

Thickness trends and sequence stratigraphy of the Middle Devonian Marcellus Formation, Appalachian Basin: Implications for Acadian foreland basin evolution

Gary G. Lash and Terry Engelder

ABSTRACT

Analysis of more than 900 wireline logs indicates that the Middle Devonian Marcellus Formation encompasses two third-order transgressive-regressive (T-R) sequences, MSS1 and MSS2, in ascending order. Compositional elements of the Marcellus Formation crucial to the successful development of this emerging shale gas play, including quartz, clay, carbonate, pyrite, and organic carbon, vary predictably within the proposed sequence-stratigraphic framework. Thickness trends of Marcellus T-R sequences and lithostratigraphic units reflect the interplay of Acadian thrust-load-induced subsidence, short-term base-level fluctuations, and recurrent basement structures. Rapid thickening of both T-R sequences, especially MSS2, toward the northeastern region of the basin preserves a record of greater accommodation space and proximity to clastic sources early in the Acadian orogeny. However, local variations in T-R sequence thickness in the western, more distal, area of the basin may reflect the reactivation of inherited Eocambrian basement structures, including the Rome trough and northwest-striking cross-structural discontinuities, induced by Acadian plate convergence. Episodes of block displacement locally warped the basin into northeast-southwest-trending regions of starved sedimentation and/or erosion adjacent to depocenters in which regressive systems tract deposits were ponded. Block movement

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appears to have initiated in late Early Devonian time, resulting first in thinning and local erosion of the Oriskany sandstone in northwest Pennsylvania. This study, in addition to providing the basis for a predictive sequence-stratigraphic model that can be used to further Marcellus exploration, tells of a foreland basin more tectonically complex than accounted for by simple flexural models.

INTRODUCTION

The much studied Hamilton Group of the Appalachian Basin is an eastward and southeastward thickening wedge of marine and nonmarine shale, siltstone, and sandstone (Cooper, 1933, 1934; Rickard, 1989). To the west of the Hudson Valley, the Hamilton succession is defined by an increasing abundance of fossiliferous black and gray marine shale interbedded with several thin laterally extensive limestones (Cooper, 1930, 1933; de Witt et al., 1993). Hamilton Group deposits, part of the Catskill delta succession, accumulated in an elongate foreland basin that formed in response to the Acadian oblique collision of the Avalonia microplate and Laurentia (Figure 1; Ettensohn, 1985, 1987; Fail, 1985; Ferrill and Thomas, 1988; Rast and Skehan, 1993). Throughout much of the basin, the Middle Devonian Marcellus Formation (Hall, 1839), the basal unit of the Hamilton Group, comprises two black shale intervals separated by a sequence of limestone, shale, and lesser sandstone of variable thickness (Clarke, 1903; Cate, 1963; de Witt et al., 1993). Recent advances in well stimulation of, and production from, Devonian–Mississippian black shale gas deposits of the eastern and southern United States, including the Barnett, Fayetteville, and Woodford shales, have yielded similarly promising results in the Marcellus Formation (Engelder et al., 2009). Continued successful exploration and exploitation of the Marcellus necessitates an enhanced understanding of its stratigraphy, including thickness trends and compositional attributes of the component units, throughout the core region of Marcellus production beneath the Appalachian Plateau.

Results reported on in this article are based on our analysis of more than 900 wireline logs from the Appalachian Basin of Pennsylvania, New York, northern West Virginia, eastern Ohio, and western Maryland (Figure 2). We focus on the Appalachian Plateau region of the basin for three reasons: its greater density of available wireline logs, fewer structural complications, and greater economic viability. Specific points addressed in this study include (1) the distribution and thickness trends of the two black shale members of the Marcellus Formation; (2) the distribution and thickness of the intervening limestone, an interval that could be critical to stimulation and production considerations; and (3) the stratigraphy and distribution of the organic-rich Levanna Shale Member of the Skaneateles Formation, a unit that has been confused with the Marcellus Formation.

As important as the previous points are, however, the more significant contribution of this article is a sequence-stratigraphic framework of the Marcellus Formation as deduced from publicly available wireline logs. Partington et al. (1993) and Emery and Myers (1996), among others, have demonstrated the use of some of the more common wireline log suites to the interpretation of sedimentary successions in terms of such sequence-stratigraphic elements as sequence boundaries, systems tracts, condensed sections, and maximum flooding surfaces. Such an approach serves as a means by which basin fill can be organized into unconformity (or equivalent conformable surface) bounded packages of strata that provide a framework for predictive reservoir assessment and correlation into regions of minimal or poor data control. Thickness trends of lithostratigraphic units and the sequence stratigraphy of the Marcellus Formation reveal a basin that was more tectonically active than heretofore realized. Reactivated extensional basement structures, including Eocambrian faults associated with the Rome trough, and northwest-striking wrench faults (i.e., cross-strike structural discontinuities of Wheeler, 1980), appear to have controlled sedimentation patterns of at least the upper Lower through lower Middle Devonian succession, including the Marcellus Formation, in western New York and northwest Pennsylvania.

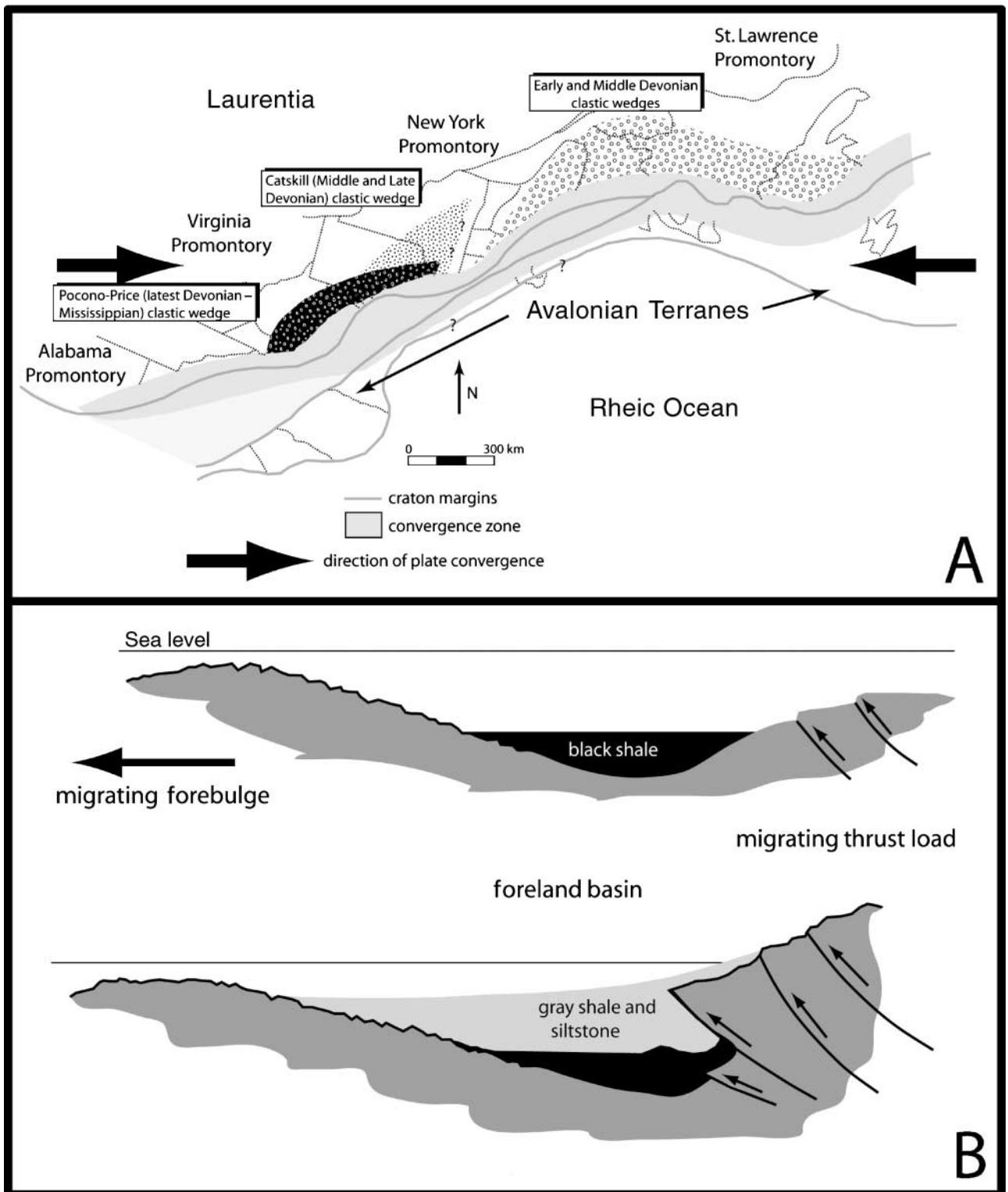


Figure 1. (A) General tectonic reconstruction of the Appalachian orogenic belt during the Acadian orogeny showing approximate locations of synorogenic deposits. Modified from Ettensohn (1992) and Ferrill and Thomas (1988). (B) Diagram illustrating the inferred relationship among thrust loading, foreland basin subsidence, black shale sedimentation, and forebulge development. Modified from Ettensohn, 1994.

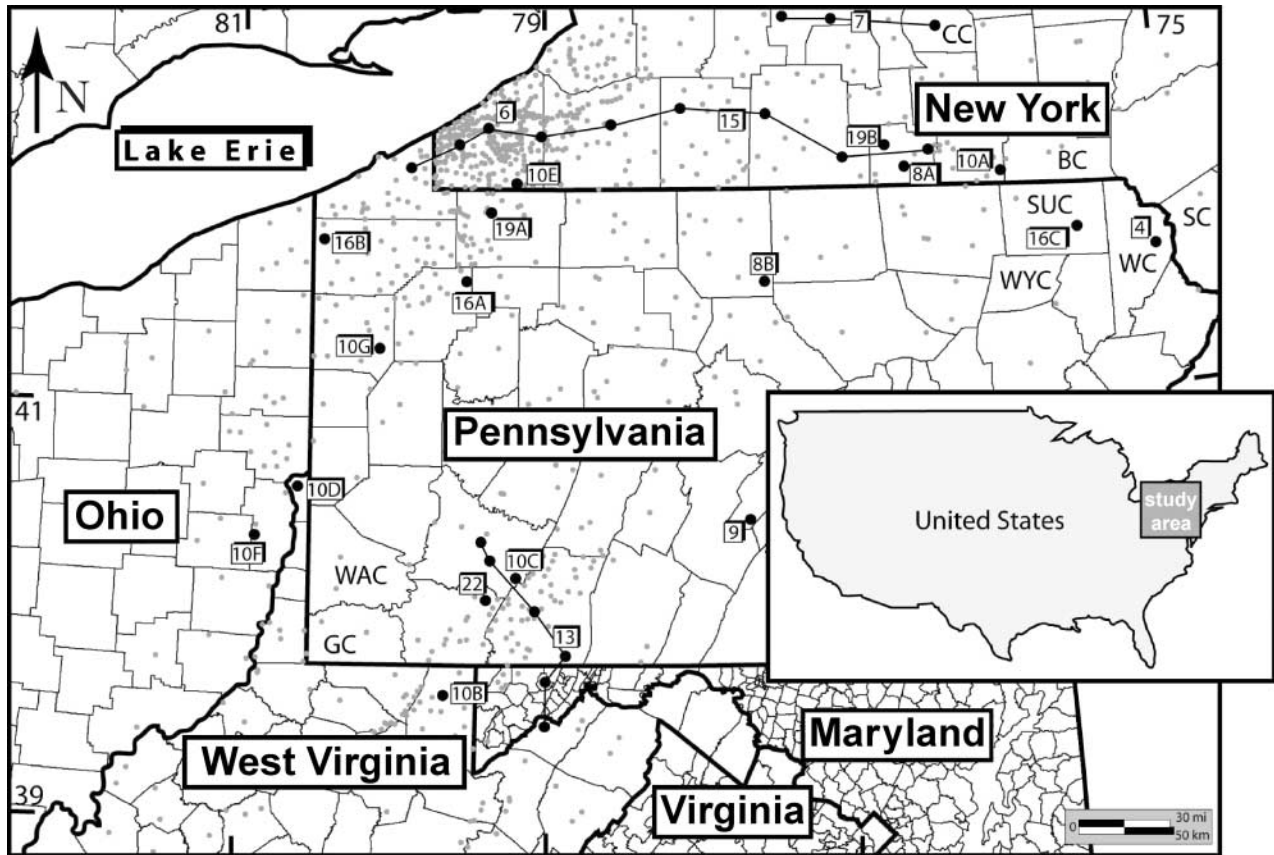


Figure 2. Base map of the core region of the Marcellus Formation basin in New York, Pennsylvania, eastern Ohio, western Maryland, and northern West Virginia. Symbols indicate the locations of wireline logs that were used in this study; numbered and lettered wells and cross-sections refer to text figure numbers. WC = Wayne County, Pennsylvania; SUC = Susquehanna County, Pennsylvania; WYC = Wyoming County, Pennsylvania; WAC = Washington County, Pennsylvania; GC = Greene County, Pennsylvania; SC = Sullivan County, New York; CC = Cayuga County, New York; BC = Broome County, New York.

central and western New York and eastern Ohio			eastern New York			central and eastern Pennsylvania			this study		
Marcellus Subgroup	Oatka Creek Formation	Berne Member	Marcellus Subgroup	Mount Marion Formation	Solsville and Pecksport members	Marcellus Formation	Dalmatia (Fisher Ridge) Member	Marcellus Formation	Oatka Creek Member		
		Cherry Valley Member Hurley Member			Berne Member		Purcell (Turkey Ridge) Member		Cherry Valley Member		
Marcellus Subgroup	Union Springs Formation	Bakoven Member	Marcellus Subgroup	Union Springs Formation	Stony Hollow Member Bakoven Member	Marcellus Formation	Shamokin Member	Marcellus Formation	Union Springs Member		

Figure 3. Comparison of the Marcellus Formation stratigraphy used in this study with the recently revised Marcellus stratigraphic nomenclature of Ver Straeten and Brett (2006).

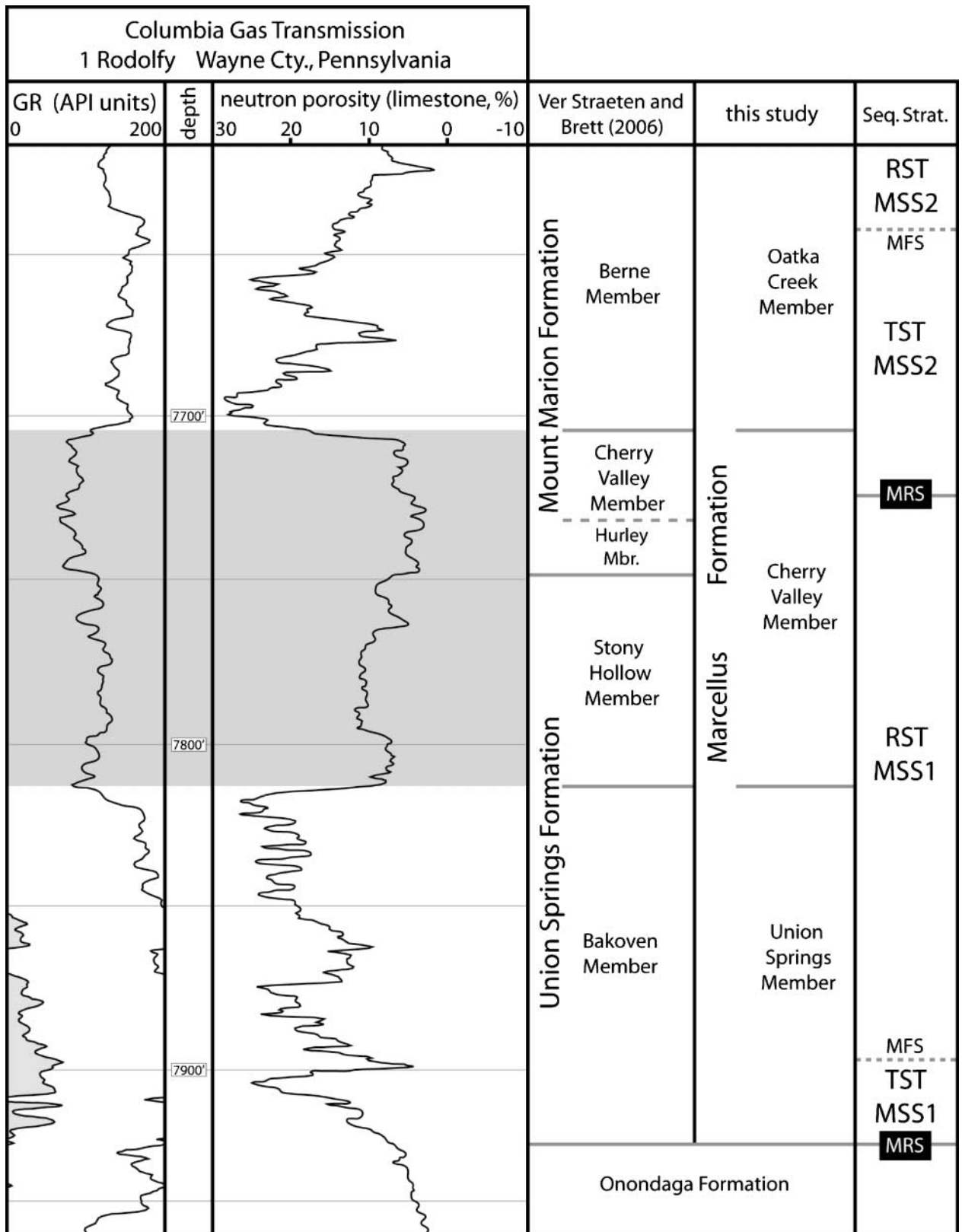


Figure 4. Wireline log of the Rodolfy well in Wayne County, Pennsylvania, illustrating (1) the Marcellus stratigraphy of Ver Straeten and Brett (2006), (2) the lithostratigraphy used in this study, and (3) the sequence stratigraphic interpretation of the log (refer to text for discussion). See Figure 2 for location of the Rodolfy well. RST = regressive systems tract; MFS = maximum flooding surface; TST = transgressive systems tract; MRS = maximum regressive surface.

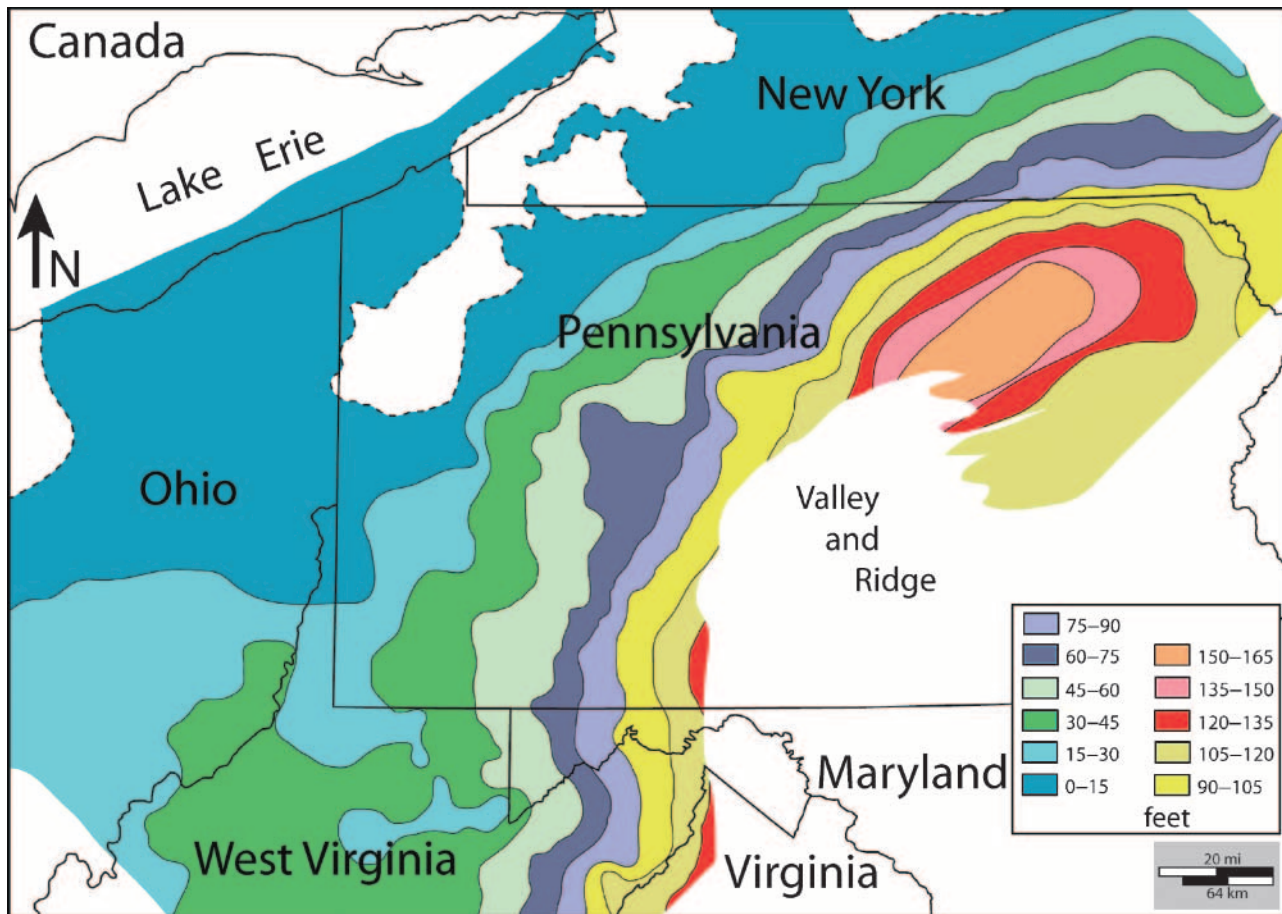


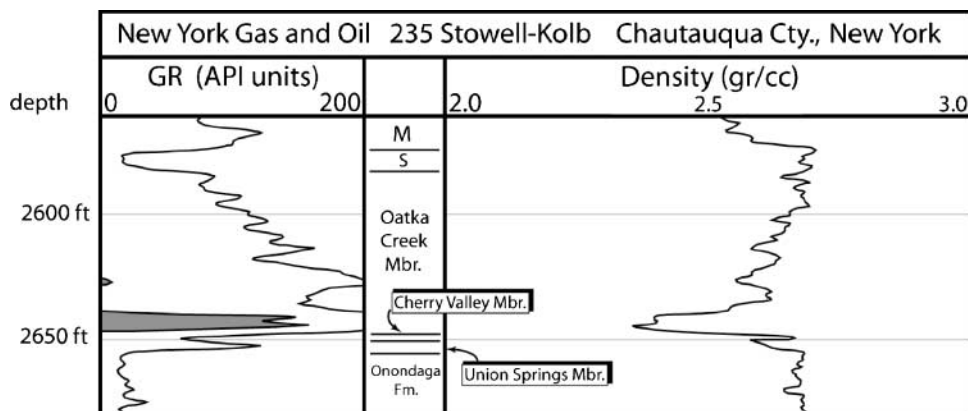
Figure 5. Isochore map of the Union Springs Member. The dashed isochore represents the mapped zero line of the Union Springs Member; elsewhere, contouring is limited by a lack of data. Isochore lines were not traced into the valley and ridge because of sparse data and structural complexities.

STRATIGRAPHIC FRAMEWORK

James Hall (1839) was the first to apply the name “Marcellus shale” to organic-rich black and gray

shale exposed near the village of Marcellus, Onondaga County, New York. Henry Darwin Rogers, a contemporary of Hall, referred to Marcellus equivalents exposed in Pennsylvania as the “Cadent Lower

Figure 6. Wireline log signature of a very thin Union Springs Member as exhibited in the Stowell-Kolb well, Chautauqua County, New York; M = Levanna Member of the Skaneateles Formation; S = Stafford Member of the Skaneateles Formation.



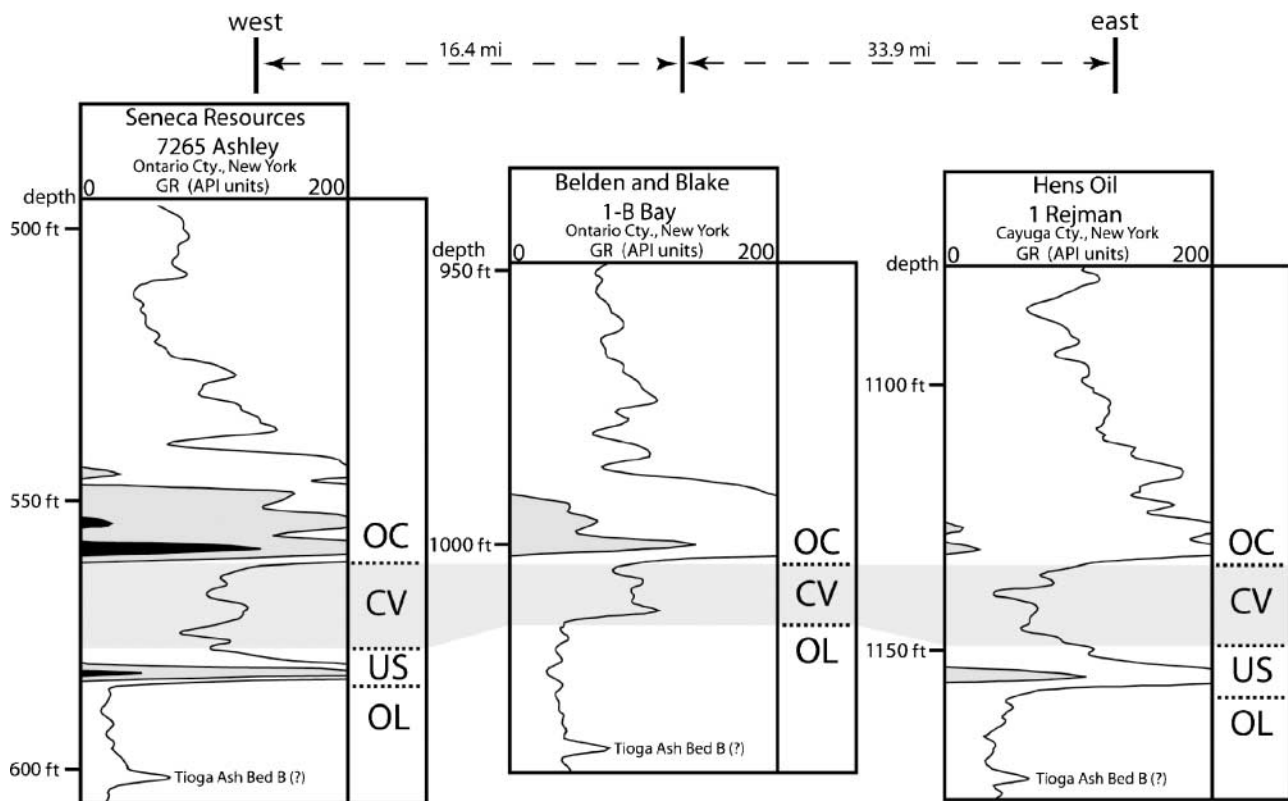


Figure 7. Cross-section illustrating lateral variations in the thickness of the Union Springs Member and overlying Cherry Valley Member; OL = Onondaga Limestone; US = Union Springs Member; CV = Cherry Valley Member; OC = Oatka Creek Member; Depth = drilling depth.

Black and Ash-Colored Slate” in his report of the First Pennsylvania Geological Survey started in 1836 (Millbrooke, 1981). Eighty years lapsed before Cooper (1930) subdivided the Marcellus Formation into the Union Springs Member and overlying Oatka Creek Member. This nomenclature, or some modification of it, largely stemming from outcrop investigations in New York, has been adopted by most workers who have studied the lower Hamilton Group in the subsurface of New York, Pennsylvania, and Ohio (e.g., Oliver et al., 1969; Van Tyne, 1983; Rickard, 1984, 1989).

In a series of articles 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 yr 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 basal black shale to nearshore sandstone and fluvial deposits. Specifically, Ver Straeten and Brett (2006) raised the Marcellus Formation to the subgroup level and the Union Springs and overlying Oatka Creek members to the formation level (Figure 3). Such a modification is consistent with the nomenclature of the overlying units of the Hamilton Group, including the Skaneateles Formation, Ludlowville Formation, and Moscow Formation, in ascending order.

The newly defined Union Springs Formation is principally composed of black shale of the Bakoven Member (Ver Straeten and Brett, 2006). The organic-rich Bakoven Member of eastern New York is overlain by the Stony Hollow Member of the Union Springs Formation (Figure 3), a succession of calcareous shale, siltstone, and fine-grained sandstone. In western New York, fossiliferous limestone and dark shale of the Hurley Member of the Oatka Creek Formation overlies the Bakoven Member

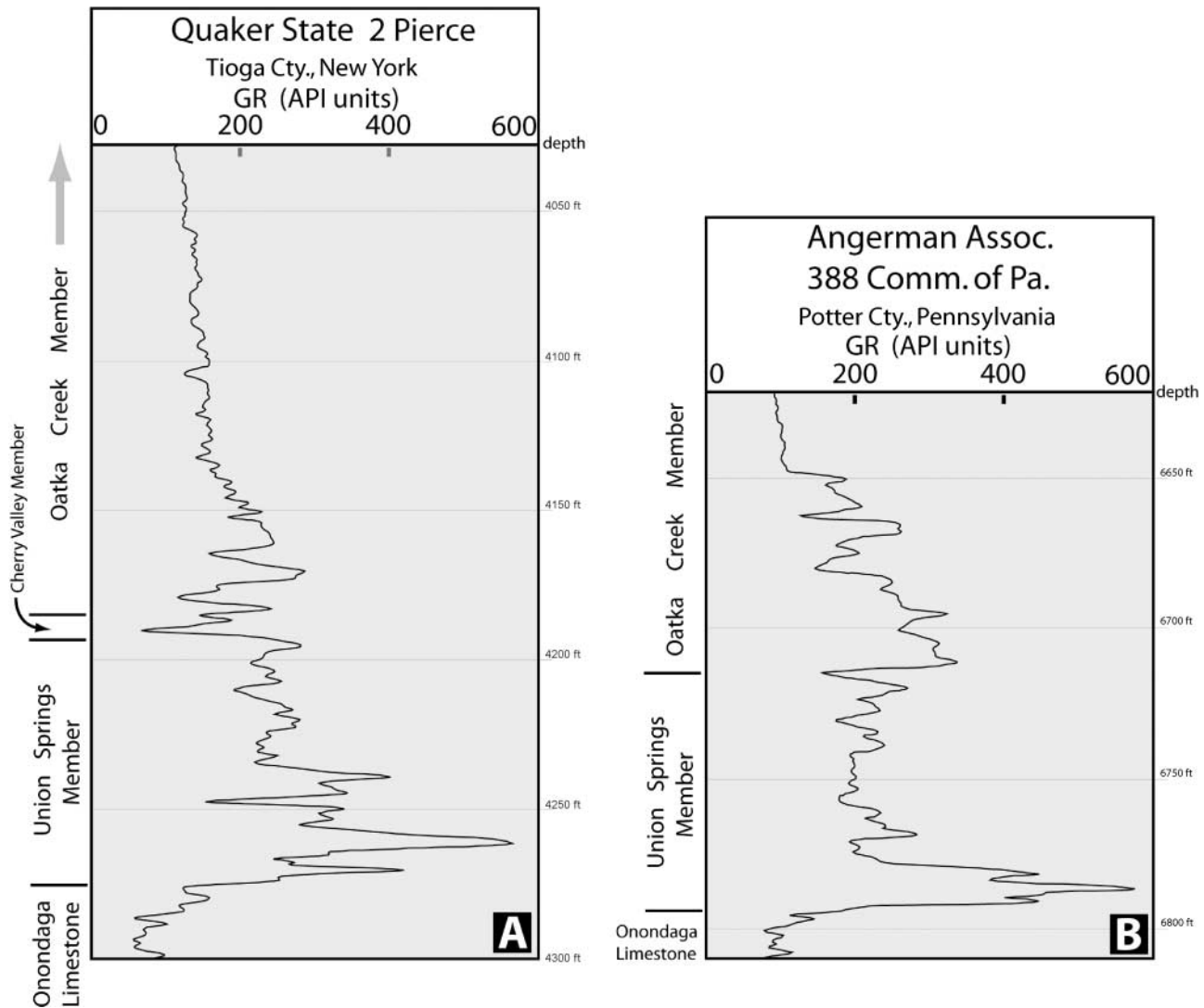


Figure 8. Wireline logs illustrating the gamma-ray signatures of the Union Springs and overlying Oatka Creek members of the Marcellus Formation: (A) Gradational contact of the Cherry Valley and overlying Oatka Creek Member; (B) Union Springs–Oatka Creek contact (Cherry Valley Member is absent).

(Figure 3) (Ver Straeten and Brett, 2006). To the east, the Hurley Member of the Oatka Creek–equivalent Mount Marion Formation overlies the Stony Hollow Member (Figure 3). The Hurley Member is overlain by bedded and nodular fine-grained limestone of the Cherry Valley Member (Ver Straeten et al., 1994; Ver Straeten and Brett, 1995, 2006). The latter thickens to the east and south to as much as 32 ft (10 m) south of Albany, New York, where it is composed of interbedded shale and bioturbated sandstone (Ver Straeten and Brett, 1995). The Purcell Member of Pennsylvania is equivalent to the Cherry Valley Member (Brett

and Ver Straeten, 1995) (Figure 3). The Cherry Valley is sharply overlain by black and gray shale of the Berne Member of the Oatka Creek and Mount Marion formations (Figure 3). The Berne Member of the more proximal eastern region of the basin passes upward into organic-lean strata of the Otsego, Solsville, and Pecksport members, in ascending order (Ver Straeten and Brett, 1995). The top of the Marcellus subgroup is defined by the base of the calcareous Stafford Member of the Skaneateles Formation in western New York and northwestern Pennsylvania and the laterally equivalent Mottville Member of central New York (Ver Straeten and Brett, 1995).



Figure 9. Interbedded black shale and limestone defining the gradational contact of the Onondaga Formation and overlying Union Springs Member of the Marcellus Formation exposed along the Norfolk Southern mainline, Hamilton-Newton, Pennsylvania. Hammer = 30 cm (11.8 in.) long.

The revised Marcellus stratigraphy is a welcome contribution to our understanding of Appalachian Basin stratigraphy. However, such a detailed level of subdivision, especially its differentiation of grossly similar units such as the Hurley and Cherry Valley members is not readily applicable to basinwide stratigraphic interpretation of publicly available wireline logs of varying quality. Indeed, only one log available to us permits a tentative identification of the Marcellus stratigraphy espoused by Ver Straeten and Brett (2006). The Marcellus Formation displayed by the Columbia 1 Rodolffy well of Wayne County, Pennsylvania, includes upper and lower black shale intervals separated by more than 110 ft (34 m) of calcareous and arenaceous rock (Figure 4). Gamma-ray and neutron porosity signatures permit the subdivision of the middle horizon into upper and lower intervals (Figure 4). The latter is composed of approximately 65 ft (20 m) of what appears to be sandy limestone, the likely equivalent of Ver Straeten and Brett's (2006) Stony Hollow Member of their Union Springs Formation (Figures 3, 4). These deposits are overlain by approximately 45 ft (14 m) of poorly radioactive low-porosity strata that include the Hurley and overlying Cherry Valley members of Ver Straeten and Brett's (2006) Mount Marion Formation (Figures 3,

4). Unfortunately, the paucity of wireline logs from this area of the basin precludes the more widespread application of this stratigraphy.

In this article, we adopt a lithostratigraphy more in line with that used by Rickard (1984, 1989) and one that lends itself to subsurface correlation of wireline log signatures. Specifically, we define our basal unit of the Marcellus Formation as the Union Springs Member, a term recognized by the U.S. Geological Survey Geologic Names Lexicon (USGS, 2008). The Union Springs Member of this study, which encompasses the Bakoven Member of Ver Straeten and Brett (2006), is overlain by the Cherry Valley Member (Figure 3). Our Cherry Valley Member, which is composed of variable amounts of interlayered carbonate, shale, and sandstone, correlates with the Stony Hollow Member of the Union Springs Formation and the Hurley and Cherry Valley members of the Oatka Creek and Mount Marion formations of Ver Straeten and Brett (2006) (Figure 3). Finally, we use the name Oatka Creek Member, also recognized by the Geologic Names Lexicon, for the succession of black and gray shale and lesser siltstone and limestone that underlies the Stafford and Mottville members of the Skaneateles Formation (Figure 3).

MARCELLUS FORMATION: SUBSURFACE LITHOSTRATIGRAPHY AND THICKNESS TRENDS

The subsurface stratigraphy of the Marcellus Formation has been addressed by a number of workers. Harper and Piotrowski (1978, 1979) and Piotrowski and Harper (1979) published isopach maps of Devonian units in Pennsylvania, including the Marcellus Formation. Their maps present the net thickness of the radioactive facies of the Hamilton Group, principally the Marcellus Formation. Rickard (1984, 1989) mapped the Marcellus in the subsurface of New York, northeastern Ohio, and the northern two thirds of Pennsylvania. He presented isopach maps for what he termed the lower Marcellus Formation (Union Springs Member–Cherry Valley Member interval), the upper Marcellus Formation (Oatka Creek Member and equivalent units), and the total

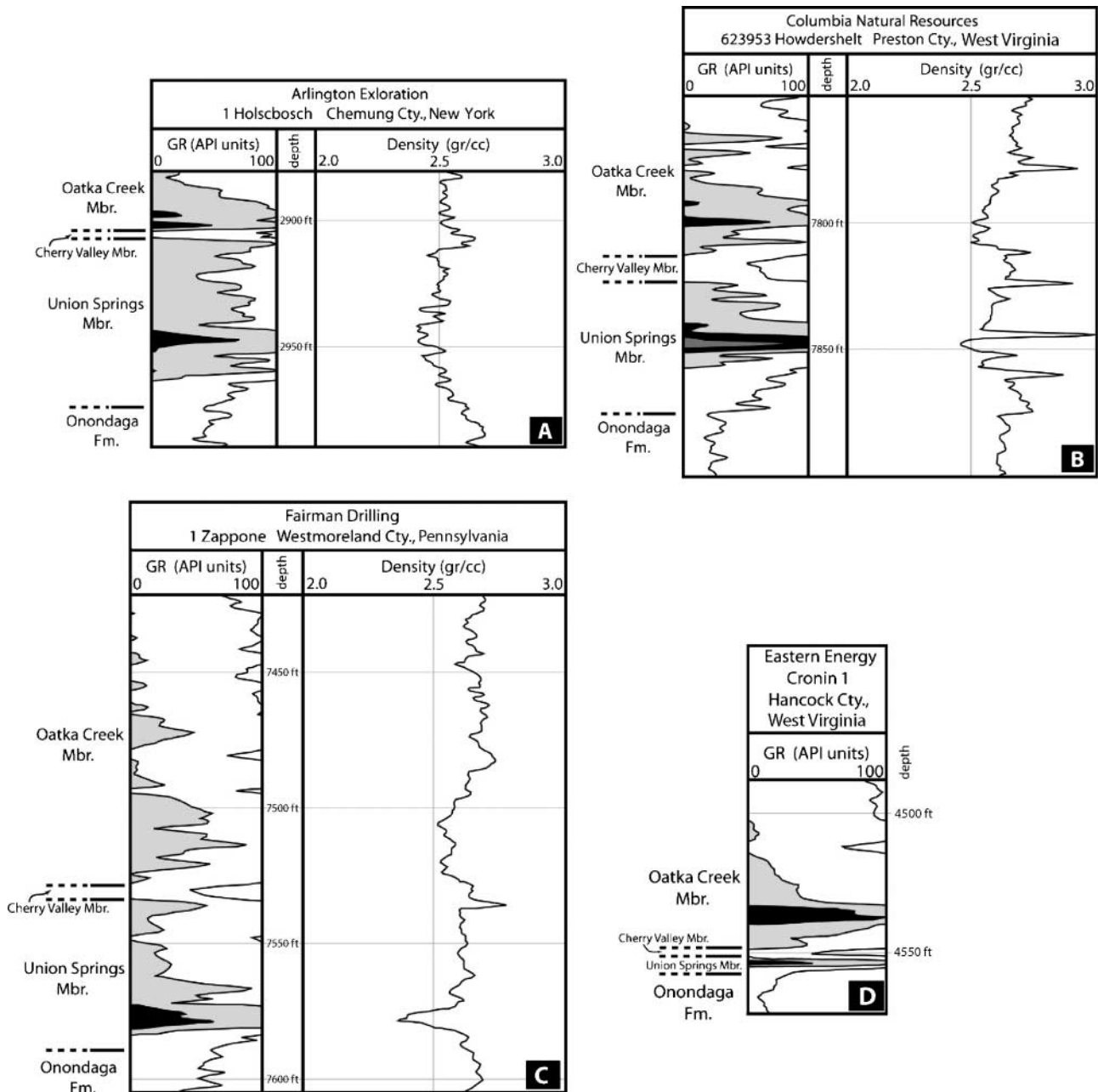


Figure 10. Variations in the Onondaga Limestone–Marcellus Formation contact across the Appalachian Basin; (A–C) gradational Onondaga–Union Springs contact; (D) sharp Onondaga–Union Springs contact; (E–G) = sharp Onondaga–Oatka Creek contact. Note that the wireline log D is a gamma-ray log only.

Marcellus Formation. Last, a series of stratigraphic cross-sections and isopach maps of the Devonian clastic succession of the Appalachian Basin, including the Marcellus Formation, were published as part of the Eastern Gas Shales Project of the late 1970s and early 1980s (e.g., de Witt et al., 1975; West, 1978; Roen et al., 1978; Kamakaris and Van Tyne, 1980).

Union Springs Member

The Union Springs Member of this report thickens to the east and southeast from western New York; it is especially thick in northeastern Pennsylvania where it exceeds 160 ft (49 m) (Figure 5). Particularly intriguing, though, is the local absence of the Union Springs along a northeast-southwest-trending

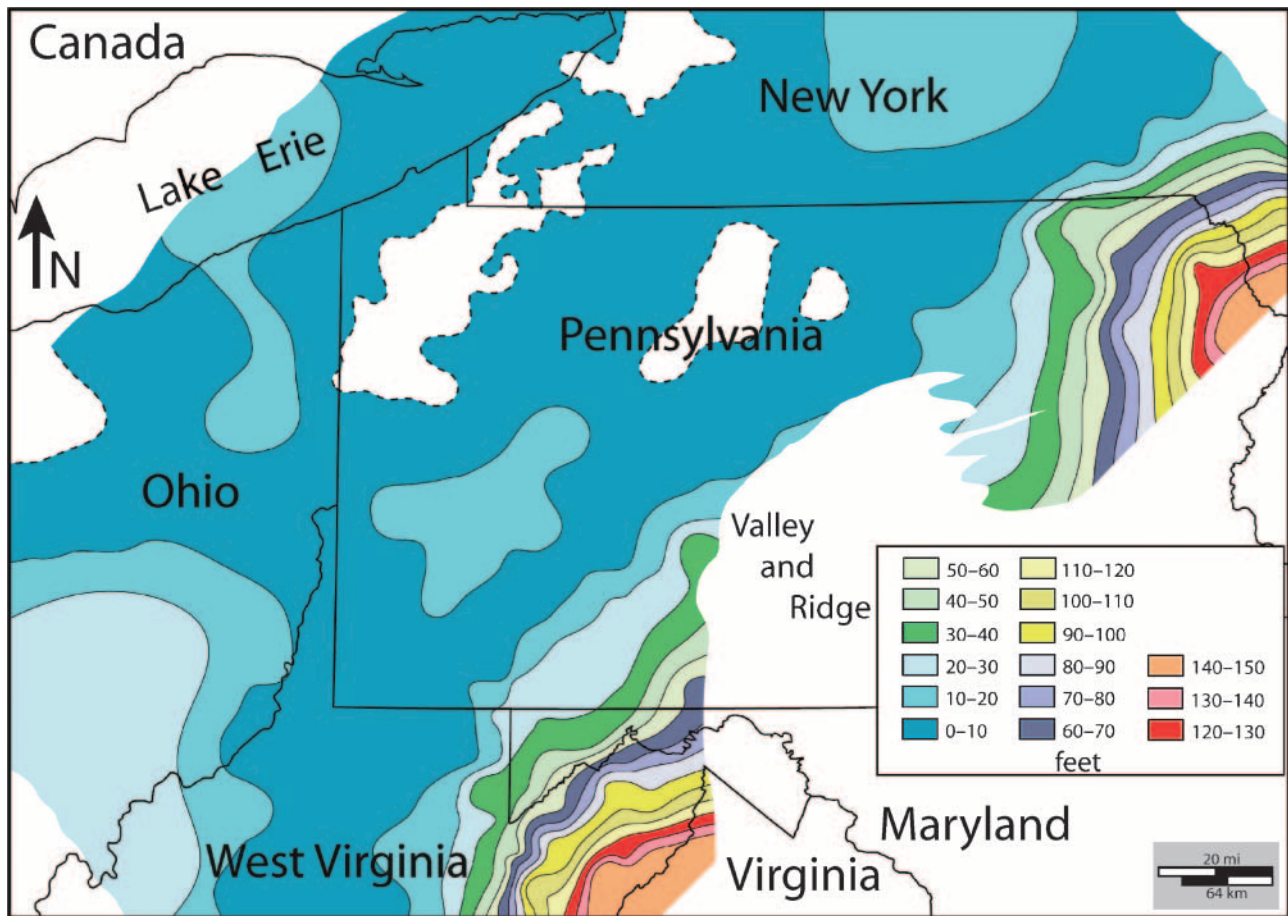


Figure 11. Isochore map of the Cherry Valley Member. The dashed isochore represents the mapped zero line of the Cherry Valley Member; elsewhere, contouring is limited by a lack of data. Isochore lines were not traced into the valley and ridge because of sparse data and structural complexities.

Union Springs Member (Figure 4), the Onondaga–Marcellus contact is unconformable. The Onondaga–Union Springs contact appears to cut progressively deeper into the Onondaga Limestone from central New York eastward to the Hudson Valley (Rickard, 1989). Oliver (1954, 1956), however, described the exposed contact in Cayuga County, New York, as an approximately 7-ft (2.1 m)-thick interval of interbedded shale and limestone, suggestive of a more gradational contact. Moreover, gamma-ray and bulk density logs reveal a seemingly gradational contact extending across central and eastern New York and south through Pennsylvania into northern West Virginia (Figure 10A, B, C). However, the Onondaga–Union Springs contact of the West Virginia northern panhandle and eastern Ohio is very sharp, although not necessarily unconformable (Figure 10D).

Cherry Valley Member

The Union Springs Member of western New York is overlain by the Cherry Valley Member, the correlative of the Purcell Limestone of Pennsylvania and West Virginia, an interval of bedded and nodular limestone, shale, and siltstone (Cate, 1963; Dennison and Hasson, 1976). As noted previously, the Cherry Valley Member of this report includes the interval encompassing the Stony Hollow Member of the Union Springs Formation of Ver Straeten and Brett (2006) and the Hurley and Cherry Valley members of Ver Straeten and Brett’s (2006) Mount Marion Formation (Oatka Creek Formation equivalent) of the more proximal eastern region of the basin (Figure 2).

The Cherry Valley Member increases from less than 10 ft (3 m) thick in western New York and

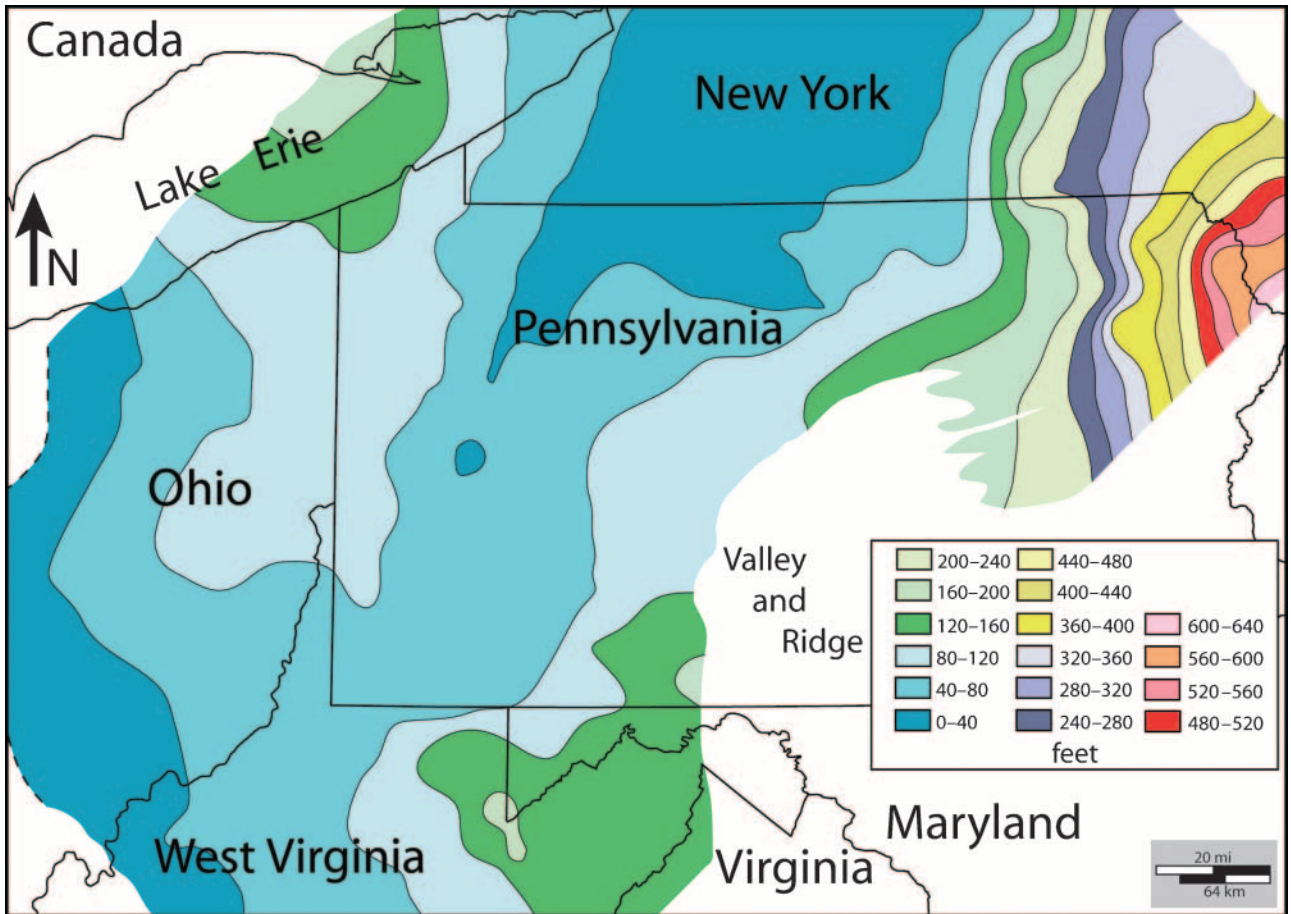


Figure 12. Isochore map of the Oatka Creek Member. The dashed isochore represents the mapped zero line of the Oatka Creek Member; elsewhere, contouring is limited by a lack of data. Isochore lines were not traced into the valley and ridge because of sparse data and structural complexities.

northwestern Pennsylvania to more than 140 ft (43 m) thick in Wayne County, northeastern Pennsylvania, and Sullivan County, southeastern New York, as well as northeastern West Virginia (Figure 11). The rate of change of thickness is especially rapid in southeastern New York, a likely reflection of the presence of Ver Straeten and Brett's (2006) Stony Hollow Member (Figures 4, 11). Indeed, gamma-ray and bulk density log signatures indicate that the Cherry Valley interval becomes more arenaceous to the east (see Figure 4), an observation consistent with field relationships documented from southeastern New York (Ver Straeten et al., 1994). Moreover, an ongoing coring program to sample the Marcellus Formation in the Pennsylvania Valley and Ridge confirms the presence of an arenaceous Purcell Limestone in this region of the basin.

Although present throughout much of the basin, the Cherry Valley Member displays rather irregular thickness trends (Figure 11), similar to what can be observed at the outcrop scale (e.g., Ver Straeten et al., 1994). Note that the Cherry Valley is absent along a northeast-southwest-trending region of western New York and northwestern Pennsylvania, coincident with that area of the basin from which the Union Springs Member is thin or absent (compare Figures 5, 11). Locally, the Cherry Valley Member overlies the Onondaga Limestone, the intervening Union Springs Member absent because of erosion or nondeposition (Figure 7). In this case, the Cherry Valley is recognized as a plateau on the gamma-ray signature immediately beneath the radioactive basal interval of the overlying Oatka Creek Member (Figure 7).

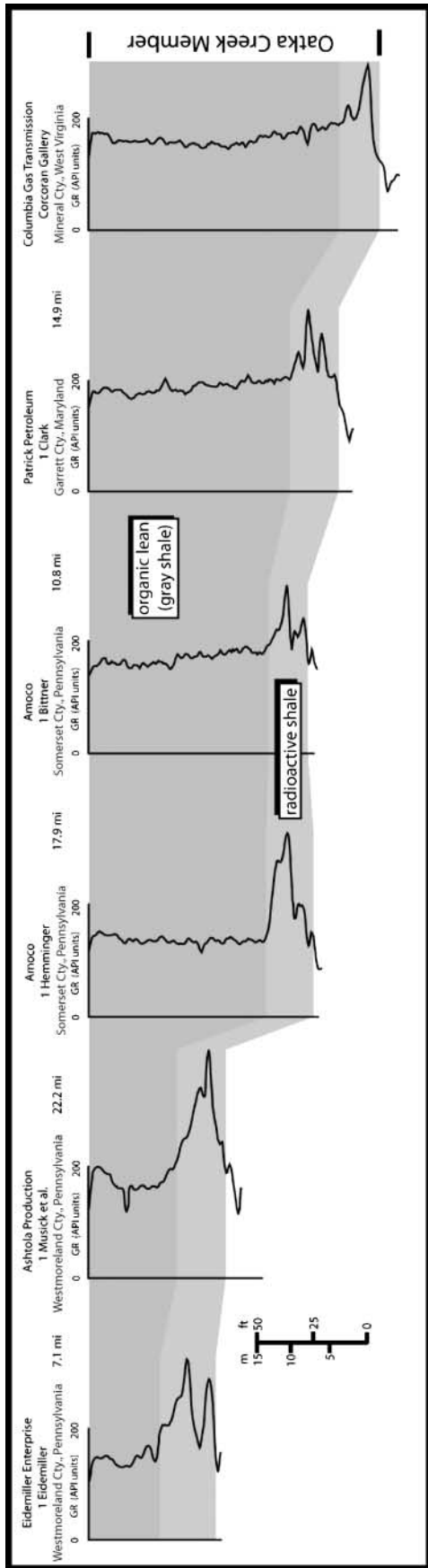


Figure 13. Cross-section illustrating eastward thickening of the Oatka Creek Member. The Oatka Creek is subdivided into an organic-rich (radioactive) basal interval overlain by organic-lean shale (low radioactivity). The cross-section is hung on the top of the Oatka Creek Member.

Oatka Creek Member

The Cherry Valley Member is overlain by the Oatka Creek Member, recognized by a radioactive basal interval that passes upward into a higher density, less radioactive (lower TOC) shale succession containing occasional carbonate layers (Figure 8A). The organic-rich basal interval of the Oatka Creek is generally less radioactive than the basal interval of the Union Springs Member (Figure 8A, B). The contact of the Oatka Creek and underlying Cherry Valley is readily defined in wireline logs; locally, the contact appears to be gradational (Figure 8A). However, in the absence of the Cherry Valley (or perhaps a limestone interval thinner than tool resolution), the contact is placed a short distance below the gamma-ray peak in the radioactive basal interval of the Oatka Creek Member (Figure 8B). The Oatka Creek rests disconformably on the Onondaga Limestone in those areas of the basin absent in the Union Springs and Cherry Valley members. Indeed, in almost every example studied, well-log signatures indicate a very sharp contact (Figure 10E, F, G), quite unlike the seemingly gradational Onondaga Limestone–Union Springs Member contact described from much of the basin.

The Oatka Creek Member, present throughout the core region of the basin, thickens to the east, most rapidly along a north-south line east of the meridian that defines the western edge of Broome County, New York, and the western boundaries of Susquehanna and Wyoming counties, Pennsylvania (Figure 12). It exceeds 550 ft (168 m) thick in eastern Wayne County, Pennsylvania, into Sullivan County, New York (Figure 12). However, thickening of the Oatka Creek Member is manifested principally by a marked increase in the thickness of the organic-lean upper part of the unit (Figure 13). That is, the basal radioactive interval displays an eastward increase in thickness, but at a rate far less than that of the overlying higher density (organic-lean) shale.

The Oatka Creek Member thins to less than 30 ft (9 m) along a northeast-southwest-oriented axis in western New York, extending into Pennsylvania (Figure 12). This area is displaced to the east of the similarly oriented region of the basin over



Figure 14. Contact of the Oatka Creek Member, Marcellus Formation, and the overlying Stafford Member of the Skaneateles Formation exposed on Oatka Creek, village of LeRoy, New York. TOC of the Oatka Creek Member immediately beneath the contact is 4%. Note heavily jointed nature of the organic-rich shale.

which the Union Springs and Cherry Valley members are absent (compare Figures 5 and 11 with 12). Thinning of the Oatka Creek Member is confined to the organic-lean shale interval; that is, there is no concomitant thinning of the radioactive low-density basal interval of the Oatka Creek across this structure (Figure 14). Indeed, the organic-rich facies of the Oatka Creek Member thickens across the region over which the unit thins (Figure 15).

Stafford and Levanna Members, Skaneateles Formation

The upper contact of the Oatka Creek Member is defined by the base of the Stafford (Mottville) Member, an easily recognized marker in the subsurface at the base of the Skaneateles Formation (Oliver et al., 1969; de Witt et al., 1993) (Figure 16A). The bulk of the Stafford is present in a northeast-southwest-oriented region of western New York and northwestern Pennsylvania; however, outliers of the Stafford are present in southwestern Pennsylvania and western Maryland (Figure 17). The limestone appears to pass laterally into shale.

The Stafford Member is overlain by the Levanna Member of the Skaneateles Formation, a relatively geographically restricted little-described

carbonaceous shale (Figure 18). The subsurface Levanna Member, unlike the Union Springs and Oatka Creek members of the Marcellus Formation, is recognized by increasing gamma-ray response upsection (Figure 19A). We arbitrarily place the top of the Levanna at the stratigraphically highest gamma-ray peak equal to or greater than 20 API units in excess of a gray shale baseline established in the overlying nonsource succession of the Skaneateles Formation (Figure 19A), an approach similar to that adopted for use by the Eastern Gas Shales Project (de Witt et al., 1993). The most organic-rich facies of the Levanna Member passes laterally into undifferentiated organic-lean shale of the Skaneateles Formation (Figure 19B). Thus, although Levanna outcrops can be observed outside our isochore region, it is likely that these deposits represent the organic-lean lateral equivalent of the Levanna Member mapped in the subsurface. The bulk of the Levanna occupies a generally rectangular region of the basin extending from western New York to southwestern Pennsylvania, thinning rapidly to the east and southeast, and more gradually to the west and southwest (Figure 18).

Although not part of the Marcellus Formation, the Levanna Member supplements the Middle Devonian source rock inventory where it is present.

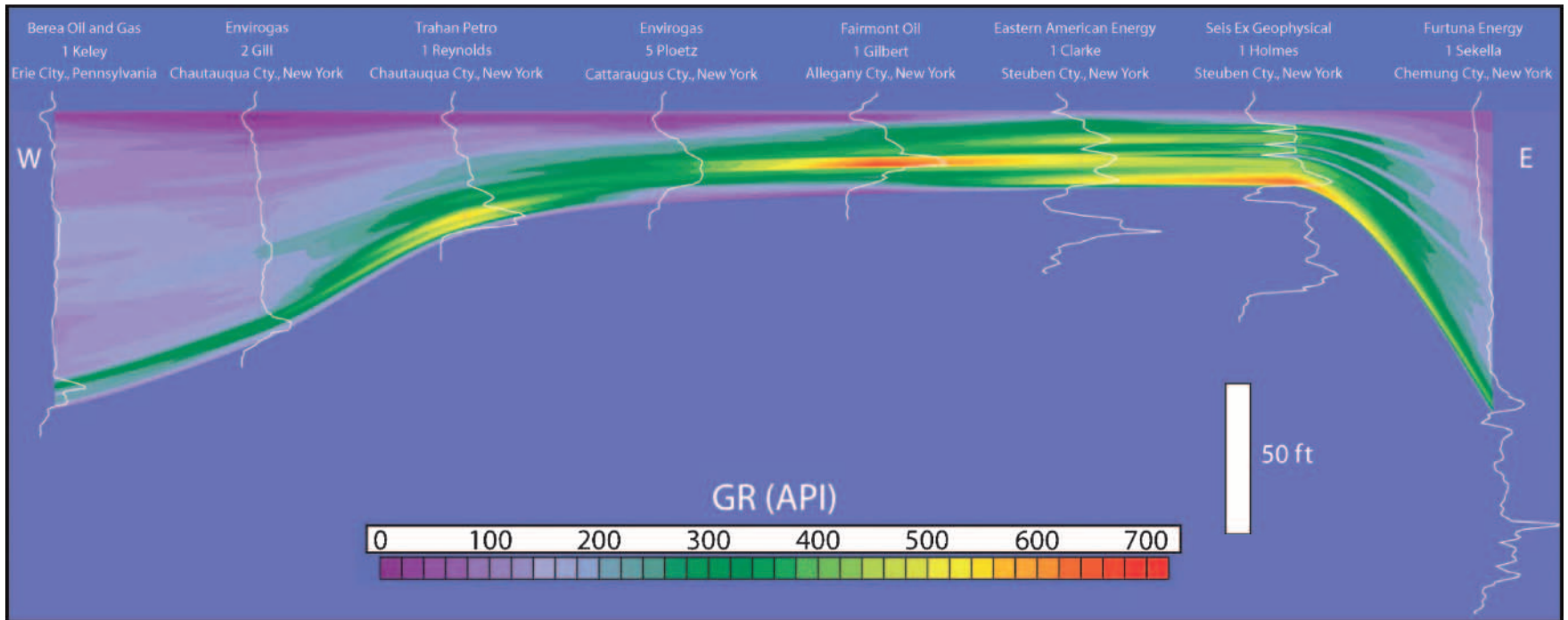


Figure 15. Interpretive fill cross-section of gamma-ray values. Note onlapping of organic-lean deposits on both sides of the central region of thinning.

Indeed, in those regions of the basin lacking the Stafford Member, the Levanna can be mistaken for the Marcellus Formation (de Witt et al., 1993). The organic-rich facies of the Levanna Member appears to pass rapidly eastward into organic-lean low-radioactivity shale (Rickard, 1989); the Levanna thins westward into Ohio where it aids in the recognition of the top of the Oatka Creek. In the absence of the Stafford Member, the Oatka Creek Member–Skaneateles Formation contact is placed at a density maximum and/or gamma-ray minimum (Figure 16B) or a subtle reduction (~10–15 API units) in gamma-ray signature (Figure 16C) that can be traced laterally into the Stafford Member.

SEQUENCE-STRATIGRAPHIC FRAMEWORK OF THE MARCELLUS FORMATION

Mapping of lithostratigraphic units in the subsurface is necessary to basin analysis and exploration considerations. However, it is the application of the sequence-stratigraphic approach that enables one to subdivide basin fill into a framework of systems tracts and bounding and internal surfaces based on depositional models and asymmetric (actualistic) base level curves. The resulting framework of reservoir properties, including compositional attributes, can be used to reduce risk in frontier regions of the basin or areas of poor data control. The use of wireline logs to sequence-stratigraphic analysis has been demonstrated by a number of authors, including Van Wagoner et al. (1990), Embry and Johannessen (1992), Embry (1993, 2002), Partington et al. (1993), Emery and Myers (1996), Brown et al. (2005), and Singh et al. (2008). Moreover, combinations of wireline logs can be used to interpret rock properties, including organic richness, in terms of a sequence-stratigraphic framework (e.g., Passey et al., 1990; Creaney and Passey, 1993).

T-R Sequences: Background

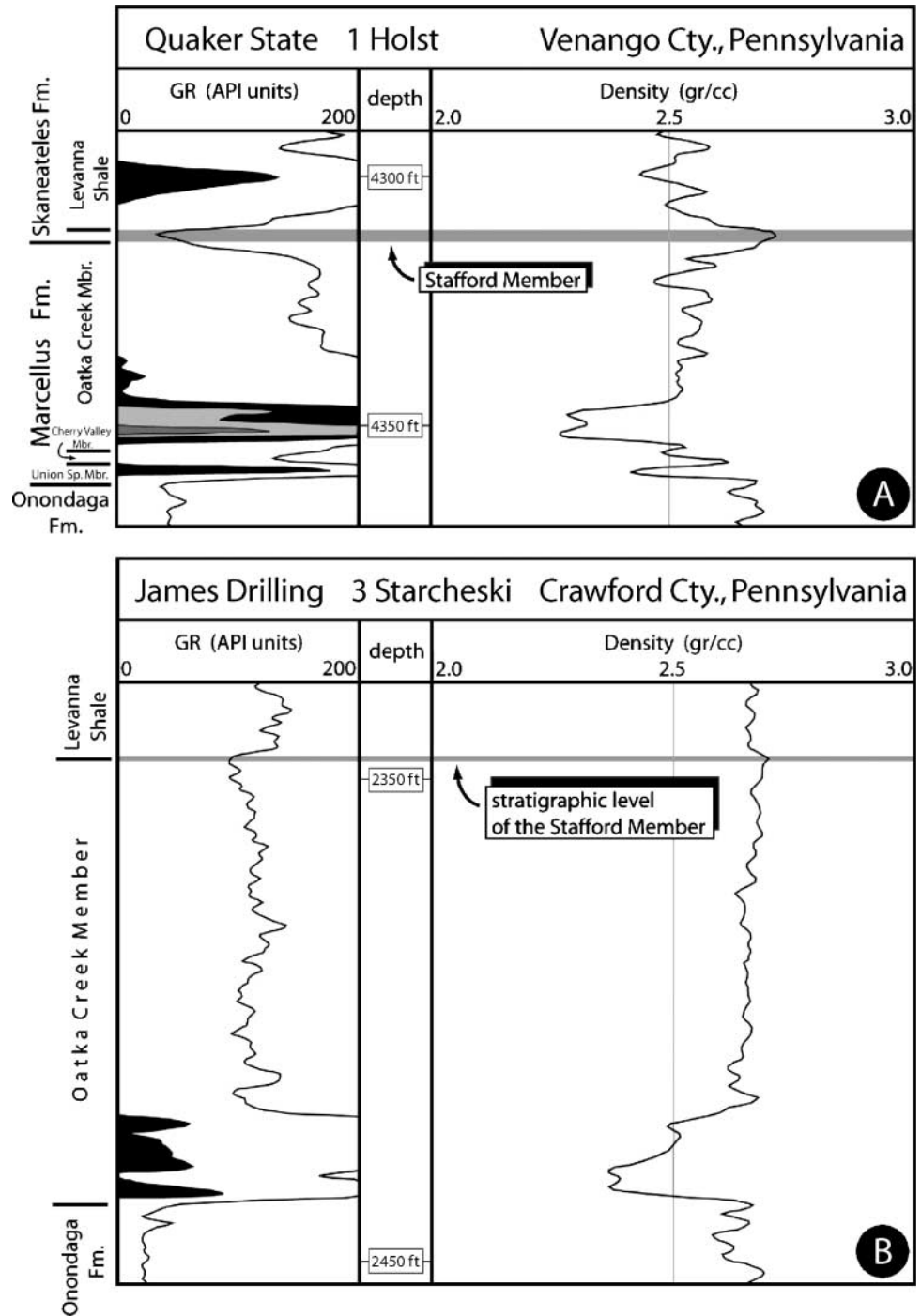
Sequence stratigraphy seeks to define and correlate changes in depositional dynamics that reflect a single base level cycle. This article adopts the T-R sequence described by Johnson et al. (1985) and

Embry and Johannessen (1992) and further refined by Embry (1993, 1995, 2002). Indeed, Johnson et al. (1985) first applied the T-R sequence concept to the Devonian succession of the Appalachian Basin a quarter of a century ago. A single T-R sequence comprises a transgressive systems tract, a deepening-up succession that records rising base level, overlain by regressive systems tract deposits that accumulated during falling base level and consequent reduced accommodation space (Embry and Johannessen, 1992; Embry, 1993, 2002). Delimiting T-R sequences requires the identification of minimally diachronous sequence-boundary surfaces (Embry, 2002; Mancini and Puckett, 2002). Embry (2002) demonstrated that those surfaces most conducive to defining T-R sequences include the subaerial unconformity, the unconformable shoreline ravinement, and the maximum regressive surface.

The previously mentioned surfaces are best thought of in terms of a single asymmetric base-level cycle initiated by a rapid rise in base level (Embry et al., 2007). Early in this base-level cycle, the sediment flux to the shoreline may be high enough so that the shoreline continues to advance toward the basin. Soon, however, the rate of base level rise at the shoreline exceeds the rate of sediment supply, resulting in a landward or transgressive movement of the shoreline. The maximum regressive surface marks this change in depositional regime from regression to transgression (Embry, 2002; Embry et al., 2007). Consideration of asymmetric base level curves defined by a rapid initial rise in base level suggests a coincidence of the maximum regressive surface and the onset of the rise in base level (Embry et al., 2007).

Rapid transgression is accompanied by wave-induced erosion, much of the sediment being transported seaward. The resulting scoured surface, a shoreline ravinement, may or may not have incised through an underlying subaerial unconformity produced during the preceding regression. In the former case, the shoreline ravinement is termed an unconformable shoreline ravinement and is overlain by a deepening-upward marine succession, the transgressive systems tract, which reflects a reduced supply of sediment to the marine environment as the shoreline migrates landward. The presence of

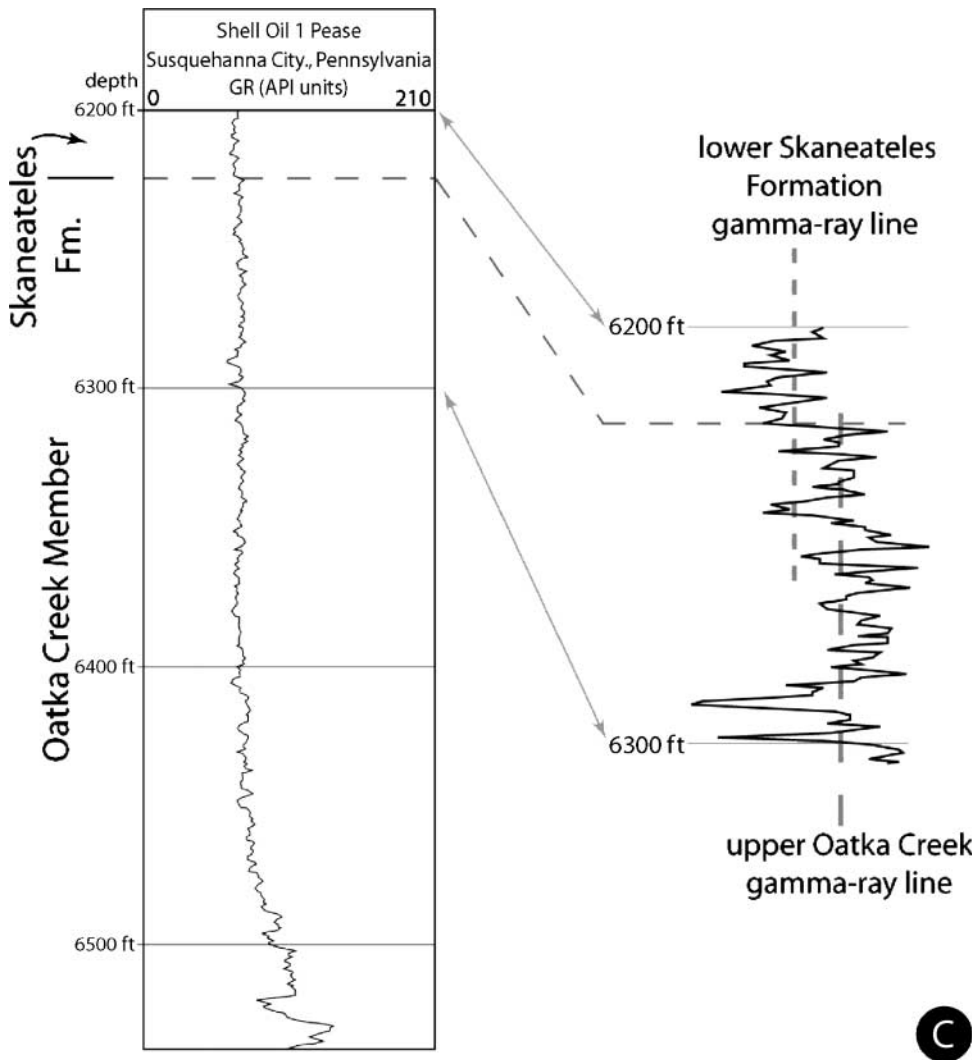
Figure 16. Wireline logs illustrating the nature of the top of the Oatka Creek Member–Skaneateles Formation contact in the presence (A) and absence (B) of the Stafford Member. Panel (C) illustrates the approach taken in this study to define the contact in more proximal regions of the basin.



deepening-upward marine deposits immediately above both the unconformable shoreline ravinement and the maximum regressive surface suggests that these surfaces are minimally diachronous lateral equivalents, the former along the flanks of the basin and the latter in the more basinal conformable succession (Embry, 2002; Embry et al., 2007).

As the rate of base level rise slows to such a point that the sediment flux at the shoreline exceeds the rate of sediment removal by wave action, ravinement development stops and the shoreline begins a regressive seaward migration. Increased sediment supply results in progradation of coarser sediments across the shelf thereby replacing the

Figure 16. Continued.



deepening-upward trend with a shallowing-upward succession that comprises the regressive systems tract. The surface defining the change from deepening-upward strata, the transgressive systems tract, to shallowing-upward deposits of the regressive systems tract is the maximum flooding surface (Embry, 2002; Embry et al., 2007). The maximum flooding surface does not necessarily record the base level maximum. Indeed, regression can be initiated during a base level rise if the rate of rise is less than the rate of sediment supply as might be expected of tectonically driven base level change (Embry et al., 2007).

Increased accommodation space and the reduced clastic sediment flux that attends transgression during rising base level culminates in the maximum flooding surface, which may correspond with,

or pass distally into, a condensed section (Emery and Myers, 1996; Jervey, 1988; Van Wagoner et al., 1988, 1990; Partington et al., 1993). Condensed sections are aerially extensive and can be composed of abundant organic matter (high TOC) and authigenic/diagenetic minerals (Loutit et al., 1988; Posamentier et al., 1988; Sarg, 1988; Liro et al., 1994; Emery and Myers, 1996). Abundant planktonic fossils common to condensed sections (Loutit et al., 1988; Posamentier et al., 1988) reflect the markedly reduced supply of clastic detritus to more basal environments (Partington et al., 1993).

Some have argued that transgressive systems tract deposits, especially the condensed section, have the greatest source rock potential (e.g., Vail, 1987; Emery and Myers, 1996). Indeed, the reduced clastic flux that attends a base level high favors the

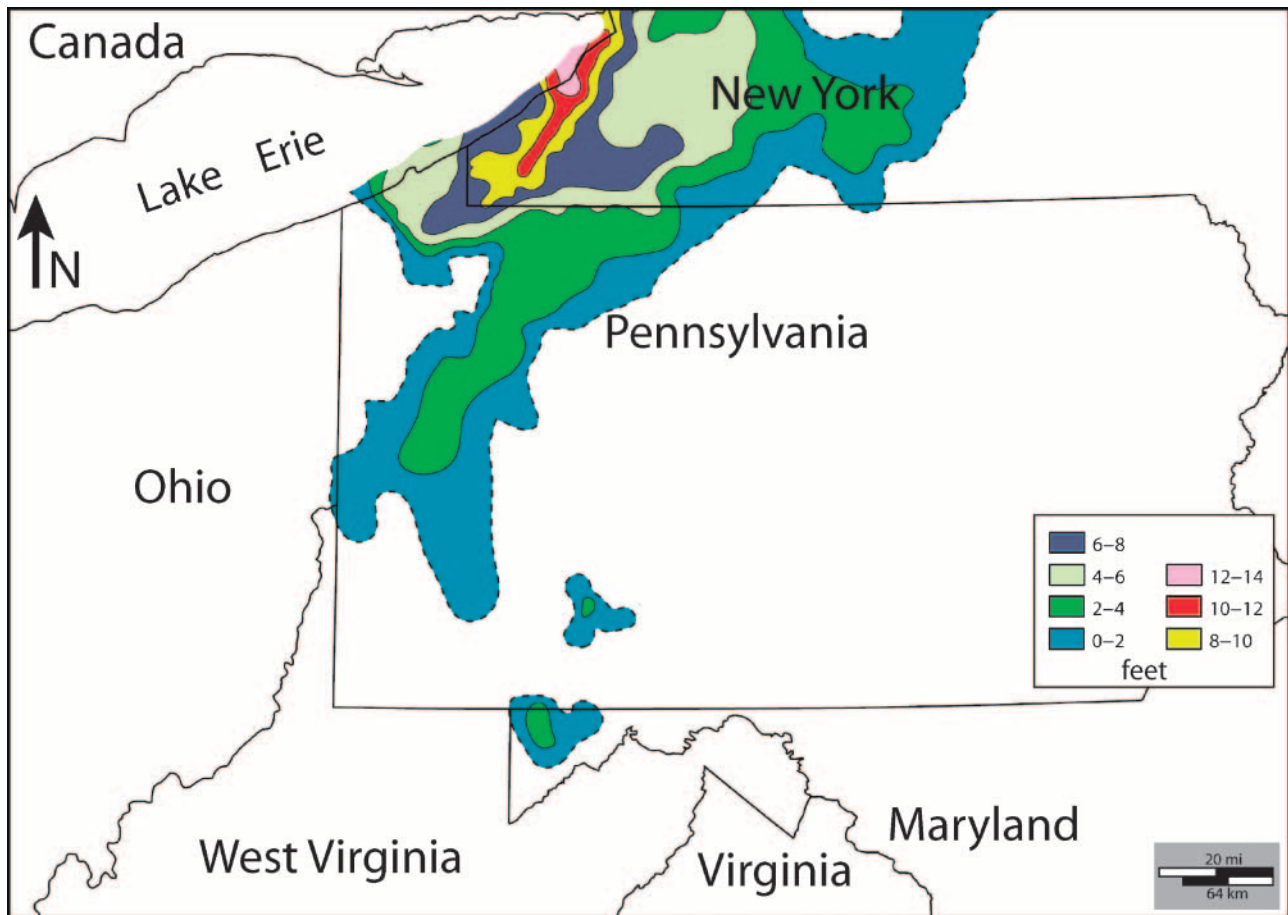


Figure 17. Isochore map of the Stafford Member of the Skaneateles Formation. The dashed isochore represents the mapped zero line of the Stafford Member; elsewhere, contouring is limited by a lack of data. Isochore lines were not traced into the valley and ridge because of sparse data and structural complexities.

concentration of oil-prone marine organic matter (Creaney and Passey, 1993). Moreover, the landward shift of environments accompanying a rise in base level induces the shoreward translation of organic-rich facies, resulting in accumulation of the condensed section close to or at the top of the transgressive systems tract. The described facies shift is manifested by a general increase in TOC upward from the base of the transgressive systems tract (Creaney and Passey, 1993). Still, organic-rich source rocks may continue to accumulate as part of the regressive systems tract as relative base level begins to drop. Further reduction of base level and accommodation space, however, is normally accompanied by an increase in clastic sediment flux and consequent dilution of TOC (Creaney and Passey, 1993).

The general relationship of condensed section deposits and source rock potential, although confirmed by a number of studies (e.g., Lambert, 1993), is somewhat more complex than described (e.g., Mancini et al., 1993). Passey et al. (1991) demonstrated that the greatest source potential interval in the Upper Cretaceous Tocito Sandstone and associated Mancos Shale of the San Juan Basin, New Mexico, is not found in the condensed section within the shale, but rather 82 ft (25 m) below, in the Tocito Sandstone. Similarly, Leckie et al. (1990) demonstrated that the highest source rock potential in a succession of shallow water Cretaceous deposits in western Canada exists within the transgressive systems tract, but below the condensed section. However, Curiale et al. (1991) observed that the most promising source rock interval in the

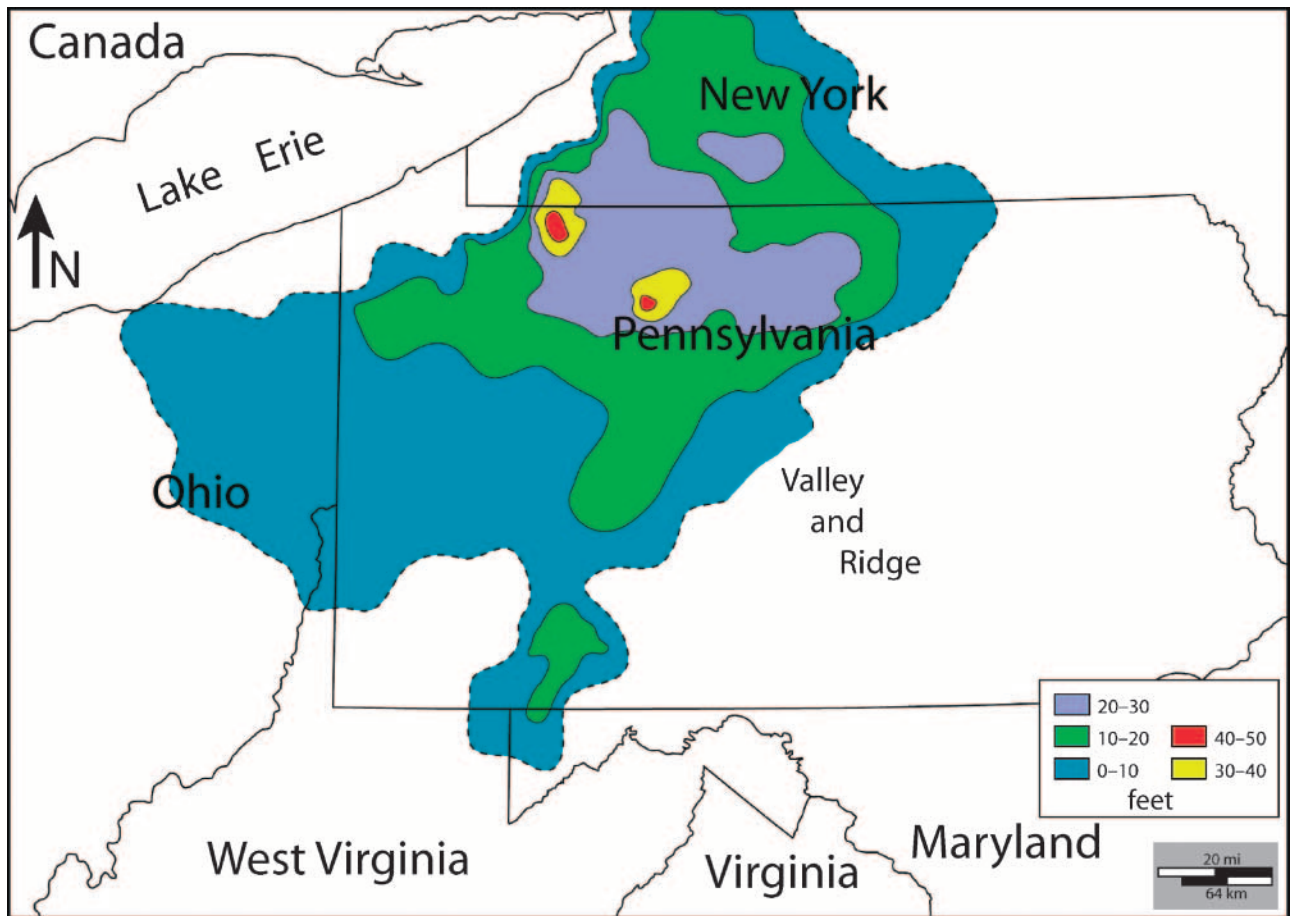


Figure 18. Isochore map of the Levanna Member of the Skaneateles Formation. The dashed isochore represents the mapped zero line of the Levanna Member; elsewhere, contouring is limited by a lack of data. Isochore lines were not traced into the valley and ridge because of sparse data and structural complexities.

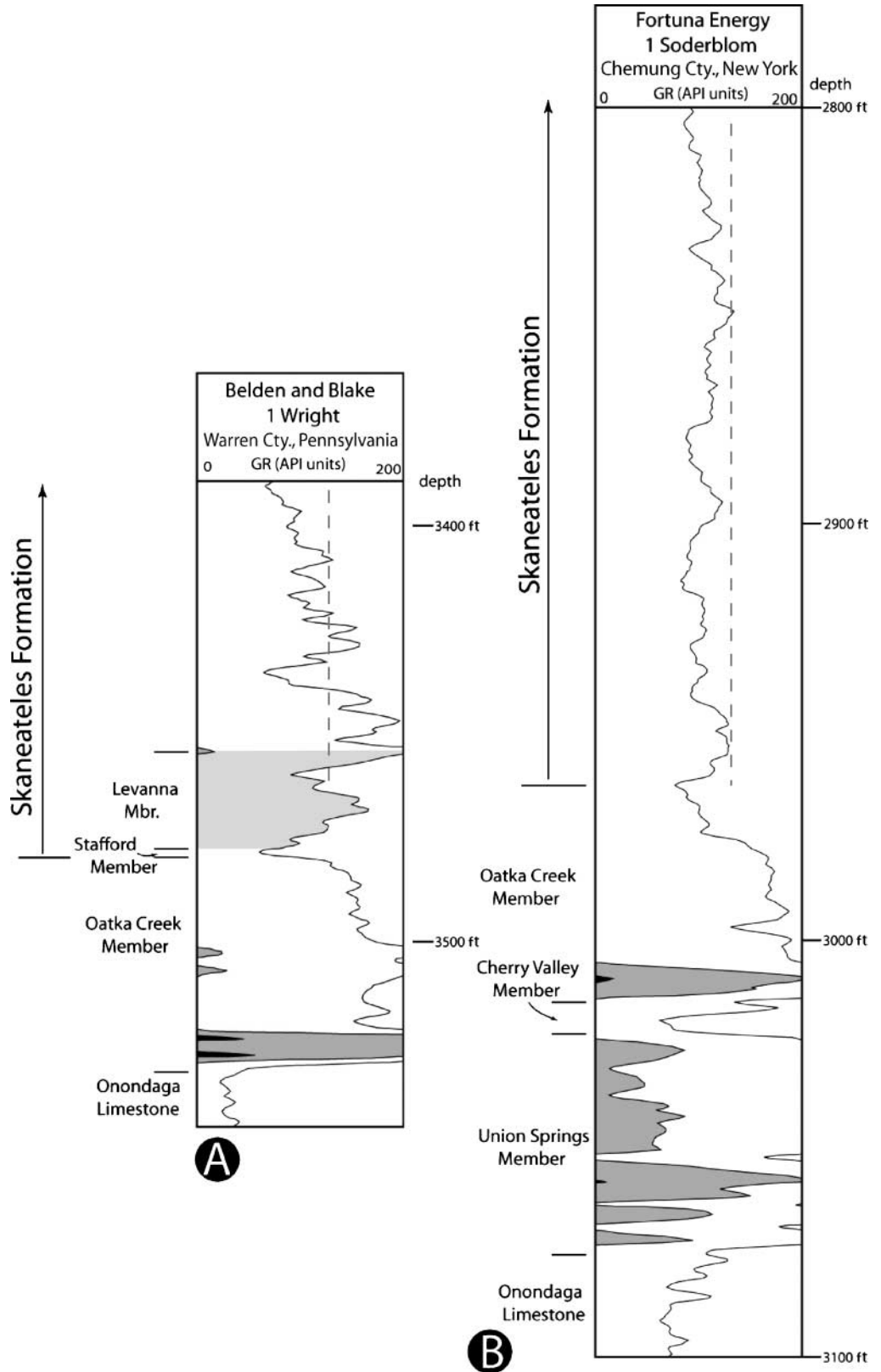
Cenomanian–Turonian succession of New Mexico lies above the condensed interval. Palsey et al. (1991) suggested that in some cases, sedimentation rates within those environments favorable to accumulation of condensed section deposits are too low to preserve abundant hydrogen-rich organic matter.

Discussion

Our sequence-stratigraphic analysis of the Marcellus Formation begins with the generation of a composite wireline log as described by Brown et al. (2005) (Figure 20). Referred to as a site-specific sequence-stratigraphic section (S^5) benchmark chart (Brown et al., 2005), the composite log places the Marcellus into a sequence-stratigraphic framework

that can be used for basinwide correlation. Indeed, our composite wireline log serves as a type section of the Marcellus Formation in the subsurface. The Marcellus encompasses the bulk of two T-R sequences herein referred to as MSS1 and MSS2, in ascending order (Figure 20). These sequences, approximate equivalents of Johnson et al.'s (1985) T-R cycles Id and Ie and Ver Straeten's (2007) Eif-2 and Eif-3 sequences, span approximately 1.8 m.y., extending from the upper "costatus" conodont zone through the "hemiansat" zone (Kaufmann, 2006; Ver Straeten, 2007). The relatively short duration of MSS1 and MSS2 is consistent with their reflecting third-order base level cycles (Mitchum and Van Wagoner, 1991) that occurred within a second-order cycle, encompassing much of the Middle and Upper Devonian succession (Johnson et al., 1985). Embry (1995) advocated a hierarchical system of

Figure 19. Example of the approach used in this study to determine the thickness of the Levanna Member; (A) gamma-ray log displaying the Levanna Member and (B) gamma-ray log from a well lacking the organic-rich Levanna Member. Refer to text for discussion.



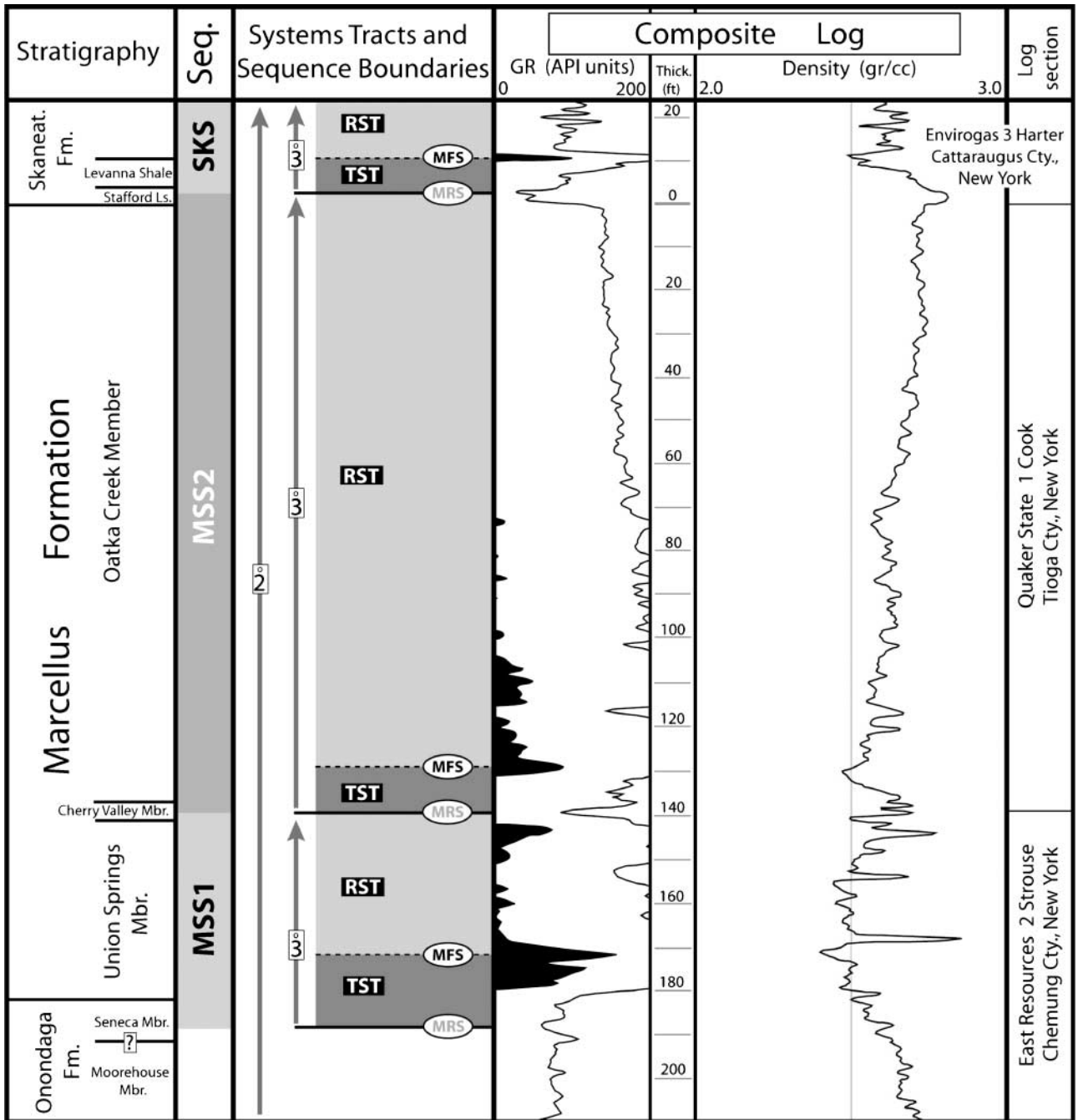


Figure 20. Sequence-stratigraphic type section of the Marcellus Formation that encompasses the upper part of the underlying Onondaga Formation and the lower interval of the Skaneateles Formation. TST = transgressive systems tract; RST = regressive systems tract; MFS = maximum flooding surface; MRS = maximum regressive surface. Refer to text for discussion.

T-R sequences based on sequence boundary characteristics rather than duration. The generally basin-wide 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).

The maximum regressive surface that defines the base of T-R sequence MSS1 is placed at a gamma-ray minimum and/or bulk density maximum close to or at the top of the Onondaga Limestone (Figure 20).

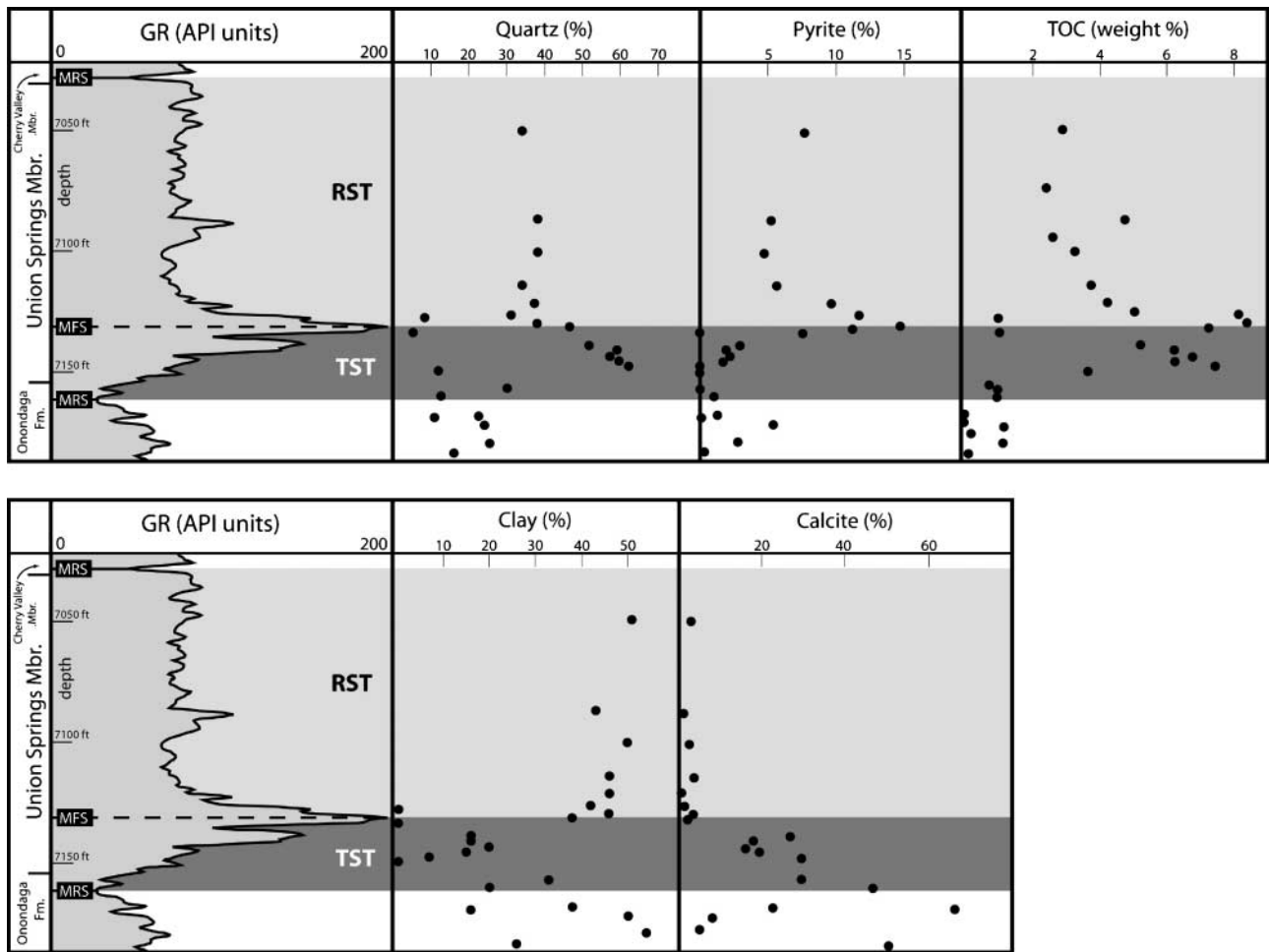


Figure 21. Mineralogy and total organic carbon (TOC) trends through the MSS1 depositional sequence. Data based on a suite of sidewall core samples recovered from a Marcellus Formation well; the location is proprietary. MRS = maximum regressive surface; MFS = maximum flooding surface; RST = regressive systems tract; TST = transgressive systems tract.

Overlying MSS1 transgressive systems tract deposits display upward-increasing (“dirtying”) gamma-ray and upward-decreasing bulk density log signatures (Figure 20), principally reflective of decreasing grain size and increasing TOC (e.g., Singh et al., 2008) related to increasing base level. In the more distal northwestern region of the basin, however, transgressive systems tract deposits appear to be very thin or absent, the contact of the Onondaga Limestone and overlying Union Springs Member being sharp (see Figure 10D). These relations suggest that the rate of base level rise in this region of the basin far exceeded clastic sediment flux.

The top of the MSS1 transgressive systems tract, the maximum flooding surface, is placed at a gamma-ray peak a short distance above the maximum regressive surface (Figure 20). The maxi-

imum flooding surface is roughly coincident with a condensed section defined by abundant pyrite and thin carbonate layers, both evident in bulk density and photoelectric index wireline logs as well as in core. Mineralogic analysis of a suite of sidewall core samples recovered from MSS1 transgressive systems tract deposits reveals an abundance of quartz well in excess of that observed in overlying regressive systems tract deposits (Figure 21). Note that the clay content of the quartz-rich interval is relatively low (Figure 21), likely a reflection of the rapid landward shift of marine environments at this time (e.g., Liro et al., 1994). Thin section and scanning electron microscopic examinations reveal that the bulk of the quartz in the MSS1 transgressive systems tract (condensed section) is microcrystalline, likely derived from the dissolution of silica tests.

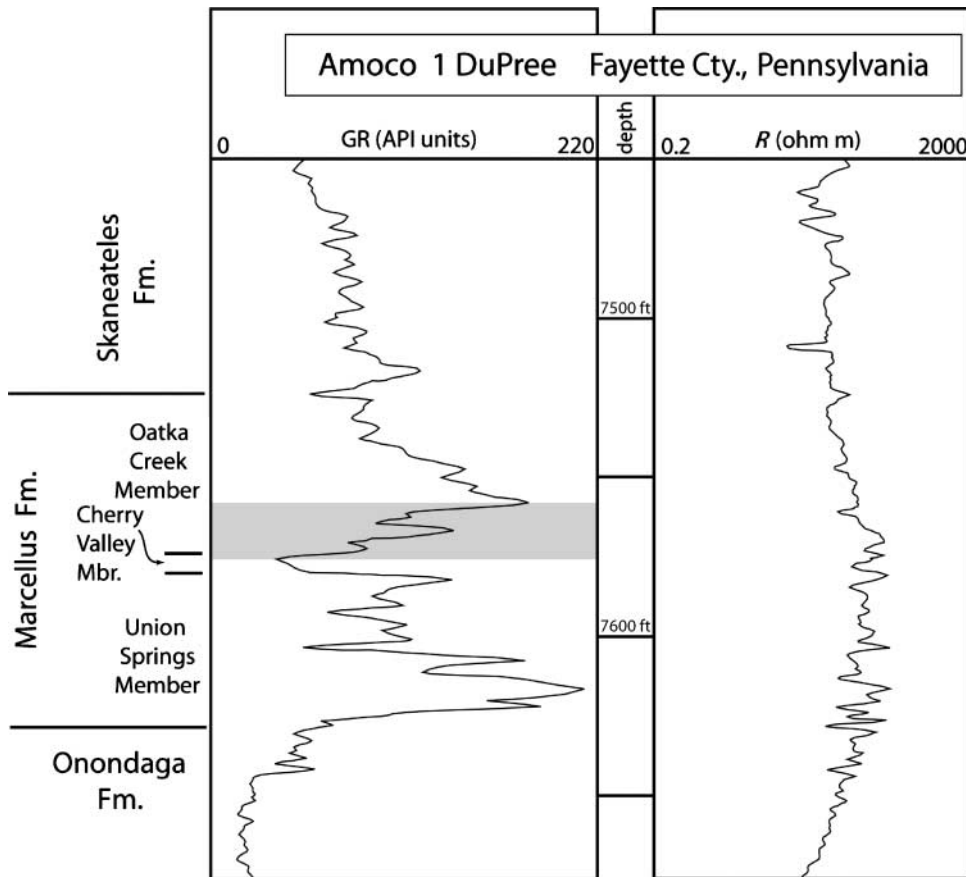
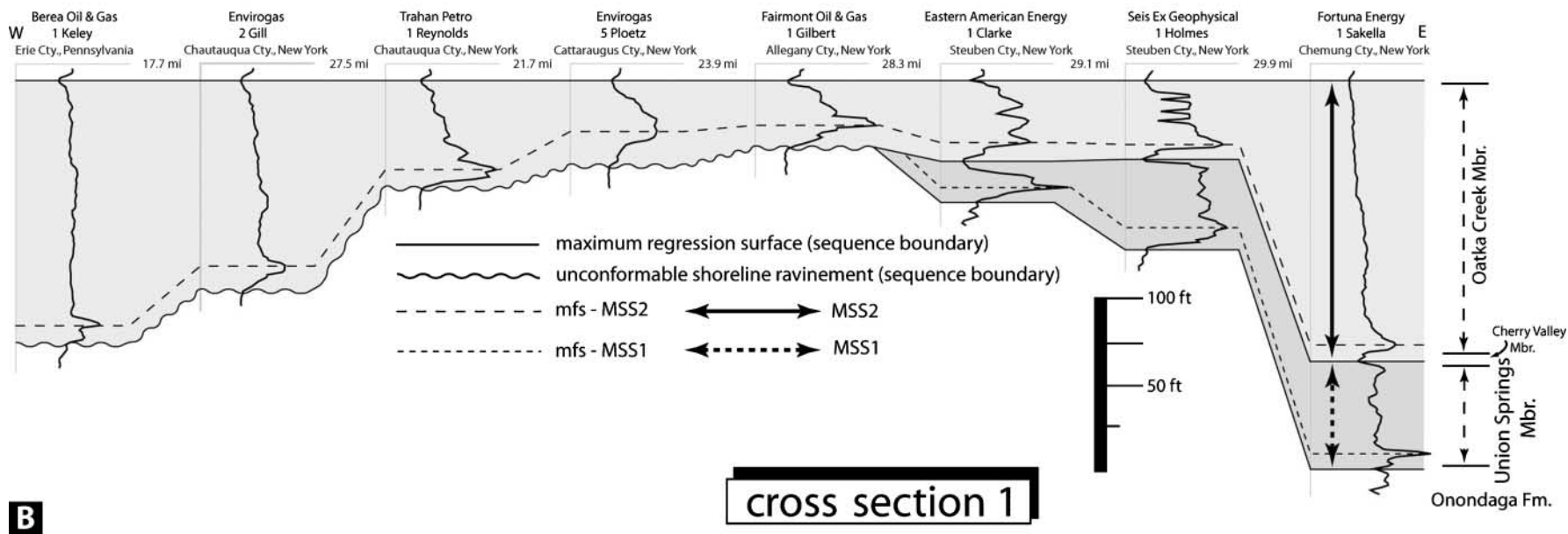
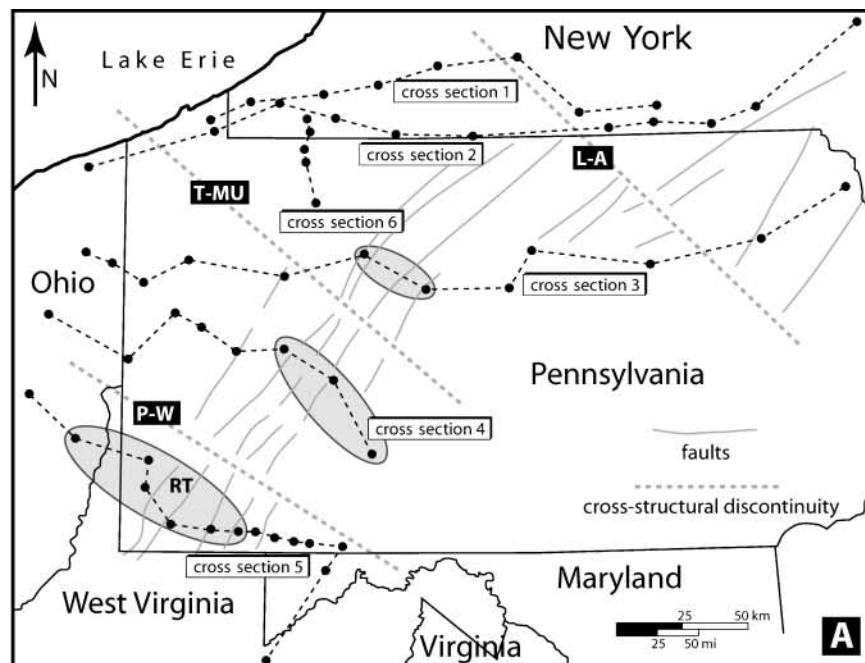


Figure 22. Gamma-ray log (left) showing the gradational contact (shaded interval) of the Cherry Valley Member and overlying Oatka Creek Member.

Much of the quartz lines pore throats or forms irregular microcrystalline aggregates that coat detrital clay grains. Occasional angular detrital quartz and feldspar grains are probably windblown detritus. Calcite is as much as three times as abundant in transgressive systems tract deposits as in the overlying regressive systems tract (Figure 21). Most calcite occurs as single crystals or patches of microspar and microcrystalline aggregates that originated from styliolinid fragments. Last, peaks in pyrite and TOC are coincident with the inferred maximum flooding surface (Figure 21). At this time, conditions conducive to the preservation of organic matter, perhaps fully euxinic bottom conditions related to salinity and density stratification of the water column (e.g., Ettensohn and Elam, 1985; Werne et al., 2002), were established. Alternatively, the relatively rapid rate of sedimentation during the time of Marcellus deposition, perhaps enhanced by a basin shallower than generally presumed, may have combined to diminish the rate of oxidation of organic matter in the water column.

The MSS1 maximum flooding surface defines the top of the transgressive systems tract. Immediately overlying regressive systems tract deposits record the slow reduction of base level and/or increased sediment flux relative to base level rise (Figure 20). Overlying strata display a gradual increase in bulk density and reduced gamma-ray response (Figure 20) reflecting reduced organic carbon and increasing clay content. Diminished total quartz (Figure 21) probably records dilution of the biogenic contribution by increasing amounts of clastic detritus, principally clay, as nearshore environments were displaced seaward and accommodation space diminished. The reduction in base level reflected in the MSS1 regressive systems tract culminated in accumulation of carbonate deposits of the Cherry Valley Member across much of the basin (Figure 11). Werne et al. (2002) maintain that bioclastic debris and detrital carbonate mud that comprise the Cherry Valley was derived, in part, from exposed carbonate platform areas bordering the basin to the west, including the Cincinnati and Algonquin

Figure 23. Map (A) showing basement structural elements (modified from Alexander et al., 2005) and location of sequence stratigraphic cross-sections (B-G) across the core region of the Marcellus Formation basin. RT = Rome trough; L-A = Lawrenceville-Attica cross-structural discontinuity; T-MU = Tyrone-Mount Union cross-structural discontinuity; P-W = Pittsburgh-Washington cross-structural discontinuity. Cross-sections are hung on the top of the MSS2 T-R sequence. Shaded ovals on the map denote cross-section segments (with shaded labels) that display evidence of syndepositional faulting. Logs are gamma-ray logs; maximum American Petroleum Institute (API) count = 700. Refer to text for discussion.



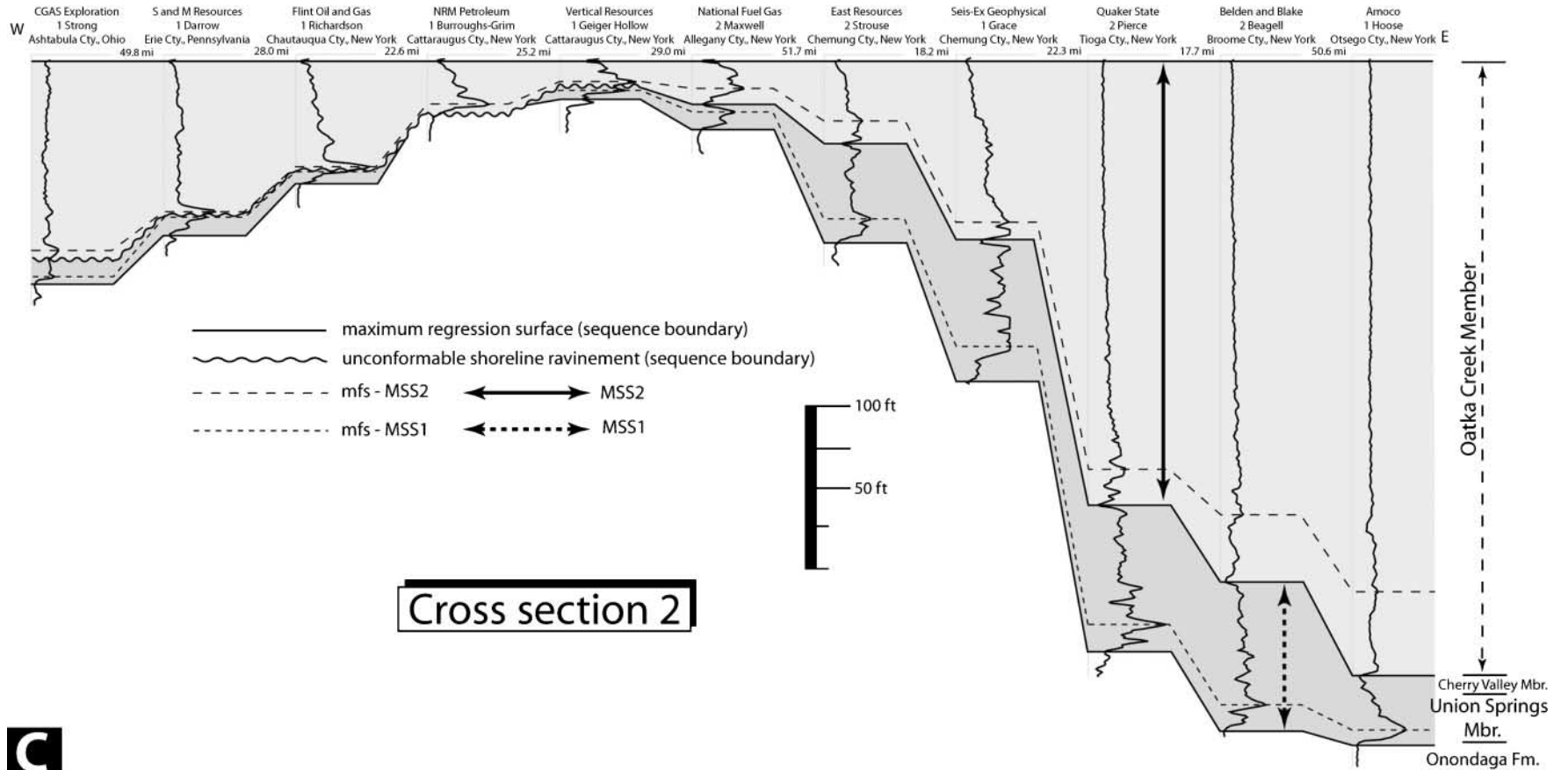
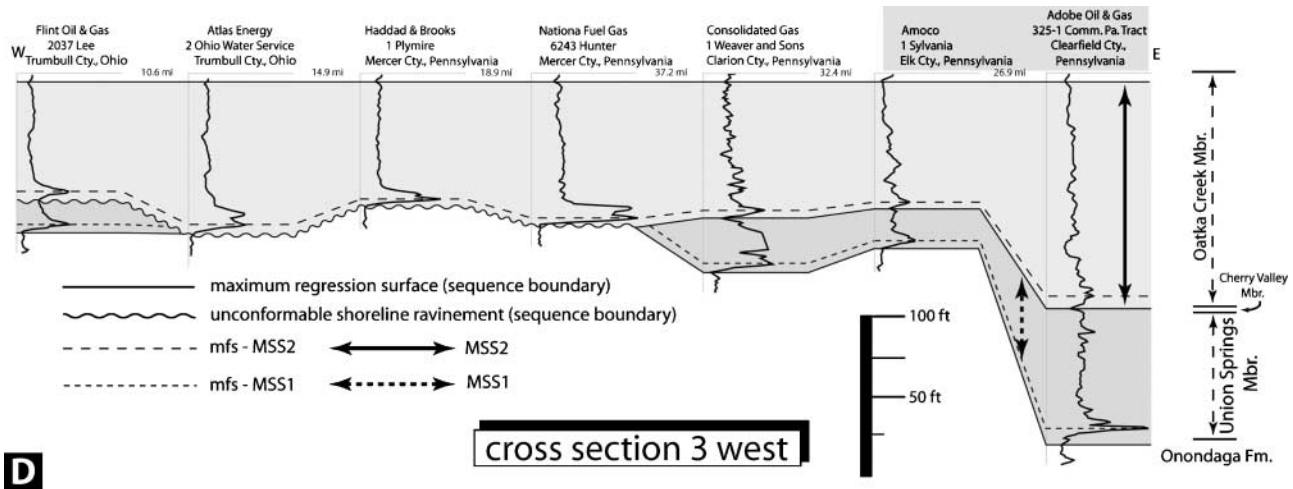


Figure 23. Continued.





D Figure 23. Continued.

arches. Similarly, Brett and Baird (1996) have interpreted bioclastic carbonates higher in the Hamilton Group to be lowstand deposits. The maximum regressive surface that defines the top of MSS1, then, is placed at a gamma-ray minimum and/or bulk density maximum within the Cherry Valley Member (Figure 20).

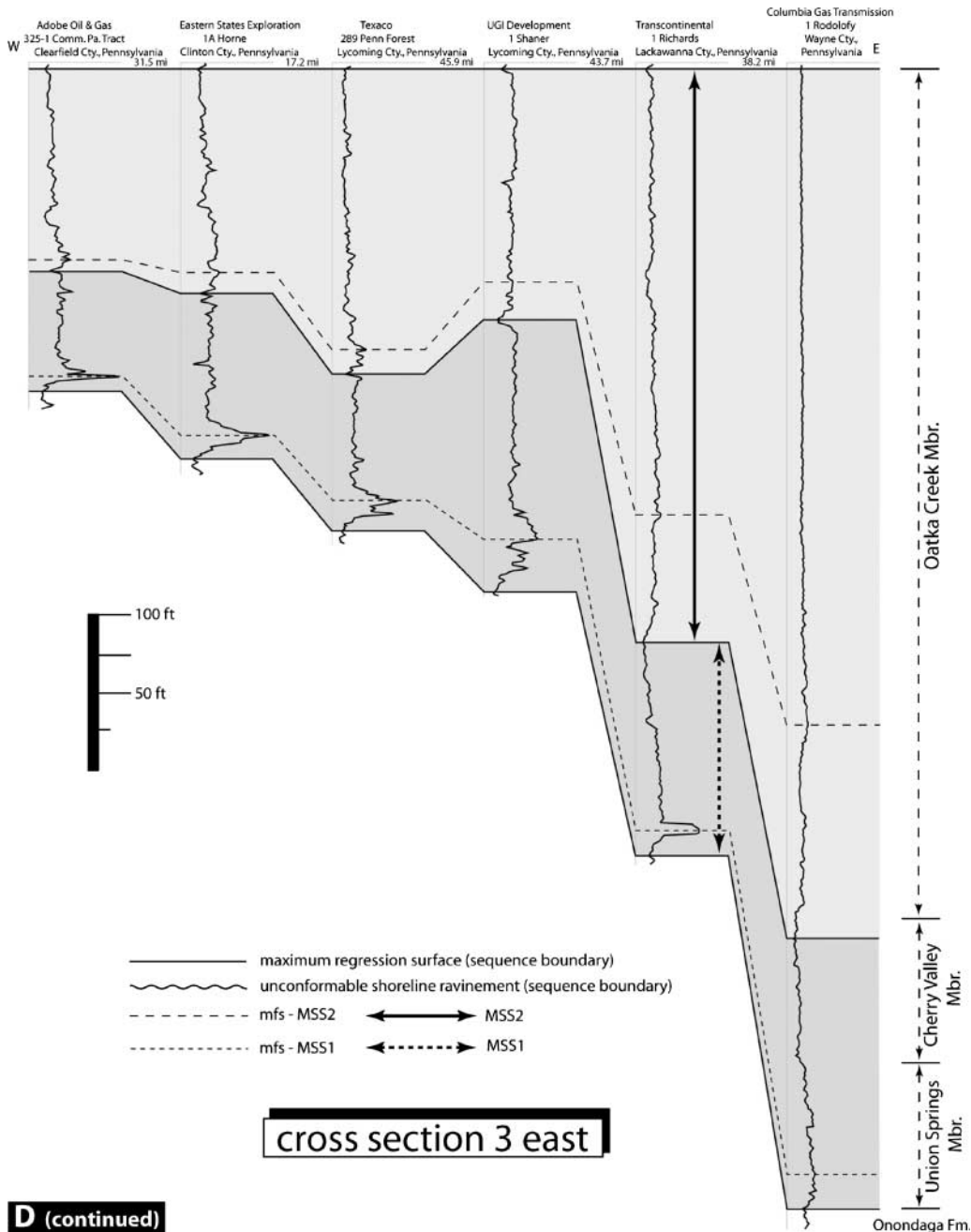
Note that the Columbia 1 Rodolffy well in Wayne County, Pennsylvania, penetrated approximately 65 ft (20 m) of what appears to be arenaceous limestone at the top of the regressive systems tract interval (see Figure 4). These deposits, included in the Cherry Valley Member of this study (Figures 3, 4), are interpreted to comprise a paralic succession (e.g., Emery and Myers, 1996) that had prograded in response to reduced base level. This arenaceous facies of the Cherry Valley Member in the subsurface of northeastern and central Pennsylvania and probably northern West Virginia, as well as in exposure near Kingston, New York (Ver Straeten and Brett, 2006), approximately 50 mi (80 km) east of the Columbia 1 Rodolffy well, reflects the effects of reduced base level at the end of MSS1 time.

Well-log signatures suggest a normally rapid transition from the Cherry Valley Member into the overlying Oatka Creek Member (Figure 22). The upper part of the Cherry Valley represents transgressive carbonate deposits recording the initiation of the MSS2 base level rise. Ver Straeten et al. (1994) interpret a prominent bone bed described

from the top of the Cherry Valley Member in central western New York as a lag deposit formed during maximum transgression, consistent with asymmetric base level curves typified by a rapid initial rise in base level (e.g., Embry et al., 2007). The wave-worked carbonate deposits are overlain by the increasingly organic-rich (increasing gamma-ray and decreasing bulk density) basal interval of the Oatka Creek Member. These deposits comprise the bulk of the MSS2 transgressive systems tract (Figure 20).

A condensed section associated with the MSS2 maximum flooding surface, which is placed at a gamma-ray maximum and bulk density minimum (Figure 20), includes carbonate and pyritiferous layers evident on bulk density and photoelectric index logs. Werne et al. (2002) and Sageman et al. (2003), as part of a geochemical investigation of Oatka Creek samples recovered from two cores drilled in western New York, observed a strong correlation between TOC maxima and intervals of sediment starvation. However, sidewall core data of the present study and gamma-ray and bulk density log signatures (see Figure 8) suggest that MSS2 transgressive systems tract (condensed section) deposits are not as organic rich as equivalent deposits of the MSS1 T-R sequence. The seemingly less organic nature of the MSS2 transgressive systems tract (condensed section) may reflect the existence of a better established terrestrial drainage network capable of diluting the organic flux by

Figure 23. Continued.



D (continued)

this time. Alternatively, the magnitude of the base level rise recorded by MSS2 transgressive systems tract deposits may not have been great enough to shift facies belts far enough landward to effectively isolate the basin from clastic detritus.

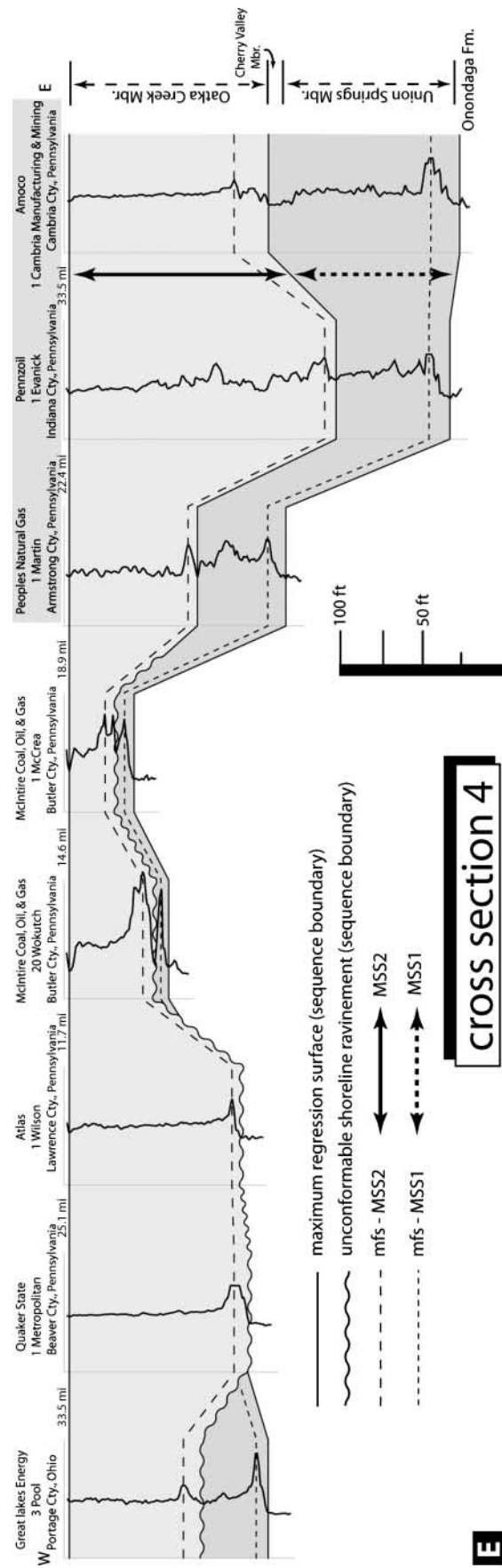
Regressive systems tract deposits immediately above the MSS2 maximum flooding surface record the slowing of base level rise and increasing sediment supply and consequent seaward shift of the shoreline. The increasing supply of clastic de-

tritus to the basin and consequent dilution of organic material raining out of the water column is reflected in diminishing TOC upsection recorded by a progressive increase in bulk density and reduced gamma-ray count (Figure 20). Furthermore, some gamma-ray logs display a facies change upward from relatively carbonaceous lower regressive systems tract deposits into an overlying organic-lean regressive systems tract interval (Figure 20). In places, however, this facies distinction is better made on

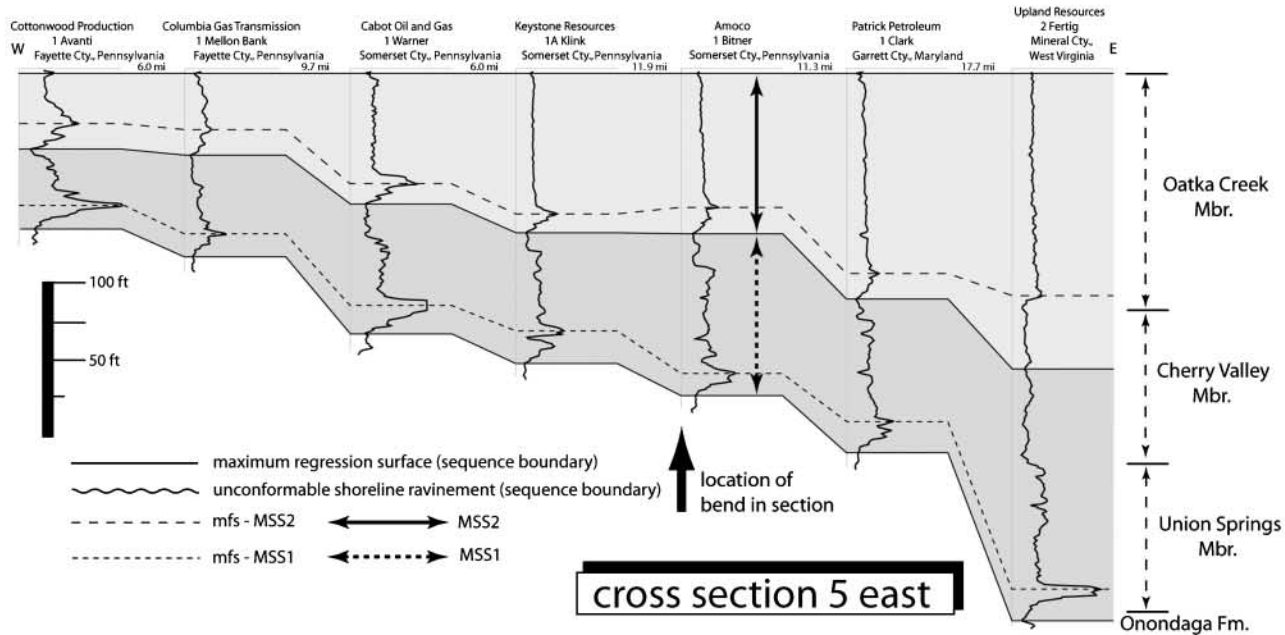
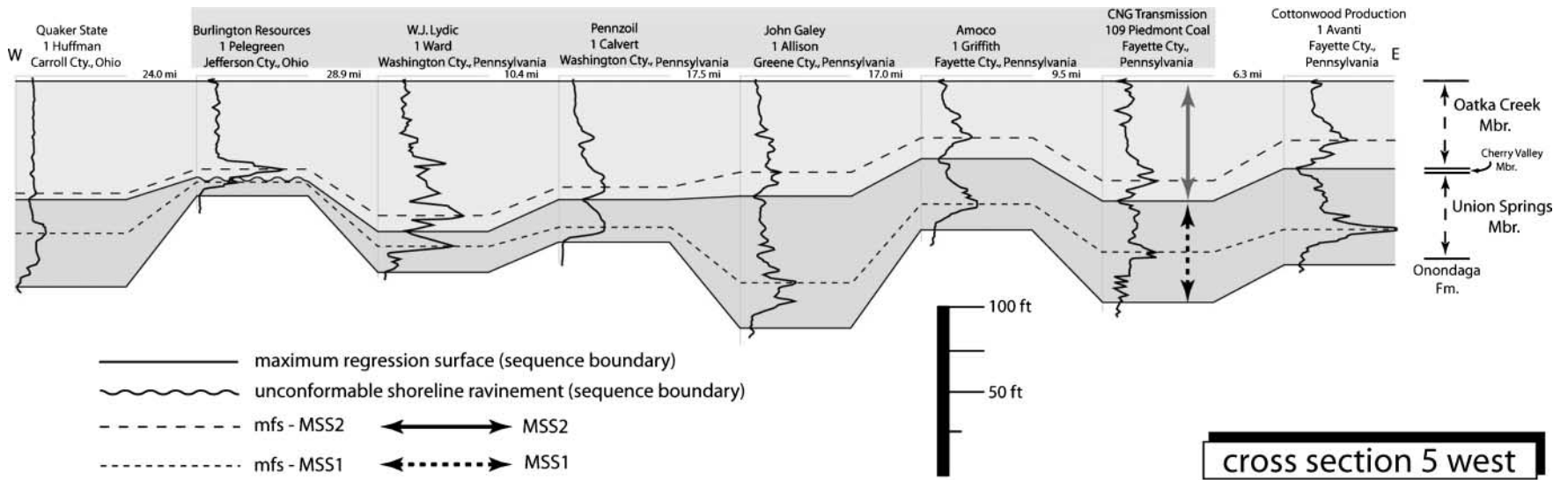
bulk density logs (Figure 20), the response of which appears to provide a truer indication of organic carbon content (e.g., Schmoker, 1979, 1980, 1981).

The MSS2 base level cycle culminated with accumulation of carbonate deposits of the Stafford Member of the Skaneateles Formation, a likely lowstand carbonate (e.g., Brett and Baird, 1996). In the absence of limestone, the maximum regressive surface delimiting the top of MSS2 is placed at a gamma-ray minimum and/or bulk density maximum (see Figure 16B). Such a contact separating a “cleaning-up trend” from an overlying “dirtying-up trend” has been described from several modern basins, including the Upper Jurassic Ute Field, offshore Norway (Emery and Myers, 1996).

A series of sequence-stratigraphic cross sections through the core region of the basin (Figure 23A) reveals several significant aspects of the stratigraphic architecture of the Marcellus Formation. MSS1 is thickest in northeastern Pennsylvania and southeastern New York where a thick regressive systems tract interval includes arenaceous limestone (Figure 23B, C). Similarly, depositional sequence MSS2 thickens toward the northeastern region of the basin, the bulk of the thickening restricted to the regressive systems tract (Figure 23B–D). Moreover, gamma-ray log responses suggest that MSS2 transgressive systems tract deposits are less organic rich in this region of the basin (Figure 23B, C), a likely reflection of the dilution of these deposits by a high clastic flux derived from the Acadian highland source region. This interpretation is buttressed by the increasing thickness of MSS2 transgressive systems tract deposits to the northeast (Figure 23B, C). Impressed upon the northeastward thickening trends of both MSS1 and MSS2 are local variations in thickness that may reflect the effects of basement tectonics (e.g., Harper, 1989). Indeed, the imprint of Rome trough-related extensional tectonics in southwestern Pennsylvania appears to be manifested by variations in the thickness of MSS1 between Washington and Fayette counties (Figure 23F). Furthermore, variations in the thickness of Marcellus Formation T-R sequences are found in the western region of Pennsylvania (Figure 23D, E, especially F), proximal to the projected region of Rome trough-related faulting (Figure 23A).



E Figure 23. Continued.



F

Figure 23. Continued.

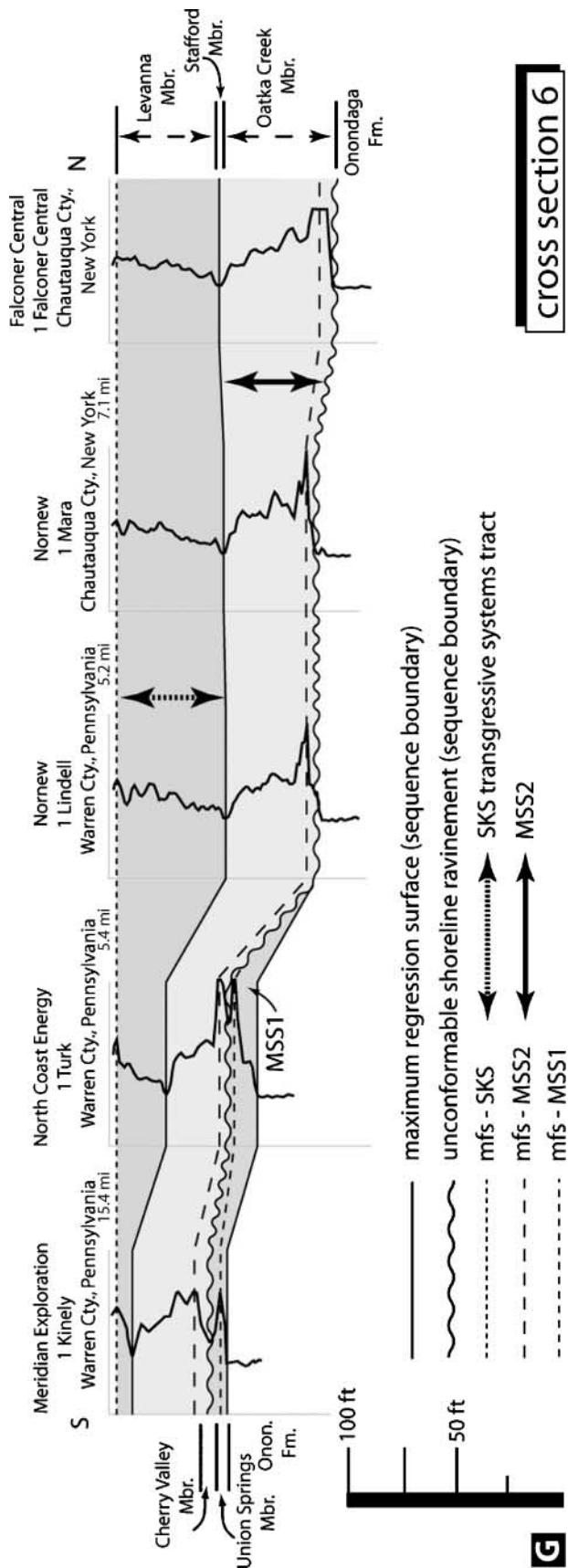


Figure 23. Continued.

Variations in the thickness of MSS1 regressive systems tract deposits are suggestive of local erosion. Notably, the base of MSS2 cuts progressively deeper into MSS1 westward from eastern New York (Figure 23B, C). Moreover, the absence of MSS1 in some regions of the basin indicates that erosion has occurred. In this case, the base of MSS2 is interpreted to be an unconformable shoreline ravinement that passes laterally into a maximum regressive surface in the conformable succession (Figure 23B–D, G). Elsewhere, the MSS2 maximum flooding surface is no more than a few meters or so above the MSS1 maximum flooding surface (Figure 23B–G), again, indicative of erosion. The thinning and/or local absence of MSS1 in western New York and Pennsylvania may reflect the effects of both local uplift and concomitant lowering of base level. In this scenario, uplift began soon after MSS1 regressive systems tract deposits started to accumulate. This, in tandem with the drop in base level that followed MSS1 maximum flooding, resulted in the accumulation of a thin, perhaps starved, regressive systems tract across the uplifted region of the basin. However, the local absence of Cherry Valley lowstand carbonates (e.g., Figure 8B) indicates that T-R sequence MSS1 was eroded.

Regressive systems tract deposits of MSS2, like their counterparts of T-R sequence MSS1, exhibit local variations in thickness (Figure 23D–F). Moreover, the upper boundary of MSS2 cuts well down into the sequence in western New York (Figure 23B, C, G). Locally, the maximum regressive surface at the top of MSS2 lies within several meters of the maximum flooding surface (Figure 23C, G). However, MSS2 regressive systems tract deposits thicken into northeastern Ohio before thinning to the west (Figure 23C). The marked thinning of MSS2 in western New York, locally evident in the subsurface of western Pennsylvania (Figure 23E), appears to be a consequence of uplift and related sediment starvation across this region of the basin rather than local erosion. The presence of the complete MSS2 T-R sequence, including lowstand carbonate rocks of the Stafford Member, indicates that erosion is not responsible for the marked thinning of MSS2 in this region of the basin. Instead, thinning is likely

a consequence of reduced sedimentation rate, perhaps related to local uplift. Note that the gamma-ray signature of MSS2 regressive systems tract deposits across the area of thinning in western New York is indicative of higher TOC relative to correlative deposits to the east and west (compare Figures 15, 23B).

The Stafford and overlying Levanna members of the Skaneateles Formation comprise the transgressive systems tract of T-R sequence SKS (Figure 20). Rising base level is manifested by an upward-increasing gamma-ray response and decreasing bulk density; the SKS maximum flooding surface is placed at the gamma-ray peak/bulk density minimum within the organic-rich Levanna Member (Figure 20). Note that the SKS maximum flooding surface in central western New York is only approximately 20 ft (6 m) above the upper maximum regressive surface of T-R sequence MSS2, whereas farther to the west, the SKS maximum flooding surface is approximately 60 ft (18 m) above the base of SKS. Similarly, the SKS maximum flooding surface in northwestern Pennsylvania is only approximately 8 ft (2.5 m) above the top of MSS2. However, in western New York, the same surfaces are separated by approximately 35 ft (11 m) (Figure 23G). These relationships reflect the onlapping of SKS transgressive systems tract deposits to the west and north.

BASIN DYNAMICS: THE ROLE OF BASEMENT STRUCTURES ON MARCELLUS FORMATION SEDIMENTATION PATTERNS

The Middle and Upper Devonian succession of the Appalachian Basin records the cratonward advance of the Catskill Delta complex in response to the Acadian oblique collision of the Avalonia microplate and Laurentia (Ettensohn, 1987). Rapid eastward thickening of MSS1 and MSS2 (Figure 23C, D) toward the Acadian highlands, the remnant of which is found in New England, is characteristic of foreland basin deposits (DeCelles and Giles, 1996). Superimposed on this protracted shallowing trend is a hierarchy of black shale-based depositional sequences that reflect shorter term base level oscillations (e.g., Johnson et al., 1985; Brett and Baird, 1986, 1996; Van Tassel, 1994; Ver Straeten, 2007).

Ettensohn (1985, 1987, 1994) proposed a tectono-stratigraphic model for the Acadian orogeny that entails four tectophases. Within this framework, black shale accumulated as a consequence of thrust-load-induced subsidence and rapid deepening of the foreland basin. Still, the conodont-based correlation of some black shale intervals of the Appalachian Basin with depositional sequences in Europe, Morocco, and the Cordilleran region of the United States suggests a eustatic signature (Johnson et al., 1985; Johnson and Sandberg, 1989). Both mechanisms—eustasy and thrust-induced subsidence—likely shaped the Middle and Upper Devonian stratigraphic architecture of the core region of Marcellus exploration in the Appalachian Basin (Werne et al., 2002). However, it is difficult at best to assess the relative contributions of each driving factor—tectonism and eustasy—separately (Burton et al., 1987).

Lithospheric-flexure models predict that the onset of a foreland thrust-load event and consequent elastic basin subsidence is accompanied by development of a forebulge on the cratonward margin of the foreland basin (Quinlan and Beaumont, 1984; Tankard, 1986a, b; Crampton and Allen, 1995; DeCelles and Giles, 1996). Viscoelastic (Quinlan and Beaumont, 1984; Beaumont et al., 1987) and elastic flexure (Flemings and Jordan, 1990; Jordan and Flemings, 1991) models envisage both cratonward and hinterlandward migration of foredeeps and bulges that reflect different loading and subsequent relaxation histories (e.g., Ettensohn and Chesnut, 1989; Ettensohn, 1992, 1994; Giles and Dickinson, 1995). However, the Appalachian foreland basin, at the onset of the Acadian orogeny, hosted a network of intermittently active basement structures inherited from the breakup of Rodinia, the Precambrian supercontinent. Chief among these structures is the Rome trough, which enters Pennsylvania near its southwest corner (Scanlin and Engelder, 2003), although its location throughout much of the state remains elusive (Harper, 1989). Furthermore, the core region of Marcellus exploration includes several northwest-striking basement wrench faults (Figure 23A) (Parrish and Lavin, 1982; Rodgers and Anderson, 1984; Harper, 1989). These faults, referred to as cross-strike structural discontinuities

Figure 24. Maps showing cross-structural discontinuities and (A) the location of the Oriskany no-sand area, Kane gravity high, and the area of the basin absent in the Union Springs Member of the Marcellus Formation; (B) the area of the basin absent in the Cherry Valley Member of the Marcellus Formation; (C) the region of the basin over which the Oatka Creek Member of the Marcellus Formation thins; (D) the distribution of the Stafford Member of the Skaneateles Formation; (E) the distribution of the Levanna Member of the Skaneateles Formation illustrating 0 ft, 5 ft (1.5 m), and 20 ft (6.1 m) isochores. Cross-structural discontinuities: L-A = Lawrenceville-Attica; B-B = Blairsville-Broadtop; H-G = Home-Gallitzin; T-MU = Tyrone-Mount Union; P-W = Pittsburgh-Washington (modified from Parrish and Lavin, 1982).

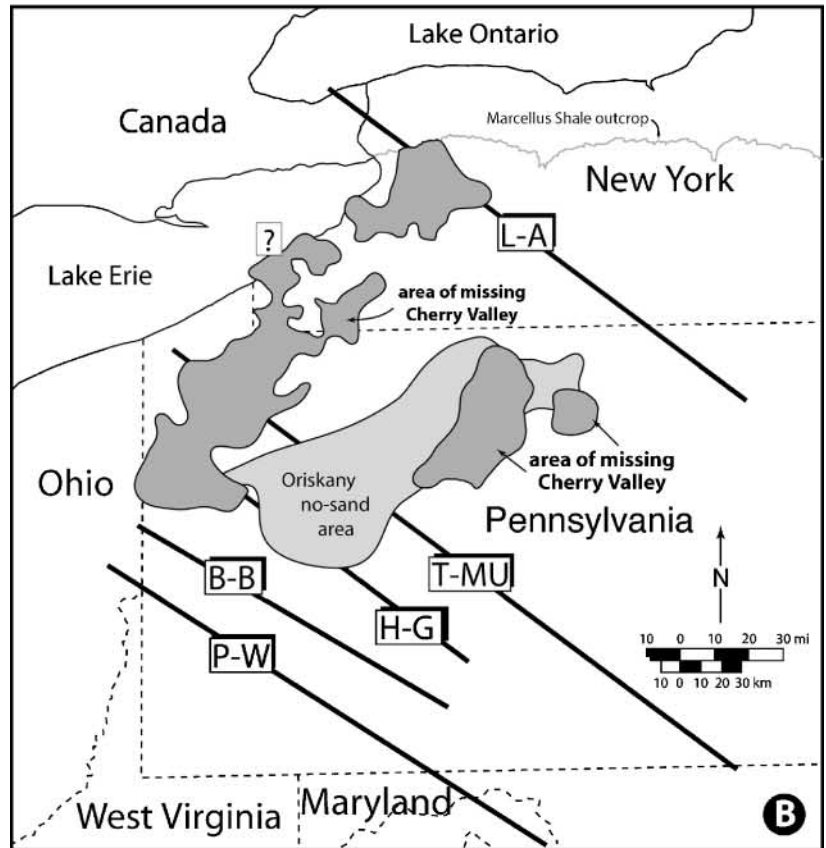
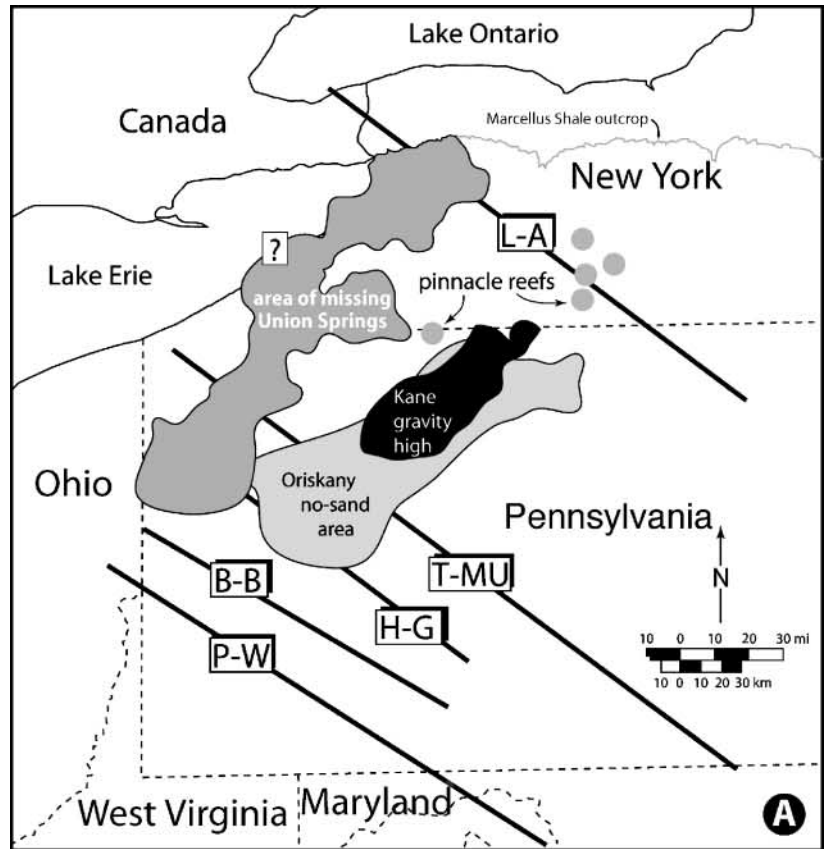


Figure 24. Continued.

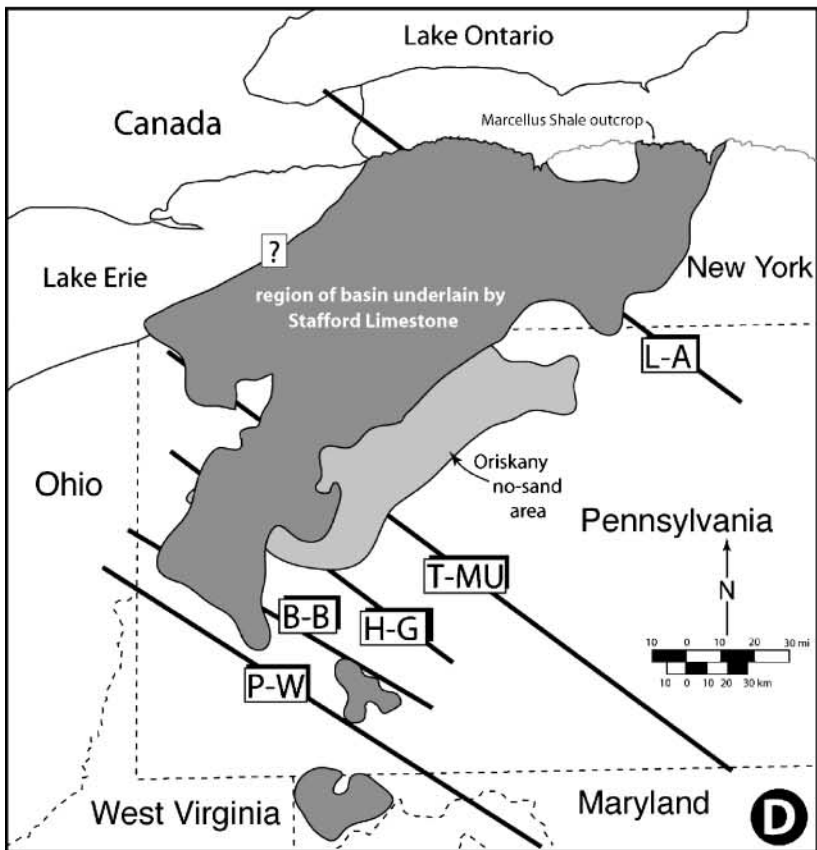
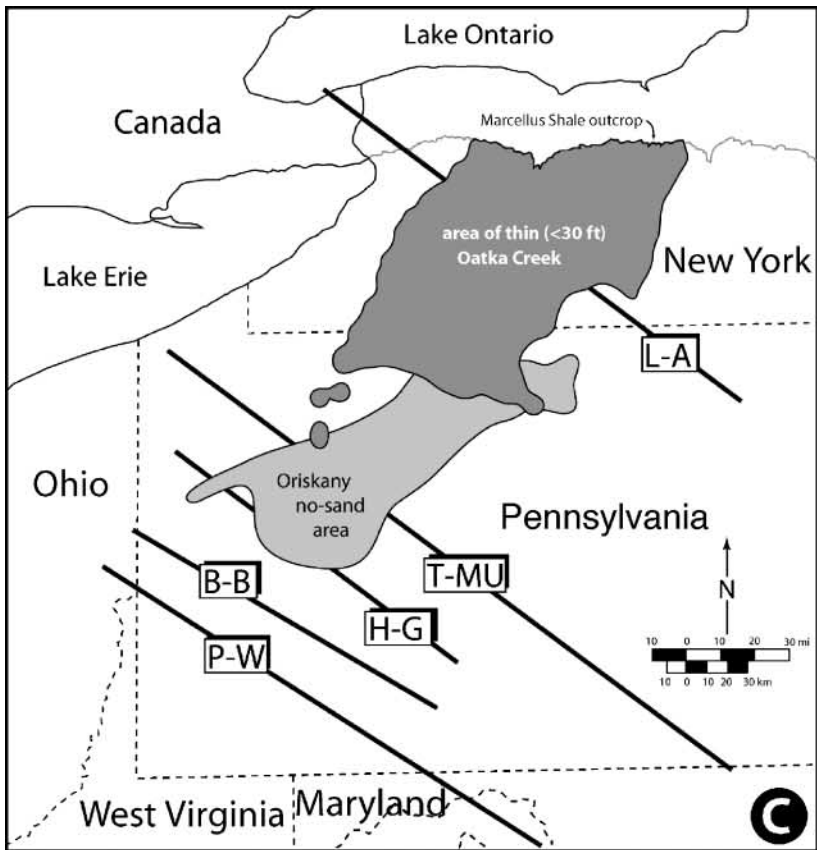
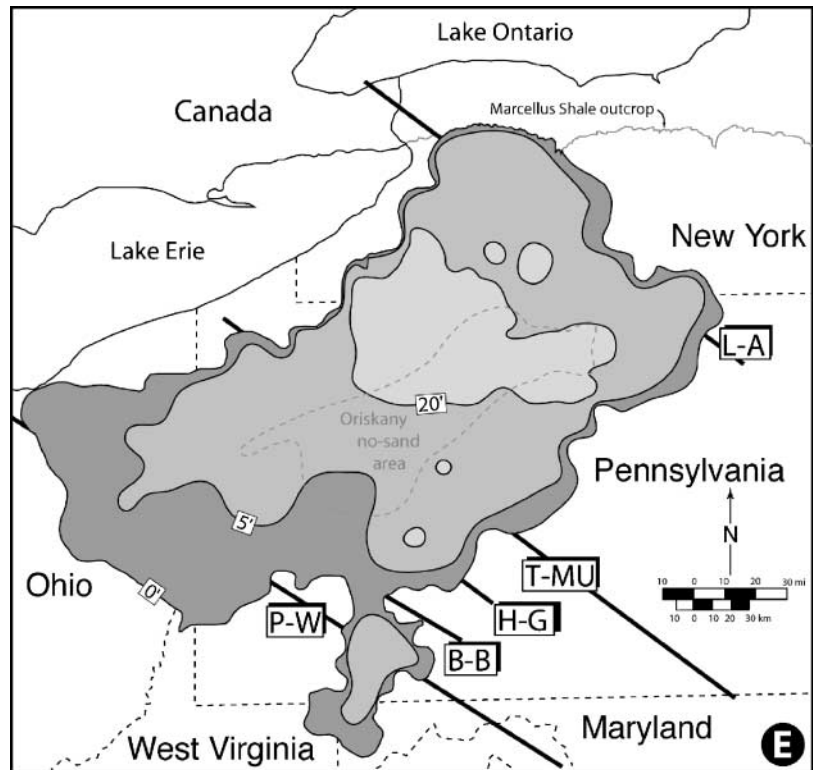


Figure 24. Continued.



([CSD] Wheeler, 1980), likely originated as strike-slip faults related to the Late Precambrian formation of the proto-Atlantic Ocean (Thomas, 1977). Some authors relate the apparent segmentation of the Rome trough to slip on CSD (Lavin et al., 1982; Harper, 1989); however, it appears that strike-slip displacement was replaced by dip-slip displacement by early Paleozoic time (Wagner, 1976). Although the regional architecture of the Acadian foreland basin was a consequence of load-induced subsidence, it is difficult to imagine that inherited basement structures, including those previously described, did not, in some way, affect foreland basin evolution and sedimentation patterns during the Acadian orogeny. Reactivation of preexisting faults during foreland flexure can partition the basin into regions of fault-controlled uplift and depocenters (e.g., Tankard, 1986a; DeCelles and Giles, 1996).

An early indicator of forebulge-like dynamics induced by Acadian convergence in the Appalachian Basin is the thinning and local absence of the late Early Devonian Oriskany Sandstone along a northeast-southwest-trending region of north central and northwestern Pennsylvania (Figure 24A)

(Fettke, 1938). Williams and Bragonier (1974), commenting on the coincidence of the so-called “Oriskany no-sand area” with the Kane gravity high (Parrish and Lavin, 1982) (Figure 24A), speculated that the local absence of the Oriskany was a consequence of Early Devonian basement uplift. Etensohn (1985) subsequently attributed the late Emsian (late Early Devonian) unconformity at the base of the Onondaga Limestone to basinward migration of a forebulge caused by tectophase I thrust loading. Similarly, Ver Straeten and Brett (2000) related northeastward time transgressive pinnacle reef development described from the upper part of the Onondaga Limestone to retrogradational (eastward) migration of a flexural welt at the end of tectophase I of the Acadian orogeny. However, the subparallel orientation of the Oriskany no-sand area, which is roughly mimicked by Onondaga pinnacle reef formation (Figure 24A), and the oblique Acadian plate convergence direction (Etensohn, 1987; Ferrill and Thomas, 1988) is inconsistent with erosion of the Oriskany as a straightforward flexural response to thrust loading in the hinterland (i.e., Turcotte and Schubert, 1982). Ver Straeten

and Brett (2000) suggested that the inferred positive feature on which Onondaga pinnacle reefs formed does not comport with regional forebulge models.

Note that the elongate Oriskany no-sand area and Kane gravity high are roughly centered between the Lawrenceville-Attica and Home-Gallitzin CSD (Figure 24A). Furthermore, Onondaga pinnacle reef development appears to have terminated close to the trace of the Lawrenceville-Attica CSD (Figure 24A). It is an intriguing possibility that thickness trends of the Lower Devonian Oriskany Sandstone and Onondaga reef development in this region of the basin reflect episodes of vertical displacement of crustal blocks bounded by the Home-Gallitzin, Tyrone-Mount Union, and Lawrenceville-Attica CSD (Figure 24A). Such block displacement, which would have overprinted smooth elastic forebulge dynamics, partitioned the foredeep basin into subtle ridges and depocenters (DeCelles and Giles, 1996). We have no data to suggest that the faults were anything but blind (i.e., did not intersect the Earth's surface) during the time of Marcellus depositions. Still, enough displacement may have occurred such that related base level changes produced local sites of wave base sediment reworking (i.e., unconformable shoreline ravinement) adjacent to depocenters. Some authors have suggested that the Marcellus Formation accumulated in deep oxygen-deficient waters (Potter et al., 1981), yet a relatively shallow water depth, perhaps only 100 to 130 ft (30–40 m), would have reduced the vertical displacement necessary to bring base level to such a position that local sediment starvation or even erosion could have occurred within the limited time framework over which the Marcellus accumulated. Indeed, Harper (1999) has suggested that the subtropical Marcellus Basin could have been less than 165 ft (50 m) deep if the water column had been sufficiently stratified to preclude mixing of warm oxygenated surface water with anoxic or, as suggested by Werne et al. (2002), euxinic bottom water.

The organic-rich Union Springs Member reflects a sharp increase in base level that resulted in accumulation of a transgressive systems tract at the onset of tectophase II of the Acadian orogeny (Ettensohn, 1985, 1994). In more distal regions of

the basin, however, farther removed from clastic sources, rising base level was not accompanied by deposition of an obvious transgressive systems tract. Instead, the rise in base level is manifested only by a sharp contact of the Onondaga Formation and overlying Union Springs Member (see Figure 10D). Reactivated basement structures, including Rome trough-like extensional structures, appear to have influenced the thickness of MSS1 and MSS2 from southwestern into central Pennsylvania (Figure 23D–F). The Union Springs Member, which comprises the bulk of MSS1 transgressive and regressive systems tract deposits, as well as the Cherry Valley Member, much of which encompasses the upper part of the MSS1 regressive systems tract, are absent along a northeast-southwest-oriented region of western New York and northwest Pennsylvania (Figures 5, 11, 24B) that parallels the Oriskany no-sand area to the south (Figure 23A, B). As with the Oriskany no-sand area, the western and eastern limits of the region absent in the Union Springs and Cherry Valley (MSS1) deposits are roughly coincident with the Home-Gallitzin and Lawrenceville-Attica CSD, respectively (Figure 24A, B). The thinning and local absence of MSS1 deposits likely reflects the effects of local warping or flexing of the basin induced by displacement on the Lawrenceville-Attica and Home-Gallitzin CSD soon after the MSS1 base level maximum. The result was a local reduction of base level of such a magnitude to cause erosion of MSS1.

Transgressive systems tract deposits of MSS2, the upper reworked horizon of the Cherry Valley Member and radioactive basal interval of the overlying Oatka Creek Member, record a rapid base level rise. Overlying regressive systems tract strata (upper Oatka Creek Member and lower interval of the Stafford Member of the Skaneateles Formation) thicken markedly toward the northeastern region of the basin (Figure 23B, C), a likely consequence of increasing proximity to the thrust load and clastic sources (e.g., DeCelles and Giles, 1996). Accommodation space at this time would have been created by relaxation of the tectonic load (Ettensohn, 1994). However, like MSS1, the MSS2 sequence thins along a northeast-southwest-trending region of western New York and Pennsylvania (Figures 12,

23B). Note also that the region of thin MSS2 deposits, notably the Oatka Creek Member, parallels the Oriskany no-sand area and is roughly bounded on the west by the Tyrone-Mount Union CSD (Figure 24C). The northern and eastern limit of the region of Oatka Creek thinning appears to have been lost to erosion (Figure 24C). The presence of the complete MSS2 T-R sequence in the areas of thinning in western New York (Figure 23B, C, G) indicates that thinning was a consequence of reduced sediment flux rather than erosion. Moreover, the thin MSS2 sequence, especially the regressive systems tract deposits, is more organic rich than laterally adjacent thicker MSS2 deposits (compare Figures 15, 23B). We attribute local thinning of MSS2 to accumulation of a starved or condensed T-R sequence across a northeast-southwest-trending topographic high. In this scenario, organic-rich sediment raining out of the water column was concentrated on the high even as base level dropped, whereas organic-lean sediment, perhaps deposited from hyperpycnal flows, ponded in adjacent bathymetric lows.

The story that emerges from the described thickness trends is one of local basement control on Marcellus sedimentation patterns that began at the end of Early Devonian time in the northwestern Pennsylvania region of the basin with erosion of the Oriskany Sandstone. Syndepositional movement on what were likely blind faults resulted in the creation of local depocenters and subtle ridges that influenced sedimentation and erosion patterns in this region of the greater Acadian foreland basin that had formed as a consequence of thrust loading. The block delimited by the Home-Gallitzin and Lawrenceville-Attica CSD experienced episodes of uplift, resulting in localized erosion and/or sediment starvation across a topographic high followed by subsidence, the creation of accommodation space, and consequent sediment accumulation. Block activation appears to have shifted to the northwest and then to the east during the time of Marcellus deposition. Furthermore, the history of crustal block movement, apparently linked to oblique plate convergence, continued beyond accumulation of the Marcellus Formation. The Stafford Member, the basal unit of the Skaneateles Forma-

tion, records the base level minimum at the top of the MSS2 T-R sequence. The isochore pattern of the Stafford, too, may reflect the influence of fault-induced warping of the basin (Figure 17). That is, the southwestern edge of the Stafford Member zeroes close to the inferred trace of the Tyrone-Mount Union CSD; to the east, the Lawrenceville-Attica CSD appears to have exerted little, if any, control on accumulation of the carbonate lowstand deposits (Figure 24D). However, the Stafford is thickest in the center of the block bounded by the Tyrone-Mount Union and Lawrenceville-Attica CSD (Figures 17, 24D) perhaps revealing the shallowest region of the block at this time. Last, the southwestern extent of the Stafford may reflect the influence of the Blairsville-Broadtop CSD (Figure 24D). Conceivably, a complex displacement history on both the Tyrone-Mount Union and Blairsville-Broadtop CSD influenced sedimentation patterns at the western edge of the Stafford depocenter.

The base level rise after accumulation of MSS2 regressive systems tract deposits is reflected in the Levanna Member of the Skaneateles Formation, transgressive systems tract deposits of the SKS T-R sequence (Figure 20). The especially intriguing aspect of the Levanna isochore pattern is its sharp eastern termination close to the inferred trace of the Lawrenceville-Attica CSD (Figure 24E). Relatively rapid thinning of the organic-rich Levanna to the east (Figure 18) is suggestive of down-to-west displacement along the Lawrenceville-Attica fault, creating a black shale depocenter. The western limit of the Levanna is more gradual (Figures 18, 24E). The thicker region of the Levanna (20-ft [6.1 m] isochore) terminates close to the Tyrone-Mount Union CSD (Figure 24E). However, the effects of this fault on basin morphology appear not to have been great enough to preclude accumulation of carbonaceous sediment well to the southwest of the Tyrone-Mount Union CSD. The southwestern limit of Levanna sedimentation may have been influenced by down-to-east displacement on the Pittsburgh-Washington CSD (Figure 24E). Progressive subsidence of the crustal block defined by the Lawrenceville-Attica and Tyrone-Mount Union CSD in tandem with rising base level is indicated by westward and northward onlapping of

the most organic-rich facies of the Levanna Member (Figure 23G).

The generally limited distribution of the organic-rich Levanna Member relative to the more widespread Union Springs and Oatka Creek members of the Marcellus Formation appears to have been a consequence of reactivated blind basement structures and consequent warping of the foreland basin. Specifically, displacement along the Lawrenceville-Attica, Tyrone-Mount Union, and Pittsburgh-Washington basement wrench faults created the accommodation space necessary for accumulation of carbonaceous sediment in a relatively restricted region of the greater Acadian foreland basin, which, at this time, was receiving otherwise organic-lean sediment of the Skaneateles Formation and equivalents.

CONCLUSIONS

The stratigraphy and thickness trends of the Middle Devonian Marcellus Formation of the Appalachian Basin reflect the interplay of Acadian thrust loading of the Laurentian craton and short-term base level fluctuations. These events shaped the stratigraphic architecture of the Marcellus in ways that can impact exploration and production strategies of this emerging shale gas play. Analysis of more than 900 wireline logs indicates that the Marcellus Formation encompasses two third-order T-R sequences, MSS1 and MSS2, in ascending order. These deposits are overlain by T-R sequence SKS, which comprises at least the lower part of the Skaneateles Formation. The Marcellus sequence stratigraphy offers a predictive framework for reservoir assessment that can be extrapolated into areas of poor data control. Compositional attributes that influence such critical reservoir properties as porosity and brittleness, including quartz, carbonate, clay, and pyrite, vary predictably as a consequence of base level oscillations. The sequence-stratigraphic framework of the Marcellus Formation presented in this study demonstrates that transgressive systems tract and early regressive systems tract deposits contain the greatest abundance of malleable organic matter. However, these same deposits are enriched in those components that enhance reservoir brittle-

ness, including quartz, calcite, and pyrite. Assessing the relative importance of these compositional elements to production and stimulation strategies becomes a reservoir engineering issue.

Variations in the thickness of lithostratigraphic units of the Marcellus Formation and immediately overlying deposits of the Skaneateles Formation as well as the MSS1 and MSS2 T-R sequences across the core region of the basin reflect the complex relationship among Acadian thrust loading, fluctuations in base level, recurrent basement structures, and proximity to clastic sources. Acadian thrust loading of Laurentia played a first-order role in creating the accommodation space necessary for accumulation of both Marcellus T-R sequences. Indeed, accommodation space was greatest in the northeast region of the basin, proximal to the Acadian thrust load and clastic sources. However, marked local variations in the thickness of both T-R sequences, especially regressive systems tract deposits, are likely a consequence of displacement along reactivated blind basement faults, including those associated with the Rome trough, which warped the foreland basin into low relief ridges and depocenters.

The possible influence of structural control on the thickness of the Marcellus Formation is not a new concept. Piotrowski and Harper (1979) presented evidence that the Laurel Hill, Negro Mountain, and Chestnut Ridge anticlines of southwest Pennsylvania were active during accumulation of the Marcellus Formation. More recently, Scanlin and Engelder (2003) demonstrated that the thickening of Marcellus black shale in the hinges of anticlines in southwest Pennsylvania was a consequence of Alleghanian folding. The present study extends the concept of basement control on sedimentation and erosion patterns over a greater expanse of the basin. The northwest-striking Tyrone-Mount Union, Lawrenceville-Attica, and Home-Gallitzin wrench faults, most likely blind, appear to have been especially active in the central to western New York and western Pennsylvania region of the Appalachian Basin during the time of Marcellus deposition, producing subtle ridges and depocenters manifested by local erosion and increases in thickness of the MSS1 and MSS2 sequences.

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