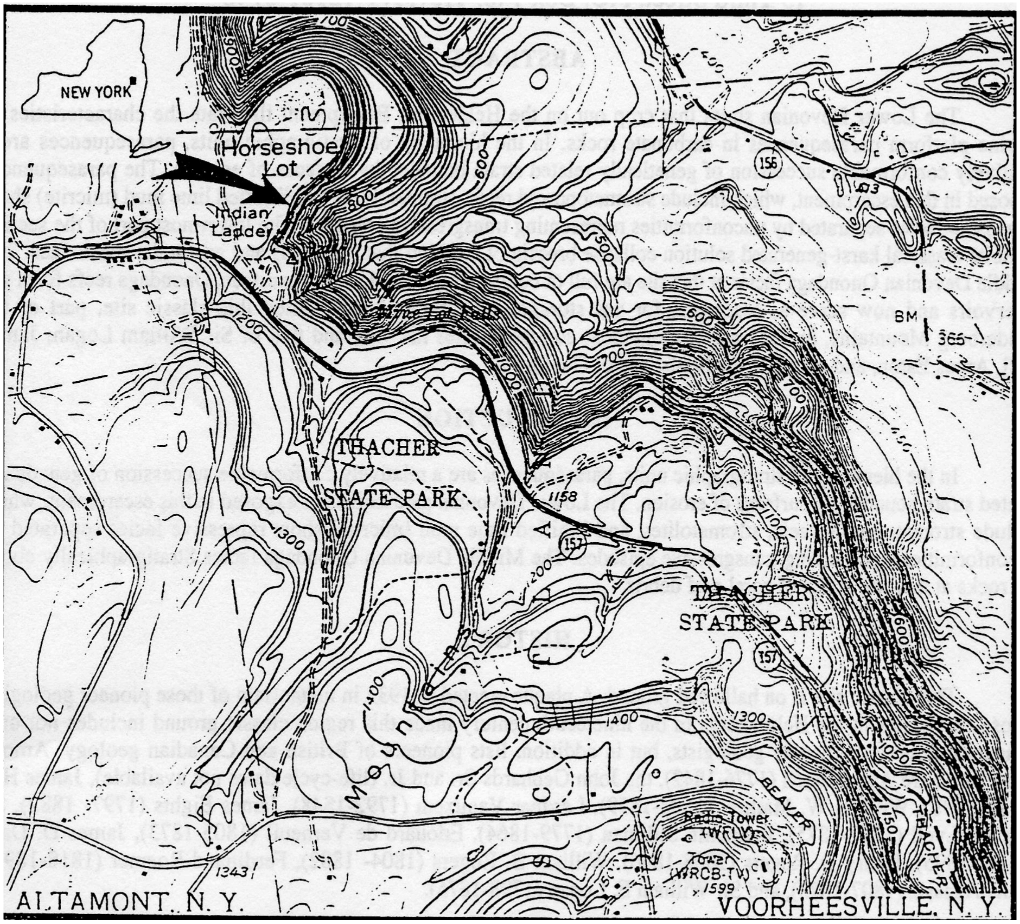


Figure 2A. Location of John Boyd Thacher-State Park (Fisher, 1987). The Indian Ladder Trail is located on Route 157.

more concerned in his account with the "Helderberg war" between Van Rensselaer and his tenants. On his return to the "Helderberg Mountains" in May 1846 the "Helderberg war" absorbed him again because he states that "the anti-renters have not only set the whole militia of the state at defiance, but have actually killed a sheriffs officer, who was distraining for rent." (Lyell, 1849, p. 260). The definitive studies of the Lower Devonian carbonates of the Helderberg Escarpment exposed at John Boyd Thacher Park date to the early New York State Geological Survey and were written by Vanuxem (1842), Mather (1843), and Hall (1843). Their reports were supplemented and complemented later in the nineteenth century.



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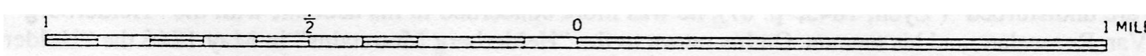

Figure 2B: Topographic map of John Boyd Thacher-State Parking showing location of Indian Ladder Trail.

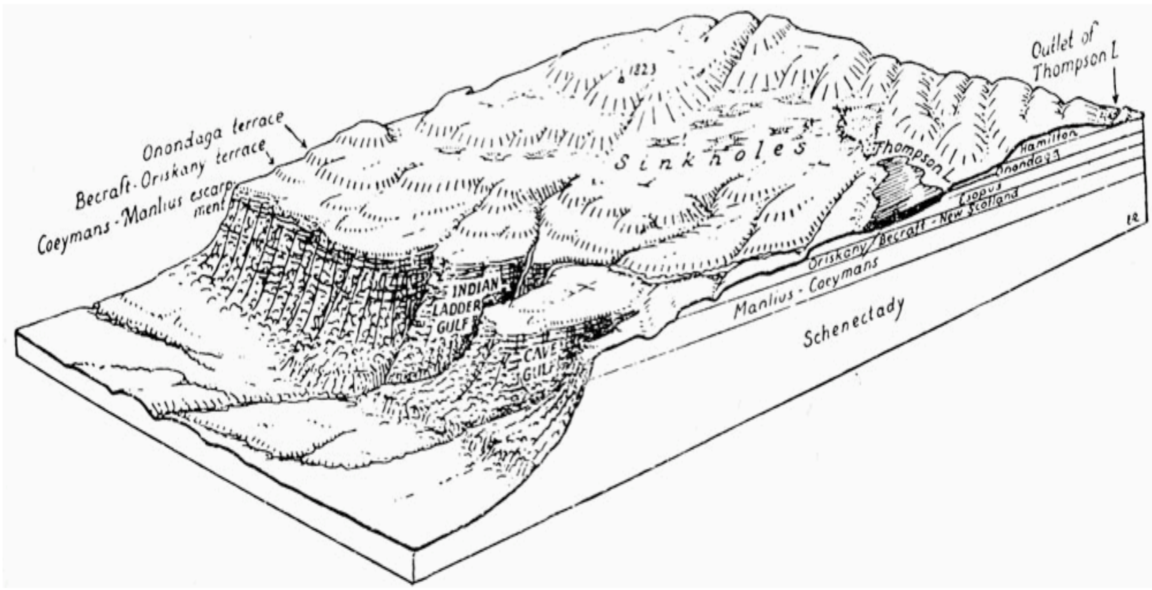

Figure 3. Block diagram of Helderberg Escarpment showing Indian Ladder location (labeled Indian Ladder Gulf), dip slope of Paleozoic strata, prominent terraces, and sinkhole topography (H.F. Cleland, 1930: modified by Winifred Goldring, 1935).

The Indian Ladder Trail site provides an unusual opportunity to study a vertical cliff of limestone strata: a vertical exposure of approximately 80 ft or 24 m exposed in the cliff is accessible by stairway and footpath; hand railings assure safety. One can view the entire sequence of the rocks at close quarter, including by hand lens; comparable physical settings in quarries never allow such close inspection.

Why the name Indian Ladder? Verplanck Colvin, one of the earliest men to write about the Helderbergs, in 1869 wrote:

"What is this Indian ladder so often mentioned? In 1710 this Helderberg region was a wilderness; nay all westward of the Hudson River settlement was unknown. Albany was a frontier town, a trading post, a place where annuities were paid, and blankets exchanged with Indians for beaver pelts. From Albany over the sand plains ... "Schenectada " (pine barrens) of the Indians ... led an Indian trail westward. Straight as the wild bee or the crow the wild Indian made his course from the white man's settlement to his own home in the beauteous Schoharie valley. The stern cliffs of these hills opposed his progress; his hatchet fells a tree against them, the stumps of the branches which he trimmed away formed the round of the Indian ladder."

The trail ended where the cliff did not exceed twenty feet in height. Here stood "the old ladder." In 1820 this ladder was still in daily use (Goldring, 1935). The modem stairway crosses the old Indian ladder road which ran to the top of the escarpment where the trail begins.

LOCATION

Figure 2 shows the location of the John Boyd Thacher State Park, where the Indian Ladder Trail reveals the vertical sequence of Lower Devonian limestones that rest unconformably on the Ordovician Indian Ladder beds and Schenectady Formation which can be seen, locally, in gullies below the escarpment (Fig. 3). Entering Thacher State Park from Albany on Route 157 stop at the "Cliff Edge Overlook" for a view of the Taconic and Berkshire Mountains, Adirondacks, Hudson River, and City of Albany; then drive to the next parking lot which has a sign La Grange Bush Picnic Area - Indian Ladder Trail. The trail is open from May 1 to November 1, weather conditions permitting. Descend here for study of the Lower Devonian carbonate fades. Examine also the memorial plaque near the cliff edge at the Mine Lot Creek parking lot which has been attached to a vertical rockwall. It says "in memory of those pioneer geologists whose researches in the Helderbergs from 1819 to 1850 made this region classic ground." The names of these pioneers have been cited under History.

THE SEQUENCE STRATIGRAPHIC COLUMN OF THE HELDERBERG GROUP

The cliff face exposes an excellent case history of sequence stratigraphy. Lower Devonian limestone of the Helderberg Group reveal sets of parasequences which may be recognized among the exposed formations (Rondout, Manhus, and Coeymans formations) (Fig. 3). Parasequences are the building blocks of vertical sequences. A parasequence is defined as a relatively conformable succession of genetically related beds bounded by surfaces (called parasequence surfaces) of erosion, nondeposition, or their correlative conformities (Van Wagoner, 1985). Each sequence is initiated by a eustatic fall in sea level rapid enough to overcome subsidence (Van Wagoner, 1985) or by epeirogenic upward motion. A parasequence surface commonly is an unconformity surface. Below the Indian Ladder Trail, where a waterfall known as Minelot Falls spouts across the path, sandstones and shales of the Middle Ordovician Schenectady Formation are mostly concealed beneath a cover of blocks of Devonian limestone forming a talus slope. At the waterfall, a major unconformity just below the trail separates the Ordovician strata from the Rondout Formation, exposed at the base of the cliff. The nonfossiliferous Rondout Formation is latest Silurian or earliest Devonian (Fisher, 1987). The rocks of this formation consist of brecciated dolostone cemented by gypsum. They display spectacular karst-generated solution-collapse features of the kind that hosts oil and gas in the subsurface. Evaporite minerals, now all dissolved, were present in this deposit. Note the concentration of springs which relates to the pore space created by ground-water dissolution of the evaporites.

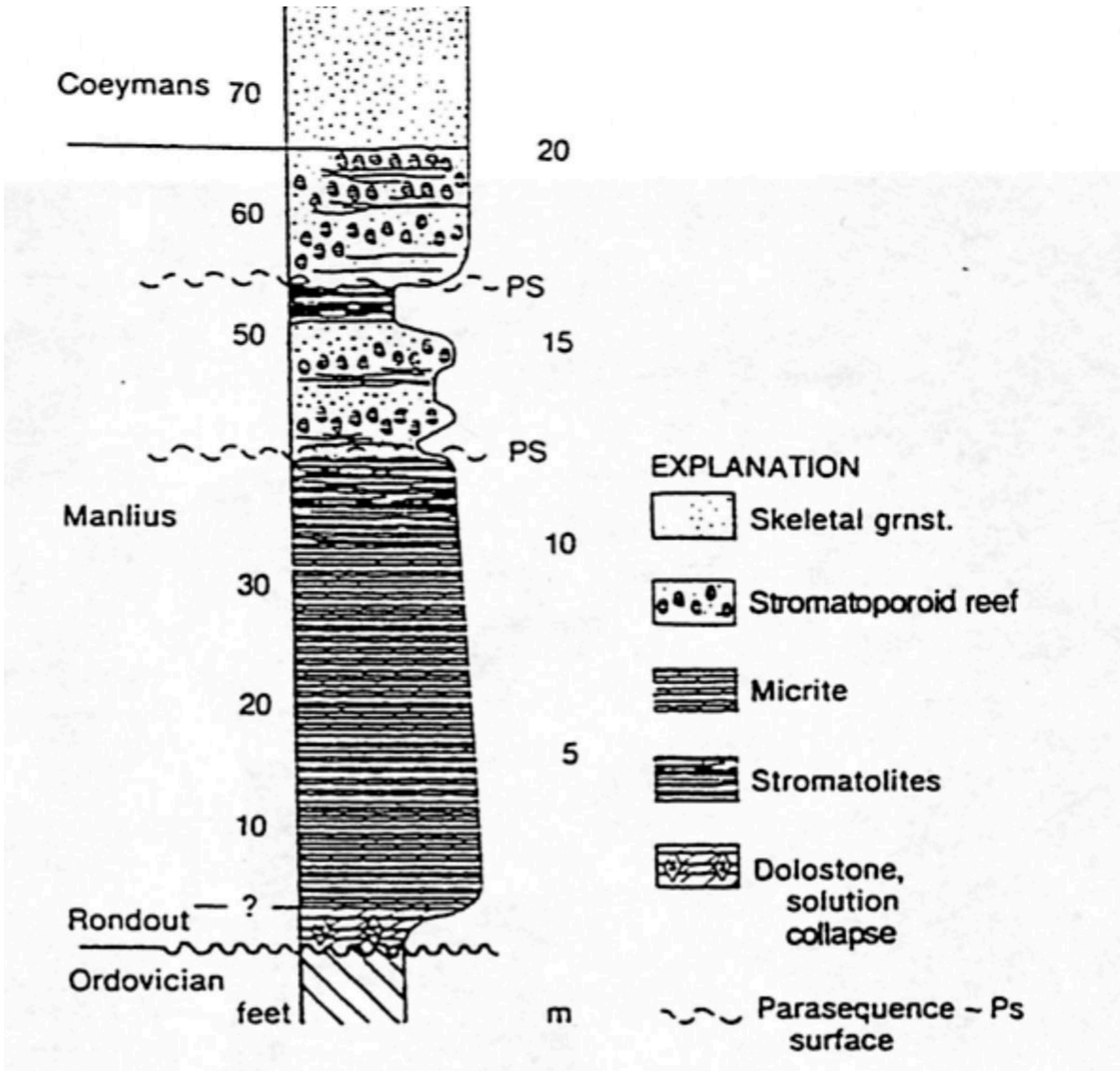

Figure 4a. Columnar stratigraphic section, parasequence surfaces, and facies distribution of the Lower Devonian carbonates. The Rondout Formation at the base is overlain by the Manlius Formation, and the top of the section extending to the break in slope at the top of the cliff is occupied by the Coeymans Formation. The stratigraphic section exposed on this trail is the type locality for the Thacher Member of the Manlius Formation, proposed by Rickard (1962). The Ordovician deposits at the base of the section are Indian Ladder beds- Schenectady Formation (see Fig. 3).

The Lower Devonian strata that crop out on the escarpment at John Boyd Thacher State Park illustrate the characteristics of marine parasequences in carbonate rocks (Figure 4). Two of the several parasequences are constituted as follows: a skeletal grainstone is overlain by stromatoporoid reefs and these, in turn, are overlain by interreef grainstone [interval between the upper PS (parasequence surface) and the scale mark for 20 m on Figure 4]. An underlying parasequence consists of skeletal grainstone that grades up into stromatolites (algal-laminated mudstone). Each of these two parasequences consists of strata that were formed when a depositional slope prograded seaward. The surfaces bounding the parasequences (labeled PS in Figure 4) are inferred to have resulted from rapid submergence. A set of several repeating parasequences, as shown in Figure 3, is known as a parasequence set, defined as "a relatively conformable succession of genetically related parasequences bounded by surfaces (called parasequence set boundaries) of erosion, non-deposition, or their correlative conformities."

Concepts identical to those just set forth, and developed independently of the definition of parasequence in seismic stratigraphy were formulated under the name of Punctuated Aggradational Cycles (PACs). What have been named PACs are thin (1-5 m) upward-shoaling cycles whose boundaries are defined by the depositional products of episodes of rapid submergence (Goodwin and Anderson, 1982).

OUTCROP GUIDE

Studies of vertical sequences should normally be worked from the base of the section upward. However, at this exposure it is best to work the section downward following the stairway from the edge of the cliff.

The top of the section is composed of skeletal grainstone (locally skeletal packstone) in which fossils, especially brachiopods, and crinoids, are evident (Fig. 4a); the pentamerid Gypidula coeymanensis is prevalent. This facies is part of the non laminated Coeymans Formation (Fig. 4b). Its lower contact is sharp and obvious in the field. Below this contact is the Manlius Formation which underlies most of this escarpment. A stromatoporoid reef with locally intercalated skeletal grainstone represents the top of this formation (Fig. 3). The stromatoporoids show their distinctive globular concentric structures resembling cabbage heads. Previous authors (Fisher, 1987; Rickard, 1962) have termed this reef facies a biostrome, presumably because its geometry in outcrop is sheetlike rather than mound- shaped. In my experience with reefs of all ages, I have observed that most large reefs are Hat on top and bottom, especially on the scale of this exposure. Other geologists share this experience, thus Shaver and Sunderman (1989) note "virtually all large reefs seen on outcrop have eroded, flattened tops, whereas smaller reefs that were not naturally aborted and that were unaffected by erosion as seen on outcrop have convex-upward rounded tops."

Figure 4b. This facies is part of the non laminated Coeymans Formation.

Close examination of the reef facies reveals a fine-grained matrix between the ramework-building stromatoporoids. This matrix resembles micrite, a lithified lime mud; hence this facies may be misinterpreted as representing a low-energy setting. However, in modern reefs, cement forms millimeters to centimeters beneath the living part which, in thin section, is finely crystalline (cryptocrystalline) and semi-opaque. Hence the matrix in such reefrock looks just like low-energy micrite (Friedman et al., 1974). Case histories abound where unwary geologists have confused high-energy reefrock with a supposed "low-energy" lime-mud facies (Friedman, 1975). Therefore, the observation of a fine-grained matrix between the framework builders does not deter, in fact confirms, the interpretation that this part of the section formed as a high-energy reef facies, and not in a low-energy setting. The stromatoporoids are massive which in the ecologic zonation of Devonian reefs represents the shallowest-water zone of a subtidal setting. Below the reef facies occurs a stromatolitic (finely laminated) facies which is recessed back creating a near cavelike morphologic feature (Fig. 5). This recessed feature can be traced throughout Thacher State Park and is known as "Upper Bear Path". By analogy with modem environments the stromatolitic facies represents a low-energy mtertidal or supratidal setting. The sharp contact between the intertidal or supratidal low-energy stromatolitic facies and overlying subtidal high-energy reef facies represents a parasequence surface (Fig. 4a).Downward from the stromatolites, a stromatoporoid reef facies is present, separated by bedded skeletal grainstone from the stromatolites.
in fact, the reef facies is present twice (Fig. 6). Hence downward the setting changes from intertidal or supratidal to subtidal shallow water. Below this double-reef section, the

change is again interpreted to be intertidal or supratidal stromatolites. Hence, once again, a parasequence surface separates the subtidal high-energy reef facies from the underlying intertidal to supratidal stromatolites (Fig. 4a). Interestingly, this stromatolitic facies is resistant to erosion (Fig. 6). and it projects outwards in the cliff face, whereas the upper stromatolite facies is recessed almost cavelike Below this lower stromatolite facies the lithology and facies are that of a low-energy, thin-bedded micrite with local

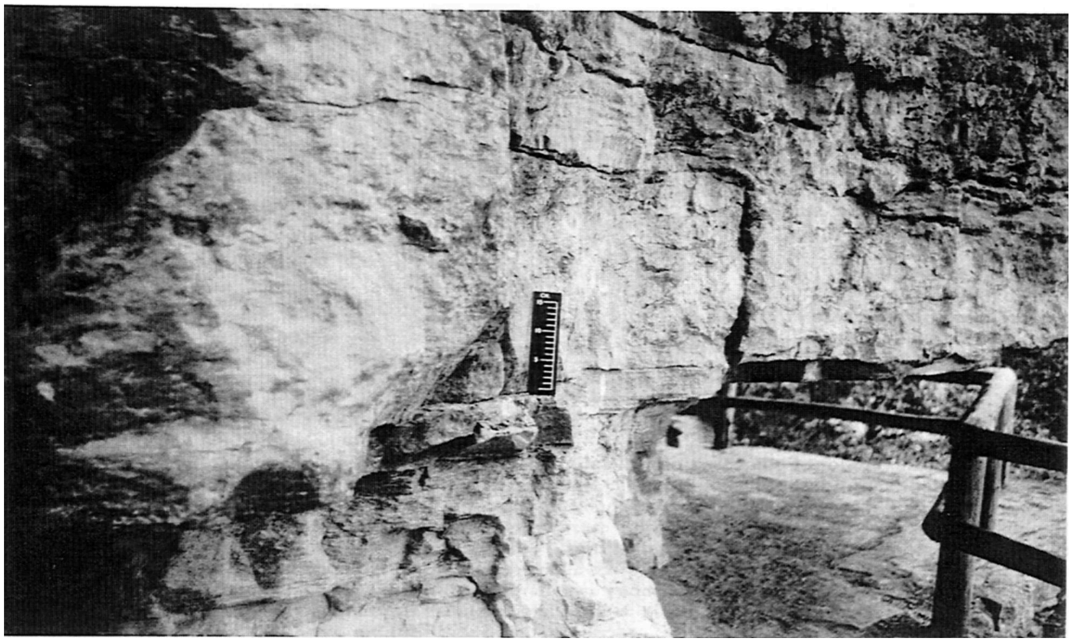

Figure 5. Photograph showing recessed underlying stromatolitic (finely-laminated) facies and overlying stromatoporoid reef facies of the Manlius Formation. Sharp contact between the two facies on which scale rests is parasequence surface (see Fig.4).

skeletal grainstone occurring as finely interbedded couplets, scour-and-fill structures, local cross-bedding, and some beds containing abundant spiriferid brachiopods, tentaculitids, ostracodes, and bryozoans. Near the base of the Manlius Formation occur several thicker beds, up to about 20 cm in thickness. Near the base of the section is the Rondout Formations, a recessed zone at the base of the cliff characterized by breciated carbonate rock cemented by gypsum. Its exact contact with the overlying Manlius Formation is subject to debate. In the columnar section (Fig. 4a) the Rondout Formation is identified where solution-collapse features are prominent and the lithology changes to dolomitic, especially dolomitic stromatolites, with sporadic intercalated calcitic laminae and shale laminae, an interpreted supratidal facies. Clasts of solution-collapse breccia are prominent together with gypsum-filled veins. The angular clasts of the collapse breccia may have resulted from collapse and brecciation of overlying carbonate strata when evaporites underlying them were dissolved. It represents a karst setting. Springs and caves, which are present here, are a function of dissolution of evaporites by groundwater. Karst- type openings were created during subaerial emergence. The Rondout Formation is commonly known as Rondout Waterlime. Its base is at or below the trail.

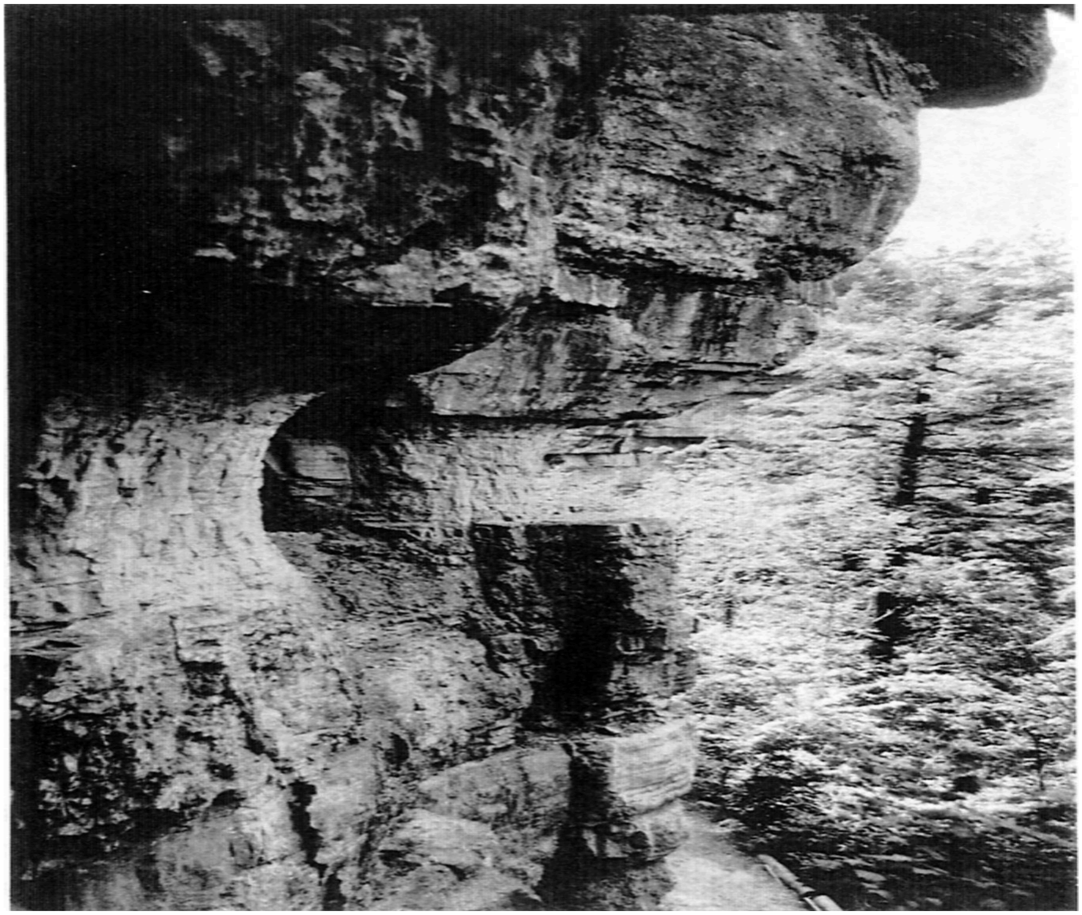

Figure 6. Photograph showing from top downward: massive projection reef separated by parasequence surface from underlying recessed stromatolites ("Upper Bear Path"). Below recessed stromatolite facies note in descending order bedded skeletal grainstone, non-bedded reef, bedded skeletal grainstone, reef, parasequence surface, and non-recessed stromatolite fades recognizable as a well-bedded fades. For detail compare with Figure 3.

THE ONONDAGA FORMATION

In John Boyd Thacher State Park the Onondaga Terrace (Fig. 3) exposes carbonate rocks of the Onondaga Formation which in the subsurface produces gas and serves as gas-storage reservoirs. Take Rt. 157 southwest to exit of park and continue to Indian Ledge Road (right turn-off before NY 85) (Fig. 1). Turn right on to Indian Ledge Road and drive 0.7 mile, park on right of road just above uppermost limestone outcrop. The outcrop exposes the Edgecliff Member of the Onondaga Formation. This limestone in not reefal. It is a skeletal limestone and ranges in this exposure from micrite to skeletal gramstone. Cnnoids and coral fragments are abundant. Of interest in this outcrop is the coarse-grained moldic facies in mid-section. Dissolution of fossils has led to high-porosity fades of a kind that would make excellent reservoirs for oil or gas. Decide for yourself whether this high-porosity zone continues into the subsurface or is only a surface feature. This porous facies vies in its porosity with the best of reservoir facies in the subsurface The accumulation of this coarse debris resulted from an episodic event, perhaps a storm or even a tsunami Note the prominent erosional surface which underlies this deposit (Fig. 7). Lindemann (1979) described the biofacies of this exposure.

Figure 7. Photograph of exposure of Onondaga Formation, Indian Ledge Road, Onondaga Terrace (Fig. 2). Note prominent erosional truncation on mid-section. High-porosity moldic storm coquinite overlies surface of truncation

Driving Directions To John B. Thacher State Park

You may use Google Maps to get the directions to John B. Thacher State Park, New Scotland, New York.
Park Information Telephone Number: (518) 872-1237

TRIP LOG

From: Syracuse University, NY 13202
To: John B. Thacher State Park, New Scotland, NY 12186
Total 139 Miles about 2 hours and 46 minutes.

1. Head north on Crouse Dr toward University Pl	0.4 mi
2. Take the 1st left onto University Pl	381 ft
3. Take the 1st right onto Irving Ave	0.3 mi
4. Turn left onto Harrison St	0.2 mi
5. Take the 1st right onto Almond St	59 ft
6. Merge onto I-81 N via the ramp on the left to I-690 W	4.2 mi
7. Take exit 25A to merge onto I-90 E toward Albany	
Toll road	89.4 mi
8. Take exit 29 toward US-10/Canajoharie/Sharon Springs	
Toll road	0.2 mi
9. Continue straight	
Toll road	0.2 mi
10. Turn left onto NY-5S E/E Main St	
Continue to follow NY-5S E	2.5 mi
11. Take the NY-162 N ramp	0.2 mi
12. Merge onto NY-162 S	13.8 mi
13. Continue onto New York 30A S	3.8 mi
14. Turn left onto New York 30A S/NY-7 E	1.1 mi
15. Turn right onto New York 30A S/Zicha Rd	
Continue to follow New York 30A S	1.2 mi
16. Continue onto NY-30 S	1.4 mi
17. Turn left onto NY-443 E	10.5 mi
18. Keep left at the fork	413 ft
19. Turn right to stay on NY-443 E/Helderberg Trail	3.4 mi
20. Turn left onto Co Rd 252/Thacher Park Rd	
Continue to follow Thacher Park Rd (157)	5.1 mi
21. Turn right onto John Boyd Thacher State Park	75 ft
22. Take the 1st right to stay on John Boyd Thacher State Park	0.3 mi
23. Continue to the main entrance to the Indian Ladder Trail	

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Trip B-3

SILURIAN AND DEVONIAN EURYPTERID HORIZONS IN UPSTATE NEW YORK

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The upper Salina Group consists of a minimum of 70 feet of argillaceous beds (Camillus Formation) that weather rather rapidly. Most beds are mudstones or dolomitic mudstone and shale. The succeeding Bertie Group begins with the deposition of hypersaline „algal‟ mounds and a eurypterid fauna – the Fort Hill Waterlime (see Ciurca, 1973, 1990). Overlying this is the Oatka Shale – dolomitic mudstone with no fossils currently known.

The initiation of Bertie Group sedimentation with a unit of eurypterid-bearing, stromatolitic waterlime set the stage for deposition of overlying units all, in one way or another, deposited during a regime of extensive microbialite sedimentation in the Late Silurian. This, and the numerous eurypterid faunas, typifies the Bertie Group sequence across upstate New York and southwestern Ontario, Canada.

Ciurca (1990), in a redefinition of the Bertie Group, added similar lithologic units that occur above the Akron- Cobleskill to the Group (e.g. Moran Corner Waterlime) considering them part of depositional cycles occurring as the result of the shifting of paleoenvironments to and fro during the close of the Late Silurian of the region. We may never know if latest Silurian strata occur in New York as many of the distinguishing zonal fossils may never be found within our sequences. In the meantime, it has been continuously suggested that *Eurypterus* characterizes the Late Silurian and that *Erieopterus* characterizes our Early Devonian (at least in the northeast U.S. and Ontario, Canada). The earliest invasion of the Helderbergian transgression may have removed our latest Late Silurian sediments. See Ciurca 1990, Fig. 5, Distribution of Eurypterid-bearing Waterlimes.

Except for the *'Eurypterus' pittsfordensis*, found at the base of the Salina Group, true *Eurypterus* are found throughout the Bertie Group and readily distinguish the beds from units such as the Early Devonian Manlius Group or somewhat equivalent beds (Honeoye Falls Dolostone or the Clanbrassil Formation of southwestern Ontario, Canada) – these are replete with the eurypterid, *Erieopterus*.

Figure 1 Algal mounds (microbialites) in the Fort Hill Waterlime at the base of the Oatka Formation downstream from the NY 96 bridge over Flint Creek, Phelps, N.Y. The dam/falls can be seen in the background. The mounds rest on a layer of dolostone bearing abundant evidence of hypersalinity (e.g. salt hoppers).

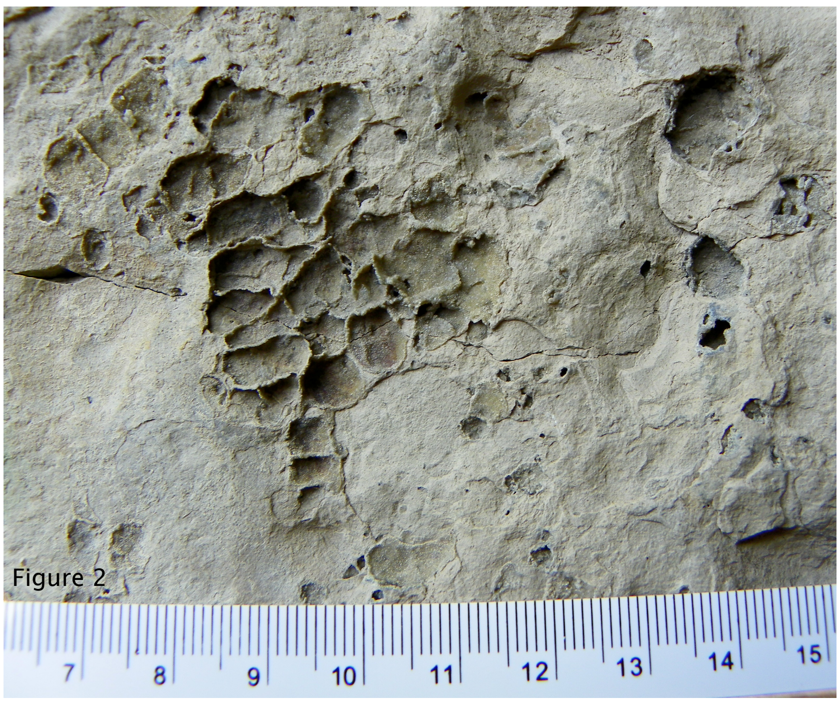
STRATIGRAPHY

Selected Strata and Localities

Fiddlers Green Formation, Litchfield Town Hall

The Fiddlers Green Formation is exposed in a roadcut opposite the Litchfield Town Hall, Litchfield, New York. Two members are exposed, the Phelps Waterlime constitutes the upper portion and the rest is Victor Dolostone. Hundreds of specimens of eurypterid remains (and associated fauna) were obtained by this author and over the years (since the 1960s) fossil collectors have found many more.

The Phelps Waterlime here is about 1.5 m thick, gray in color but much darker when fresh. It breaks with conchoidal fracture and is characterized at this site with a zone (~0.25 m thick) of stromatolites and mudcracks at the top. Recently, interesting features have been observed in the uppermost transition to the overlying Forge Hollow Formation (shaly, platy dolostone near the contact with the Phelps). These include “boxwork” (Figure 2 – note optical illusion) and small-scale mudcracks associated with algal mats at the top of the stromatolite beds. (Figure3).



The boxwork occurs on the underside of the rock and, presumably, represents the structures left by the removal of gypsum nodules. To the west, the Forge Hollow Fm. contains thick beds of gypsum that were formally extensively quarried (Syracuse area and Cayuga Lake region). Mudcracks are a common structure at the top of the Phelps Waterlime, but at this site they are intimately meshed with the stromatolite beds. The small-scale mudcracks illustrated above are associated with algal mudflats that were present just prior to deposition of the Forge Hollow Fm. They have also been observed at Flint Creek at Phelps, N.Y. but there are overlain by the Ellicott Creek Breccia.

Cobleskill Formation

While the Cobleskill Formation in its limestone facies (type section westward to Cayuga Lake) is replete with a Silurian marine fauna (see Berdan 1972 and Stock 1979, for example, for faunal listings), the unit is much more complicated in structure and distribution than generally realized (Ciurca 2005). Eurypterids occur below (Williamsville-Oxbow), within and above most facies across upstate New York.

Of particular interest to this author is the occurrence of eurypterids above the main portions of the Martisco Reef Complex (MRC) (Ciurca 2003, 2005) from Rock Cut Gorge westward to Auburn. *Eurypterus* sp. (*dekayi*- type), a dolichopterid, ostracods, brachiopods and a single conulariid have been retrieved thus far from a thin zone above the reef. Within the fossil-bearing bed, about 0.25 m thick, salt hoppers occur indicating still another eurypterid bed that formed under hypersaline conditions. Salt hopper structures are particularly abundant in some of the eurypterid-bearing waterlimes in New York and Ontario, Canada.

Recently, another small structure (Figure 4) has been located north of the MRC and is interpreted to be another example of a highly dolomitized patch reef (North Reef). Silicified stromatoporoids occur here also.

Figure 4 North Reef – Note massive bedding, apex, and wedging. Rock above reef is slightly cherty and is mostly of stromatolitic origin and overlain by more typical Chrysler Formation.

Eurypterid remains are not easy to extract from the waterlime in the sequence. However, many specimens have been obtained through persistent search. Figure 5 shows the characteristic eurypterid found in this zone.

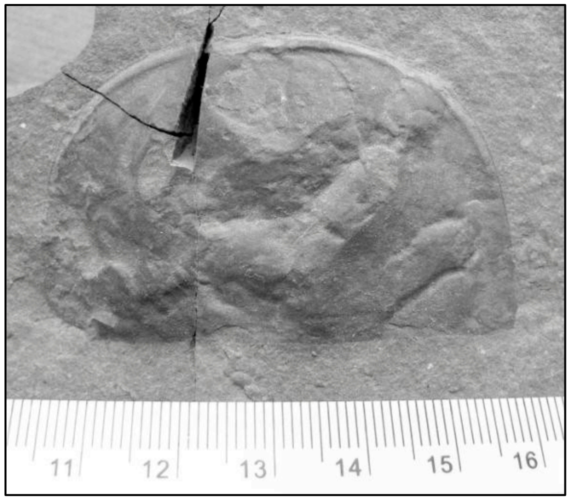


Figure 5 Wide eurypterid carapace of the *Eurypterus dekayi* type. Species is undetermined. Note small ostracods and trace fossil.

At Owasco Outlet, Auburn, New York a different "Cobleskill" Limestone is present with a eurypterid horizon slightly older than that of the MRC. This is treated below as the "Canoga Waterlime" – still another eurypterid- bearing unit within the Late Silurian of New York. New York probably has more eurypterid horizons than any place in the world (for now).

Canoga Waterlime (new)

Immediately overlying the Cobleskill Limestone at Auburn, New York are about 2 m of thin- to thick-bedded, fine-grained dolostone (waterlime). Owasco Outlet is the suggested type section of the Canoga Waterlime. Thus far, the unit is limited in occurrence to northern Auburn, Cayuga Lake and west to Seneca Falls.

At Owasco Outlet, the Canoga Waterlime bears a *Eurypterus* fauna with small leperditiid ostracods. Windrows, consisting of very finely-comminuted carbonaceous material, have been observed and stylolites are common.

How this dolomitic unit has gone unnoticed for so long is perplexing. In 1909, D. Dana Luther noted that waterlime occurred above the stromatoporoid-bearing Cobleskill Ls. at the old McQuan‟s Quarry south of Seneca Falls and even noted the occurrence of eurypterid fragments:

> "Overlying the stromatopora layer in the McQuan quarry there is a bed of dark somewhat shaly magnesian limestone 9 feet thick, some parts of which are dolomitic. It is the only exposure on this quadrangle (Geneva-Ovid) of the Rondout waterlime, a formation 40 feet thick in the eastern part of the state…" and "segments of Eurypterus have been found 2 or 3 feet below the top of the bed."

The region about Auburn and Seneca Falls is anomalous (herein the ASF Anomaly). Normally, the Canoga Waterlime (occurring above the Cobleskill limestone facies as it does) would be considered part of the Chrysler Formation. However the „other‟ Cobleskill Fm. occurs higher in the section (see discussion of Cobleskill Formation).

Correlation of the Canoga Waterlime with other eurypterid-bearing units is not currently possible as it is sandwiched between the two „Cobleskill‟ units mentioned above. The contact with overlying units has not been precisely observed, but a hard, argillaceous (presumably dolomitic) unnamed unit does occur slightly above.

Luther"s observation over a hundred years ago was keen. The unit he observed is the Canoga Waterlime as defined above and we now finally know more about the eurypterid fauna characteristic of the unit, i.e. *Eurypterus* sp. (Figures 6 and 7).

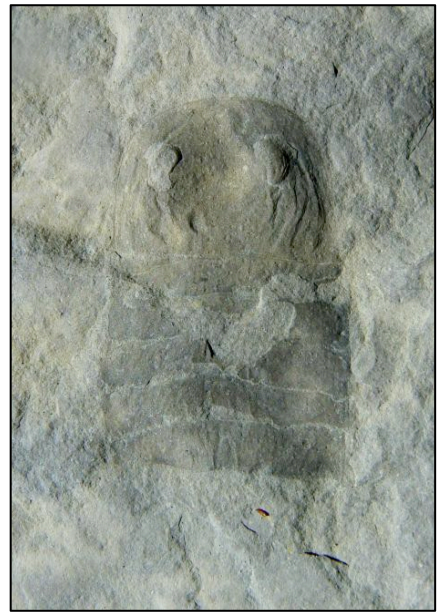


Figure 6 (left) and Figure 7 (right): Somewhat thick beds of the Canoga Waterlime (~2 m thick). Below this are beds of Cobleskill Limestone. The eurypterid found in the Canoga Waterlime is a type of *Eurypterus* sp. and detailed comparisons with eurypterids in other horizons have not been made.

Flint Creek (Phelps) and Mud Creek (East Victor)

A comparison of sections shows stromatolite layers of the Phelps Member at Flint Creek grading into typical eurypterid-bearing waterlime facies. The mudcracks at the top of the Phelps Waterlime appear to grade into a stromatolitic facies bearing occasional desiccation features among algal clasts (this situation also occurs over
100 miles to the east at Litchfield Town Hall – see STOP 1). At Mud Creek, East Victor, New York, the

mudcracked beds are at least 0.35 m thick, more typical of the distribution of this marker horizon east to Passage Gulf. At Mud Creek, the mudcracks appear beneath the Ellicott Creek Breccia (ECB), but to the west disappear either due to unconformity, or more likely by grading into a zone of stromatolite mounds of the ECB or the underlying (upper) Victor Dolostone.

At the Neid Road Quarry, northeast of LeRoy, such stromatolite mounds are common and eurypterids are found between and above the mounds in hypersaline settings (salt hoppers are found that are up to a foot on a side). See Ciurca 2005, Nudds and Selden 2008 for illustrations.

A peculiar feature at Mud Creek is worth noting. The "topographic waterlime" (so characteristic of the upper ECB in Canada – Figure 8) is present, presumably only as large clasts (Figure 9) as if most of the unit had been removed before deposition of the overlying Scajaquada Fm. The ECB at Mud Creek is rubbly-looking and heavily brecciated. It is believed to represent part of a widespread Late Silurian paleoseismic event, attesting to a violent origin. Clasts (within the breccia) are often vertical to the bedding. Topographic waterlime is so unusual that it can be recognized even as pieces.

Figure 8 Topographic Waterlime: Part of a large block from the topmost stromatolite zone of the Ellicott Creek Breccia Member, Fiddlers Green Formation at Fort Erie, Ontario, Canada (RQS quarry). The specimen is wet to bring out details of this highly unusual type of eurypterid-bearing waterlime. The microbialite mounds in this area are measured in meters. Note incipient brecciation.

Figure 9 Ellicott Creek Breccia (Fiddlers Green Formation) at Mud Creek, East Victor, New York. Note the very large clast of topographic waterlime to the right of the hammer.

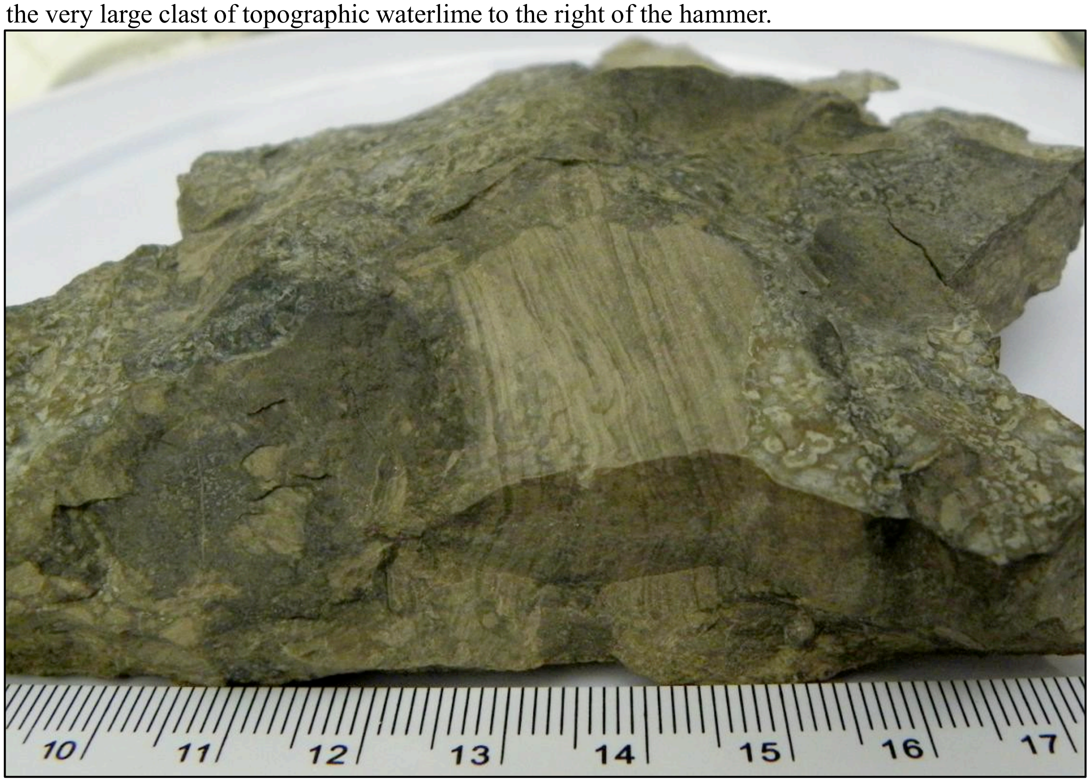


Figure 10 Chunk of Ellicott Creek Breccia with large clast of "algal" topographic waterlime in center. Note the clast is perpendicular to the bedding – Flint Creek at Phelps, New York.

At Flint Creek, the Ellicott Creek Breccia is exposed for an unusually long section of the creek from south of the William Street Bridge to the north sections where it is seen in the wall of the exposures leading to the dam at NY 96. Various other lithologies are represented in the clasts including what looks like black shale – provenance unknown.

At the type section, at Ellicott Creek in Williamsville, the ECB is a tripartite unit with stromatolites concentrated in the middle and is about 2.5 m thick. The *Eurypterus remipes* Fauna is present and occurs throughout the distribution of the unit. Other than eurypterids, fossils are rare except for ostracods and occasional cephalopods. Salt hopper structures are relatively common at many sites. See Figure 85 in Nudds and Selden 2008 for a color photograph.

Chittenango Falls, New York

The Silurian rocks exposed on the east side of NY 13, south of the falls, present an unusually thick section of dolomitic strata („Cobleskill", Chrysler) with overlying Helderbergian limestones. See Rickard, 1962 for detailed stratigraphy.

Massive, fine-grained dolostone constitutes most of the "Cobleskill" at the base of the cliff. Rare stromatoporoids have been observed and fragmentary eurypterid remains, some unusually large pieces.

For many years, Celestite crystals and associated minerals were eagerly sought after by collectors (Ciurca 1962). The source was the Chrysler Fm. which appears in the cliff above the "Cobleskill" as a series of thin- to medium-bedded dolostones often with mudcracked surfaces. The crystals line prolific vugs and even the mudcracks. Fossils are not common, but *Erieopterus* occurs in beds within the upper half of the Chrysler Fm. (above the Celestite beds). The Silurian-Devonian boundary is placed approximately at the base of a thick waterlime (Figure 10). The (time) interval between deposition of the "Cobleskill" here and the appearance of the *Erieopterus* Fauna represents long emergence to the atmosphere with multiple layers of sun-baked sediments in what must have been a playa-like setting.

Figure 11 Chittenango Falls NY 13 Section: Massive beds at base are „Cobleskill" Formation with Chrysler dolostones above. Dark patches (excavations) are where collectors obtained Celestite crystals over the years. Arrow marks approximate position of the Silurian-Devonian boundary.

The section at Chittenango Falls lacks the tripartite division seen at Clockville that has a limestone unit in the middle with abundant, though poorly preserved fossils. At Clockville, though, a large *Eurypterus* is present in the upper third associated with brachiopods. How these sections are related to the Martisco Reef Complex to the west in not understood. What is evident, however, is the continuous shift in paleoenvironments as Late Silurian sedimentation occurred throughout the region, layer-cake in presentation but much more complex in realization. Traced eastward, the „Cobleskill" becomes interbedded with fascinating eurypterid beds in limestone and dolomitic facies that are most unusual and little-understood. Though they all contain Bertie-type eurypterid faunas, the faunas are still little-studied compared to the marine biota. Collections of some of the material obtained thus far are in the research collections of the Peabody Museum of Natural History in New Haven, Connecticut.

CONCLUDING OBSERVATIONS

It would appear that many factors contributed to the deposition of Bertie Group sediments. These include proximity to a northern/eastern shoreline, degree of carbonate buildups (e.g. stromatoporoids) and restriction that allowed for the formation of hypersaline lagoons (what I call, alglagoons) over most of the region. It is assumed here that restriction was caused not only by biostromes and localized patch reefs, but also by great expanses of microbialites formed in shallow waters of Late Silurian bays/seas. Further, carbonate shoals contributed to restriction due to great piles of shell material (e.g. *Whitfieldella*) along with the microbialite mounds. Cross-bedding observed in such units as the Victor Dolostone demonstrates active currents (not a static environment) in which shell debris is piled up into „dunes" that form shoals that finally become micritized and blended into the surrounding sediment.

Figure 12 Victor Dolostone, Marcellus Falls, N.Y. Crinkled stromatolitic beds with low amplitude stylolites (irregular black lines) with abundant micritized *Whitfieldella* (brachiopods) and cross-bedding and microbreccia.

All of these factors must have influenced salinity in what are interpreted as near-shore lagoons for there we find salt hopper structures in great abundance associated with prolific fragmentary eurypterid remains and other biota transported into them during episodes of storm activity (tempestites).

What has recently been made clear to this author is that the more one looks at the Bertie Group, with its multiplicity of eurypterid horizons, the more various morphotypes of microbialites become evident – a fertile area for study. The Late Silurian, at least in upstate New York and Ontario, Canada, was a strange place.

ACKNOWLEDGMENTS

Mark Wade was invaluable in the field as we reinvestigated some of the important exposures of several of the Bertie Group units with special emphasis on the Phelps Waterlime and the Ellicott Creek Breccia. Many new observations were made and some of these will be the subject of future publication.

ROAD LOG FOR SILURIAN-DEVONIAN EURYPTERID-HORIZONS, UPSTATE NEW YORK

CUMULATIVE MILEAGE	MILES FROM LAST POINT	ROUTE DESCRIPTION
0.0	0.0	Exit 30 Toll booth, NYST I-90 at Herkimer
0.1	0.1	Turn left, Follow NY-28 south through Mohawk to Illion
0.7	0.6	NY-28 (continue through town, W. Main St.
1.5	0.8	Entering Illion
2.2	0.7	Remington Arms Factory on left
2.5	0.3	Turn left (head south NY-51)
4.0	1.5	Entering Illion Gorge (Steele Creek) ENTERING ORDOVICIAN TIME ZONE
5.7	1.7	Town of Litchfield
6.1	0.4	ENTERING SILURIAN TIME ZONE
8.1	2.0	Vernon Red Shale
8.9	0.8	Syracuse Formation on left (above creek)
10.2	1.3	Cedarville
10.6	0.4	Turn right at stop sign, follow Cedarville Road
11.3	2.0	STOP 1 Litchfield Town Hall (on right)

STOP 1. EXPOSURES: THE VICTOR AND PHELPS MEMBERS, FIDDLERS GREEN FORMATION

The type section of the Fiddlers Green Fm. is along Butternut Creek, north of Jamesville, N.Y., where the upper beds form the top of a waterfall. While the rock there is much darker in appearance, the divisions seen here are recognizable there. At the type section there is an upper, very fine-grained dolostone (Phelps Waterlime) with eurypterids and salt hoppers and below, .massive beds of the Victor Dolostone

When I first discovered this locality in the 1960s, I called it Con Loc 1 (or Ciurca Locality 56) as we were trying to keep a hoard of collectors from visiting the site in the early years. The site was glacially polished and striated and quite resistant to intrusion, but with weathering and working it slowly over several years, it produced a large number of eurypterid specimens and associated fauna. After a brief interval of several years, when the site was mostly covered with locally-collected drainage ditch material, the site became popular with eurypterid collectors who continued to excavate the rock layers to the state the exposure is today. The research collection I made over the years is available for study at the Peabody Museum of Natural History in New Haven, Connecticut.

Return to the New York State Thruway, I-90 and head west to the Canastota Exit (Exit 34).

69.4	43.0	Toll booth – Canastota.
69.6	0.2	Turn right on NY 13 and stop at McDonalds.
69.7	0.1	Leave McDonalds, turn right and follow NY-13 south.
70.3	0.6	Erie Canal Museum on right.
71.0	0.7	NY-5, continue south on Oxbow Road.
72.9	1.9	Cotton Road, Clockville (continue south).
73.7	0.8	STOP 2 Roadcut on east side of Oxbow Road (pull into parking area).

STOP 2. EXPOSURES: LATE SILURIAN AND EARLY DEVONIAN STRATA (CLOCKVILLE)

This is an important section, stratigraphically. Rickard (1962) gives a detailed description of the exposed strata. At the base is the uppermost Fiddlers Green Formation (Bertie Group) with eurypterids and salt hoppers. Overlying strata include the Forge Hollow, 'Cobleskill,' Chrysler, Thacher and part of the Olney Formations. At least four eurypterid horizons are present in the section. The Silurian/Devonian boundary occurs within the Chrysler Fm. here. At Chittenango Falls, portions of the Chrysler Fm. are rich in horizons of Celestite – a strontium mineral (STOP 3).

Return north to NY 5 and turn left following NY 5 to Chittenango.

76.4	2.7	NY-5 (head west to Chittenango.
82.4	6.0	NY-13 Chittenango, follow NY-13 south.
87.4	5.0	Parking area on right, opposite large cliff face (STOP 3, see Figure 11).

STOP 3. EXPOSURES: LATE SILURIAN AND EARLY DEVONIAN STRATA (Celestite locality)

For many years, this was a famous mineral collecting locality (mostly Celestite and Calcite). See text for a description of the stratigraphy. Note the massive beds at the base of the cliff – these are the dolomitic facies of the 'Cobleskill' Formation. Slightly farther up the road is Chittenango Falls State Park with one of the most beautiful cascading-type waterfalls. Trails take you to the bottom and are well worth seeing. A side canyon just to the north shows what appears to be large 'algal' reefs in the Thacher Fm.

87.4	5.0	Turn around and head back to NY-5.
96.9	9.5	Fayetteville (continue west on NY-5.
99.5	2.6	Turn right, I-481 (ramp) and head north to I-690.
101.0	1.5	Exit 4 to I-690.
111.0	10.0	Exit 6 to I-695 south.
113.2	2.2	Right lane for Camillus west, NY-5 (to Auburn).
114.1	0.9	NY-5 west.
117.1	1.0	Exit to NY-5 (Camillus, Warners).
117.4	0.3	Turn left, head to Camillus.
118.1	0.7	Turn right, village of Camillus.
118.2	0.1	Turn left, NY-174, head south through a beautiful valley.
121.4	3.2	Marcellus Falls.
121.9	0.5	STOP 4 Park at Marcellus Paper Company.

STOP 4. MARTISCO REEF COMPLEX – 'COBLESKILL' FORMATION

This is one of the most unusual facies variations of the 'Cobleskill' Formation. Sedimentary structures here are interpreted to be part of a highly dolomitized reef complex with silicified fossils, wedging of beds, unusual black chert deposits and succeeding stromatolite beds with breccia and eurypterids. See Ciurca 2005 and http://eurypterids.net/MartiscoReefComplex.html for additional comments. Note: Parts of the overlying Chrysler Fm. are well-displayed here.

One other stop is suggested for those interested on this field trip. From this site, take the road back to NY-5 and head west through Auburn and continue to Waterloo, New York. Turn right on NY-96 and head for Phelps where a bridge right in the village goes over Flint Creek (see text). Looking north from the bridge is the Oatka Shale, a rare outcropping. Looking south, below the dam, is lowermost Fiddlers Green Fm. and above the dam is upper Fiddlers Green Fm. including a suggested paleoseismite represented by the Ellicott Creek Breccia. We can collect samples of this for those interested. Many eurypterid specimens have been retrieved from the Phelps Waterlime of this area. Another interesting feature of this locality is the four foot bed of sandstone at the Silurian/Devonian unconformity (just below the Onondaga Limestone).

END OF FIELD TRIP

Note: From Phelps, there are many routes heading to the NYS Thruway or directly to Rochester.

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Trip B-4

PALEOECOLOGY OF MIDDLE DEVONIAN BLACK AND GRAY SHALES OF CENTRAL NY

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INTRODUCTION

The dysaerobic zone is a broadly defined biozone that encompasses all reduced oxygen taxonomic occurrences. However, because reduced oxygen conditions include a wide range of bottom water oxygen levels, from the edge of metazoan habitability to near normal marine conditions, a single biofacies is not appropriate to describe all reduced oxygen settings. Further, reduced oxygen conditions in modern oceans are common in a broad range of physical settings and were likely common in an even more diverse range of settings in the geologic past (Tyson and Pearson, 1991; Calvert and Pedersen, 1993; Lyons, 1997; Raiswell and Canfield, 1998; Hurtgen, et al., 1999; Lyons et al, 2003; Levin, et al, 2003). The Finger Lakes region of central New York provides a unique opportunity to examine a range of interpreted relative oxygen levels preserved in black shales. Within these units, taxonomic and paleoecologic variations through the dysaerobic zone are recognized. On this trip we will explore localities representing a range of conditions through the dysaerobic zone and highlight the range of organisms adapted to low oxygen conditions and the additive pattern of diversity and life habits through these intervals.

GEOLOGIC SETTING

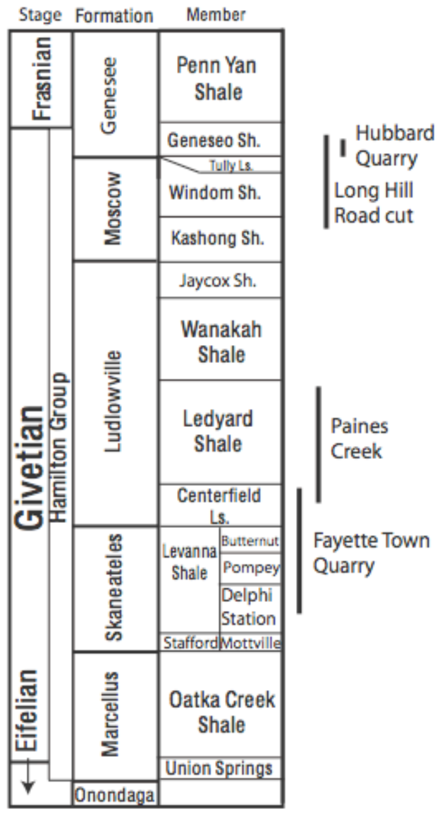
The Middle-Upper Devonian (Eifelian-Frasnian) sequence of New York State is an eastward thickening package of dominantly clastic rock that represents deposition in a shallow, tropical eperic sea (Rickard, 1975; Brett, 1986). This clastic wedge of eroded material resulting from the Acadian Orogeny is known as the Catskill Delta complex and was deposited into a northeast trending forearc basin that covered much of modern-day New York, Pennsylvania, and parts of Ohio (Ettensohn, 1985a, b, 1987; Kent, 1985; Woodrow and Isley, 1983; Woodrow and Sevon, 1985; Brett, 1986; Brett and Baird, 1996). The central, deepest part of the paleo-basin, likely 100-200 m in depth at maximum flooding (McCollum, 1987; Brett, et al., 1991), trends through the area of the modern-day Finger Lakes in west-central New York (Fig. 1.1). The strata in this area are interpreted to record several transgressive-regressive cycles, marked by the deposition of fine grained clastic units, punctuated by carbonate strata interpreted to be deposited during lowstand events (Kirchgasser, et al., 1988; Brett, et al., 1991). Shale intervals are interpreted to coincide with relative deepening events due to combined tectonic and eustatic influences within the Appalachian fordeep and are interbedded with silty mudstones, siltstones, and sandstone intervals interpreted to represent deposition under shallower water depths (Ettensohn, 1985a, b, 1987; Woodrow, et al., 1989; ver Straeten and Brett, 1995; Murphy, et al., 2000 a,b; Werne, et al., 2002; Sageman, et al., 2003).

During the Middle Devonian, central New York was under a broad shallow epeiric sea. The basin was commonly occupied by dysoxic to anoxic waters throughout much of the Devonian, resulting in the deposition of numerous black to gray shales that are punctuated by carbonate beds (see Brett et al, 1991).

Mudstones examined are listed in ascending order: the Levanna member of the Skaneatles Formation, the Ledyard member of the Ludlowville Formation, the Windom Shale of the Moscow Formtaion, and the Genesee member of the Geneseo Formation (Fig 1). These mudstones are lithologically very similar as grey to black mudstones with variable silt composition (Boyer and Droser, 2009). Clays within these units are dominantly illite with some clorite and typically <10% carbonate composition.

DYSAEROBIC FAUNA

The dysaerobic zone is defined from ecological patterns of infaunal and macrofaunal benthos observed in modern settings that can be largely generalized by decreasing diversity, abundance, and size of individuals correlated with decreasing bottom water oxygen levels (Rhodes and Morse, 1971; Thompson et al., 1985;.Savrda and Bottjer, 1986; Savrda, 1992; Wignall, 1994). Based on these patterns, a tripartite model recognizing relative oxygen levels as aerobic, dysaerobic, and anaerobic was developed and is widely used today (Rhoads and Morse, 1971; Byers, 1977).



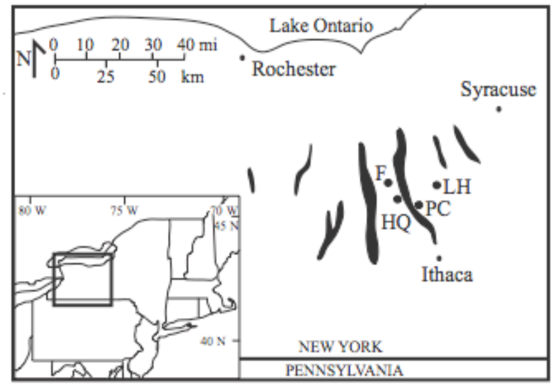


Figure 1. Stratigraphy of Middle and Upper Devonian aged field trip localities and locality map of stops in central New York. (F=Fayette Town Quarry, HQ= Hubbard Quarry, PC=Paines Creek, LH=Long Hill Road cut)

Studies in modern reduced oxygen settings reveal that particular clades, patterns of diversity, and life habits are associated with specific levels of bottom water oxygen with polychaetes, echinoderms, and mollusks generally as the taxa most tolerant to oxygen stress (Levin and Gage, 1998; Wu, 2002; Levin, 2003). Soft bodied organisms, preserved as trace fossils in the rock record, are typically the last organisms to be excluded as oxygen levels decrease to zero (Savrda, et al., 1984; Thompson, et al, 1985; Savrda, 1992). Species richness is strongly correlated with bottom water oxygen levels from a range of modern reduced oxygen settings (Diaz and Rosenberg, 1995; Levin and Gage, 1998; Wu, 2002; Levin, 2003). In modern settings, variability in dominant taxa is observed through the dysaerobic zone in association with changing levels of oxygen stress and it is unusual for one group to be dominant throughout the entire range of the dysaerobic zone (Allison, et al., 1995; Levin, 2003). Another robust ecological trend in modern settings is the persistence of deposit feeding as the dominant life habit under the most reduced oxygen levels (Wu, 2002). These established ecological patterns demonstrate the sensitivity of the biological signal to change in bottom water chemistry and allow for the recognition and interpretation of reduced oxygen conditions from the rock record (e.g. Savrda and Bottjer, 1986; Wignall, 1990; Savrda, 1992; Allison, et al., 1995; Sageman and Bina, 1997; Brett, 1998; Martin, 2004).

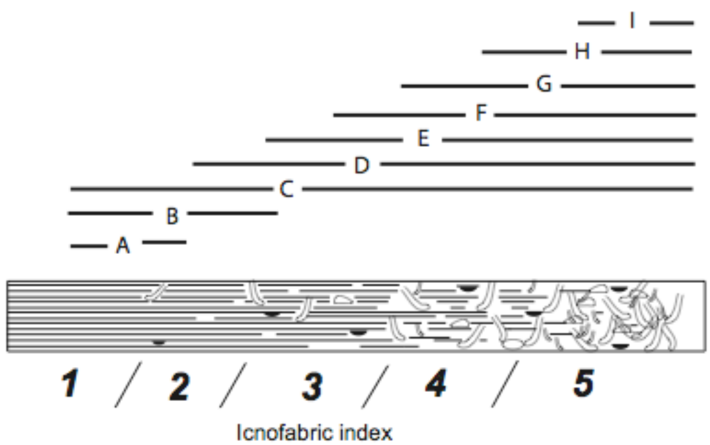
DEVONIAN DYSAEROBIC TAXA

Taxonomic Distribution

These black shale intervals are dominated by common dysaerobic taxa including species of Camarotoechioidea and Chonetoidea, *Pterochaenia*, *Lingula*, and *Orbiculoides* (Thompson and Newton, 1987; Brett, et al., 1991; Brower and Nye, 1991; Allison, et al., 1995, Boyer and Droser, 2007, 2009). At these localities, Camarotoechid brachiopods are by far the dominant group and include two common species, *Leiorhynchus quadracostata* and *Eumetabolotoechia multicostata*. They occur uncommonly in relatively dense aggregations on single bedding planes, sometimes fragmented and disarticulated indicating some reworking, but more commonly are dispersed as disarticulated, unfragmented molds. Chonetoidea brachiopods are also common at these localities. Several groups of linguliform brachiopods occur in these units including species of *Lingula* and *Orbiculoidea*. They are all relatively small, flat taxa typically less than 1 cm in length. Several other brachiopod groups are recognized at these localities with increasing diversity and abundance correlative with inferred relative bottom water oxygen levels.

At least 8 taxa of bivalves are recognized at these dysaeoribic sites, but *Pterochaenia fragilis* is the most common. Species of *Nuculites* and *Nuculoidea* are also commonly preserved but are not found as abundantly as *P. fragilis*.

Trilobites, including *Hollandclarkeia jennyae* and *Eldregeops rana*, are not common at these localities and gastropods occur rarely.



The distribution of benthic macrofauna at these localities has been recognized to vary in correlation with bottom water oxygen levels. Within dark shales at these localities two patterns are identified as taxa that are adapted to fully oxygenated conditions with variable, but predictable tolerance to oxygen stress, and taxa that are adapted to the lowest bottom water oxygen conditions. Figure 2 shows a schematic representation of taxonomic distribution through the dysaerobic zone with individual taxon given letter designations. This figure demonstrates patterns observed in these units that most of the taxa are fully adapted to fully oxygenated conditions, under decreasing oxygen levels, taxa with what are interpreted as successively higher oxygen thresholds are found to be excluded in a predictable order. As a result, specific combinations of taxa, rather than the occurrence of a specific taxon, are diagnostic of specific oxygen levels. This additive pattern has been recognized to persist across the paleo-basin through most of the classic Hamilton Group (Boyer and Droser, 2009). This trend of progressive loss of taxa with decreasing bottom water oxygen levels breaks down at the lowest end of the dysaerobic zone as numerous taxa are fully adapted to extremely reduced oxygen conditions. This is recognized in part by the common occurrence of monospecific assemblages (see below). *Eumetabolotoechia multicostata* has been recognized to have an extremely broad oxygen tolerance present in laminated sediments to high diversity assemblages. This is highly unusual compared to occurrences in modern marine settings.

Monospecific Assemblages

At these localities bedding planes that expose a single benthic species are common. Monospecific assemblages have been identified within these and other black shale units of the Hamilton group (Boyer and Droser, 2007). Several different taxa including *Leiorhynchus quadracostata*, *Eumetabolotoechia multicostata*, *Pterachaenia fragilis*, as well as *Lingula* and *Orbiculoides* species occur in monospecific assemblages. Interestingly, these monospecific assemblages are commonly found in association with laminated strata. The abundance of phylogenetically disparate groups that are found preserved in laminated sediments, interpreted to represent conditions at the edge of metazoan habitability, supports that numerous groups were equally well adapted to extreme oxygen stress. This is likely a result of repeated times of extremely reduced oxygen levels that are preserved as repeated black shale intervals through the middle and upper Devonian.

Table 1. Dysaerobic taxa identified at field trip localities. (LH=Long Hill Road Cut, PC= Paines Creek, F=Poormon Road/Fayette Town Quarry, HQ=Hubbard Quarry)				
	LH	**PC**	**F**	**HQ**
Brachiopoda				
Camarotoechids	X	X	X	X
Chonetids	X	X	X	X
Orbiculoids	X	X	X	X
Lingulids	X	X		X
Other	X		X	
Bivalvia				
Pterochaenia	X	X	X	
Nuculids	X	X		
other	X			
Trilobita				
Eldredgeops rana	X			
Hallandclarkeia jennyae	X	X		
Other				
Gastropod	X			

<u>Life Habit Distributions</u>

Within these units a range of life habits are recognized. The Rhynchonelliform brachiopods, the most abundant group in these units, are obligate epifaunal filter feeders, as well as *Orbiculoides* and the reclining Chonetes groups. Other common life habits observed at these localities include the infaunal filter feeding Lingulids, the shallow endobyssate *Pterochaenia*, the infaunal deposit feeders *Nuculoidea* and *Nuculites*, deep endobyssate bivalve groups, and epifaunal herbivore and/or deposit feeders including trilobites and rare gastropods.

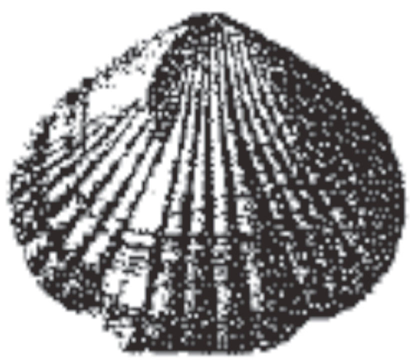
The abundance of epifaunal filter feeders at the lowest inferred oxygen levels is in contrast to modern low oxygen settings in which deposit feeding is the dominant life habit at the lowest bottom water oxygen levels (Wu, 2002). Those groups that are restricted to the lowest inferred bottom water oxygen levels are consistently of the filter feeding life habit. The infaunal deposit feeders, deeply embedded endobyssate, and mobile epifaunal groups (grazers and deposit feeders) are excluded from the lowest oxygen settings, and are found in increased abundance with increased bottom water oxygen levels. Ultimately a pattern of increased number of life habits with increasing oxygen in a predictable order is recognized and as a result life habit can be used independent of taxonomic association to identify relative variation in bottom water oxygen through the dysaerobic zone.

Black shale intervals at all of these localities preserve dysaerobic depositional conditions, but within and between these localities a range of relative oxygen levels can be recognized based on the distribution of taxa and associate life habits. Although these localities broadly represent a range of oxygen levels increasing through the trip (Hubbard Quarry extremely reduced, Paines Creek and Fayette Quarry moderately reduced, and Long Hill Road cut slightly reduced oxygen conditions), relative oxygen levels have been recognized to fluctuate on a sub-cm scale at these localities (Boyer and Droser, 2009, 2011).

BRACHIOPOD FAUNA OF THE DYSAEROBIC FACIES

Ambocoelia umbonata

Arcuaminetes scitulus



Eumetabolotoechia multicostata

Lingula spatulata

Leiorhynchus quadricostata

Orbiculoides newberryi

ROAD LOG FOR DYSAEROBIC BLACK SHALE LOCALITIES OF CENTRAL NY

CUMULATIVE MILEAGE	MILES FROM LAST POINT	ROUTE DESCRIPTION
0.0	0.0	Start at **HEROY LAB**, Syracuse University (Adjacent to Carrier Dome)
0.3	0.3	Head north on Crouse Dr. toward University Pl.
0.4	0.1	Take the 1st left onto University Pl.
0.9	0.5	Take the 1st right onto Irving Ave
1.3	0.4	Turn left onto E Genesee St
1.5	0.2	Turn right onto S Townsend St
1.7	0.2	Turn left onto Erie Blvd E
1.8	0.1	Erie Blvd E turns slightly right and becomes Oswego Blvd
1.9	0.1	Turn left onto James St
5.6	3.7	Continue onto NY-5 W/W Genesee St
5.8	0.2	Take the NY 5 W ramp to I690/NY695/Fairgrounds/Auburn
11.6	5.8	Merge onto NY-5 W
39.5	27.9	Slight right onto NY-5 W
63.2	23.7	Turn left onto NY-89 S

STOP 1: HUBBARD QUARRY will be on the left side of the road just before the second Lucas Vineyards sign; quarry sits below the road and is obscured by high grass.

This small abandoned quarry is one of the rare exposures of the Geneseo Shale, the lowest unit of the Genesee Formation, in this area. Over 3 meters of black shale are exposed. The Geneseo at this locality is very black with laminated intervals that are clearly anoxic. Silt laminae range from <1 to several mm in thickness and there are some intervals of thick silt ripples preserved at this locality. There are several concretionary layers that are exposed along the floor of the quarry that preserve carbonate-concretions up to 30 cm in diameter. At this locality bedding plane assemblages, most commonly monospecific, reveal dispersed to clumped concentrations of *Leiorhynchus quadracostata*, *Lingula spatulata*, *Orbiculoidea lodensis*, and rare *Chonetes sp*.. Individuals of *L. quadracostata* can be preserved inflated and even articulated at this locality and are typically 1.5-3 cm in length. Conversely, individuals of the other brachiopod species are all consistently small (<1 cm). The stop illustrates the low end of the dysaerobic zone, although evidence of bioturbation as individual burrows or breaks in lamination are common.

63.3	0.1	Head south on NY-89 S toward Interlaken Beach Rd
63.4	0.1	Take the 1st right onto Interlaken Beach Rd
63.5	0.1	Turn right onto County Rd 141
64.0	0.5	Take the 1st left onto County Rd 150
64.6	0.6	Continue onto Co Rd 150/Co Route 150
71.6	7.0	Turn right onto NY-96 N
74.1	2.5	Turn right onto NY-414 N/NY-96 N
74.3	0.2	Slight right onto NY-414 N/Ovid St
81.0	6.7	Continue straight onto NY-414 N/Ovid St
81.4	0.4	Turn left onto Poormon Rd

STOP 2: POORMON ROAD – FAYETTE TOWN QUARRY will be on the left

This stop exposes the uppermost Skaneateles Formation, the Levanna Shale, and the more resistant, fossiliferous Centerfield Limestone member of the Ludlowville Formation. We will concentrate on the depauperate fossil assemblages of the black shales as an example of intermediate dysaerobic conditions (moderately reduced relative oxygen levels). At this locality, *Eumetabolotoechia multicostata*, *Arcuaminetes scitulus, Devonchonetes coronatus*, *Orbiculoidea newberryi* and *Pterachaenia fragilis* are commonly found in low diversity and monospecific assemblages on bedding planes. Rare individuals of *Allanella tullia, Emanuella subumbona* and even *Conularia sp.* have been identified from this interval. Pelagic "*Styliolina fissurella*" are abundant on some bedding planes. At some horizons, there are clearly reworked lag deposits comprised mostly of fragmented remains of *E. multicostata*, but most other individuals are well preserved molds that are likely not transported.

The Levanna Shale at this locality preserves intervals of anoxia as barren laminated shales, as well as intervals of significantly higher relative oxygen levels as indicated by horizons with moderate diversity and abundance of benthic macrofauna. Detailed trace fossil analysis supports these interpretations.

81.8	0.4	Head east on Poormon Rd toward NY-414 N/Ovid St
88.5	6.7	Turn left onto NY-414 N/Ovid St
89.1	0.6	Continue onto Cayuga St
93.6	4.5	Continue onto NY-5 E/US-20 E/Auburn Rd
111.3	17.7	Turn right onto NY-90 S
111.7	0.4	Turn left onto Moonshine Rd. Park at the end of the road

STOP 3: PAINES CREEK will be down the hill

Paines Creek is the type section for the Ledyard Member of the Ludlowville Formation and can be accessed below the closed bridge. This black shale overlies the Centerfield Limestone Member that is exposed downstream as the cap of Moonshine Falls. At this locality, approximately 4 meters of black shales of the Ledyard Member are exposed in the creek bed. Fossils are dispersed and include *Eumetabolotoechia multicostata*, *Arcuaminetes scitulus, Devonchonetes coronatus*, *Orbiculoidea newberryi, Lingula spatulata*, *Pterachaenia fragilis, Nuculites oblongatus, Nuculoidea sp.* and *Hollandclarkeia jennyae*. Benthic macrofauna at this locality are found dispersed within sediments that range from laminated through almost completely bioturbated. Monospecific occurrences are present at this locality, but there are only rare densely packed associations.
This locality again records a range or bottom water oxygen conditions from anoxic through moderately oxygenated based on the variable diversity of benthic macrofaunal, which also corresponds with ichnological data. This locality was selected to reflect intermediate conditions of oxygen stress.

112.1	0.4	Head northwest toward NY-90 S
129.5	17.4	Turn left onto NY-90 S
133.1	3.6	Turn left onto NY-38 N/Main St
134.2	1.1	Turn left onto Aurora St/County Route 43C Continue to follow County Route 43C

STOP 4: LONG HILL ROAD CUT; parking will be on the right at a gravel turnoff. Locality is on the south side of the road.

This section exposes uppermost portion of the Moscow Formation, including a near complete succession of the Windom Shale and Tully Limestone, and the overlying Geneseo Shale on the south side of Long Hill Road. The black shales of the Geneseo are barren at this locality. The Windom Shale preserves several coarsening upward shale-siltstone cycles through an overall shallowing succession that is overlain by the thick Tully Limestone. Within the grey shales of the Windom are moderately diverse fossil assemblages, with some dense concentrations that are interpreted to be storm events. Intervals within this succession are interpreted to represent deposition under reduced oxygen conditions at the upper end (more oxygenated) of the dysaerobic zone based on body fossil diversity and trace fossil composition.

Common benthic macrofauna within these intervals include, the Rhynchonelliform brachiopods *Eumetabolotoechia multicostata*, *Arcuaminetes scitulus, Devonchonetes coronatus*, *Allanella tullia*, *Ambocoelia umbonata* the “inarticulate” brachiopods *Orbiculoidea newberryi, Lingula punctata*, bivalves *Pterachaenia fragilis, Nuculites oblongatus,* and *Nuculoidea sp.* and the trilobites *Eldregeops rana* and *Hollandclarkeia jennyae*.

This locality represents the upper end of the dysaerobic zone in which oxygen conditions are not at levels interpreted to be normal marine levels due to reduced diversity and ichnological features, but are higher when compared with other black shale intervals seen on this trip.

END OF TRIP.

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