Trip A-2

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

CHUCK VER STRAETEN
New York State Museum/Geological Survey, 3140 Cultural Education Center, Albany, NY 12230
GORDON BAIRD
Dept. of Geosciences, SUNY-Fredonia, Fredonia, NY 14063
CARLTON BRETT
Geology Department, University of Cincinnati, Cincinnati, OH 45221-0013
GARY LASH
Dept. of Geosciences, SUNY-Fredonia, Fredonia, NY 14063
JEFF OVER
Department of Geological Sciences, SUNY-Geneseo, Geneseo, NY 14454
CEREN KARACA
TERESA JORDAN
Department of Earth and Atmospheric Sciences, Snee Hall, Cornell University, Ithaca, NY 14853-1504
RANDY BLOOD
EQT Production, 625 Liberty Ave, Ste 1700, Pittsburgh, PA 15222

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.

****

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

<u>Geological Setting.</u> 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^{o} 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

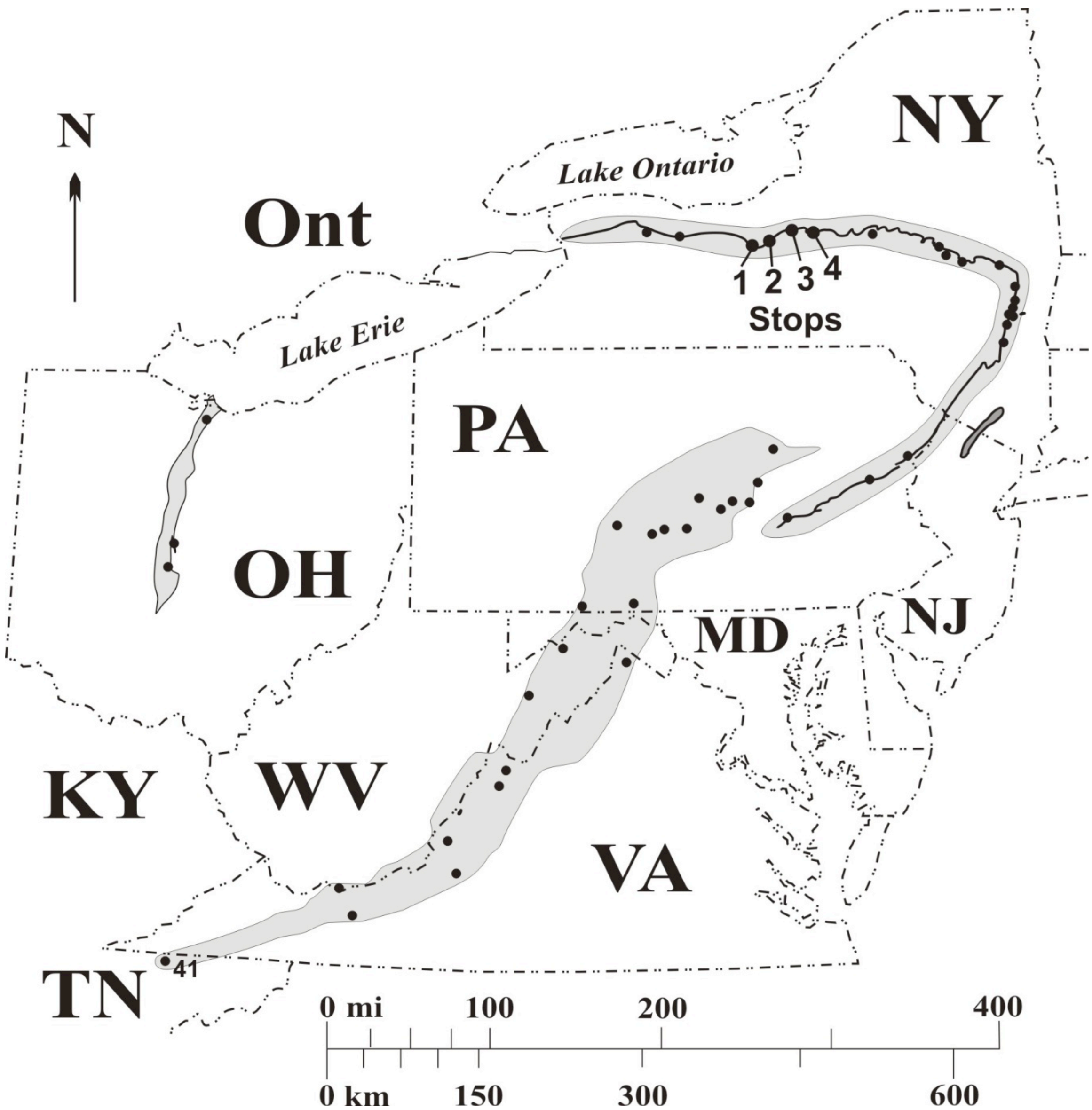


Figure 1-1. Outcrop map of the Marcellus subgroup of New York, and correlative strata in the Appalachian Basin, U.S. Marcellus rocks are found in the subsurface in between. Dots mark significant outcrops. In New York to eastern Pennsylvania, Marcellus outcrops occur in a narrow outcrop belt (dark line). 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).

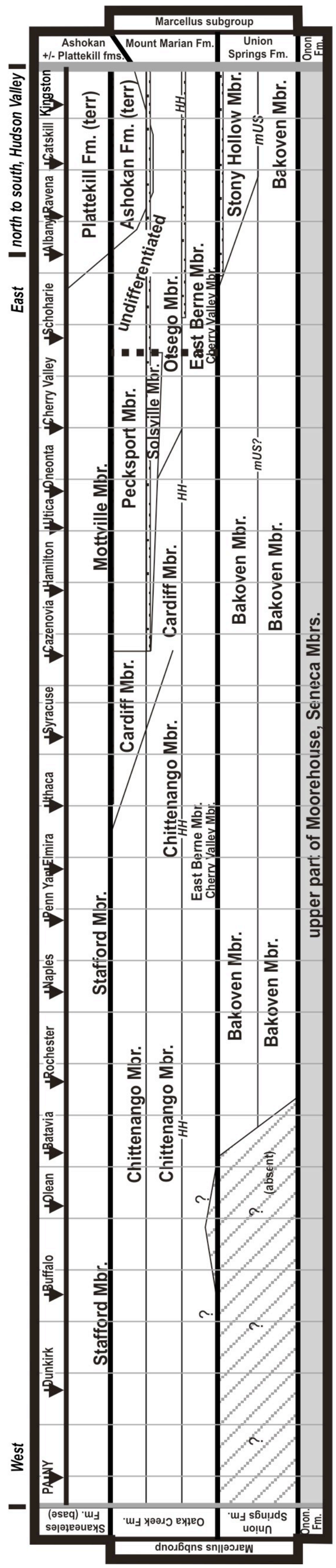


Figure 1-2

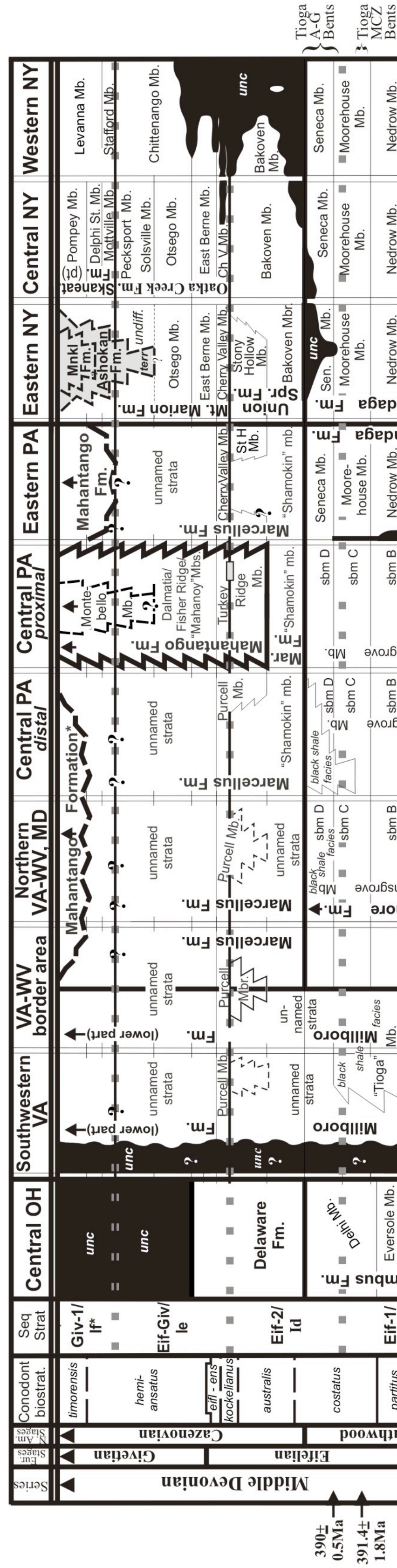


Figure 1-3

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.

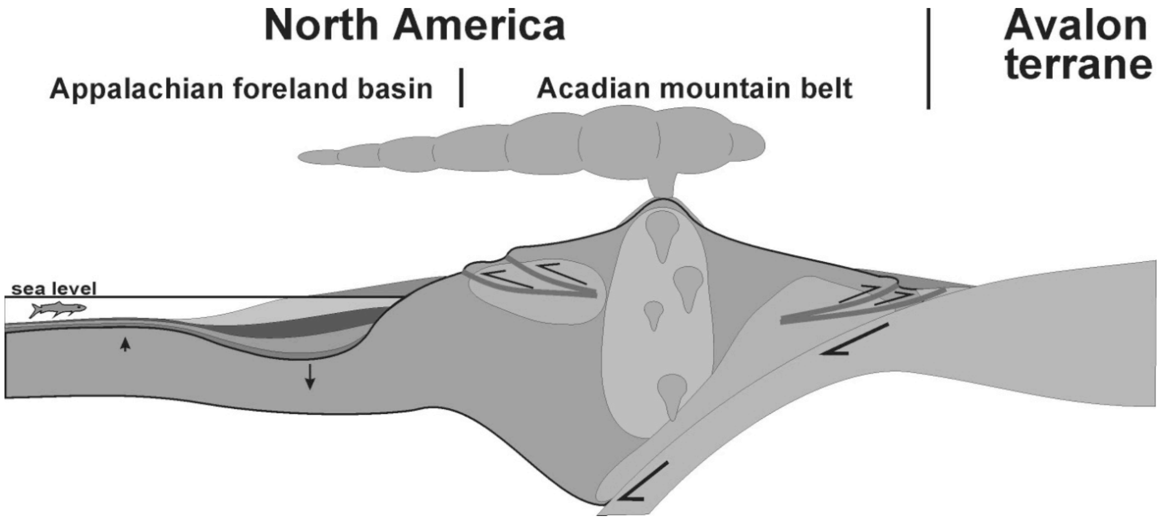


Figure 1-4. Idealized cross-section of Acadian orogen, and Acadian retroarc foreland basin system across central New England and New York into Ontario.

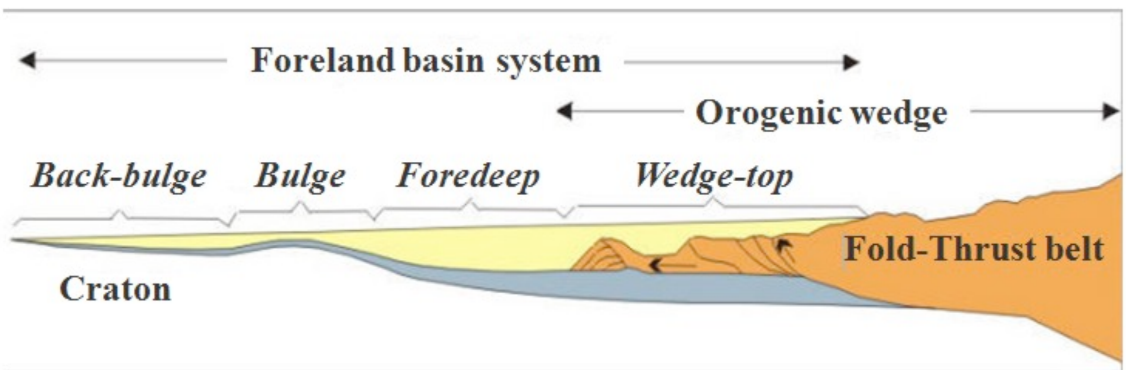


Figure 1-5. Idealized cross-section of an orogenic belt and adjacent foreland basin system. Modified after DeCelles and Giles (1996).

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.

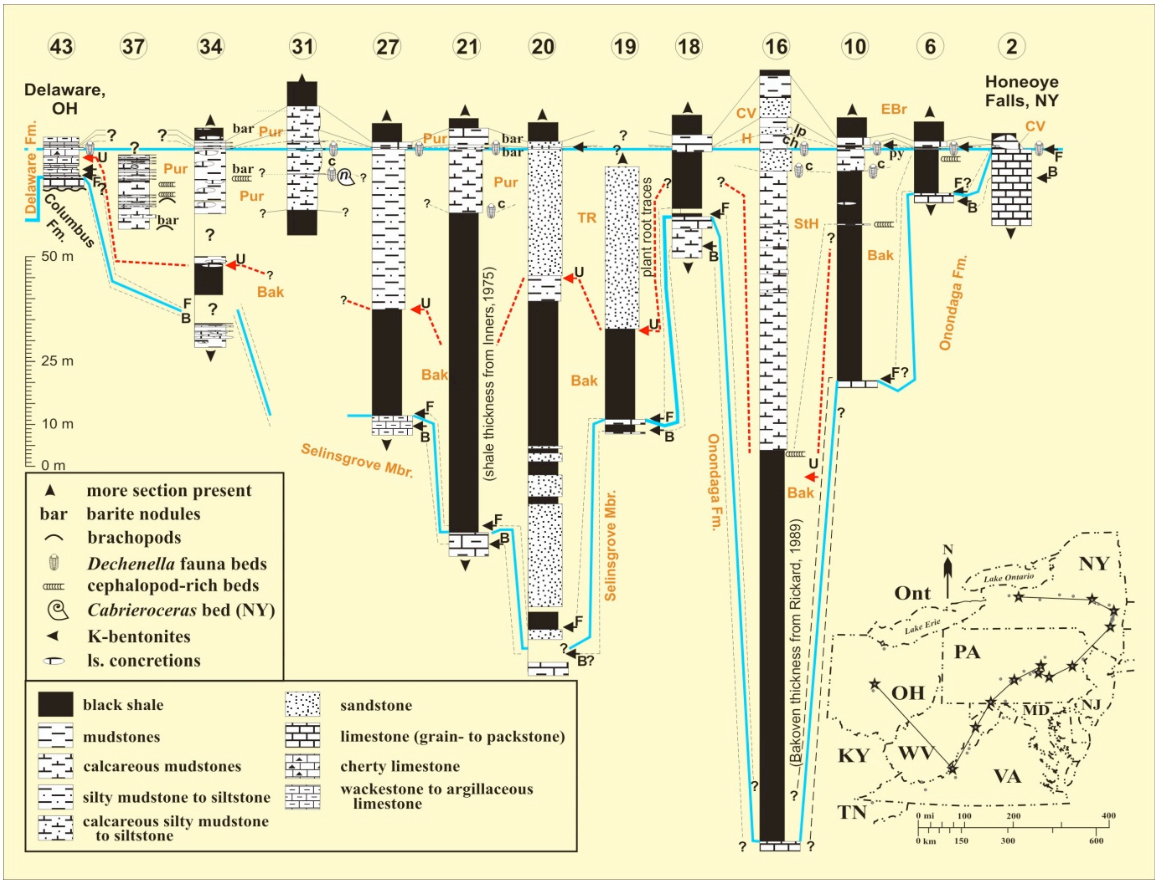


Figure 1-6. Basinwide correlations of the Marcellus subgroup and correlative strata, Appalachian Basin, New York to Virginia and Ohio. Dashed and solid lines show correlations. Bold solid lines separate New York formations and correlative horizons across basin. Non-bold solid lines separate member-level 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.

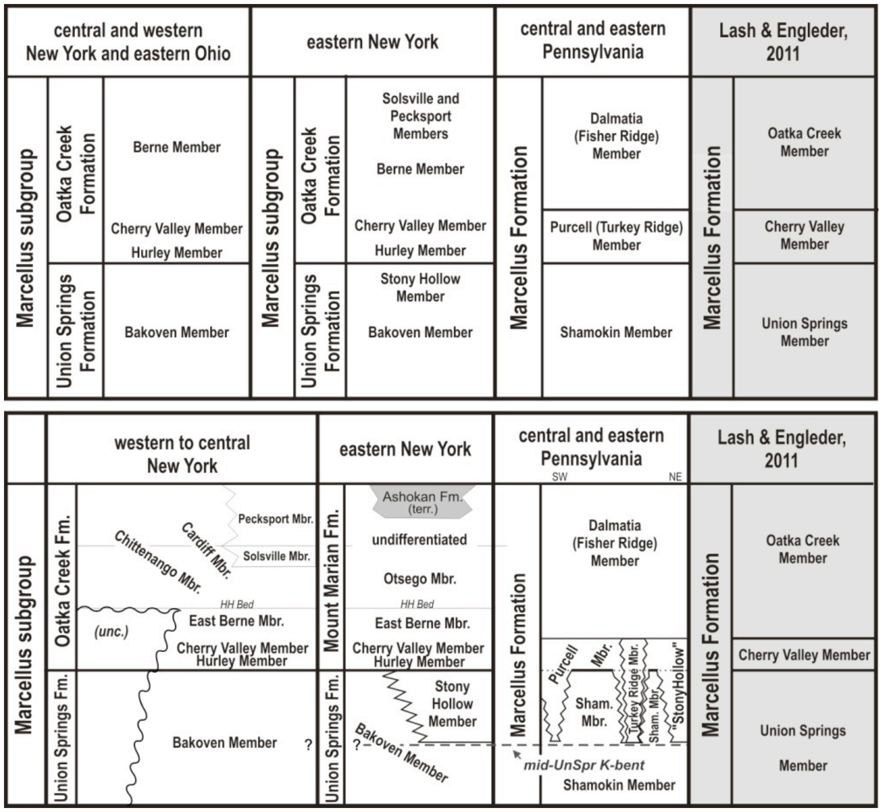


Figure 1-7. a: Comparison of Marcellus outcrop stratigraphy in New York and Pennsylvania as portrayed by Lash and Engelder (2011); b) Marcellus stratigraphy as outlined by Ver Straeten (2007), and for post-East Berne strata above. Note complex time-correlative relationships of strata, especially in central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

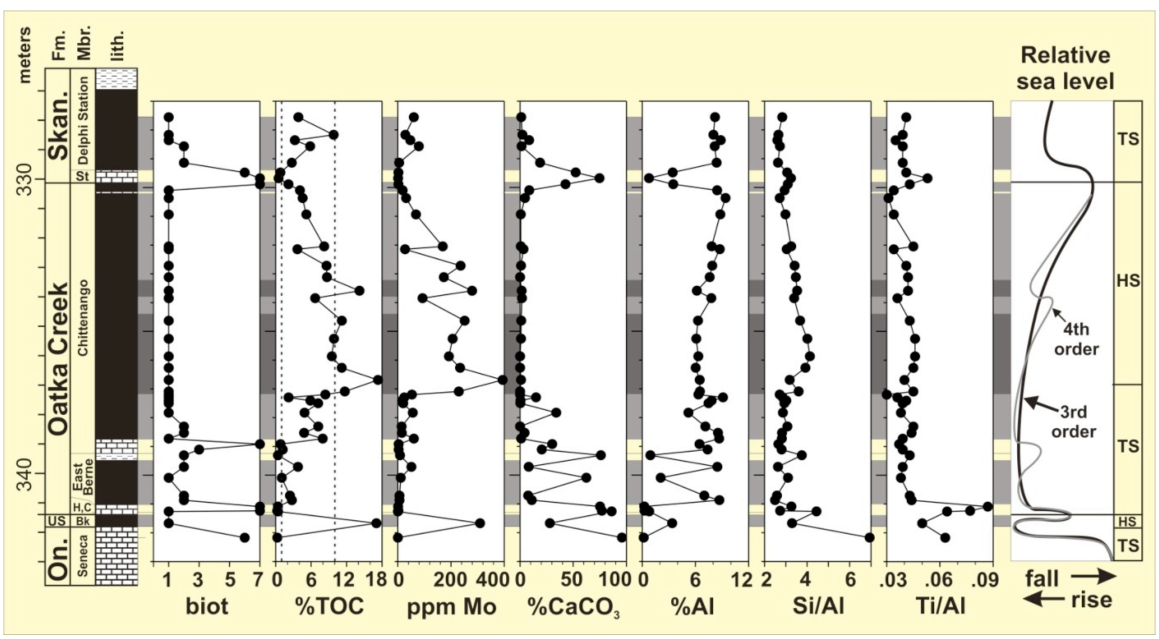


Figure 1-8. Development of Sequence Ie/Eif-Giv in black shale-dominated facies, as outlined by multi-proxy data analyses. Sedimentologic, paleobiologic and geochemical data of the Oatka Creek Formation from a subsurface core, Genesee Valley, western New York. %Al functions as a proxy for clays; Si/Al 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.

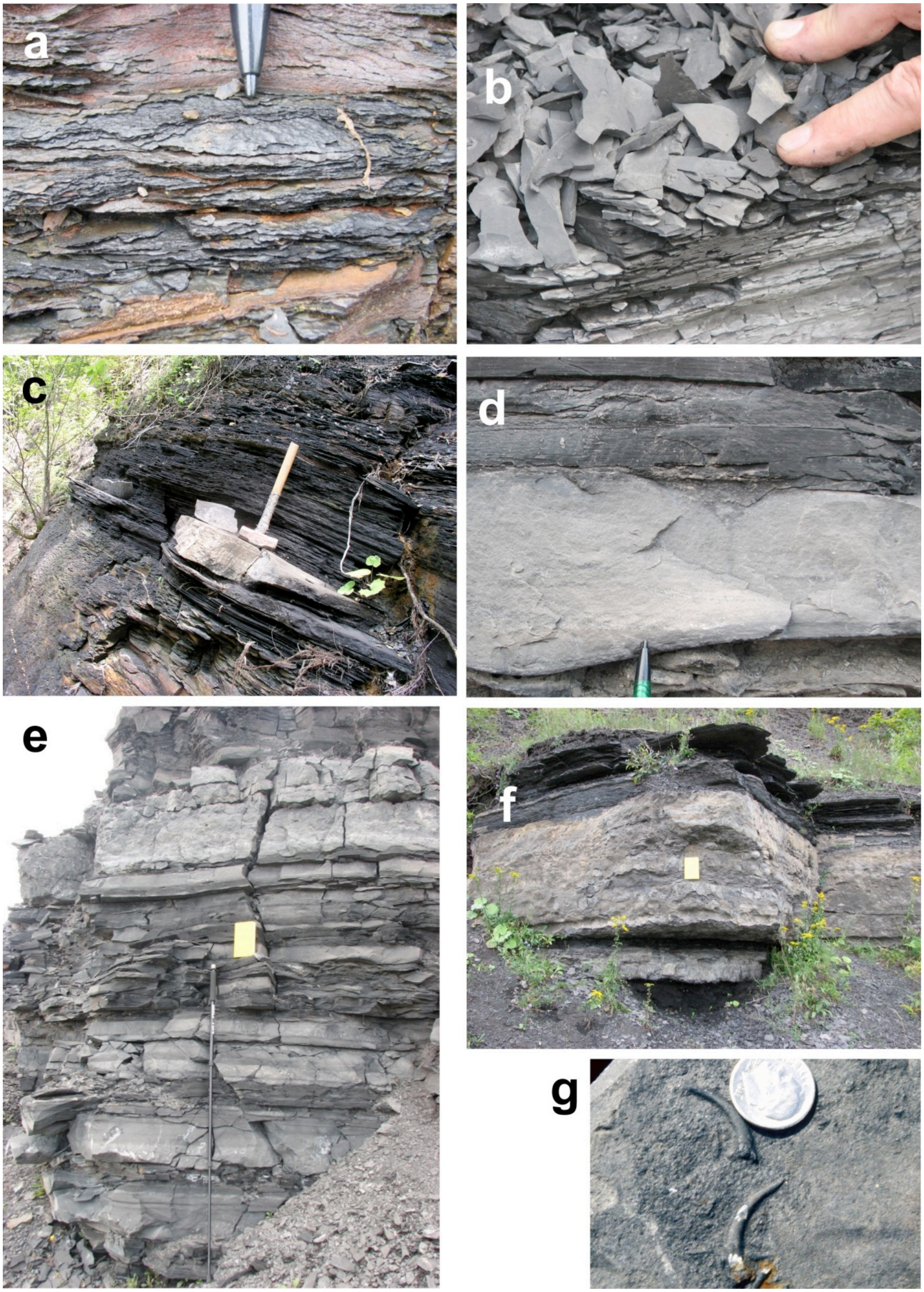

a
b
c
d
e
f
g

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 (5^{th} to 6^{th} 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

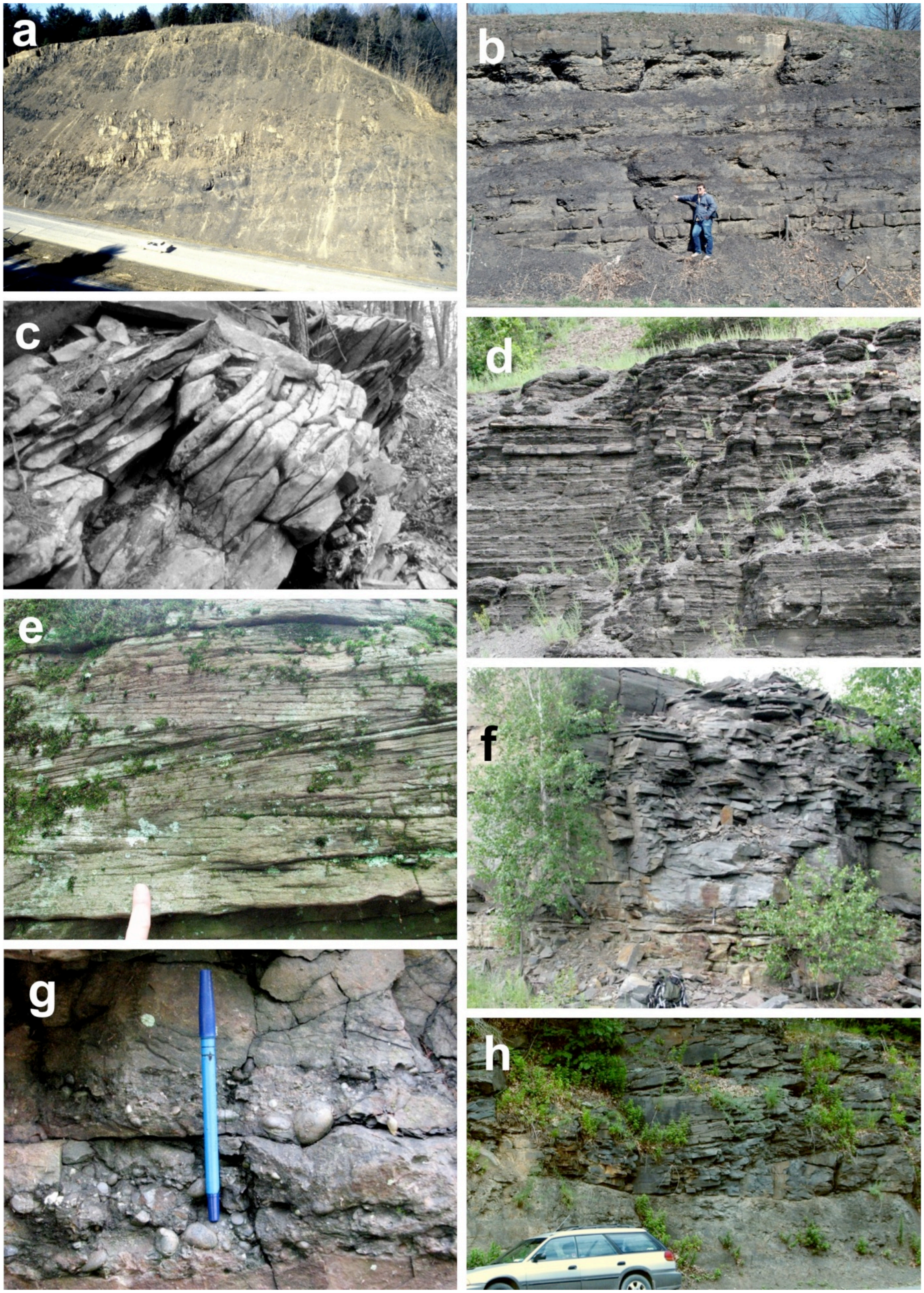

a
b
c
d
e
f
g
h

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).

---

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.

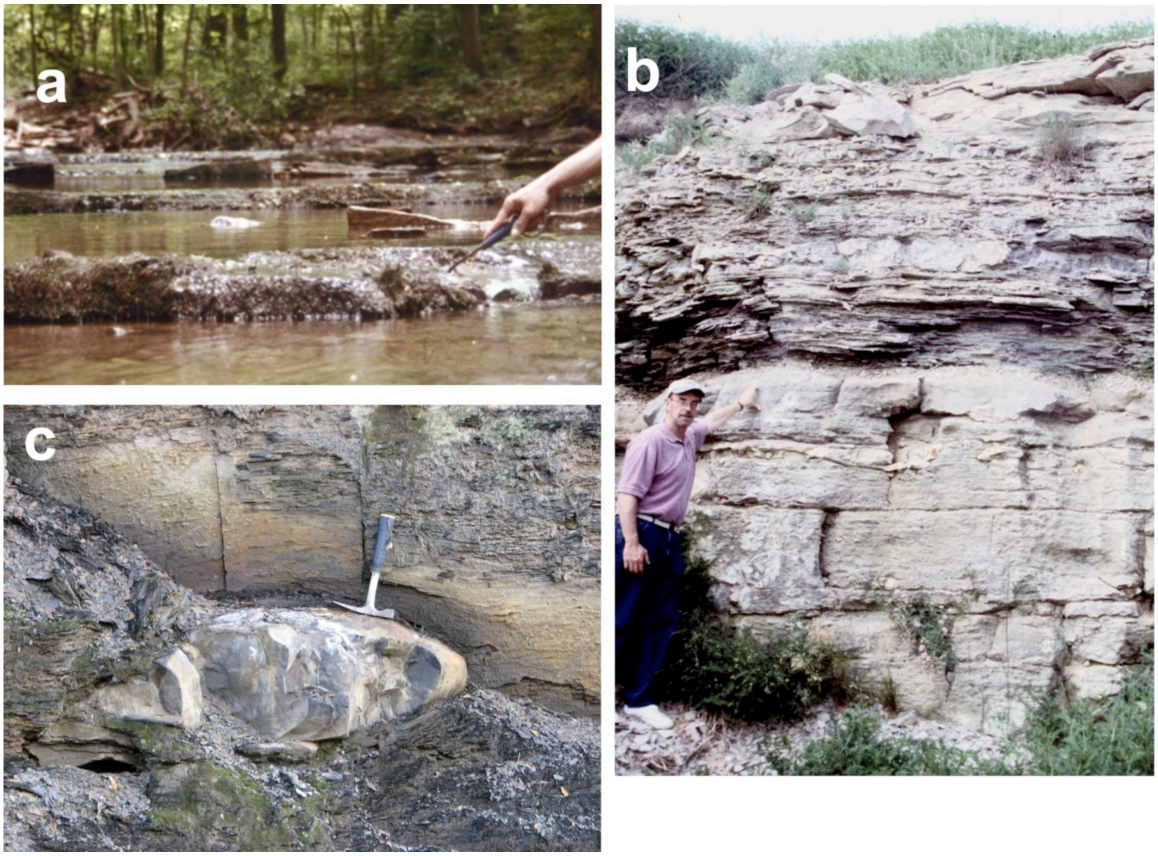


Figure 1-11. Photos of the Marcellus *carbonate-dominated facies association*. a) Hurley and Cherry Valley mbrs.-equivalent strata in the Delaware Limestone (Delaware, OH); b) shallow to deep ramp limestones and argillaceous limestones. The latter, in the center of the photo, represent the maximum transgression 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.

****
****

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

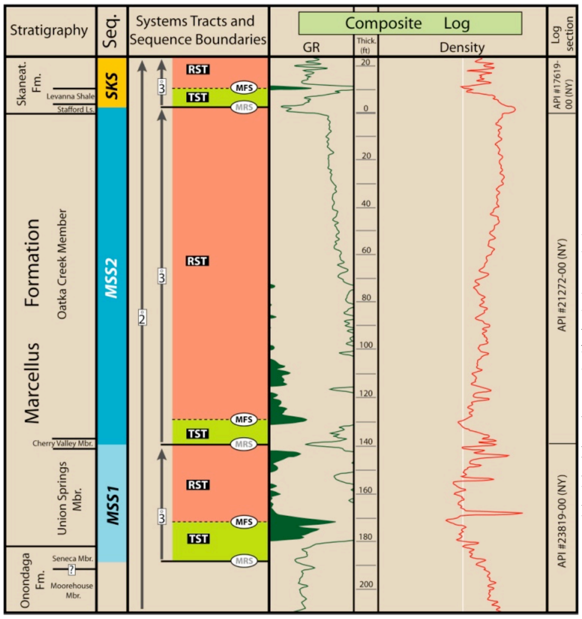


Figure 2-1: Sequence stratigraphic "type section" of the Marcellus Formation that encompasses the upper part of the underlying Onondaga Formation and the lower interval of the Skaneateles Formation; TST = transgressive systems tract; RST = regressive systems tract; MFS = maximum flooding surface; MRS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

<u>Gas-in-Place and Organic Carbon</u>

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

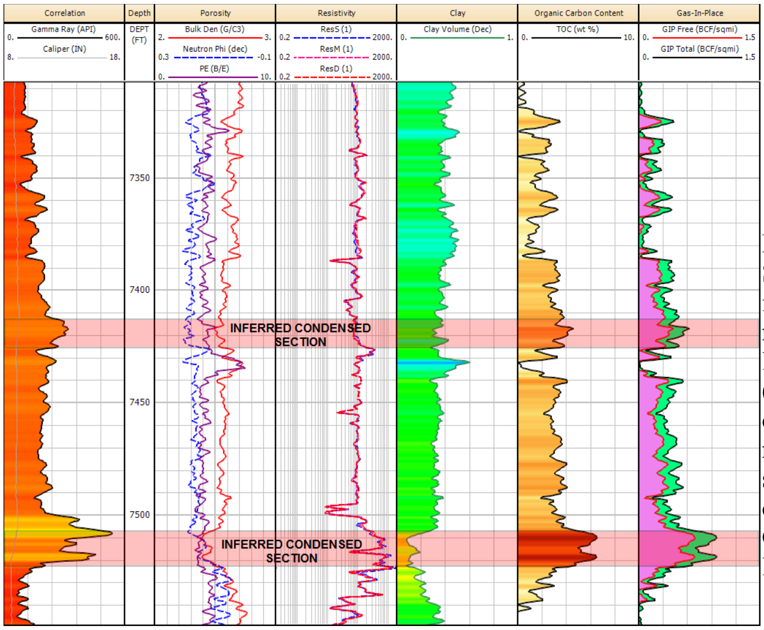


Figure 2-2: Marcellus Shale log suite, including bulk density, neutron, and resistivity logs. Clay volume (third from right) is calculated from the neutron-density log separation. Core-calibrated, log-derived GIP is illustrated in the log on the right.

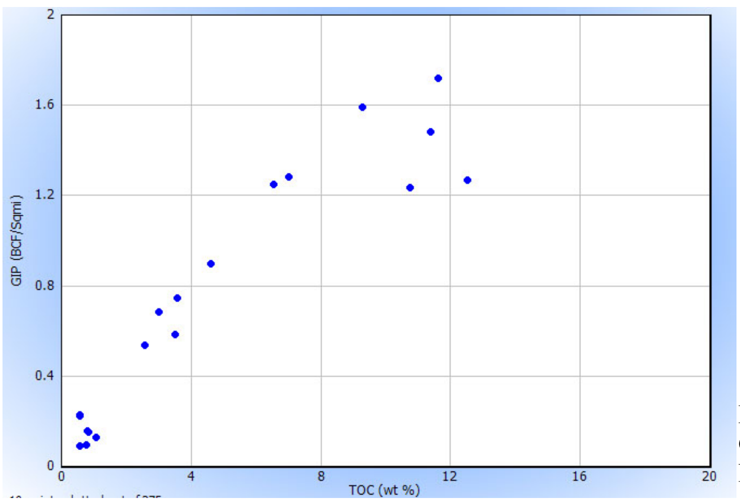


Figure 2-3: TOC versus GIP; data from a Marcellus well.

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).

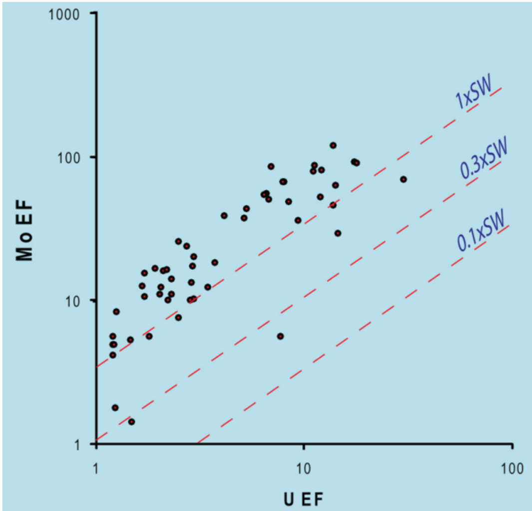


Figure 2-4: Mo_{EF} versus U_{EF} for the Marcellus Formation based on XRF data obtained from a Marcellus core. EF (enrichment factor) enables one to compare sample elemental abundances normalized to aluminum relative to elemental abundances of an "average" shale. Dashed red lines show Mo/U molar ratios 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).

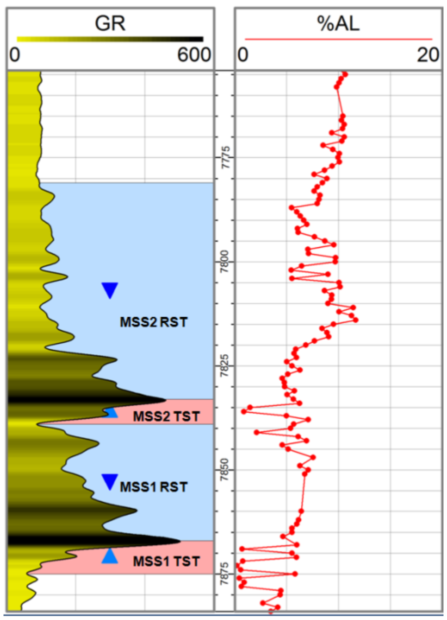


Figure 2-5: Gamma-ray (left) and %aluminum (right) logs through the Marcellus Formation (including the T-R sequences). Note the (1) general increase in Al through the unit and (2) markedly reduced Al in TST deposits.

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.

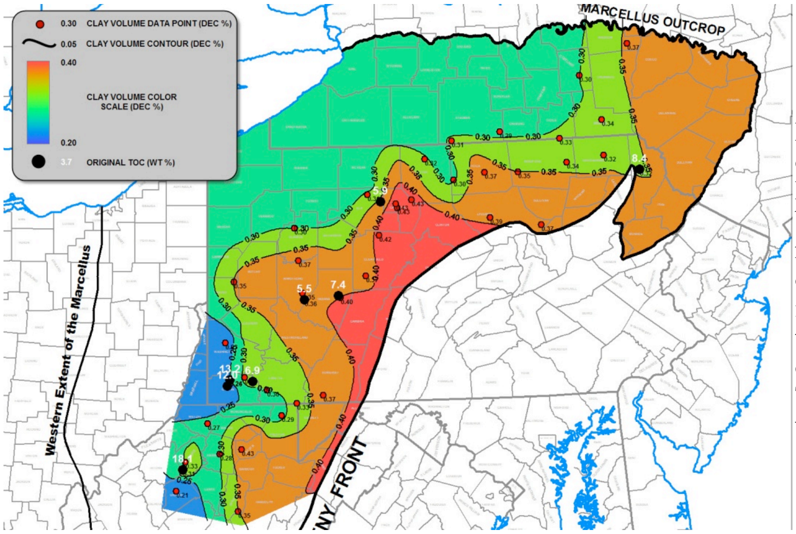


Figure 2-6: Percent clay map for the Marcellus Shale based on use of a neutron-density double clay indicator calibrated to resistivity. Note calculated original TOC values in white.

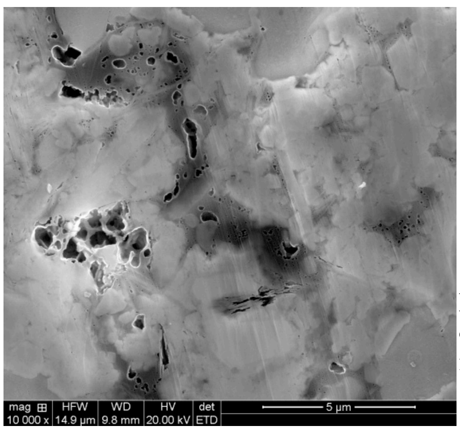


Figure 2-7: Intraparticle porosity development in organic grains (dark patches). Sample is argon ion milled.

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).

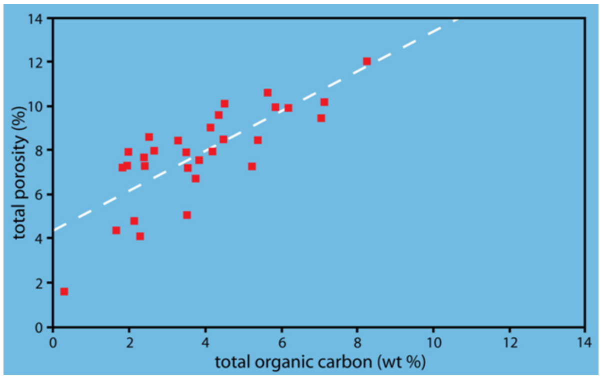


Figure 2-8: TOC – porosity cross-plot, Marcellus Formation; %Ro >1.4%.

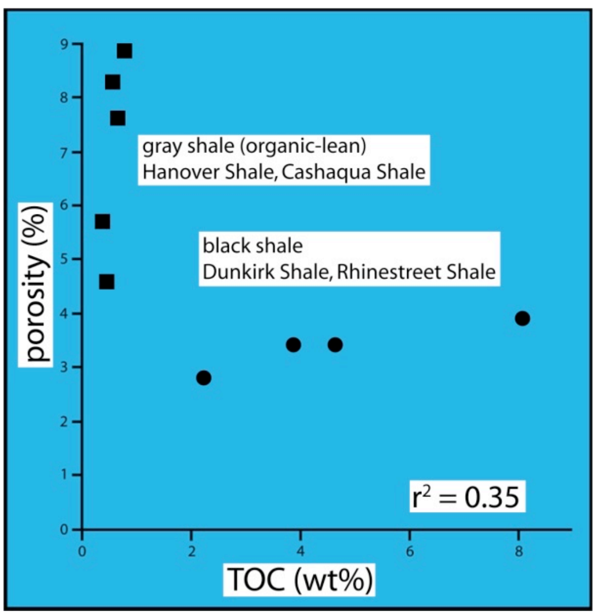


Figure 2-9: TOC – porosity cross-plot, Upper Devonian outcrop samples, western New York; %Ro = 0.6-0.74%.

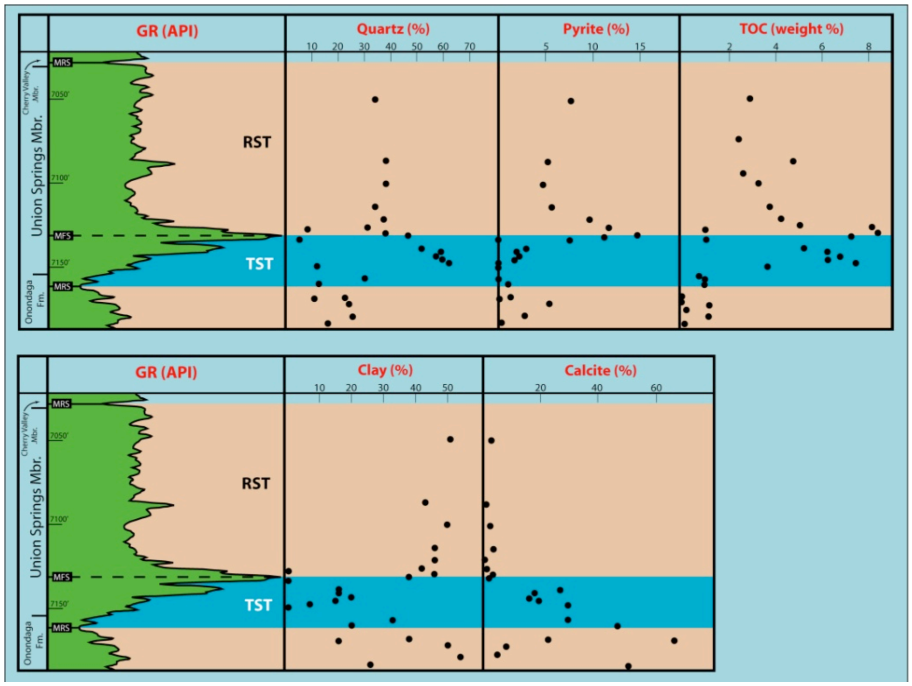


Figure 2-10: Mineralogy and TOC trends through the MSS1 depositional sequence. Data based on X-ray diffraction analysis of a suite of sidewall core samples recovered from a Marcellus well.

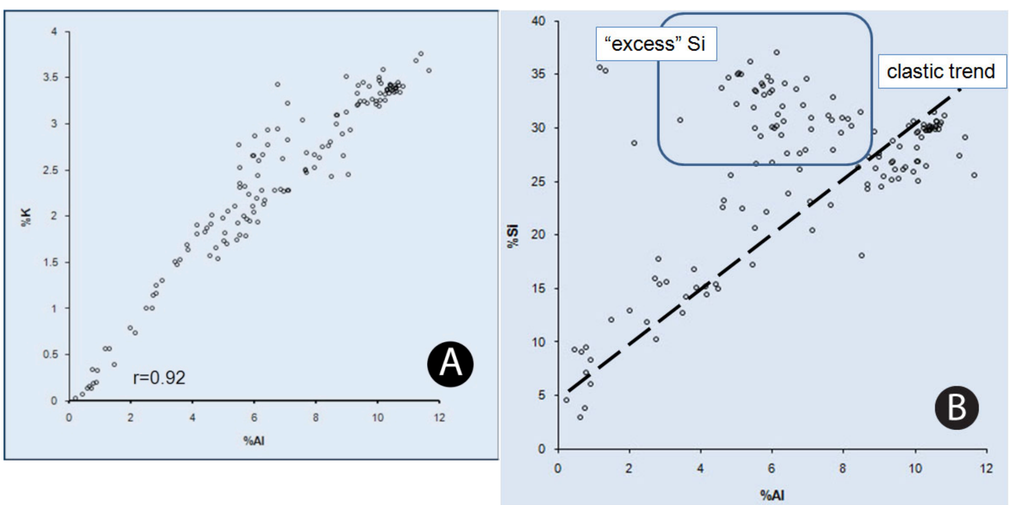


Figure 2-11: %K-%Al (A) and %Si-%Al (B) cross-plots; data obtained by handheld XRF analysis of a Marcellus Formation core.

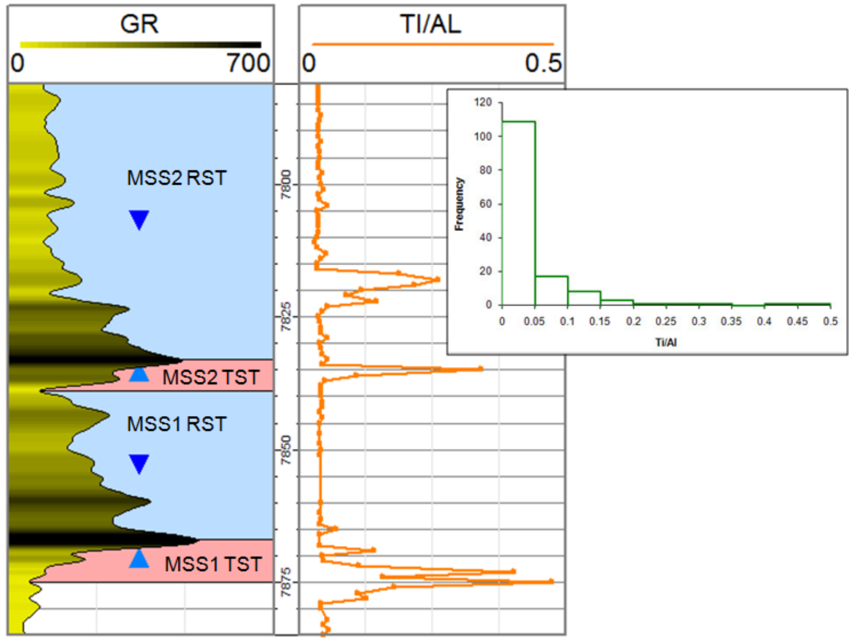


Figure 2-12: Gamma-ray log and corresponding plot of Ti/Al for a Marcellus well and core; inset shows a frequency plot of Ti/Al values (n=142). Note the great number of Ti/Al values less than or equal to 0.05. Note also the Ti/Al excursion in MSS2 RST between 7,816 and 7,822 feet.

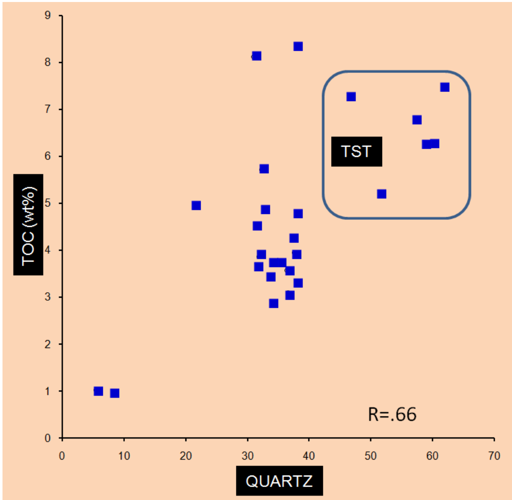


Figure 2-13: TOC – quartz cross-plot; data from a suite Marcellus core samples. The box labeled "TST" denotes those samples recovered from transgressive systems tract deposits

The distribution of clay and its bearing on mechanical rock properties

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.

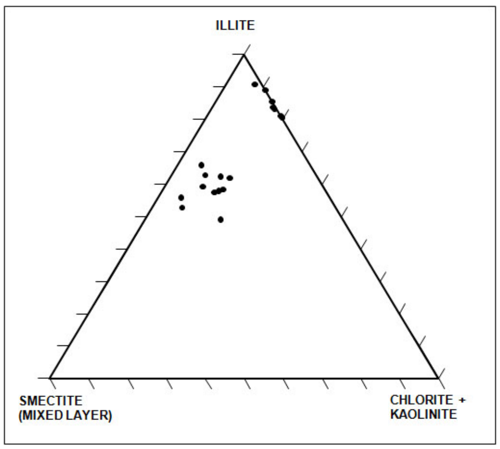


Figure 2-14: Clay mineralogy plot for data from a Marcellus Formation well, northeastern Pennsylvania.

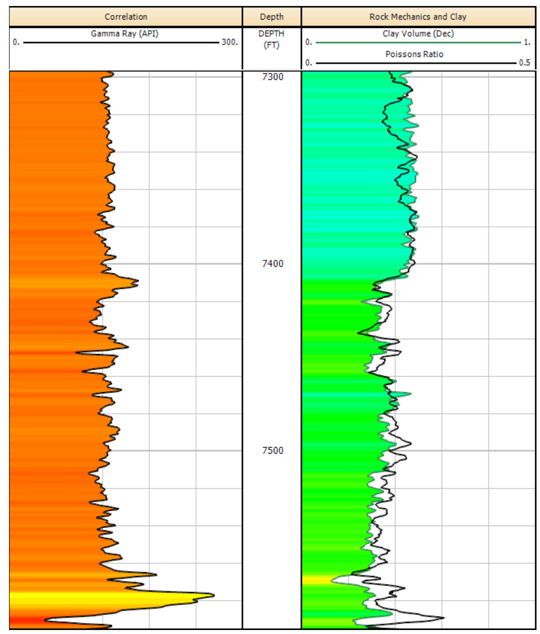


Figure 2-15: Gamma-ray log (left) and combined rock mechanics (Poisson's Ratio) and clay volume logs (right).

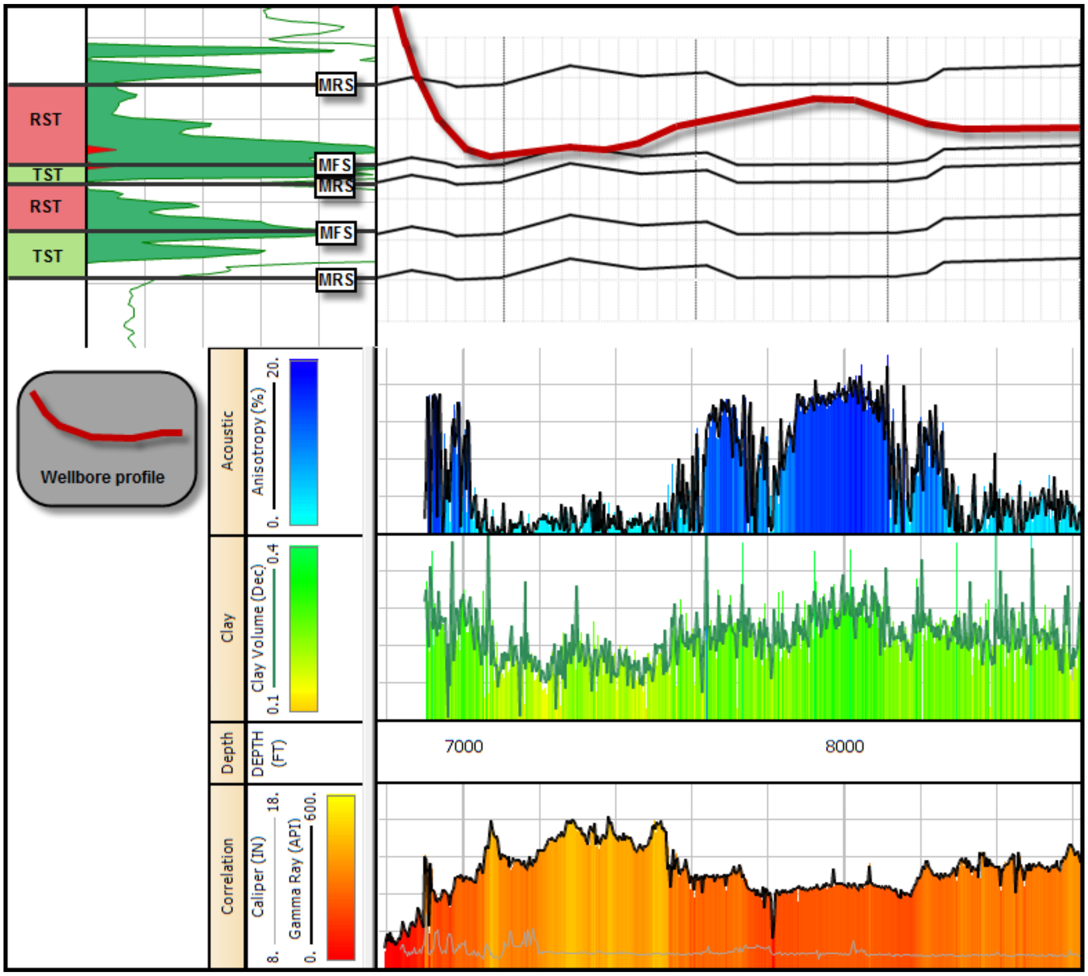


Figure 2-16: Correlation of acoustic shear wave anisotropy (top horizontal curve) and clay volume (middle curve) calculated from a suite of geophysical well logs run along a horizontal wellbore. Note the increase in clay volume and shear wave anisotropy as the lateral drifted upward away from the MSS2 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.

****

****

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

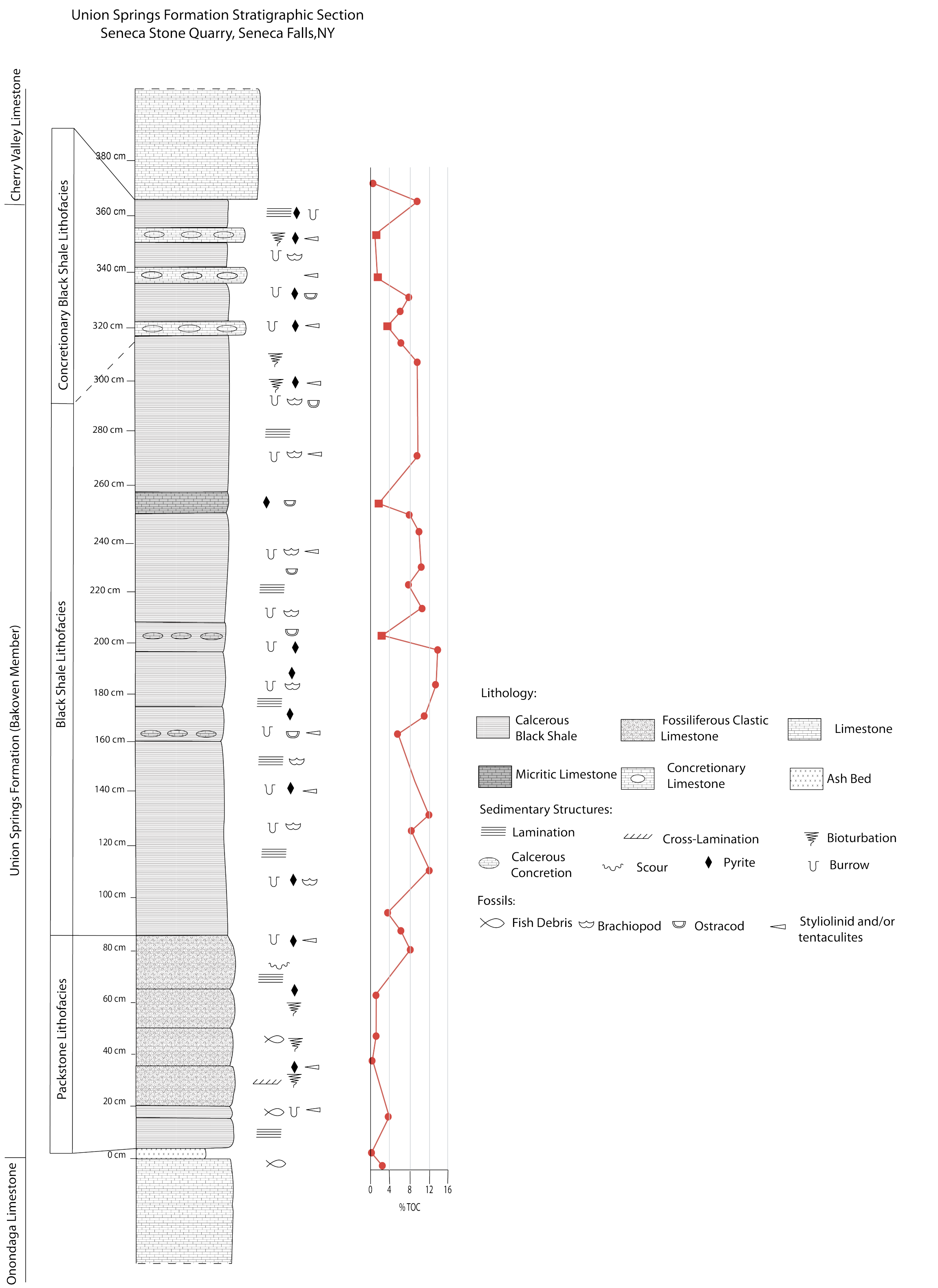


Figure 3-1: Stratigraphic section of Union Springs Formation with associated TOC values. Filled circles represent samples from shale whereas filled boxes represent samples from concretions.

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%.

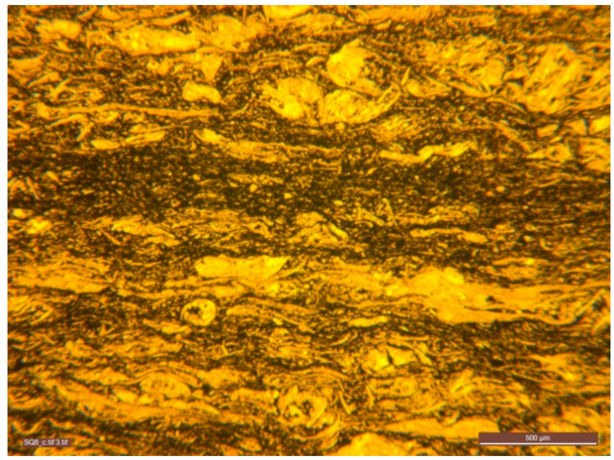


Figure 3-2: Transmitted photomicrograph of stylolaminated packstone. Dark layers are pressure solutions that are formed due to loading or tectonism. Scale bar is 500 µm.

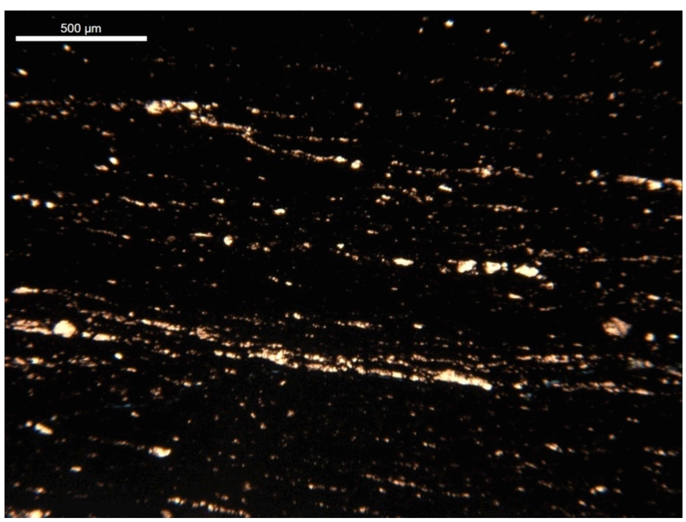
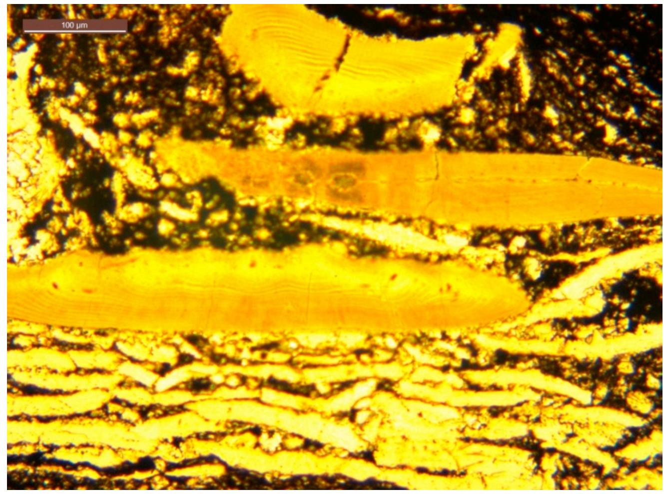


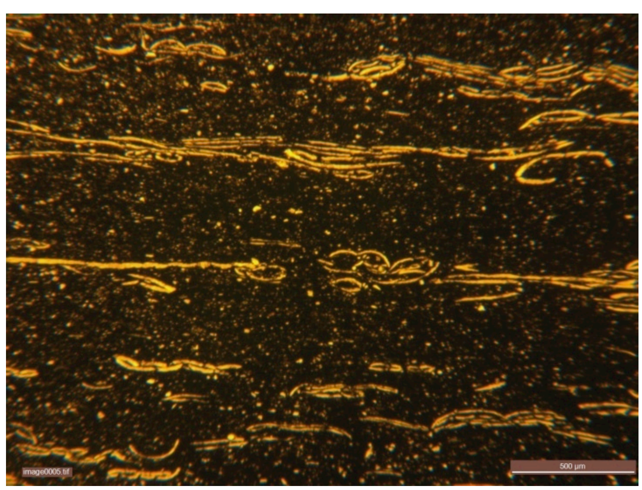
Figure 3-3: Transmitted photomicrograph of organic rich layer showing cross laminations. The sample is from the lowest level of the Bakoven member. The uppermost lamination, which parallels the top edge of the thin section, is the horizontal reference. Scale bar is 500 µm.

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.



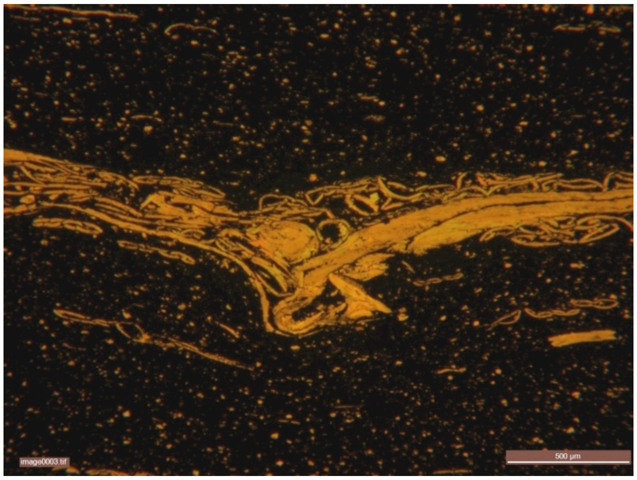


Figure 3-6: Transmitted light photomicrograph showing the load structure under the shell load. Scale bar is 500 µm.

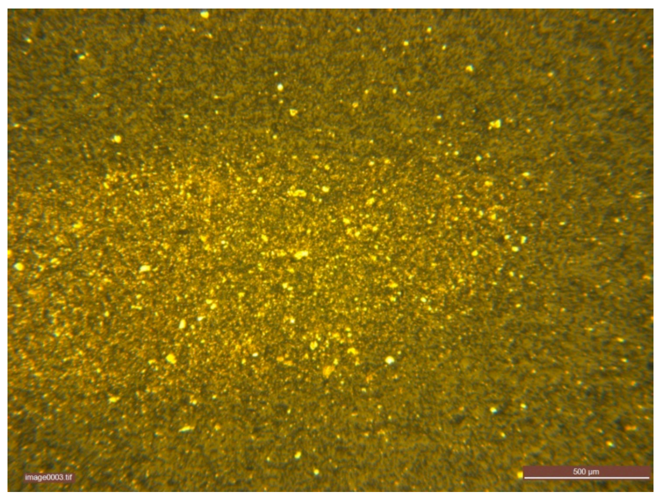


Figure 3-7: Reflected light photomicrograph showing a pyritized burrow. Probably the organic material in the burrow is replaced by pyrite. Scale bar is 500 µm.

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.

****
****

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.

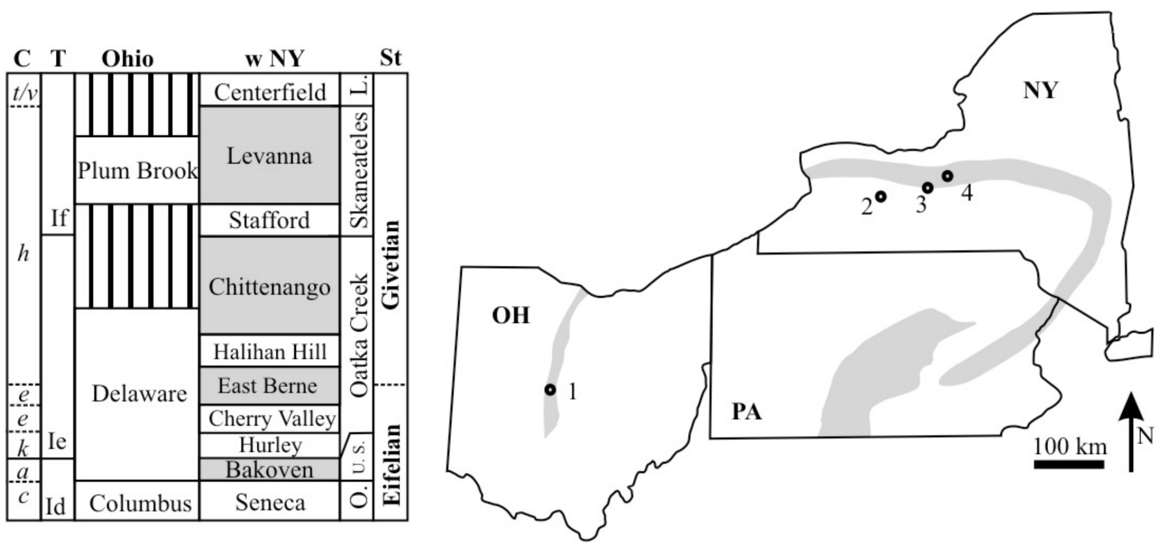


Figure 4-1 – A. General stratigraphy, conodont zonation, and major T-R cycles of the Marcellus subgroup (Unions Springs and Oatka Creek formations) and equivalents. O = Onondaga Formation; US = Union Springs; L = Ludlowville Formation; *c = costatus; a = australis; k = kockelianus; e = eiflius; e = ensensis; 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).

<u>Marcellus subgroup Depositional Packages in Western New York</u>

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.

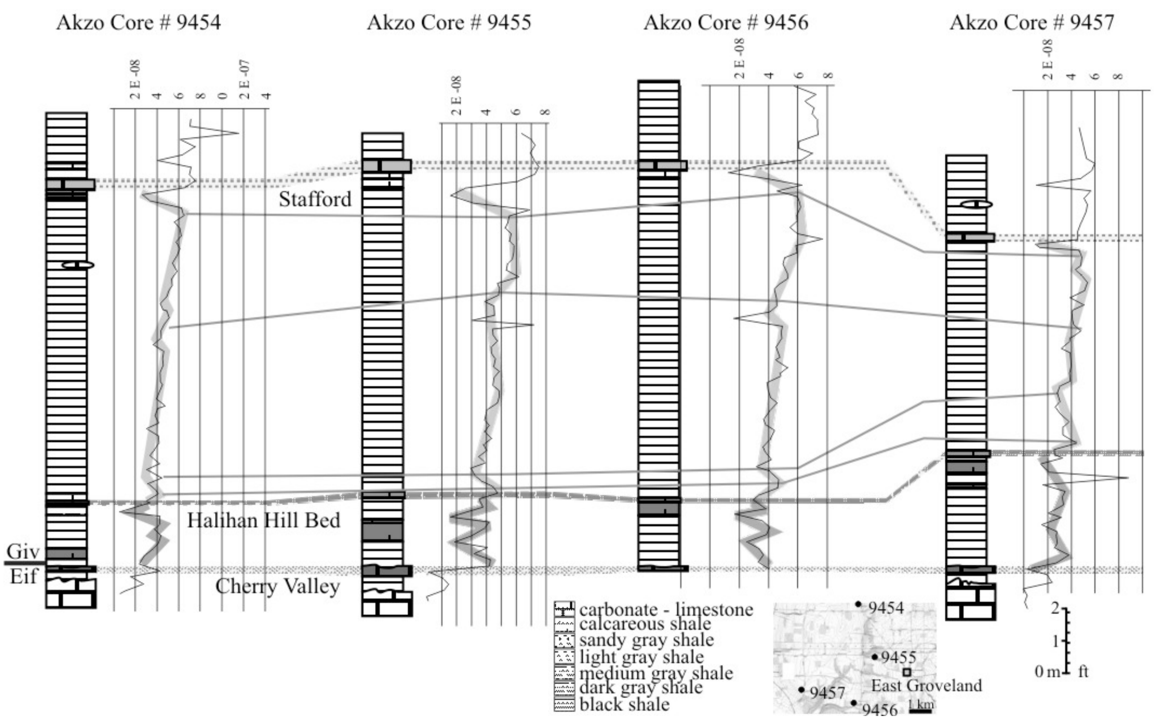


Figure 4-2 – Lithology and magnetic-susceptibility (M3/kg) of four drill cores in the Genesee River Valley near East Groveland. Key beds are correlated by patterned intervals – datum = top of the Cherry Valley Member. MS values are shaded by splined trends which are correlated by light gray lines.

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).

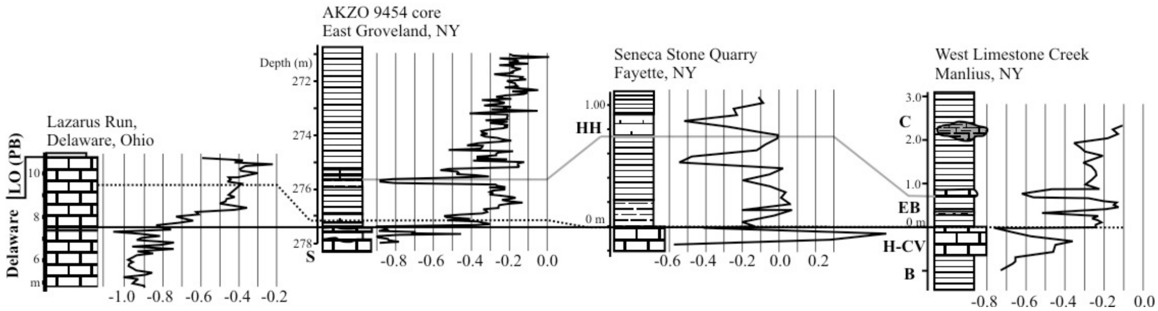


Figure 4-3 – Lithology and magnetic susceptibility (δMS) of Marcellus subgroup strata and equivalents in Ohio, western New York, and central New York – note different scales of measured sections. Datum = top of the Cherry Valley Limestone and equivalent; the Eifelian-Givetian Boundary is marked by a 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.

Chapter 4 Acknowledgements

Thanks to the Donors to the Petroleum Research Fund of the American Chemical Society and the Geneseo Foundation for financial support to DJ Over and Genseo students, and the American Rock Salt Company for access to core material.

****
****
****
****

ACKNOWLEDGEMENTS (ENTIRE PAPER AND FIELD TRIP)

All of the leaders and co-authors of this field trip would like to thank Jed Day and Jon Inners for their reviews of this paper, the management and employees of Seneca Stone Corporation and Hanson Aggregates-Jamesville, and the Lockwood family for permission to visit their properties during this fieldtrip. We would also like to thank the organizers of NYSGA meeting this year: Syracuse University Department of Earth Sciences and the Central New York Association of Professional Geologists, especially Gerry Gould and Jeff Karson for allowing us to do this trip, and for their patience.

REFERENCES

Algeo, T.J., and Tribovillard, N., 2009, Environmental analysis of paleooceanographic systems based on molybdenum-uranium covariation. Chemical Geology, v. 268, p. 211-225.

Arthur, M. A., and Sageman, B. B., 1994, Marine black shales: A review of depositional mechanisms and significance of ancient deposits. Annual Review Earth and Planetary Science, v. 22, p. 499-551.

Arthur, M.A., and Sageman, B.B., 2005, Sea Level Control on Source Rock Development: Perspectives from the Holocene Black Sea, the mid-Cretaceous Western Interior Basin of North America, and the Late Devonian Appalachian Basin. *In* Harris, N.B., and Pradier, B., eds., The Deposition of Organic Carbon-rich Sediments: Models, Mechanisms and Consequences, SEPM Special Publication No. 82, p. 35-59.

Baird, G.C., Brett, C.E., and Ver Straeten, C.A., 1999, The first great Devonian flooding episodes in Western New York: Reexamination of Union Springs, Oatka Creek, and Skaneateles Formation successions (upper Eifelian to lower Givetian) in the Buffalo-Seneca Lake region. *In* Lash, G.G., ed., New York State Geological Association, 71st Annual Meeting Guidebook, p. A1-A44.

Baird, G.C., Brett, C.E., and Ver Straeten, C.A., 2000, Facies and fossils of the lower Hamilton Group (Middle Devonian) in the Livingston County-Onondaga County Region. New York State Geological Association, 72nd Annual Meeting, Field Trip Guidebook.

Bartholomew, A.J., Schramm, T., Ver Straeten, C., Over, D.J., 2009, Refining the timing of faunal turnover in the Middle Devonian Appalachian Basin: paleoecological analysis of the earliest Hamilton fauna and a revision of the base of the Givetian Stage in eastern North America. New York State Geological Association 81st Annual Meeting, Field Trip Guidebook, New Paltz, NY, p. 10.1–10.20.

Beaumont, C., 1981, Foreland basins. Geophysical Journal of the Royal Astronomical Society, v. 65, p. 291-329.

Bertrand, P., et al., 1996, The glacial ocean productivity hypothesis: the importance of regional temporal and spatial studies. Marine Geology, v. 130, p. 1-9.

Bhattacharya, J.P, and McEachern, J.A., 2009, Hyperpycnal rivers and prodeltaic shelves in the Cretaceous seaway of North America. Journal of Sedimentary Research, v. 79, p. 184-209.

Biscaye, P.E., and Eittreim, S.L., 1977, Suspended particulate loads and transports in the nepheloid layer of the abyssal Atlantic Ocean. Marine Geology, v. 23, p. 155-172.

Blood, D.R., 2011, Sequence stratigraphy key to Marcellus lateral placement. American Oil and Gas Reporter, August 2011, p. 52-60.

Bohacs, K.M., Grabowski, G.J., Carroll, A.R., Mankiewicz, P.J., Miskell-Gerhardt, K.J., and Schwalbach, J.R., 2005, Production, destruction, and dilution – the many paths to source-rock development. SEPM Special Publication 82, p. 61-101.

Boyer, D.L., and Droser, M.L., 2009, Paleoecological patterns within dysaerobic biofacies; examples from Devonian black shales of New York State. Palaeogeography, Palaeoclimatology, and Palaeoecology, v. 276, p. 206-216.

Boyer, D. L. and Droser, M. L., 2007, Devonian monospecific assemblages: new insights into the ecology of reduced oxygen depositional settings. Lethaia. Vol. 40, p. 321-334.

Boyer, D.L., Owens, J.D., Lyons, T.W., and Droser, M.L., 2011, Joining forces: Combined biological and geochemical proxies reveal a complex but refined high-resolution palaeo-oxygen history in Devonian epeiric seas. Palaeogeography, Palaeoclimatology, Palaeoecology, v. 306, p. 134-146.

Boyle, E.A., 1983, Chemical accumulation variations under the Peru Current during the past 130,000 years. Journal of Geophysical Research, v. 88, p. 7667-7680.

Bradley, D.C., Tucker, R.D., Lux, D.R., Harris, A.G., and McGregor, D.C., 2000, Migration of the Acadian orogen and foreland basin across the Northern Appalachians of Maine and adjacent areas. U.S. Geological Survey, Professional Paper 1624, 49 p.

Brett, C. E., and G. C. Baird. 1995. Coordinated stasis and evolutionary ecology of Silurian to Middle Devonian faunas in the Appalachian Basin. In D. H. Erwin and R. L. Anstey, eds. New approaches to speciation in the fossil record. Columbia University Press, New York, p. 285-315.

Brett, C. E., and Baird, G. C., 1996, Middle Devonian sedimentary cycles and sequences in the northern Appalachian Basin. *In* Witzke, B. M., Ludvigson, G. A., and Day, J., (eds.) Paleozoic Sequence Stratigraphy: Views from the North American Craton, Geological Society of America Special Paper 306, 213-241.

Brett, C. E., and Ver Straeten, C. A., 1994, Stratigraphy and facies relationships of the Eifelian Onondaga Limestone (Middle Devonian) in western and west central New York State. *In* Brett, C. E., and Scatterday, J., eds., New York State Geological Association 66th Annual Meeting Guidebook, p. 221–269.

Brett, C. E., Dick, V. B., and Baird, G. C., 1991, Comparative taphonomy and paleoecology of Middle Devonian dark gray and black shales facies from western New York. *In* Landing, E, and Brett, C. E. (eds.), Dynamic Stratigraphy and Depositional Environments Of The Hamilton Group (Middle Devonian) in New York State, Part II, New York State Museum Bulletin 469, p. 5-36.

Brett, C.E., Baird, G.C., Bartholomew, A., DeSantis, M., and Ver Straeten, C.A., 2011, Sequence stratigraphy and revised sea level curve for the Middle Devonian of eastern North America. Palaeogeography, Palaeoclimatology, Palaeoecology, v. 304, p. 21–53.

Brett, C.E., Ivany, L.C., Bartholomew, A.J., DeSantis, M.K. and Baird, G.C. 2009. Devonian ecological-evolutionary subunits in the Appalachian Basin: a revision and a test of persistence and discreteness. Geological Society of London, Special Publications, v. 314, p. 7-36.

Cate, A. S. 1963. Lithostratigraphy of some Middle and Upper Devonian rocks in the subsurface of southwestern Pennsylvania. Pennsylvania Topographic and Geological Survey, Fourth Series, General Geology Report G39, 229–440.

Catuneanu, O., 2002, Sequence stratigraphy of clastic systems: Concepts, merits, and pitfalls. Journal of African Earth Sciences, 35, 1-43.

Conant, L.C., and Swanson, V.E., 1961, Chattanooga Shale and related rocks of central Tennessee and nearby areas. U.S. Geological Survey, Professional Paper 357, 91 p.

Cooper, G. A. 1930a, Stratigraphy of the Hamilton Group of New York. American Journal of Science, v. 19, p. 116-134.

Cooper, G. A. 1930b, Stratigraphy of the Hamilton Group of New York. American Journal of Science, v. 19, p. 214-236.

Cooper, G. A. 1933, Stratigraphy of the Hamilton Group of eastern New York, part 1. American Journal of Science, 26, 537-551.

Cooper, G. A. 1941, Facies relations of the Middle Devonian (Hamilton) group along the Catskill Front. Geological Society of America Bulletin, v. 52, p. 1893.

Crick, R.E., Ellwood, B.B., El Hassani, A., Feist, R., 2000, Proposed magnetostratigraphy susceptibility magnetostratotype for the Eifelian–Givetian GSSP (Anti-Atlas Morocco). Episodes, v. 23, p. 93–101.

Day, J., Uyeno, T.C., Norris, A.W., Witzke, B.J., and Bunker, B.J., 1996, Middle- Upper Devonian relative sea-level histories of North American cratonic interior basins. *In* B.J. Witzke, G.A. Ludvigson, and Day, J., eds., Paleozoic sequence stratigraphy, views from the North American craton. Geological Society of America Special Paper 306, p. 259-276.

DeCelles, P.G., and Giles, K.A., 1996, Foreland basin systems. Basin Research, v. 8, p. 105-123.

Demaison, G.J., and Moore, G.T., 1980, Anoxic environments and oil source bed genesis. American Association of Petroleum Geologists Bulletin, v. 64, p. 1179–1209.

Dennison, J.M., and Boucot, A.J., 1974, Little War Gap at Clinch Mountain provides standard reference section for Silurian Clinch Sandstone and most nearly complete Devonian section in eastern Tennessee. Southeastern Geology, v. 16, p. 79-101.

DeSantis, M. and Brett, C.E., 2011. Late Eifelian to early Givetian bioevents: Timing and signature of the pre-Kac ák Bakoven and Stony Hollow events. Palaeogeography, Palaeoclimatology, Palaeoecology, v. 304, p. 113-135.

DeSantis, M.K., Brett, C.E., and Ver Straeten, C.A., 2007, Persistent depositional sequences and bioevents in the Eifelian (Early Middle Devonian) of eastern Laurentia: North American evidence of the Kacak Events?. *In* Becker, R.T., and Kirshwasser, W.T., eds., Devonian Events and Correlations. Geological Society of London, Special Publications, v. 278, p. 83-104.

Dickinson, W.R., 1974, Tectonics and sedimentation. Society of Economic Paleontologists and Mineralogists Special Publication No. 22, 204 p.

Dorobek, S. L. 1995. Synorogenic carbonate platforms and reefs in foreland basins: controls on stratigraphic evolution and platform/reef morphology. In Dorobek, S. L., and Ross, G. M., eds., Stratigraphic Evolution of Foreland Basins. SEPM Special Publication, v. 52, p. 127–147.

Dunham, R. J., 1962, Classification of carbonate rocks according to depositional texture, *in* Ham, W. E., ed., Classification of Carbonate Rocks. American Association of Petroleum Geologists, Memoir 1, p. 108-121.

Ellwood, B.B., Crick, R.E., El Hassani, A., Benoist, S.L., Young, R.H., 2000, The magnetosusceptibility event and cyclostratigraphy (MSEC) method applied to marine rocks: detrital input versus carbonate productivity. Geology, v. 28, p. 1134–1138.

Ellwood, B.B., Algeo, T., El Hassani, A., Tomkin, J.H., Rowe, H., 2011, Defining the timing and duration of the Kačák interval within the Eifelian/Givetian Boundary GSSP, Mech Irdane, Morocco, using geochemical and magnetic susceptibility patterns. Palaeogeography, Palaeoclimatology, Palaeoecology, v. 304, p. 74–84.

Embry, A., 1995, Sequence boundaries and sequence hierarchies: problems and proposals. *In* Steel, R., et al., eds., Sequence stratigraphy on the northwest European margin. Norwegian Petroleum Society Special Publication 5, p. 1-11.

Embry, A., 2002, Transgressive-regressive (T-R) sequence stratigraphy. *In* Armentrout, J., and Rosen, N., eds., Gulf Coast SEPM Conference Proceedings, Houston, p. 151-172.

Embry, A., 2010, Correlating siliciclastic successions with sequence stratigraphy. SEPM Special Publication No. 94, 35-53.

Embry, A., and E. Johannessen, 1992, T-R sequence stratigraphy, facies analysis and reservoir distribution in the uppermost Triassic-Lower Jurassic succession, western Sverdrup Basin, Arctic Canada. *In* Vorren, T., et al., eds., Arctic geology and petroleum potential. Norwegian Petroleum Society Special Publication 2, p. 121-146.

Emery, D., and Meyers, K. J., 1996, Sequence stratigraphy. Blackwell Science Ltd., Cambridge, MA, 297 p.

Ettensohn, F. R. 1985a, The Catskill Delta Complex and the Acadian Orogeny: a model. *In* Woodrow, D. L. and Sevon, W. D. (eds.) The Catskill Delta, Geological Society of America, Special Paper 201, p. 39-50.

Ettensohn, F.R., 1985b, Controls on development of Catskill Delta complex basin-facies. *In* Woodrow, D. L. and Sevon, W. D. (eds.) The Catskill Delta, Geological Society of America, Special Paper 201, p. 65-77.

Fisher, J. St. J., and Hudson, J.D., 1985, Pyrite geochemistry and fossil preservation in shales, *in* Whittington, H,B., and Morris, S. C., eds., Extraordinary Fossil Biotas: Their Ecological and Evolutionary Significance. Philosophical Transactions of the Royal Society of London, Series B, v. 311, p. 167-169.

Flemings, P.B., and Jordan, T.E., 1989, A synthetic stratigraphic model of foreland basin development. Journal of Geophysical Research, v. 94, p. 3851-3866.

Flemings, P.B., and Jordan, T.E., 1990, Stratigraphic modeling of foreland basins: interpreting thrust deformation and lithospheric rheology. Geology, v. 18, p. 430-435.

Flugel, E., 2010, Microfacies of Carbonate Rocks: Analysis, Interpretation and Application. Springer, Berlin, 984 p.

Griffing, D. H. 1994, Microstratigraphy, facies, paleoenvironments, and the origin of widespread, shale-hosted skeletal limestones in the Hamilton Group (Middle Devonian) of New York State. Unpublished Ph.D. dissertation, State University of New York at Binghamton, 202 p.

Griffing, D.H., and Ver Straeten, C.A., 1991, Stratigraphy and depositional environments of the lower part of the Marcellus Formation (Middle Devonian) in eastern New York State. *In* Ebert, J.R., ed., New York State Geological Association, 63rd Annual Meeting Guidebook, p. 205-249.

Hall, J., 1839, Third annual report of the Fourth Geological District of the State of New York. New York Geological Survey Annual Report, v. 3, p. 287-339.

Hatcher, R.D., Jr., 2010, The Appalachian orogen: A brief summary. *In* Tollo, R.P., Bartholomew, M.J., Hibbard, J.P., and Karabinos, P.M., eds., From Rodinia to Pangea: The Lithotectonic Record of the Appalachian Region. Geological Society of America Memoir 206, p. 1-19.

Hibbard, J.P., van Staal, C.R., and Rankin, D.W., 2010, Comparative analysis of the geological evolution of the northern and southern Appalachian orogen: Late Ordovician-Permian. *In* Tollo, R.P., Bartholomew, M.J., Hibbard, J.P., and Karabinos, P.M., eds., From Rodinia to Pangea: The Lithotectonic Record of the Appalachian Region. Geological Society of America Memoir 206, p. 51-69.

Hjulstrom, F., 1955, Transportation of debris by moving water. *In* Trask, P.D., ed., Recent Marine Sediments, SEPM Special Publication, no. 4, p. 5-31.

House, M.R., 1985, Correlation of mid-Palaeozoic ammonoids evolutionary events with global sedimentary perturbations. Nature, v. 313, p. 17–22.

Ibach, L.E.J., 1982, Relationship between sedimentation rate and total organic carbon content in ancient marine sediments. American Association of Petroleum Geologists Bulletin, v. 66, p. 170-188.

Joachimski, M. M., van Geldern, R., Breisig, S., Buggisch, W. & Day, J. 2004, Oxygen isotope evolution of biogenic calcite and apatite during the Middle and Late Devonian. International Journal of Earth Sciences (Geol. Rundschau), v. 93, p. 542–553.

Johnson, J. G., Klapper, G., and Sandberg, C. A., 1985, Devonian eustatic fluctuations in Euramerica. Geological Society of America Bulletin, v. 96, p. 567-687.

Jordan, T.E., 1981, Thrust loads and foreland basin evolution, Cretaceous, western United States. American Association of Petroleum Geologists Bulletin, v. 65, p. 2506-2520.

Kaufmann, B., 2006, Calibrating the Devonian time scale: a synthesis of U-Pb ID-TIMS ages and conodont stratigraphy. Earth Science Reviews, v. 76, p. 175–190.

Kinley, T.J., Cook, L.W., Breyer, J.A., Jarvie, D.M., and Busbey, A.B., 2008, Hydrocarbon potential of the Barnett Shale (Mississippian), Delaware Basin, west Texas and southeastern New Mexico. American Association of Petroleum Geologists Bulletin, v. 92, p. 967-991.

Klapper, G. 1981, Review of New York Devonian conodont biostratigraphy. *In* Oliver, W. A., Jr., and Klapper, G. (eds.) Devonian Biostratigraphy of New York, Part 1. International Union of Geological Sciences, Subcommission on Devonian Stratigraphy, p. 57-76.

Lash, G.G., and Engelder, T., 2005, An analysis of horizontal microcracking during catagenesis: an example from the Catskill delta complex. American Association of Petroleum Geologists Bulletin, v. 95, p. 61-103.

Lash, G.G., and Engelder, T., 2011, Thickness trends and sequence stratigraphy of the Middle Devonian Marcellus Formation, Appalachian Basin: implications for Acadian foreland basin evolution. American Association of Petroleum Geologists Bulletin, v. 89, p. 1433-1449.

Lindemann, R. H., 2002, Dacryoconarid bioevents of the Onondaga Formation and the Marcellus Subgroup, Cherry Valley, New York. *In* McLelland, J., and Karabinos, P., eds., New England Intercollegiate Geological Conference and New York State Geological Association, Fieldtrip Guidebook, p. B7.1-7.17.

Loucks, R. G., Reed, R. M., Ruppel, S. C., and Jarvie, D. M., 2009, Morphology, genesis, and distribution of nanometer-scale pores in siliceous mudstones of the Mississippian Barnett Shale. Journal of Sedimentary Research, v. 79, p. 848-861.

Macquaker, J.H.S., and Bohacs, K.M., 2007, On the accumulation of mud. Science, v. 318, p. 1734-1735.

Macquaker, J.H.S., Bentley, S.J., and Bohacs., K.M., 2011, Wave-enhanced sediment-gravity flows and mud dispersal across continental shelves: Reappraising sediment transport processes operation in ancient mudstone successions. Geology, v. 38, p. 947-950.

McCave, N., 1984, Mechanics of deposition of fine-grained sediments from nepheloid layers. Geo-Marine Letters, v. 4, p. 243-245.

McCave, N., 2001, Nepheloid layers. *In* Steele, J.H., Thorpe, S.A., and Turekian, K.K., eds., Encyclopedia of Ocean Sciences, Academic Press, London, v. 4, p. 1861-1870.

Miall, A.D., 1995, Collision-related foreland basins, *In* Busby, C.J., and Ingersoll, R.V., eds., Tectonics of sedimentary basins. Boston, Blackwell Science, p. 393-424.

Miller, K.B., Brett, C.E., and Parsons, K.M., 1988, The paleoecologic significance of storm-generated disturbance within a Middle Devonian muddy epeiric sea. Palaios, v. 3, p. 35-52.

Milligan, T.G., Hill, P.S., and Law, B.A., 2007, Flocculation and the loss of sediment from the Po River plume. Continental Shelf Research, v. 27, p. 309-321.

Mitchum, R. M. J., and J.C. Van Wagoner, 1991. High-frequency sequences and their stacking patterns: sequence stratigraphic evidence of high-frequency eustatic cycles. Sedimentary Geology, v. 70, p. 131–160.

Mitchum, R. M., Jr. Vail, P. E., and Thompson S., III, 1977, The depositional sequence as a basic unit for stratigraphic analysis. *In* Payton, C. E., ed., Seismic Stratigraphy - Applications to Hydrocarbon Exploration, American Association of Petroleum Geologists, Memoir 26, p. 53-62.

Mulder, T., Syvitski, J.P.M., Migeon, S., Faugeres, J.-C., and Savoye, B., 2003, Marine hyperpycnal flows: Initiation, behavior and related deposits. A review. Marine and Petroleum Geology, v. 20, p. 861-882.

Murphy, A.E., Sageman, B.B., Hollander, D.J., and Ver Straeten, C.A., 2000a, Organic carbon burial and faunal dynamics in the Appalachian basin during the Devonian (Givetian-Famennian) greenhouse: An integrated paleoecological/biogeochemical approach. *In* Huber, B., MacLeod, K., and Wing, S., eds., Warm Climates in Earth History, Cambridge University Press, p. 351-385.

Murphy, A. E., Sageman, B. B., Hollander, D.J., Lyons, T.W., and Brett, C.E., 2000b, Black shale deposition in the Devonian Appalachian Basin: Siliciclastic starvation, episodic water-column mixing, and efficient recycling of biolimiting nutrients. Paleoceanography, v. 15, p. 280-291.

North American Commission on Stratigraphic Nomenclature, 1983, North American Stratigraphic Code. American Association of Petroleum Geologists Bulletin, v. 67, p. 841-875.

Oggston, A.S., and Sternberg, R.W., 1999, Sediment-transport events on the northern California continental shelf. Marine Geology, v. 154, p. 69-82.

Oggston, A.S., Cacchione, D.A., Sternberg, R.W., and Kineke, G.C., 2000, Observations of storm and river flood-driven sediment transport on the northern California continental shelf. Continental Shelf Research, v. 20, p. 2141-2162.

Oliver, W. A., Jr. 1954, Stratigraphy of the Onondaga Limestone (Devonian) in central New York. Geological Society of America Bulletin, v. 65, p. 621-652.

Oliver, W. A., Jr. 1956, Stratigraphy of the Onondaga Limestone in eastern New York. Bulletin of the Geological Society of America, v. 67, p. 1441-1474.

Osberg, P.H., Tull, J.F., Robinson, P., Hon, R., Butler, J.R., 1989, The Acadian orogen, *In* Hatcher, R.D., Thomas, W.A., and Viele, G.W., eds., The Appalachian-Ouachita orogen in the United States, Geological Society of America, The Geology of North America, v. F2, p. 179-232.

Parsons, K.M., Brett, C.E., and Miller, K.B., 1988, Taphonomy and depositional dynamics of Devonian shell-rich mudstones. Palaeogeography, Palaeoclimatology, Palaeoecology, v. 63, p. 109-139.

Partington, M. A., B.C. Mitchener, N.J. Milton, and A.J. Fraser, 1993, Genetic sequence stratigraphy for the North Sea Late Jurassic and Early Cretaceous: distribution and prediction of Kimmeridgian — Late Ryazanian reservoirs in the North Sea and adjacent areas. *In* Parker, J.R., ed., Petroleum Geology of Northwest Europe. Proceedings of the 4th Conference, Geological Society of London, p. 347–370.

Pedersen T.F. and Calvert S.E., 1990, Anoxia vs productivity – what controls the formation of organic carbon-rich sediments and sedimentary rocks. American Association of Petroleum Geologists Bulletin, v. 74, p. 454–466.

Potter, P.E., Maynard, J.B., and Prior, W.A., 1980, Sedimentology of shale. Springer-Verlag, New York, 306 p.

Potter, P.E., Maynard, J.B., and Prior, W.A., 1982, Appalachian gas bearing Devonian shales: Statements and discussions. Oil and Gas Journal, v. 80, p. 290-318.

Quinlan, G., and Beaumont, C., 1984, Appalachian thrusting, lithospheric flexure, and the Paleozoic stratigraphy of the Eastern Interior of North America. Canadian Journal of Earth Science, v. 21, p. 973-996.

Ransom, B., Shea, K.F., Burkett, P.J., Bennett, R.H., and Baerwald, R., 1998, Comparison of pelagic and nepheloid layer marine snow: implications for carbon cycling. Marine Geology, v. 150, p. 39-50.

Rast, N., and Skehan, J.W., 1993, Mid-Paleozoic orogenesis in the North Atlantic: The Acadian orogeny. I*n* Roy, D.C., and Skehan, J.W., eds., The Acadian orogeny: Recent studies in New England, Maritime Canada, and the autochthonous foreland. Geological Society of America Special Paper 275, p. 1-25.

Rich, J.L., 1951, Probable fondo origin of Marcellus-Ohio-New Albany-Chattnooga bituminous shales. Bulletin of the American Association of Petroleum Geologists, v. 35, p. 2017-2040.

Rickard, L. V., 1975, Correlation of the Silurian and Devonian Rocks in New York State. New York State Map and Chart 24, 16 p., 4 plates.

Rickard, L.V., 1984, Correlation of the subsurface Lower and Middle Devonian of the Lake Erie region. Geological Society of America Bulletin, v. 95, p. 814-828.

Rickard, L. V., 1989, Stratigraphy of the subsurface Lower and Middle Devonian of New York, Pennsylvania, Ohio, and Ontario. New York State Museum Map and Chart 39, 59 p., 40 plates.

Rine, J.M., and Ginsburg, R.N., 1985, Depositional facies of a mud shoreface in Suriname, South America – a mud analogue to sandy, shallow-marine deposits. Journal of Sedimentary Petrology, v. 55, p. 633-652.

Robinson, P., Tucker, R.D., Bradley, D.C., Berry, H.N., IV, and Osberg, P.H., 1998, Paleozoic orogens in New England, U.S.A.. GFF (Geological Society of Sweden Quarterly Journal), v. 120, p. 119–148.

Roden, M.K., Parrish, R.R., and Miller, D.S., 1990, The absolute age of the Eifelian Tioga ash bed, Pennsylvania. The Journal of Geology, v. 98, p. 282–285.

Rodgers, J., 1967, Chronology of tectonic movements in the Appalachian region of eastern North America. American Journal of Science, v. 265, p. 408–427.

Roy, D.C., and Skehan, J.W., eds., 1993, The Acadian Orogeny: Recent Studies in New England, Maritime Canada, and the Autochthonous Foreland. Geological Society of America Special Paper 275, 171 p.

Sageman, B.B., Murphy, A.E., Werne, J.P., Ver Straeten, C.A., Hollander, D.J., and Lyons, T.W., 2003, A tale of shales: The relative roles of production, decomposition, and dilution in the accumulation of organic-rich strata, Middle-Upper Devonian, Appalachian basin, Chemical Geology, v. 195, p. 229-273.

Salvador, A., 1994, International Stratigraphic Guide. International Subcommission on Stratigraphic Classification of IUGS International Commission on Stratigraphy. International Union of Geological Sciences and the Geological Society of America, 214 p.

Schieber, J., 1994, Evidence for high-energy events and shallow water deposition in the Chattanooga Shale, Devonian, central Tennessee, USA. Sedimentary Geology, v. 93, p. 193-208.

Schieber, J., 1998a, Deposition of mudstones and shales: Overview, problems, and challenges. *In* Schieber, J., Zimmerle, W., and Sethi, P.S., eds., Shales and Mudstones, E Schweizerbart'sche Verlagsbuchhandlung (Nagele u. Obermiller), Stuttgart, vol. I, p. 131-146..

Schieber, J., 1998b, Sedimentary features indicating erosion, condensation, and hiatuses in the Chattanooga Shale of central Tennessee: Relevance for sedimentary and stratigraphic evolution. *In* Schieber, J., Zimmerle, W., and Sethi, P.S., eds., Shales and Mudstones, E Schweizerbart'sche Verlagsbuchhandlung (Nagele u. Obermiller), Stuttgart, vols. I, p. 187-215.

Schieber, J., 1999, Distribution and deposition of mudstone facies in the Upper Devonian Sonyea Group of New York. Journal of Sedimentary Research, v. 69, p. 909-925.

Schieber, J., 2002a, The role of an organic slime matrix in the formation of pyritized burrow trails and pyrite concretions. Palaios, v. 17, p. 104-109.

Schieber, J., 2002b, Sedimentary pyrite: a window into the microbial past. Geology, v. 30, p. 531-534.

Schieber, J., and Yawar, Z., 2009, A new twist on mud deposition – mud ripples in experiment and rock record. The Sedimentary Record, Society for Sedimentary Geology, v.7, p. 4-8.

Schieber, J., Zimmerle, W., and Sethi, P.S., 1998, Shales and mudstones. E Schweizerbart'sche Verlagsbuchhandlung (Nagele u. Obermiller), Stuttgart, vols. I and II, 384 and 296 p.
Schieber, J., Southard, J., and Thaisen, K., 2007, Accretion of mudstone beds from migrating floccules ripples. Science, v. 318, p. 1760-1763.
Schieber, J., Southard, J., and Schimmelmann, A., 2010, Lenticular shale fabrics resulting from intermittent erosion of water-rich muds – interpreting the rock record in the light of recent flume experiments. Journal of Sedimentary Research, v. 80, p. 119-128.
Schlager, W., 1981, The paradox of drowned reefs and carbonate platforms. Geological Society of America Bulletin, v. 92, p. 197–211,
Scotese, C. R., and McKerrow, W. S., 1990, Revised world maps and introduction. *In* McKerrow, W. S., and Scotese, C. R., eds., Paleozoic Paleogeography and Biogeography, Geological Society of London, Geological Society Memoir 12, p. 1-21.
Sevon W. D., and Woodrow D. L., 1985, Middle and Upper Devonian stratigraphy within the Appalachian basin. Geological Society of America, Special Paper 201, p. 1–7.
Singh, P., R. Slatt, and W. Coffey, 2008, Barnett Shale – unfolded: sedimentology, sequence stratigraphy, and regional mapping. Gulf Coast Association of Geological Societies Transactions, v. 58, p. 777-795.
Smith, L.B., 2010, Shallow transgressive onlap model for Ordovician and Devonian organic-rich shales, New York State. American Association of Petroleum Geologists, Eastern Section Meeting, p. 57.
Sommerfield, C.K., Oggston, A.S., Mullenbach, B.L., Drake, D.E., Alexander C.R., Nittrouer, C.A., Borgeld, J.D., Wheatcroft, R.A., and Leithold, E.L., 2007, Oceanic dispersal and accumulation of river sediment. *In* Nittrouer, C.A., Austin, J.A., Field, M.E., Kravitz, J.H., Syvitski, J.P.M., and Wiberg, P.L., eds., Continental Margin Sedimentation: From Sediment Transport to Sequence Stratigraphy, International Association of Sedimentologists and Blackwell Publishing, Malden, MA, p. 157-212.
Syvitski, J.P.M., 1991, The changing microfabric of suspended particulate matter - the fluvial to marine transition: flocculation, agglomeration and pelletization. *In* Bennett, R.H., Bryant, W.R., and Hulbert, M.H., eds., The Microstructure of Fine-grained Sediment - from Muds to Shale, Frontiers in Sedimentary Geology, Springer-Verlag, New York, p. 131-137.
Taylor, K.G., Gawthorpe, R.L., and Van Wagoner, J.C., 1995, Stratigraphic control on laterally persistent cementation, Book Cliffs, Utah. Journal of the Geological Society, v. 152, p. 225-228.
Thompson, J.B. and Newton, C.R., 1987, Ecological reinterpretation of the dysaerobic *Leiorhynchus* fauna, Upper Devonian Geneseo black shale, central New York. Palaios v. 2, p. 274–281.
van der Voo, R., 1983, Paleomagnetic constraints on the assembly of the Old Red Continent. Tectonophysics, v. 91, p. 271-283.
Van Geldern, , R. & Joachimski, M. M., et al., 2006, Carbon, oxygen and strontium isotope records of Devonian brachiopod shell calcite. Palaeogeography,Palaeoclimatology, Palaeoecology, v. 240, p. 47–67.
Van Wagoner, J. C., Posamentier, H. W., Mitchum, R. M., Vail, P. R., Sarg. K., G. Loutit, T. S., and Hardenbol, J., 1988, An overview of the fundamentals of sequence stratigraphy and key definitions. *In* Wilgus, C. K., Hastings, B. S., St. C. Kendall, C. G., Posamentier, H. W., Ross, C. A., and Van Wagoner, H. C., eds., Sea-Level Changes: An Integrated Approach, Society of Economic Paleontologists and Mineralogists, Special Publication 42, p. 39-45.
van Staal, C. R., Whalen, J. B., Valverde-Vaquero, P., Zagorevski, A., and Rogers, N., 2009, Pre-Carboniferous, episodic accretion-related, orogenesis along the Laurentian margin of the northern Appalachians. *In* Murphy, J. B., Keppie, J. D., and Hynes, A. J. eds., Ancient orogens and modern analogues, London, Geological Society Special Publication 327, p. 271-316.
Ver Straeten, C.A., 1994, Microstratigraphy and depositional environments of a Middle Devonian foreland basin: Berne and Otsego Members, Mount Marion Formation, eastern New York State. *In* Landing, E., ed., Studies in Stratigraphy and Paleontology in Honor of Donald W. Fisher, New York State Museum Bulletin 481, p. 367-380.
Ver Straeten, C.A., 1996a, Stratigraphic synthesis and tectonic and sequence stratigraphic framework, upper Lower and Middle Devonian, Northern and Central Appalachian Basin. Unpublished Ph.D. thesis, University of Rochester, 800 p.
Ver Straeten, C.A., 1996b, Upper Lower and lower Middle Devonian stratigraphic synthesis, central Appalachian Basin of Pennsylvania. Open File Report 96-47, Pennsylvania Topographic and Geological Survey, 59 p.
Ver Straeten, C.A., 2004, K-bentonites, Volcanic Ash Preservation, and Implications for Lower to Middle Devonian Volcanism in the Acadian Orogen, Eastern North America. Geological Society of America Bulletin, v. 116, p. 474-489.
Ver Straeten, C.A., 2007, Basinwide Stratigraphic Synthesis and Sequence Stratigraphy, Upper Pragian, Emsian and Eifelian Stages (Lower to Middle Devonian), Appalachian Basin. *In* Becker, R.T., and

Kirchgasser, W.T., eds., Devonian Events and Correlations. Geological Society of London, Special Publications, v. 278, p. 39-81.
Ver Straeten, C.A., 2009, The classic Devonian of the Catskill front: A foreland basin record of Acadian orogenesis. *In* Vollmer, F., ed., New York State Geological Association, Annual Fieldtrip guidebook, New Paltz, p. 7-1 to 7-54.
Ver Straeten, C.A., 2010, Lessons from the foreland basin: Northern Appalachian basin perspectives on the Acadian orogeny. *In* Tollo, R.P., Bartholomew, M.J., Hibbard, J.P., and Karabinos, P.M., eds., From Rodinia to Pangea: The Lithotectonic Record of the Appalachian Region. Geological Society of America Memoir 206, p. 251–282.
Ver Straeten, C.A., and Brett, C.A., 1995, Lower and Middle Devonian foreland basin fill in the Catskill Front: Stratigraphic synthesis, sequence stratigraphy, and the Acadian Orogeny. *In* Garver, J.I., and Smith J.A., eds., New York State Geological Association, 67th Annual Meeting Guidebook, p. 313-356.
Ver Straeten, C.A., and Brett, C.E., 2006, Pragian to Eifelian strata (mid Lower to lower Middle Devonian), northern Appalachian Basin – A stratigraphic revision. Northeastern Geology, vol. 28, p. 80-95.
Ver Straeten, C.A., Brett, C.E., and Sageman, B.B., 2011, Mudrock sequence stratigraphy: A multi-proxy (sedimentologic, paleobiologic, geochemical) approach, Devonian Appalachian basin. Palaeogeography, Palaeoclimatology, Palaeoecology, p. 54-73.
Ver Straeten, C.A., Griffing, D.H., and Brett, C.E., 1994, The lower part of the Middle Devonian Marcellus "Shale," central to western New York State: Stratigraphy and depositional History. *In* Brett, C.E., and Scatterday J., eds., New York State Geological Association, 66th Annual Meeting Guidebook, p. 270-321 (includes field trip log).
Walliser, O.H., 1996, Global events in the Devonian and Carboniferous. *In* Walliser, O.H. (Ed.), Global Events and Event Stratigraphy, Springer Verlag, p. 225–250.
Werne, J. P., Sageman, B. B., Lyons, T. W., and Hollander, D. J., 2002, An integrated assessment of a "type euxinic" deposit: Evidence for multiple controls on black shale deposition in the middle Devonian Oatka Creek Formation. American Journal of Science, v. 302, p. 110-143.
Wilgus, C. K., Hastings, B. S., Kendall, C. G., Posamentier, H. W., Ross, C. A., and Van Wagoner, J. C., 1988, Sea-level changes: An integrated approach. Society of Economic Paleontologists and Mineralogists, Special Publication 42, 407 p.
Wilkin, R.T., Barnes, H.L. and Brantley, S.L., 1996, The size distribution of framboidal pyrite in modern sediments: an indicator of redox conditions. Geochimica et Cosmochimica Acta, v. 60, p. 3897–3912.
Witzke, B. J., 1990, Paleoclimatic constraints for Paleozoic paleolatitutes of Laurentia and Euramerica. *In* McKerrow, W. S., and Scotese, C. R., eds., Paleozoic Paleogeography and Biogeography, Geological Society of London, Geological Society Memoir 12, p. 57-74.
Witzke, B.J., 2011, Estimating magnitudes of sea-level changes associated with Middle and Upper Devonian sequences in the Iowa area. Geological Society of America Abstracts with Programs, v. 43, No. 1, p. 96.

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

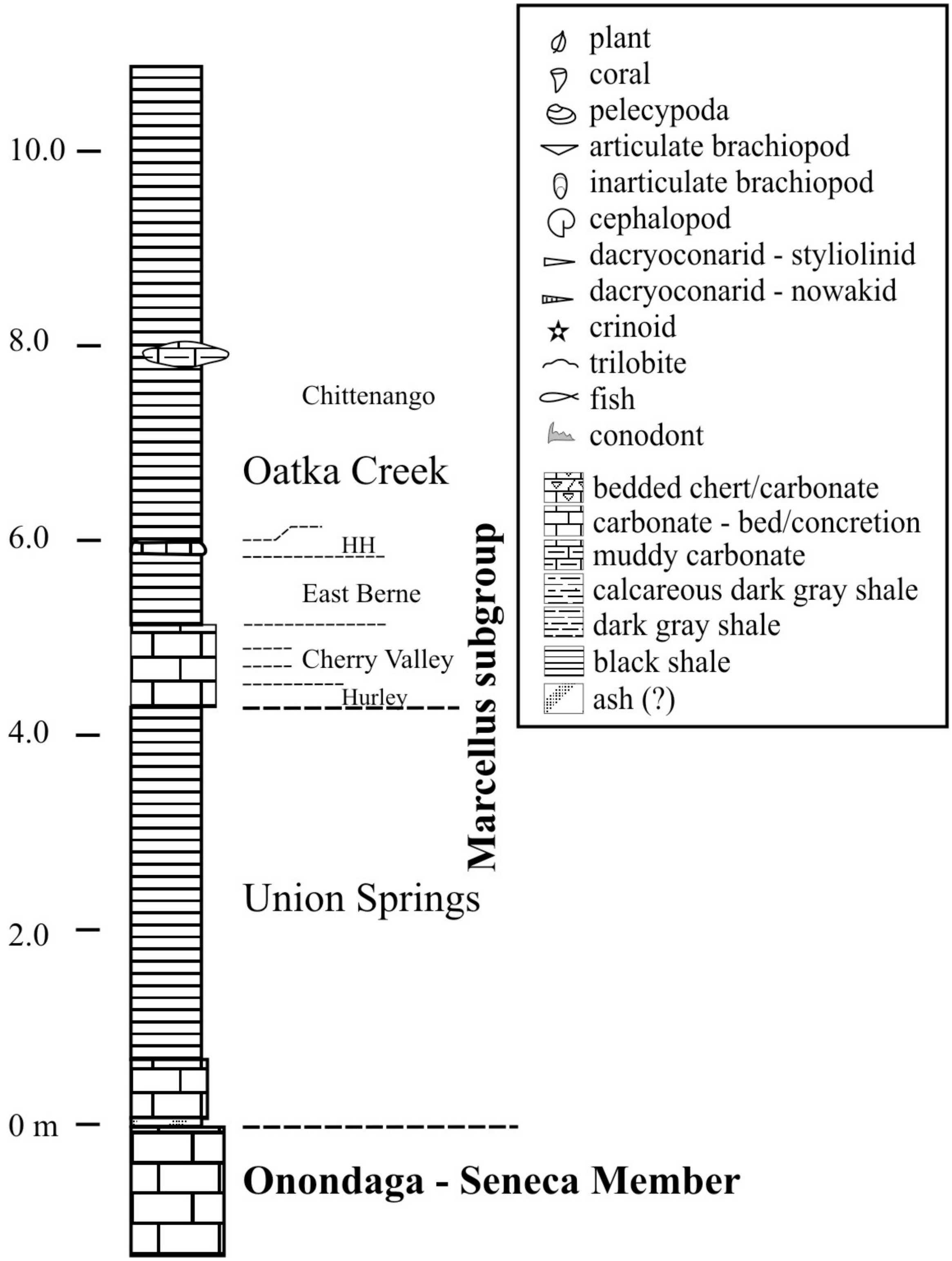


Figure RL-1A. General stratigraphy of Marcellus subgroup strata at Seneca Stone Quarryand key to symbols; HH = Halihan Hill Bed.

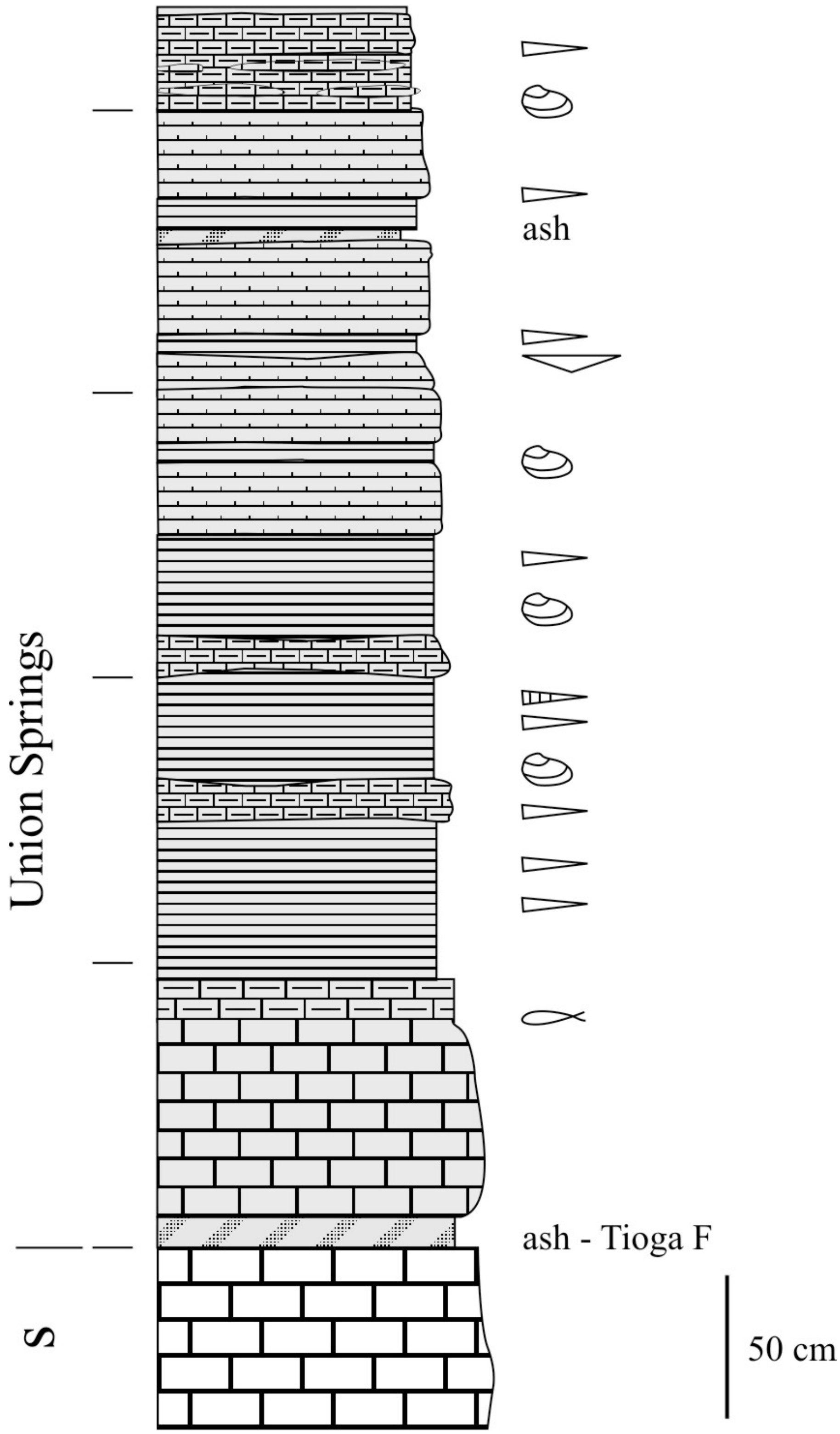


Figure RL-1B. General stratigraphy of the Union Springs Formation above the uppermost Seneca Member of the Onondaga Limestone at Seneca Stone Quarry.

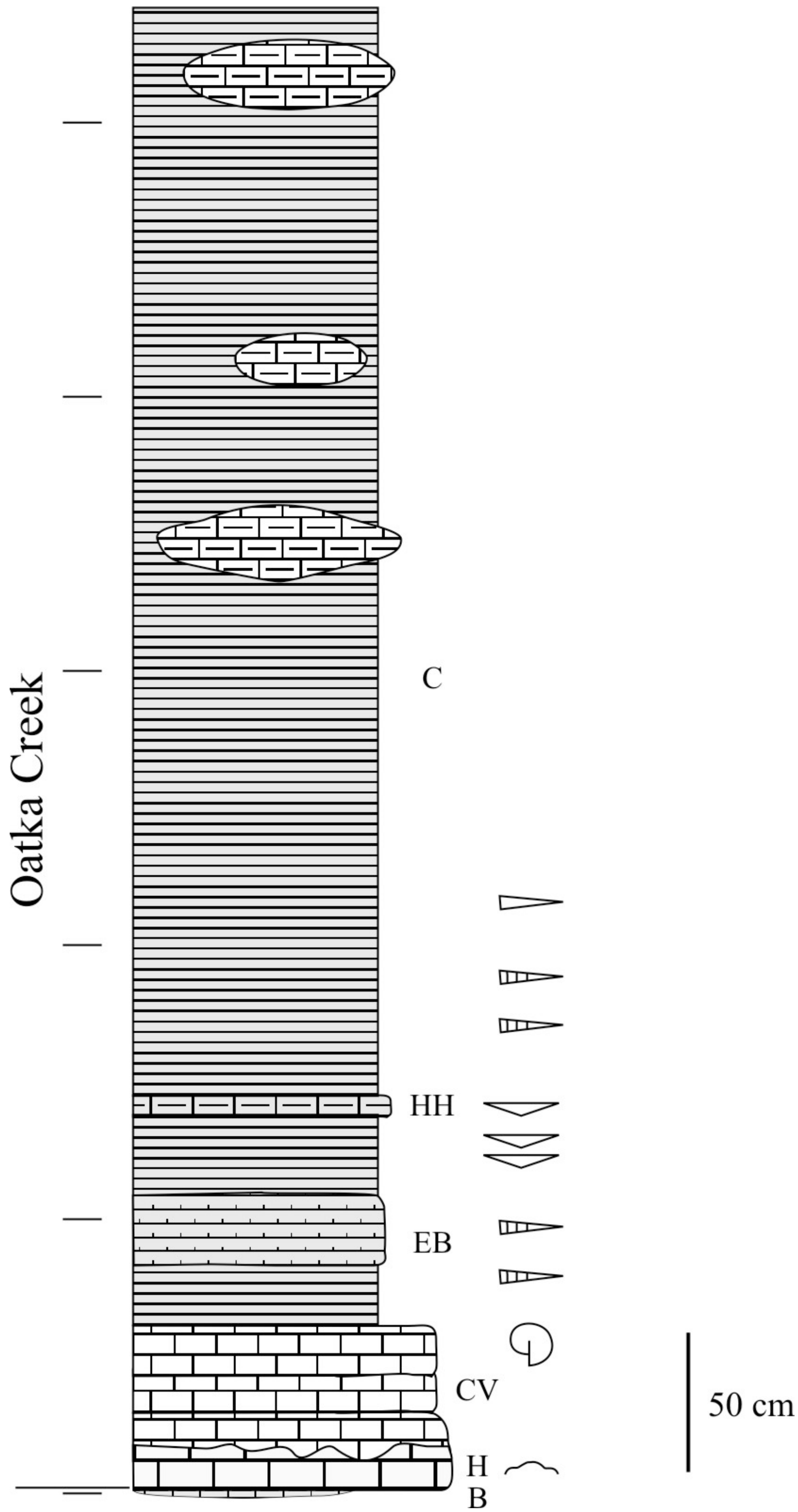


Figure RL-1C. General stratigraphy of the Oatka Creek Formation at Seneca Stone Quarry. B = Bakoven; H = Hurley; CV = Cherry Valley; EB = East Berne; HH = Halihan Hill; C = Chittenango.

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

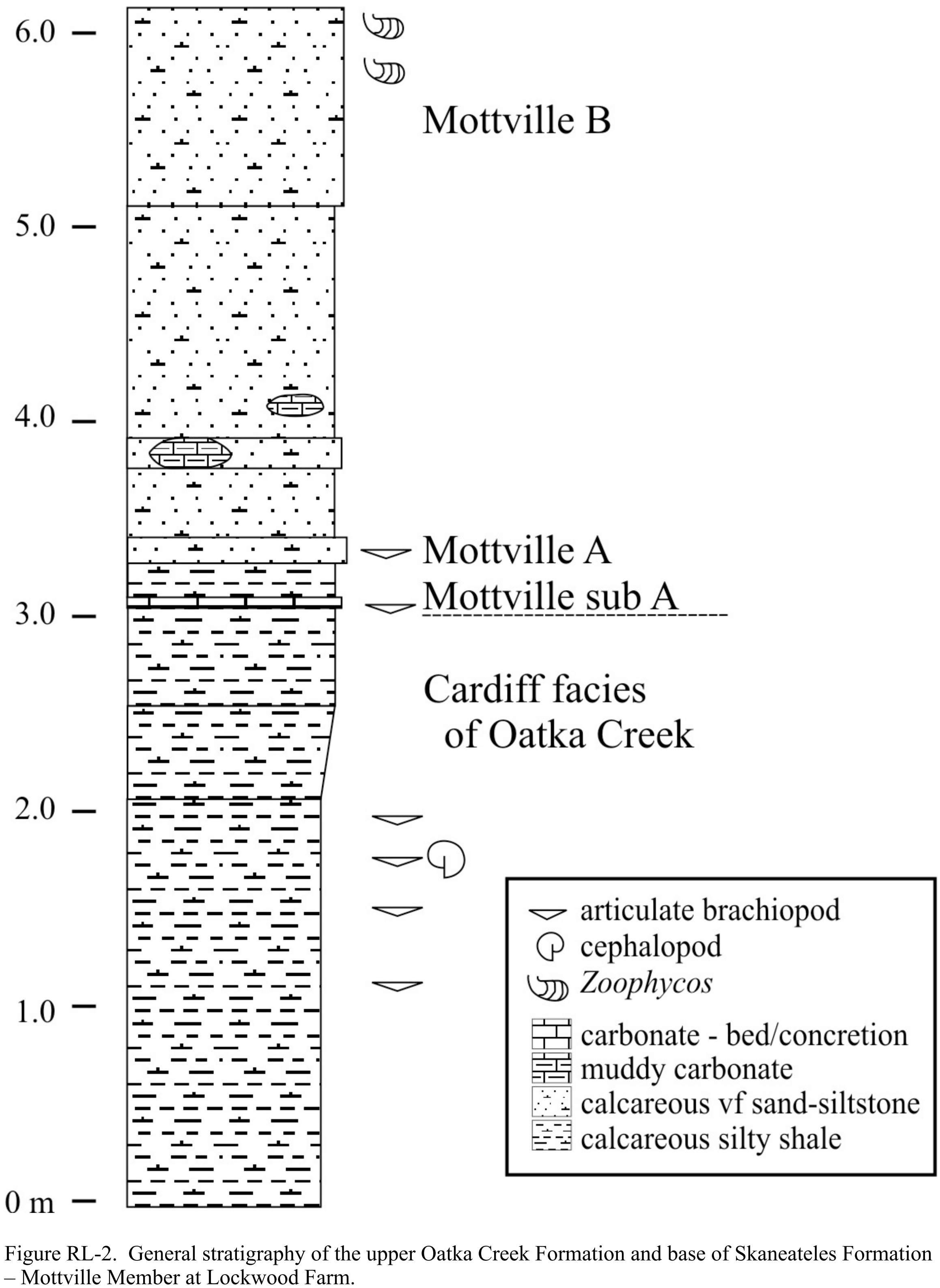


Figure RL-2. General stratigraphy of the upper Oatka Creek Formation and base of Skaneateles Formation – Mottville Member at Lockwood Farm.

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.

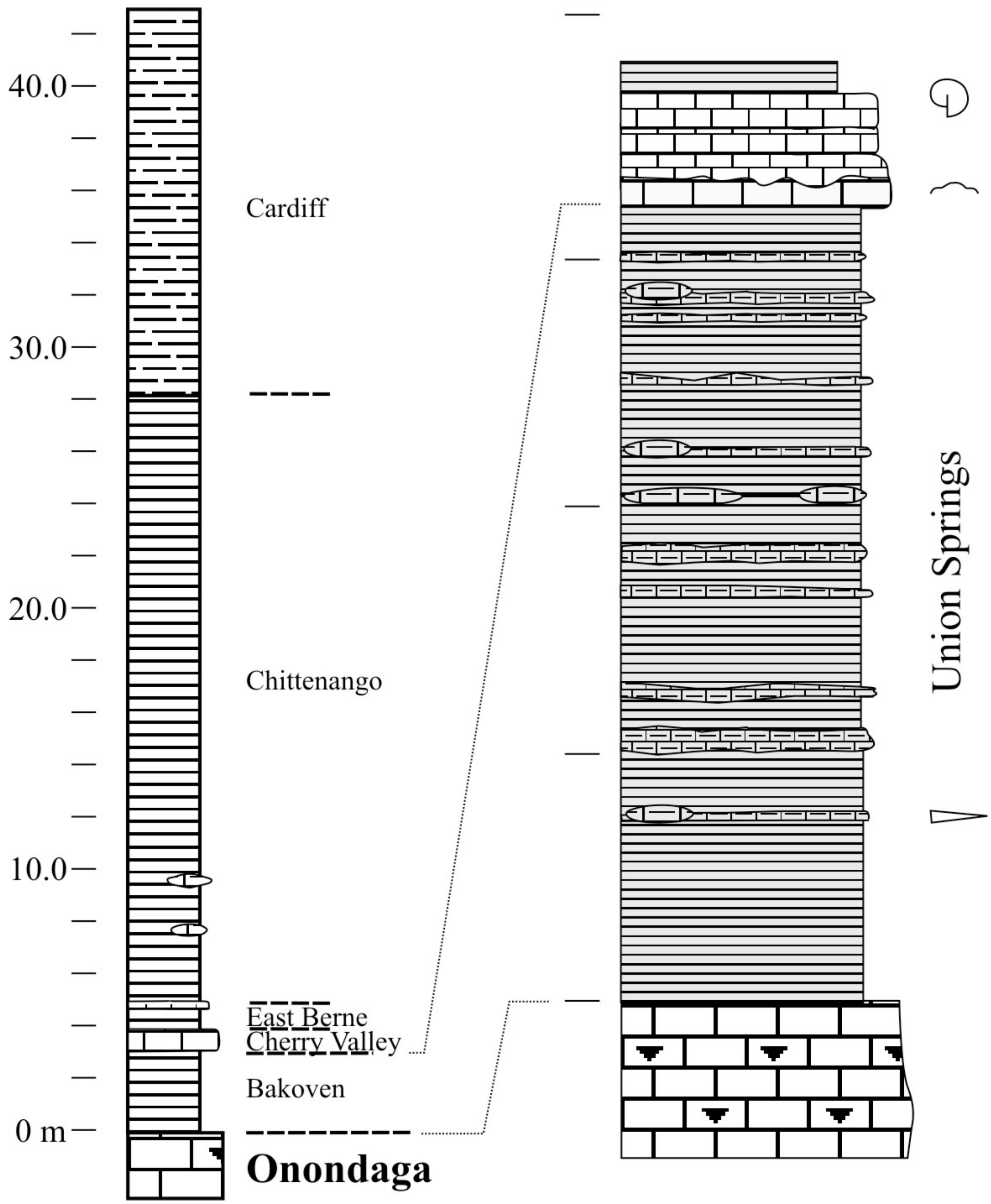


Figure RL-3. General stratigraphy of the Marcellus subgroup in the type area near Slate Hill Road, Marcellus, NY. See Figure RL-1A for key to symbols.

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.