PALEOZOIC SEDIMENTARY SUCCESSIONS OF THE VIRGINIA VALLEY & RIDGE AND PLATEAU

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INTRODUCTION

This chapter provides an overview of the Paleozoic sedimentary succession of the Appalachian Valley and Ridge and Plateau provinces of Virginia (Figs. 1 and 2). It builds on the work of Butts (1940, 1941), Colton (1970), Dennison (1970, 1976, 1985, 1996), Diecchio (1985a, b), Diecchio and Dennison (1996), Gathright et al. (2003), the many quadrangle reports from the Virginia Division of Mineral Resources, and the many student theses and dissertations on these rocks. We have attempted to provide both the formational lithostratigraphy, as well as the sequence stratigraphy as a framework, so that it will have the widest possible use. Gradstein et al. (2004) has been used as the geological time scale, which has resulted in some changes in epoch/stage boundaries from that traditionally used.

The succession has been divided into a passive margin succession, and several foreland basin successions (cf. Ettensohn, 1994). We have subdivided these tectonostratigraphic units into a sequence stratigraphic hierarchy following that of Weber et al. (1995). The sequence stratigraphic units are the result of accommodation changes, and hence the supersequences and sequences can result from eustasy, tectonics or more likely a combination of the two. We have used 2nd-order supersequences as the major sequence stratigraphic subdivision; these result from long-term (commonly 5 to 50 m.y.) changes in accommodation and are bounded by unconformities or major shallowing events and correlative conformities. The supersequences are composed of 3rd-order (0.5 to 5 m.y.) depositional sequences that result from smaller scale cycles of accommodation change. During lowered relative sea levels within a depositional sequence, the lowstand systems tracts (LST) commonly are formed on the outer shelf or during icehouse times on the slope. Transgressive systems tracts (TST) are developed during times of increased accommodation or rise in sea level. Highstand systems tracts (HST) are formed during decreasing accommodation, typically associated with sea-level fall. The depositional sequences, where composed of unconformity bound 4th-order (100 to 400 k.y.) sequences related to icehouse glacio-eustasy, are termed composite sequences. Parasequences ("cycles"), defined as small scale shallowing upward sedimentary units bounded by flooding surfaces, are the building blocks of the depositional sequences and relate to higher frequency (10 k.y. to <100 k.y.) sea-level fluctuations, commonly driven by Milankovitch orbital forcing, or autocyclic processes.

The thickness distribution (isopach maps) of supersequences or component units are shown, where possible, on palinspastic base maps, which take out the tectonic shortening due to late Paleozoic folding and thrusting. These isopach maps reflect the regional variation in sediment accumulation and subsidence of the basins. Individual data points are not shown on these maps, and the reader is referred to the original maps for this information. Where possible, we have drawn simplified cross sections to illustrate the gross relationships of units, but the reader is referred to the original papers, which in many cases show the high resolution sequence stratigraphic relations. Brief summaries of the regional facies developed (Appendix A) are given for each of the intervals discussed, along with inferred depositional settings.

Development of the Paleozoic succession is analyzed in terms of tectonics, climate and eustatic controls. The lack of a high resolution biostratigraphic, magnetostratigraphic or chemostratigraphic framework for the succession hampers correlation of local accommodation events evident in the stratigraphy to regional or global sea-level curves. Hopefully, the overview will provide a stimulus for future work on the Paleozoic rocks of the Appalachian Valley and Ridge and Plateau provinces, which are a superb repository of some 250 m.y. of tectonic, sea-level and climate change.

STRUCTURAL FRAMEWORK

Paleozoic sedimentary rocks of Virginia typically are exposed in multiple strike belts within the folded and thrusted Valley and Ridge Province and in the relatively undeformed Appalachian Plateau in the southwestern part of the state (Fig. 1). Deposition was initiated following rifting in the Neoproterozoic (Aleinikoff et al., 1995). Rifting resulted in a series of promontories/salients and reentrants/recesses along the newly formed passive margin (Fig. 2; Thomas, 1977, 1983, 2006). The promontory separated two large depocenters, one centered in Tennessee, and the other in northern Virginia extending into Pennsylvania (Fig. 1; Colton, 1970; Read, 1989a, b). These depocenters strongly influenced thicknesses of the passive margin and subsequent foreland basin deposits. Three major orogenic episodes affected the Paleozoic succession. The Taconian orogeny in the Late Ordovician was related to arc-continent collision, and converted the passive margin to a foreland basin. The Acadian orogeny, possibly related to collision of Avalonia with Laurentia, resulted in the extensive Devonian to Early Mississippian clastic wedge. The Alleghanian orogeny was caused by continent-continent collision of Laurentia and Gondwana, and besides forming the late Mississippian to Permian clastic wedge, also produced the

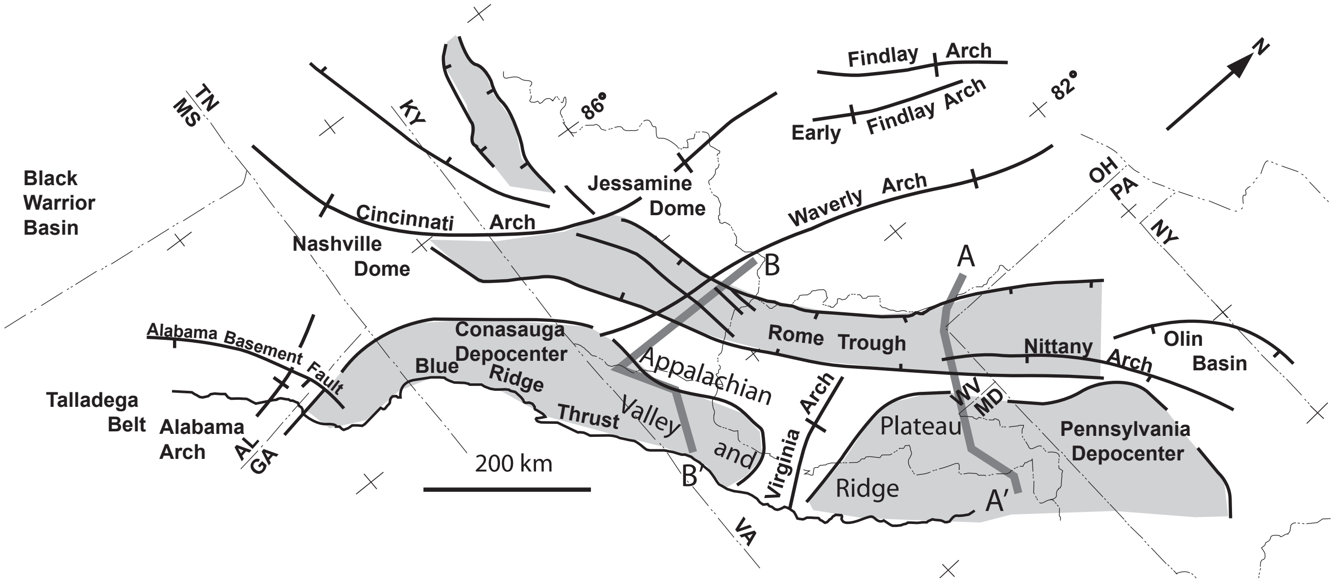


Figure 1. *Structural base map of central and southern Appalachia Valley and Ridge, showing depocenters and Rome Trough (shaded) and arches (modified from Read, 1989a). Base map is not palinspastically restored. Location of schematic cross sections A-A' and B-B' shown.*

present-day folds and thrusts of the Valley and Ridge Province. The present-day exposures within the Valley and Ridge are the result of Mesozoic and Cenozoic uplift and erosion (sometimes up to several kilometers) of the thick sedimentary successions.

CAMBRO-ORDOVICIAN PASSIVE MARGIN SUPERSEQUENCES

A regional discussion of the Cambro-Ordovician passive margin succession is given in Read (1989a, b), which provides regional isopach maps for the succession in the Appalachians and an overview of the regional stratigraphic relationships (Figs. 2A,B, 3, 4, 5A,B).

Supersequence Є-1: Lower Cambrian Chilhowee Group

The latest Neoproterozoic to Early Cambrian Chilhowee Group in Virginia is 100-1500 meters thick (Figs. 2, 3, 4A; Rankin et al., 1989) and is a 2nd-order supersequence of ~30 Ma duration (Fig. 5). The supersequence begins with the synrift Swift Run-Catoctin and lower Unicoi-Weverton succession of conglomerates, felspathic sandstone and bimodal volcanic rocks (Gathright, 1976; Simpson and Eriksson, 1989). Catoctin volcanic rocks are dated at 564+/- 9 Ma (U-Pb zircon, Aleinikoff et al., 1995). Overlying the synrift package are transgressive systems tract deposits of the upper Unicoi-Weverton Formation that are capped by maximum-flood, carbonaceous mudstones at the base of the Hampton-Harpers Formation. The Hampton-Harpers and Erwin-Antietam formations are highstand systems tract deposits (Simpson and Eriksson, 1990). Upward-shoaling siliciclastics capped by drowning surfaces, that characterize the upper Unicoi, Hampton-Harpers and Erwin-Antietam formations, are either 3rd- or 4th-order parasequences (Simpson and Eriksson, 1990). The top of the supersequence is an unconformity (type 2 sequence boundary) near the top of the Erwin Formation (Simpson and Eriksson, 1990).

In north central Virginia, the Unicoi-Weverton formations overlie Grenville basement whereas in southwestern Virginia, this package overlies older synrift volcanic and sedimentary rocks of the Mount Rogers Formation (Simpson and Eriksson, 1989). The disconformable base of the Chilhowee Group may be the Neoproterozoic-Cambrian boundary according to Southworth and Aleinikoff (2007). *Rusophycus* at the base of the upper Unicoi Formation (Simpson and Sundberg, 1987) imply that the majority of the Chilhowee Group is of Cambrian age (less than 545 Ma). The age of the top of the Chilhowee Group is more difficult to constrain, although the presence of Cambrian body fossils, notably the trilobite *Olenellus* from the Erwin-Antietam Formation, and overlying beds at a number of localities in Virginia and northeastern Tennessee (Walcott, 1891; Laurence and Palmer, 1963), imply a late Early Cambrian to early Middle Cambrian age (~510-520 Ma, Shergold and Cooper, 2004).

The Unicoi-Weverton Formation (Appendix A) consists of braided-alluvial conglomerates and feldspathic sandstone up into shallow-marine mudstone-interbedded sandstone to quartz arenite parasequences. Highstand Hampton-Harpers and Erwin-Antietam formations (Appendix A) have carbonaceous mudstone overlain by parasequences of outer- to inner-shelf mudstone grading upwards into inner-shelf to shoreface quartz arenites (Schwab, 1970, 1971, 1972; Simpson and Eriksson, 1989, 1990).

Supersequence Є-2: Lower Cambrian Shady Dolomite-Rome Formation

The Shady Dolomite (southwestern Virginia) or Tomstown Dolomite (northern Virginia) (Fig. 3) is a wedge of carbonate rock ranging in thickness from 100 m in the subsurface in the northwest of the state, to 600 m in outcrops in the southeast along the old continental shelf edge (Figs. 4B, 4C, 5A). The overlying Rome Shale is a southerly and easterly thickening wedge that is 250 m thick in the subsurface in the north (Waynesboro Formation), to 600 m thick in outcrops in the

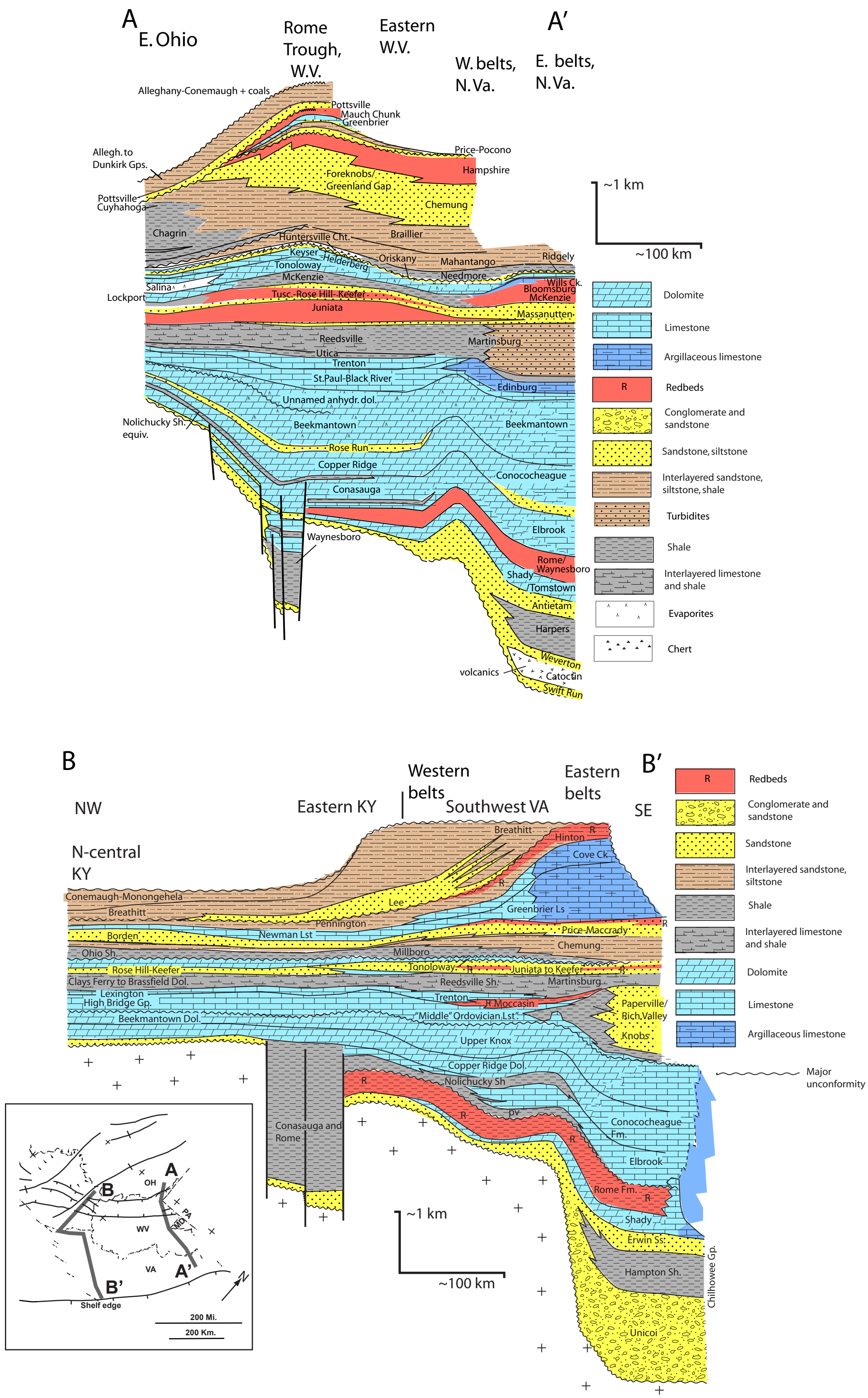


Figure 2. Schematic cross sections of the Paleozoic succession, Virginia showing general relations of units. Thicknesses and horizontal scale approximate; horizontal distances in fold-thrust belt are palinspastically restored. Top: Northern Virginia-Ohio cross section A-A'and Bottom: Southwestern Virginia-Kentucky cross section B-B'. Cross sections compiled from schematic sections of E. Rader in Cosuna Chart (Patchen et al. 1984; Ryder et al. 2009; and Read 1989a).

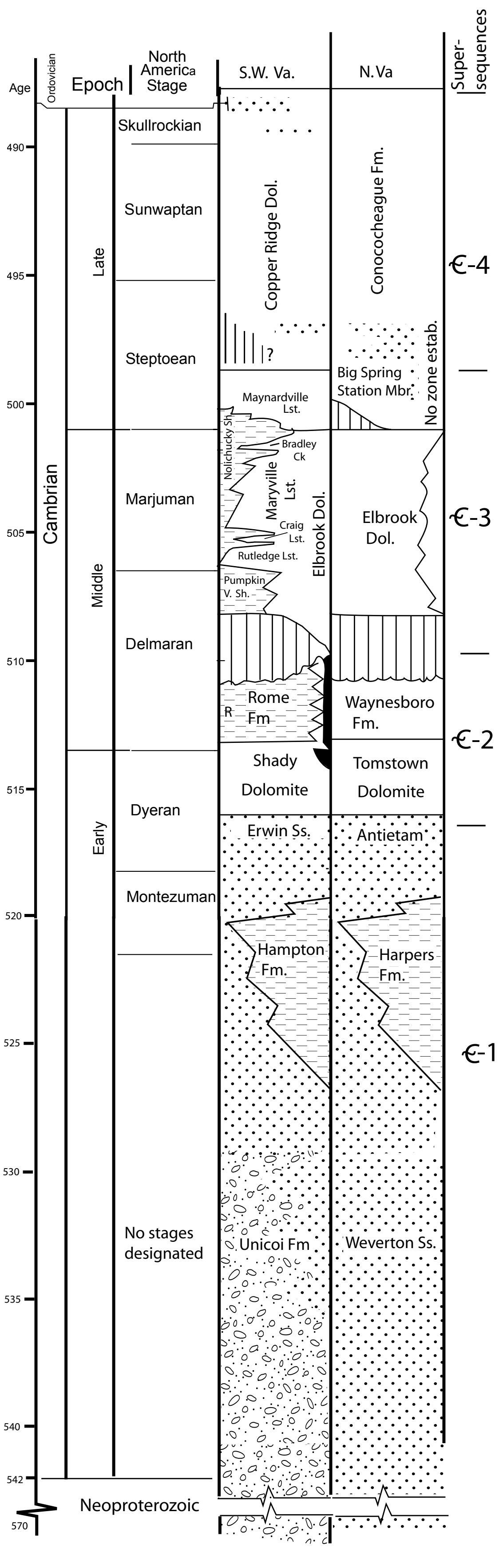


Figure 3. *Chronostratigraphic chart of Cambrian passive margin supersequences (based on Read, 1989a). Conglomerate (open circles and dots), sandstone (dots), carbonates (blank), offshelf facies (black), unconformities (vertical lines).*

south in the Conasauga depocenter. Reverse faulting makes thickness estimates uncertain.

The Shady carbonates in the southeastern belts, Virginia, have been described by Currier (1935), Byrd (1973), Pfeil and Read (1980), and in the subsurface (Barnaby and Read, 1990). Willoughby (1977) assigned the Shady Dolomite and the bulk of the Rome Formation to the late Early Cambrian Bonnia-Olenellus zone, with the uppermost Rome Formation extending into the Middle Cambrian Plagiura-Poliella biozone.

The Shady/Tomstown-Rome succession forms a supersequence, composed of two major sequences. Facies are summarized in Appendix A. The lower sequence has a basal transgressive quartz arenite (Upper Erwin Formation) overlain by 300 m thick, deeper ramp/slope Patterson Member grading up into mudmounds. The overlying Austinville Member (Appendix A) is a white locally oolitic and cyclic dolomite, with at least three thin quartzose units that mark a significant lowstand. These appear to pass downslope into deep water shaly carbonates (Taylor marker).

The Upper Shady Dolomite behind the shelf edge in southwestern Virginia consists of a 300 m thick unit of peritidal, highly cyclic dolomites and lesser, interbedded red clastics (Betzner and Read, 2007). The peritidal carbonates pass eastward into a microbial reefal rim facies which are backstepped relative to the underlying mounds. These shelf edge facies pass downslope into deep water periplatform talus (Barnaby and Read, 1990). Rome Formation cyclic redbeds of the upper sequence are widespread over the platform interior (Harris, 1964; Samman, 1975).

A significant unconformity is developed on top of the Rome-Waynesboro succession (the Hawke Bay event of Palmer and James, 1979). The unconformity spans the Middle Cambrian biozones extending from the top of the *Plaguira-Poliella*, through the Albertella and into the lower *Glossopleura* biozones (Fig. 3; Willoughby, 1977).

The ramp profile which initiated Shady deposition was inherited from the underlying Chilhowee siliciclastic shelf, which would have had a ramp-like profile. Flooding of the shelf shut down siliciclastic supply, and this, possibly coupled with plate migration into warmer latitudes (Scotese, 2001; Meert and Torsvik, 2004), resulted in deposition of the upward shallowing slope and bank margin carbonate facies of the lower sequence. Following incipient drowning, the ramp evolved into a high-relief rimmed shelf, following backstepping of the rim. This rimmed shelf appears to have had at least 1000 m of relief above the adjacent deep basin. It developed during late highstand, when continued rapid subsidence of the margin generated an aggrading rather than a prograding rimmed margin. Global greenhouse climate with high frequency, small precessional(?) sea-level changes generated the highly cyclic peritidal successions. Regional regression (Hawke Bay event) possibly coupled with uplift along the Rome Trough to the northeast, shed red siliciclastics onto the earlier carbonate platform, and culminated in the regional unconformity at the top of the Rome Formation.

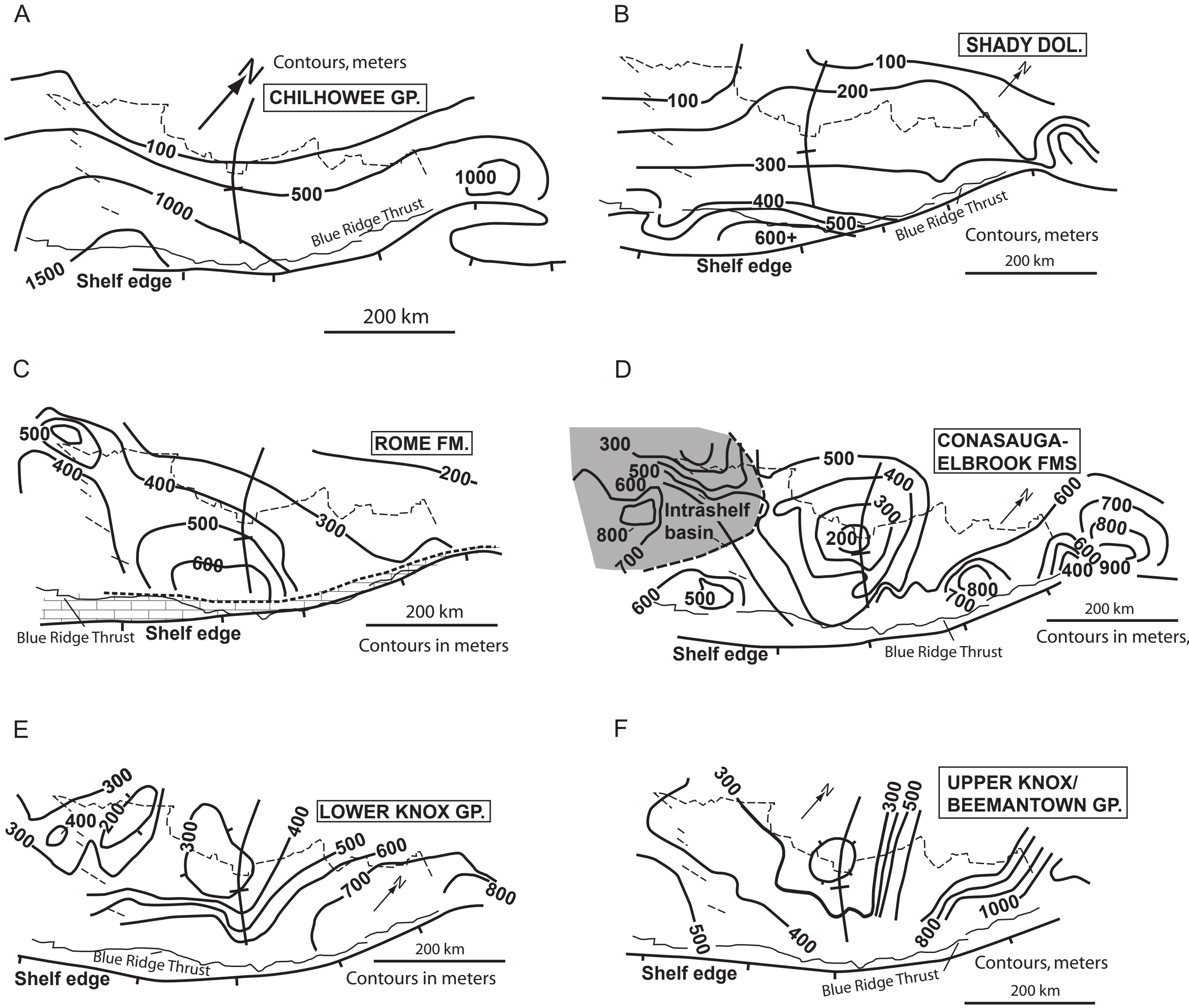


Figure 4. *Isopach maps (palinspastic base) of Cambro-Ordovician passive margin supersequences. A. Chilhowee Group, Early Cambrian. B. late Early Cambrian Shady Dolomite. C. Early to Middle Cambrian Rome Formation; brick pattern is inferred shelf edge carbonate belt. D. Middle Cambrian Conasauga Group. E. late Cambrian Lower Knox Group. F. Early Ordovician Upper Knox-Beekmantown Group. Measured section locations shown on maps in Read (1989a).*

Supersequence Є-3: Middle to Basal Upper Cambrian Elbrook-Conasauga Interval

This supersequence consists of the Elbrook Formation and the laterally equivalent Conasauga Shale (Fig. 3). It is dominated by the development of an intrashelf basin on the Middle Cambrian shelf, bordered toward the shelf edge on the southeast and along strike to the northeast by shallow, peritidal carbonates. The paleontology and stratigraphy of the succession have been described by Derby (1965), and regional facies by Harris (1964), Erwin (1981), Markello et al. (1979) and Markello and Read (1981, 1982) and Koerscher and Read (1989). The supersequence includes the *Albertella* to the *Aphelaspis* biozones (Derby, 1965).

Supersequence 3 is thin (to 200 m) over the Virginia Promontory (Thomas, 1977) but thickens rapidly to over 600 m to the northeast and southeast, into the regional depocenters on the shelf (Figs. 4D, 5). The interval also thickens westward into the Rome Trough, a Middle Cambrian rift system extending through Kentucky, West Virginia and Pennsylvania. The succession is dominated by intrashelf basin shale-prone facies (Conasauga Group) in southwestern Virginia, Honaker Dolomite (an oolitic dolomite facies fringing the basin) and cyclic peritidal carbonate facies (Elbrook Formation) along the eastern part of the shelf bordering the intrashelf basin, and extending throughout the shelf in northern Virginia. Overall, the succession shows gradual onlap onto the craton.

Large scale interfingering of the basin shale and platform carbonates is associated with six to nine 3rd-order sequences from 20 to 100 m thick. Interfingering shale and limestone units include, from base to top, the Pumpkin Valley Shale, Rutledge Limestone, Rogersville Shale (all three confined to

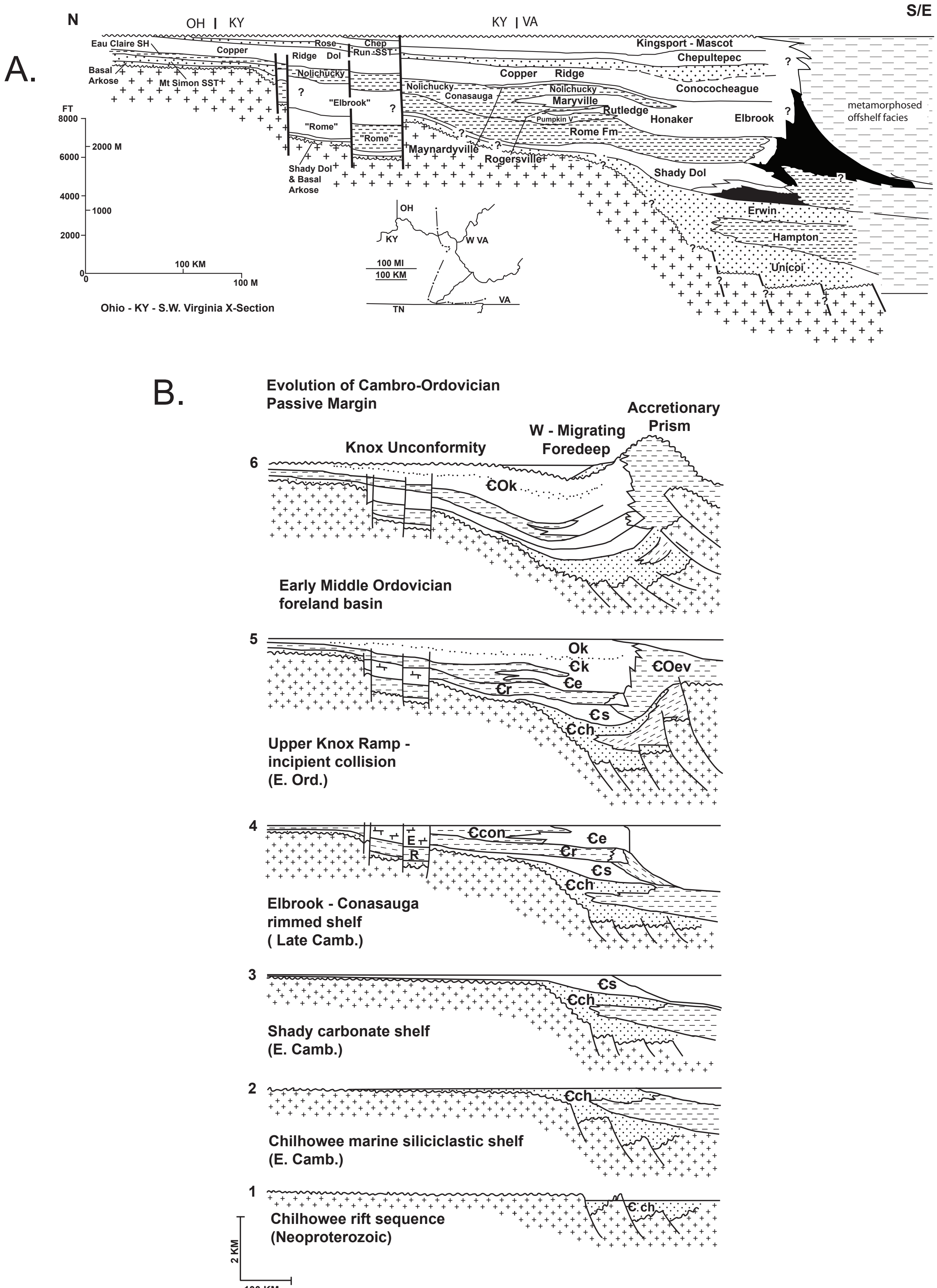


Figure 5. *A. Regional cross section of Cambro-Ordovician passive margin succession (based on Koerschner and Read 1989). B. Evolution and destruction of Cambro-Ordovician passive margin sequence. Єch is Chilhowee Group, Єs is Shady Dolomite, Єr is Rome-Waynesboro formations, Єe and Єcon are Elbrook and Conasauga formations, Єk is Lower Knox Group, Ok is Upper Knox/Beekmantown carbonates, ЄOk is lower and upper Knox Group combined, and ЄOev is Evington Group (modified from Read, 1989b).*

far southwestern Virginia), and the Nolichucky Formation (comprised in southwestern Virginia of the Maryville Limestone, the Lower Limestone, Lower Shale, Middle Limestone, Upper Shale and the Maynardville Limestone). The 3rd-order lowstand carbonates are typically thin deeper marine limestone tongues that extend out into the shale basin. The Middle Limestone is a conspicuous 3rd-order lowstand unit between the Lower and Upper Shales. The limestones backstepped during transgression, and were overlain by intrashelf basin shale. The 3rd-order highstands are manifested by progradational carbonates extending out over deeper water shales. Maximum flooding of the supersequzence probably is at the base of the Nolichucky Upper Shale, and on the platform at the base of the Widener Limestone (a muddy limestone) in the southeast. The upper part of the supersequence shallows upwards through deeper ramp Maynardville carbonates that are capped by Late Cambrian, quartzose peritidal dolomites (Copper Ridge-Conococheague formations) of the next supersequence.

Facies are summarized in Appendix A. The intrashelf basin shales (Nolichucky Lower and Upper Shales), are underlain, overlain and pass updip into deep ramp muddy carbonates (Maryville and Maynardville members). These grade laterally into an oolitic dolomite rim (upper Elbrook and Honaker formations) from 20 to 40 m thick, bordering the intrashelf basin, which in turn grades laterally into peritidal cyclic carbonates (Elbrook Formation).

Rapid downwarping of the Rome Trough to the west (Webb, 1980) and synchronous subsidence of the depocenters localized deposition of shales in the south, fed by the Kerbel "delta" system in Ohio, under the influence of paleotrade winds (Jannsens, 1973; Dalrymple et al., 1985). A long-term, 2nd-order sea-level rise (Bond et al., 1984), with superimposed 3rd-order eustatic sea-level cycles, formed the supersequence and its component sequences, similar to the Grand Cycles of the Great Basin, western United States (Palmer, 1981; Osleger and Read, 1993). Some of the parasequences evident in the basin and on the platform resulted from high frequency, orbitally forced sea-level fluctuations. Basinal shales formed during high sea level, humid phases and a stratified water column, which flooded the adjacent platform to shallow depths to form subtidal parts of parasequences. High-energy subtidal limestone caps to intrashelf basin cycles formed during falling and low sea levels and semi-arid conditions, when storm waves reworked the carbonates, and the platform cycles shallowed to tidal flats culminating in exposure of parasequences.

Supersequence Ꞓ-4: Upper Cambrian Conococheague-Copper Ridge Formations

These successions span the Late Cambrian *Aphelaspis* to *Saukia* biozones, to the base of the Early Ordovician (Fig. 3; Derby, 1965) and generally thicken toward the shelf edge to the southeast, and thin to 300 m over the Virginia Arch (Figs. 4E, 5). The units thicken into depocenters to the north (to 800 m) and to the south (600 m). Facies to the west are dominantly cyclic dolomites (Copper Ridge Dolomite), while those to the east are dominantly cyclic limestone (Conococheague Limestone).

There are four to five 3rd-order sequences of cyclic carbonates in Supersequence Ꞓ-4 (Appendix A). Lowstand cycles have thin quartz sandy intervals in laminate caps; transgressive tracts have relatively thick, dominantly subtidal and thrombolitic cycles while the highstand tracts have relatively thin cycles dominated by well developed microbial laminite caps (Koerschner and Read, 1989; c.f. Demicco, 1985). Quartz sandstone beds are common in laminite caps and at bases of cycles in the lower and upper parts of the supersequence.

By the start of Late Cambrian time, the intrashelf basin had filled to peritidal depths. The Late Cambrian succession was initiated by a major lowstand of sea level, during which basal quartz arenites were deposited, followed by a broadly onlapping and offlapping 2nd-order succession, culminating in lowstand, upper sandy dolomites. Shelf edge facies are not exposed in Virginia, but the regional geology (especially western Maryland; Demicco 1985) suggests that the margin continued to be a high relief rimmed margin into the Late Cambrian, perhaps with up to 800 m of relief (Read, 1989a, b). At least four 3rd-order sea-level cycles are developed. Higher sedimentation rates on the seaward part of the platform probably resulted in more complete shallowing to sea level here, compared to platform interior locations; thus tidal flats may have prograded back in from the margin; this is indicated by westward transition from peritidal into dominantly subtidal transgressive parts of sequences in western Maryland (Demicco, 1985). Westerly facies underwent synsedimentary dolomitization in semi-arid tidal flats and by subsequent reflux.

Supersequence O-1: Lower Ordovician Upper Knox-Beekmantown Carbonates

In Virginia, supersequence O-1 includes the Upper Knox (southwest Virginia) or Beekmantown carbonates (northern Virginia; Fig. 6). The supersequence thins markedly over the Virginia Promontory, and thickens eastward toward the shelf edge. It also thickens northeast to 1000 m and southeast to 450 m within the depocenters (Figs. 4F, 5). Depocenters at this stage extended far back onto the platform, so that isopachs of the supersequence are at high angles to the margin of the shelf. In general, the Cambrian-Ordovician boundary is relatively conformable and is generally placed above the Copper Ridge sands. The supersequence spans Early Ordovician Ibexian to Cassinian, and in northern Virginia continues up into the Middle Ordovician Whiterockian (Patchen et al., 1984, 1985; Read, 1989a, b). The supersequence is capped by the major Middle Ordovician erosional unconformity.

The Early Ordovician margin was a ramp in contrast to the underlying rimmed shelf. This low gradient ramp was characterized by a transition from platform interior cyclic carbonates, passing downslope into cyclic muddy carbonate-thrombolitic facies, and then into thin bedded deep ramp carbonates with storm beds (Appendix A; Oder, 1934; Bova and Read, 1987; Montañez and Read, 1992; Pope and Read, 1997). The lack of megabreccias and reef derived carbonates reflect this ramp morphology.

There are six to eight 3rd-order sequences developed in supersequence O1, the upper two being restricted to the northern depocenter (Read, 1989a, b). The lower three sequences include the Chepultepec/Stonehenge formations (containing the supersequence maximum flood), the middle three sequences

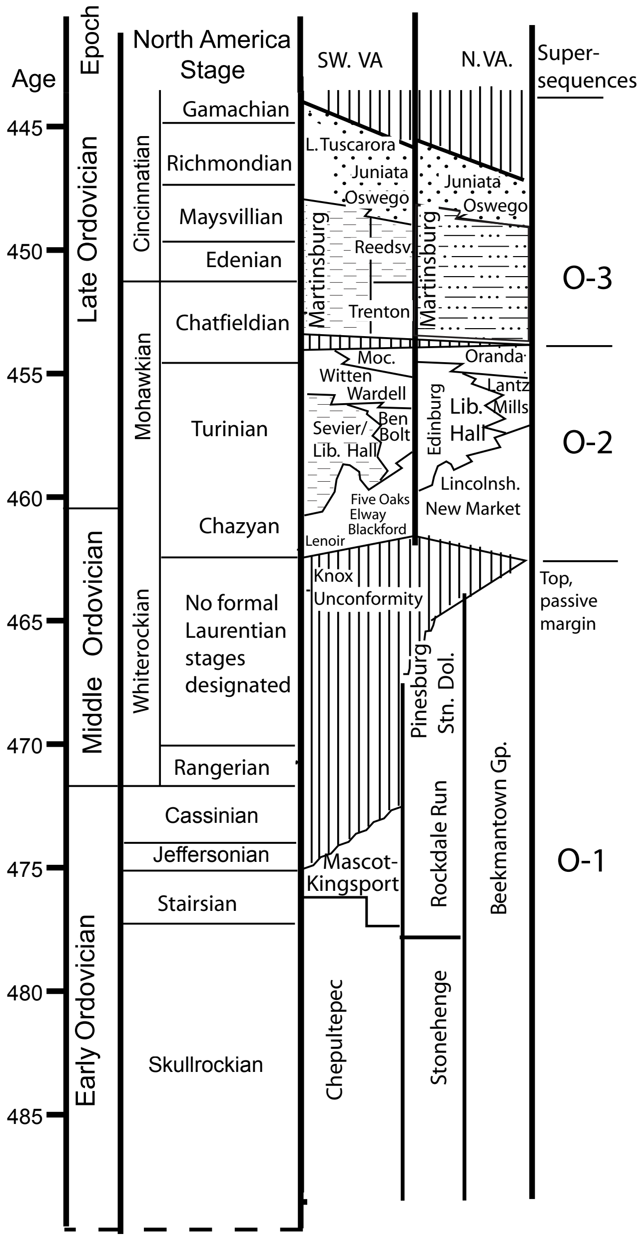


***Figure 6.** Chronostratigraphic chart of Ordovician supersequences (based on Read 1989a; Holland and Patzkowsky, 1996). Sandstone (dots), turbidites (dot-dash), deeper water shale/limestone (dashes), carbonates (blank), unconformities (vertical lines).*

include the Longview-Kingsport-Mascot units (southwest Virginia) and Rockdale Run carbonates (northern Virginia), and the upper two include the Early Middle Ordovician Pinesburg Station Dolomite (northern Virginia). The sequences commonly show sandy or silty dolomite parasequences in the lowstand, thick dominantly subtidal limestone parasequences in the transgressive tract, and predominantly thin, cyclic peritidal dolomite (and silicified evaporite) parasequences in the highstands (Appendix A; Bova and Read, 1987; Montañez, 1989).

Transition from the earlier rimmed shelf into a ramp was probably due to tectonic shallowing of the basin, due to reverse movement on earlier, rift-related listric normal faults during incipient arc-continent collision heralding the early phase of the Taconic orogeny (Rowley and Kidd, 1981; Shanmugam and Lash, 1982; Read, 1989b). The supersequence resulted from a major 2nd-order sea-level cycle, with the maximum flooding in the Chepultepec/Stonehenge interval. Water depths over the Ordovician peritidal ramp were relatively shallow and interior platform facies are low energy mudstones, with widespread tidal flats facies. Higher energy facies developed along the mid-ramp, associated with thrombolite and grainstone shoals, with a storm-influenced muddy deep ramp/slope. Deposition of the passive margin succession was terminated by incipient collision of the Laurentian margin with the Ordovician arc, uplift of the shelf, coupled with eustatic sea-level fall (Fig. 5B; Mussman and Read, 1986).

ORDOVICIAN AND SILURIAN TACONIAN FORELAND BASIN SUPERSEQUENCES

Supersequence O-2: "Middle" Ordovician Supersequence (Upper Whiterockian-mid-Mohawkian) Limestone Interval

Supersequence O-2 ("Middle" Ordovician Limestone to Moccasin interval) overlies the Knox Unconformity, which separates the underlying passive margin succession from the overlying foreland basin deposits (Fig. 6). The so-called "Middle" Ordovician limestone now extends into the Late Ordovician (Fig.6). The unconformity had paleo-relief of up to 150 m, and intense karstification and paleocave formation (Heyman, 1970; Mussman and Read, 1986; Mussman et al., 1988). The supersequence is from upper Whiterockian through the Ashbyan and Black Riveran, with the clastics at the top of the succession extending into the basal Franklinian, based largely on conodonts (Markello et al., 1979) and graptolites (Holland and Patzkowsky, 1996 and references therein). Butts (1940), Cooper and Cooper (1946) and Cooper (1944a, b, 1945, 1960), Cooper and Prouty (1943), Edmundson (1958), Rader and Biggs (1976), Rader and Henika (1978), Rader and Read (1989), Kreisa (1980) and other papers cited in Read (1980), provide the first detailed sedimentary studies of the succession. Read (1980, 1982) provides a regional depositional framework for the interval. Over 50 formation names have been used in mapping the succession in Virginia, which have confused the regional relations.

The Middle Ordovician supersequence is a large scale transgressive to regressive, southeasterly thickening wedge (200 to 650 m) that also thickens to the southwest (to over 1000 m) down the axis of the foredeep (Fig. 7A). The southern depocenter dominated the deposition of the supersequence TST, controlling southwest to northeast migration of the shoreline up the axis of the foredeep, and onlap to the west. This southern depocenter gradually filled, and the highstand tract of the supersequence was dominated by the unfilled northern depocenter. Thus, regression was both to the southeast into the basin, and to the northeast down the northern basin axis.

Up to five sequences occur in the Middle Ordovician supersequence (Holland and Patzkowsky, 1996). Facies are summarized in Appendix A. It is dominated by peritidal to deep ramp limestones on the ramp in the northwest and deep basin shales and local coarse siliciclastics in the southeast (Fig. 8A). Detrital dolomite interbedded with limestone and shale (Blackford Formation) occurs at the base of the supersequence in the west, eroded from the exposed Early Ordovician Knox carbonates.

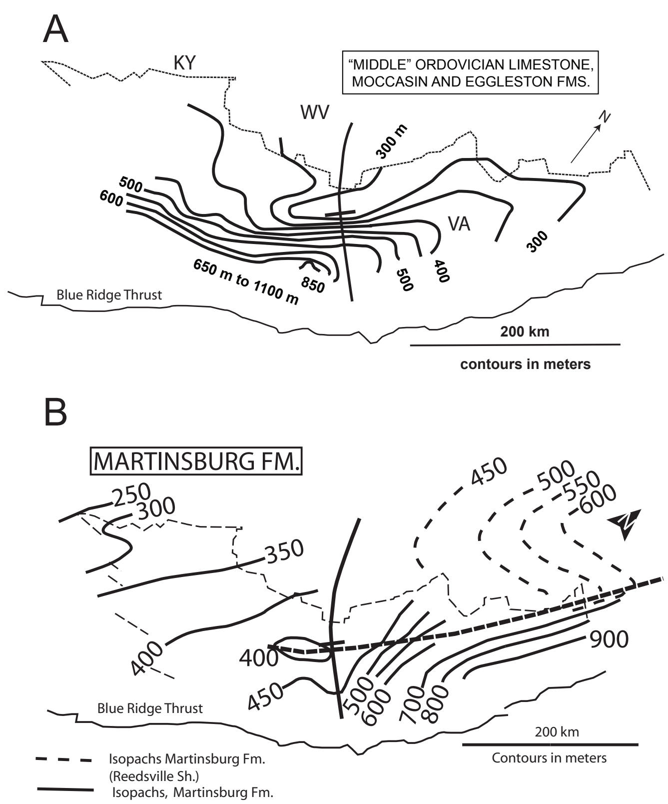


***Figure 7.** Isopach maps (palinspastic base) of Middle to Late Ordovician supersequences. A. Middle Ordovician Limestone supersequence (modified from Read, 1980). B. Later Ordovician Reedsville/Martinsburg formations; dashed lines are Reedsville Shale; continuous lines are Martinsburg Formation which includes turbidite siliciclastics in northern Virginia and shelly storm beds, siltstone and shale in southwestern Virginia. Heavy dashed line is approximate location of axis of thinning. Figure modified from Kreisa, 1981 and Woodward, 1951 (in Diecchio, 1985b).*

These facies are overlain by thick peritidal parasequences (New Market, Elway and Five Oaks limestones (Grover and Read, 1978; Read and Grover, 1977). Crinoid-bryozoan banks developed along the mid ramp and deep ramp margin in the transgressive tract. With flooding, deep ramp shaly carbonates (Ben Bolt Formation) and black slope limestone and graptolitic shale (Liberty Hall, Rich Valley, Paperville formations) were deposited. In the northwest, the highstand consists of an overall upward shallowing succession from deep ramp carbonates up into peritidal carbonates capped by redbeds (Benbolt to lower Moccasin interval). Third-order sequence boundaries in the succession occur on tops of commonly microkarsted, peritidal carbonate units such as on the Mosheim, Five Oaks, Gratton units or beneath the thin redbeds of the Bowen Formation (Holland and Patzkowsky, 1996).

Along the proximal basin margin to the southeast, in the Abingdon area, thin shallow marine limestones (Mosheim-Lenoir limestones) are overlain by basinal Paperville Shale, and then by 750 to 1300 m thick, upward coarsening successions of channelized deep water fan deposits (Knobs Formation; Raymond et al., 1979). Polymictic channel-filling conglomerates locally occur in the lower Knobs Formation and in the Fincastle Conglomerate at the top of the Liberty Hall basinal shale (Karpa, 1974; Gao and Eriksson, 1991).

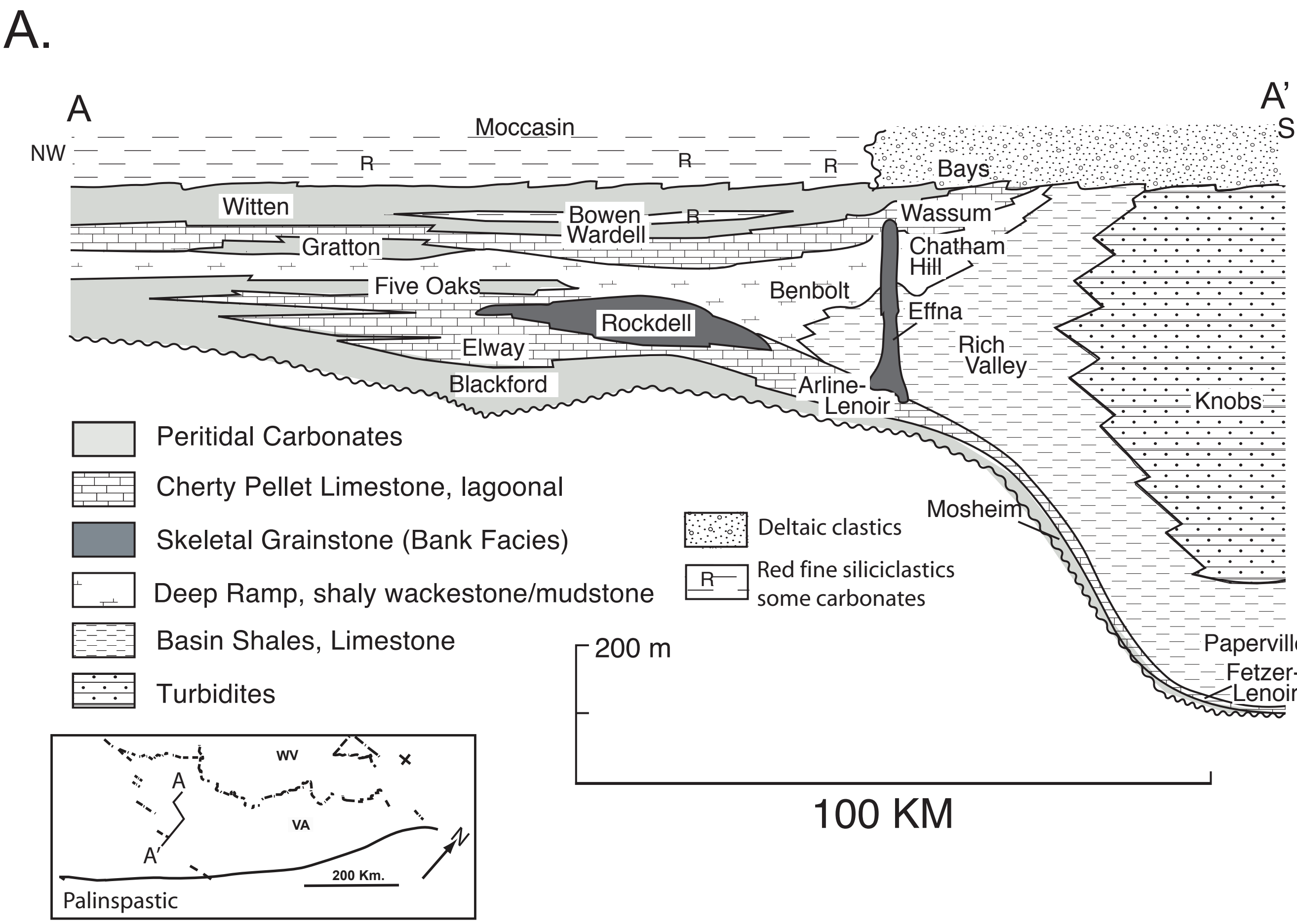
The Moccasin redbeds contain thin bentonites, and are overlain by and interfinger with nearshore shaly lime mudstone (Eggleston Formation), grading northeast into relatively deep ramp and slope facies (Lantz Mills, Liberty Hall) shallowing up into Oranda shallow water carbonate and siltstone, and eastward into locally thick deltaic immature sandstone and mudrock (Bays Formation; Hergenroder, 1966). A local disconformity is present within the Moccasin redbeds at the base of the Walker Mountain sandstone, a thin conglomerate and sandstone in southwestern Virginia (Haynes and Goggin, 1993). Based on bentonite correlations, a supersequence boundary occurs stratigraphically higher in states to the west of Virginia. This suggests that the disconformity in southwest Virginia was influenced by tectonics and is of local significance, and that the supersequence boundary is above the Moccasin-Eggleston units and equivalent peritidal carbonates to the west (Holland and Patzkowsky, 1996; Haynes et al., 2005).

Initial transgression of the highly karsted Knox surface was from southwest to northeast, as well as gradually onlapping to the west. The Virginia promontory or arch (Thomas, 1977) appears to have broadly separated the two basins, and a distinctive orogen-parallel arch (Tazewell arch, Read, 1980) was associated with flexure of the downgoing plate, as was the Cincinnati Arch, assumed to be the peripheral bulge (Ettensohn, 1994). A major recess in the ramp margin in southwest Virginia (Read, 1980) likely was associated with a basement transform or transfer fault inherited from the rift phase (cf. Thomas, 1977). The large ramp margin- and downslope-banks of the supersequence TST formed during sea-level rise and rapid subsidence; they pass southwest in the Tennessee into Holston reef complexes adjacent to the deeper basin (Ruppel and Walker, 1984).

The supersequence highstand generally prograded southeastward into the foredeep, but subsequent highstand filling of the foreland basin in Tennessee (to the southwest), was accompanied by deposition of Bays delta plain clastics and Moccasin mudflat facies that graded westward into carbonates peritidal (Kreisa, 1980). To the northeast, basin deposition continued throughout much of the highstand. The switch from the southern depocenter to the northern depocenter during deposition probably was controlled by timing of collision, with the early Blountian phase causing subsidence in the south, and the later classic Taconic collision heralding onset of rapid subsidence in the north (Rodgers, 1971; Lash, 1988). The basin was highly stratified and anoxic below about 50 m water depth, and basin deposition was dominated by siliciclastic detritus from the accretionary prism, associated with arc-continent collision, along with contribution of fine carbonate from the ramp to the west. Rebound uplift of the proximal foreland generated the sub-regional disconformity with its local conglomerates in the Moccasin Formation, which slightly predated the regional development of the supersequence boundary on top of the High Bridge Group, Kentucky (Holland and Patzkowsky, 1996; Haynes et al., 2005).

Supersequence O-3: Upper Middle Ordovician (upper Mohawkian) to Upper Ordovician (Richmondian) Martinsburg to Basal Tuscarora Formations

Supersequence O-3 in Virginia is an overall upward-coarsening succession that includes the Martinsburg (Franklinian



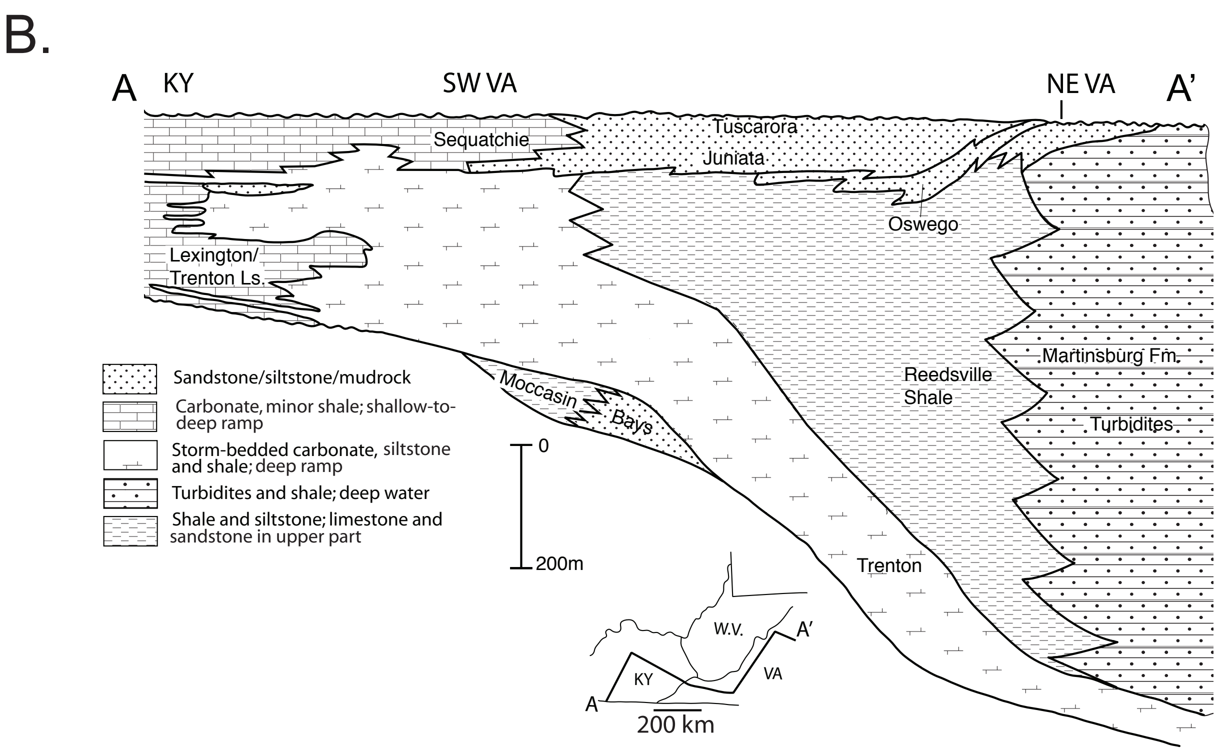


Figure 8. *Regional cross sections of Middle and Late Ordovician supersequences. A. "Middle" Ordovician Limestone supersequence (modified from Read, 1980). B. Late Ordovician supersequence (Martinsburg Formation/Trenton-Reedsville formations) (modified from data in Kreisa, 1981 and Diecchio, 1985a).*

to Maysvillian), Oswego and Juniata (upper Maysvillian to Richmondian) and "lower" Tuscarora formations (Richmondian to Gammachian?)(Fig. 6; Kreisa, 1980; Ross et al., 1982; Diecchio, 1985a, b). At the base, the supersequence also may include the transgressive upper Moccasin redbeds and overlying Eggleston near-shore limestones (Kreisa, 1980; Pope and Read, 1998), but for convenience, these are included in the underlying supersequence (Holland and Patzkowsky, 1996). Four to five major sequences are recognized in the proximal foreland basin of Virginia; one in the lower Martinsburg Formation ("Trenton"), another in the upper Martinsburg ("Reedsville") and the remainder in the Oswego/Juniata to lower Tuscarora interval (Holland and Patzkowsky, 1996; Diecchio and Broderson, 1994; Pope and Read, 1997). Facies are summarized in Appendix A.

Kreisa (1980) and Diecchio (1985a, b) have described the Martinsburg Formation in southwestern Virginia and northern Virginia, respectively. Although thicknesses in the eastern deformed belts are uncertain, the Martinsburg Formation thickens from 250 m in the southwest to 1000 m down the axis of the foreland basin to the northeast (Figs. 7B, 8B). It thins over a northeast trending arch (Shenandoah axis; Rader and Perry, 1976a, b) which separates the thick Virginia successions from those west of the arch.

The Martinsburg Formation in southwest Virginia is dominated by deep ramp storm bedded units. The lower Martinsburg Formation in western belts (Kreisa, 1981; Fig 8B) has a transgressive (0 to 40 m) proximal ramp, shelly limestone at the base, overlain by 140 m of shale-poor, calcarenite/calcisiltite storm beds. These limestone-prone units pinch out to the southeast. They are overlain by broadly onlapping, storm bedded quartz-silty calcarenite/calcisiltite and shale. In the middle Martinsburg Formation, a tongue of calcarenite-calcisiltite extends southeast, marking the correlative conformity between the two Martinsburg sequences. The upper Martinsburg sequence in the west is storm-bedded carbonate-siliciclastics. This grades southeastward into an upward coarsening succession of shale with carbonate storm beds along with sandstone and siltstone that increase up-section.

In the eastern belts of northern Virginia, basinal black limestone and shale occur in the lower Martinsburg Formation (Rader and Biggs, 1976), grading westward into shallower carbonates and shaly carbonates (Nealmont-Dolly Ridge formations). These are overlain by thick, siliciclastic turbidite facies (McBride, 1962; Diecchio, 1985a, b) that grade west and southwest into shaly limestone (Trenton equivalent), and the overlying Reedsville Shale. The Reedsville Shale consists of gray calcareous shale passing up into interbedded fossiliferous siltstone, shale and limestone, and then into immature sandstone. Upper Reedsville beds contain the distinctive *Orthorhynchula* assemblage zone and Lingula (Diecchio, 1985a).

The overlying Oswego-Juniata siliciclastics in Virginia are described by Kreisa (1980) and Diecchio (1985a, b). These units form late highstand deposits above the Martinsburg Formation throughout most of Virginia, and pass southwest into Tennessee into the Sequatchie carbonates. The Oswego Sandstone is 0 to 150 m thick, with a depocenter in the western part of northern Virginia (Fig. 6). The Oswego Sandstone interfingers with and is overlain by the Juniata Formation, except in the east, where the Juniata Formation is erosionally truncated beneath the Tuscarora Sandstone and locally has a polymictic conglomerate (Rader and Perry, 1976b).

In southwest Virginia, the offshore to nearshore transition spans the Martinsburg-Juniata boundary consisting of lower- to middle-shoreface sandstones capped by foreshore sandstones (Kreisa, 1980). The Juniata Formation is 0 to 250 m thick, thickening northward into a depocenter in northern West Virginia. It consists of red and lesser green shale, mudstone, siltstone and sandstone. The Juniata Formation is typically unfossiliferous except for some Lingula at the base, and contains abundant trace fossils. It commonly has parasequences 0.3 to 5 m thick, consisting of marine, channeled sandflats up to mudflats (Kreisa, 1980; Diecchio, 1985a, b). Transgressive surfaces commonly have bedding-plane trace fossils. Paleocurrents for the Oswego-Juniata show high dispersion but with generally westerly modes (Diecchio, 1985a).

The Sequatchie Formation (up to 90 m thick) is a peritidal limestone unit in southwestern Virginia that is a western equivalent of the Juniata siliciclastics (Kreisa, 1980). The lower two thirds consist of storm-influenced nearshore ramp, thin bedded gray red mudstone, skeletal limestone and fine siliciclastics. The upper third of the unit lacks marine biotas, and consists of peritidal mudcracked or burrowed mudstone, and thin-bedded very fine intraclastic sandstone-siltstone.

The "lower" Tuscarora Sandstone gradationally overlies the Juniata redbeds, the Tuscarora Sandstone being free of any red units. The "lower" Tuscarora Sandstone is a trough cross-bedded quartzarenite with thin mudstones and abundant Skolithos and Cruziana trace fossils (Bambach, 1987; Dorsch et al., 1994). Its top is a significant unconformity, the Tuscarora unconformity of Dorsch et al. (1994).

Martinsburg deposition was dominated by the northern depocenter which localized the basin in northern Virginia. This basin was deep, and an upward-coarsening succession of turbidites was deposited. Turbidity currents in the Martinsburg basin flowed northeastward, down the basin axis (parallel to present structural strike), presumably turning down the basin axis after coming in from the southeastern tectonic highlands (McBride, 1962; Diecchio, 1985a, b). The deep basin was bordered to the southwest and west by a storm-influenced mixed carbonate-siliciclastic ramp (Kreisa, 1981). The Martinsburg ramp in southwestern Virginia consisted of a progading siliciclastic marine shelf, with mixed carbonate-siliciclastic storm-facies forming further offshore (Kreisa, 1981). These deep ramp facies were overlain by prograding shoreface clastics and then by intertidal mudflat facies (Oswego-Juniata units).

The "lower" Tuscarora unit appears to mark a renewed transgressive phase prior to erosion. Both glacioeustatic (Dennison, 1976; Bambach, 1987) and tectonoeustatic (Dorsch et al., 1994) origins have been proposed for the origin of the unconformity on the lower Tuscarora Sandstone. It formed during glacioeustatic sea-level lowering related to the latest Ordovician (Hirnantian) glaciation and isostatic rebound following erosional removal of the Taconic thrust slab.

POST-TACONIAN FORELAND BASIN SUPERSEQUENCES

Supersequence S-1: Silurian Tuscarora/Rose Hill/ Keefer Formations

The Silurian Supersequence S-1 (Fig. 9) in Virginia ranges up to 200 m in northern Virginia, thinning to 60 m in southwestern Virginia. The supersequence consists of, in ascending order: 1) the "upper" Tuscarora Formation (lower Llandovery age) and correlative Clinch Formation in southwestern Virginia that is up to 200 m thick in northern Virginia, thinning to the southwest; 2) the Rose Hill Formation (upper Llandovery/Wenlock) that thickens from less than 20 m in the southeast to 150 m in the northwest; and 3) the "Eagle Rock/Keefer" sandstone (Wenlock/Ludlow) that is up to 125 m in the southeast and interfingers with the Mifflintown and Bloomsburg formations (up to 40 m thick) in northern Virginia (Fig. 10A; Lampiris, 1975; Diecchio, 1985a, b; Diecchio and Dennison, 1996). Biostratigraphic constraints above the lower Rose Hill Formation are based on the brachiopod *Eocoelia*, together with ostracodes and conodonts. Facies are summarized in Appendix A.

Supersequence S-1 (~20 million years duration) records drowning followed by upward shallowing. The upper Tuscarora Formation and lower half of the Rose Hill Formation, including the lower hematite member, are the transgressive systems tract of the supersequence whereas the upper Rose Hill Formation, including the upper hematite member, and the Eagle Rock/Keefer record highstand progradation (Meyer et al., 1992). Maximum flooding coincides with the *Eocoelia* C5-C6 time boundary recognized by Lampiris (1975) in the middle of the Rose Hill Formation.

The upper Tuscarora Formation rests with a marked angular unconformity on Upper Ordovician Supersequence O-3 throughout the Valley-and-Ridge Province; in successive thrust sheets towards the southeast, the upper Tuscarora Formation progressively truncates the lower Tuscarora, Juniata and Martinsburg formations (Fig. 11A; Dorsch et al., 1994). Biostratigraphic data for the Tuscarora Sandstone are lacking but the basal unconformity coincides with the Ordovician-Silurian boundary (e.g., Dorsch et al., 1994). Dennison et al. (1992) have estimated up to 11 million years of missing time in the region around Catawba. The upper Tuscarora Formation has a local basal incised-valley conglomerate (cf., Castle, 1998, 2001). These are overlain by lower shoreface sandstones (Dorsch and Driese, 1995) that are fining- and thinning-upward with diverse trace fossils. A lower shoreface environment, in which braided alluvial sediments were reworked by storm processes, is inferred by a proximal (southeasterly) braided alluvial facies belt identified in Pennsylvania which may not be represented in the outcrop belt of Virginia.

The Rose Hill is conformable on the Tuscarora Formation and consists of more offshore, shallow-marine heterolithic hematite-cemented quartz-rich sandstone, interbedded shale and siltstone, and fossiliferous shale; the sediments have abundant trace fossils and are intensely bioturbated (Dorsch and Driese, 1995). The hematite-cemented sandstones define lower and upper members that coalesce to the southeast (Fig. 11A). The Rose Hill Formation pinches out to the southeast where the Tuscarora and Eagle Rock/Keefer formations coalesce to form

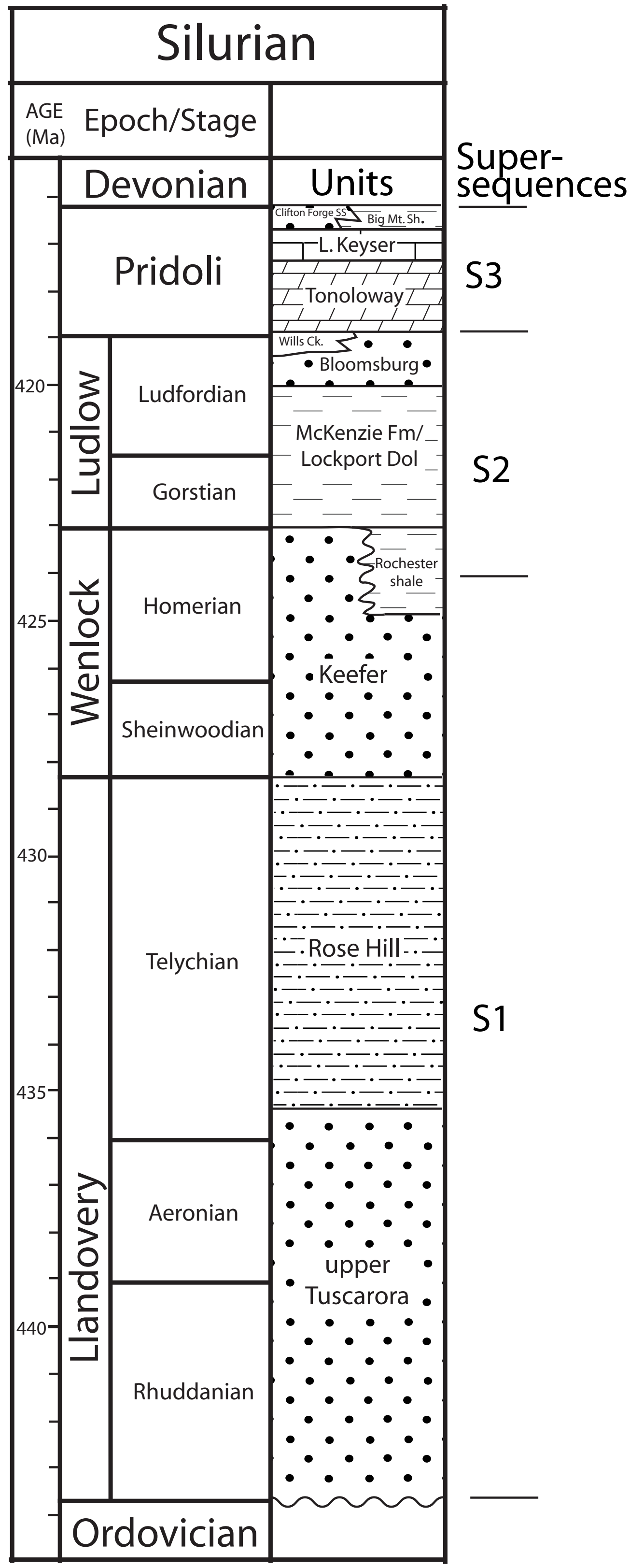


Figure 9. *Chronostratigraphic chart of Silurian supersequences (based on Sartain, 1981; Patchen et al., 1985). Sandstone (dots), mixed siliciclastics (dot-dash), shale (dashes), limestone (brick pattern), dolomite (inclined brick pattern).*

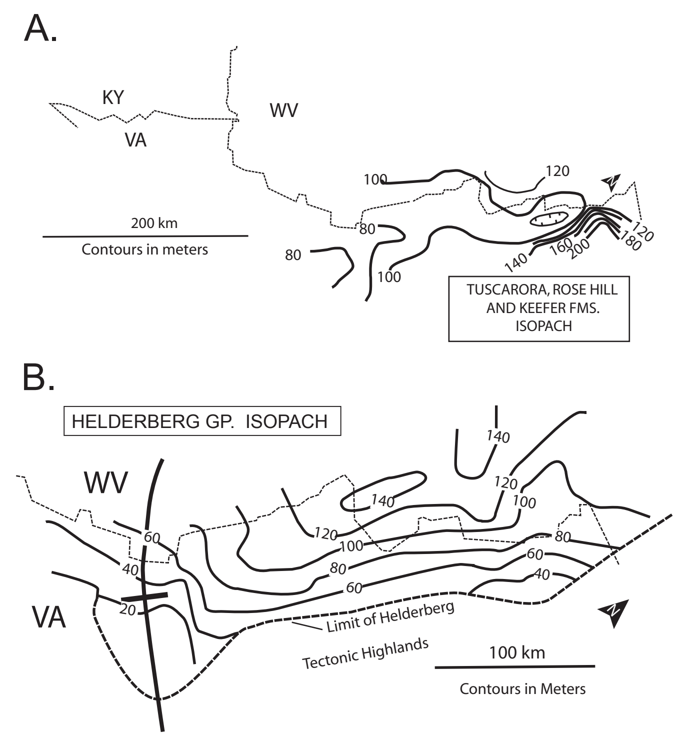


***Figure 10.** Isopach maps (palinspastic base) of Silurian-Early Devonian supersequences. A. Silurian Tuscarora-Rose Hill-Keefer formations (modified from Diecchio, 1985). B. Late Silurian-Early Devonian Keyser-Helderberg Group carbonates (modified from Dorobek and Read, 1986).*

the Massanuttan Sandstone (Diecchio and Dennison, 1996).

Quartz arenites dominate the shallow-marine Keefer/Eagle Rock formations which have *Arthrophycus*-like track ways, burrows and desiccation cracks. The Mifflintown Formation mapped in west central Virginia, is a distal facies of the Eagle Rock Formation and consists of limestone and shale with sandstone interbeds becoming common to the southeast (Lampiris, 1975; Diecchio and Dennison, 1996). The Mifflintown Formation also extends up and includes units within the McKenzie Formation elsewhere (Patchen et al., 1985).

Supersequence S-2: Upper Silurian Rochester-McKenzie-Bloomsburg interval

The Rochester Shale is a thin 0-5 m thick unit at the base of the McKenzie Shale in northern Virginia (Smosna and Patchen, 1978). The McKenzie Formation is a maximum of 45 m thick in westernmost Virginia, and consists of shale with thin limestone interbeds (Butts 1940; Smosna and Patchen, 1978). The Bloomsburg Formation is 10 to 120 m thick in northern Virginia, pinching out below a major unconformity in southwestern Virginia. It consists of non-marine red mud rocks and red sandstone, and grey-green sandstone and shale (Gathright et al., 2003). It grades southwest into sandy shoreline facies of the Williamsport Sandstone (0 to 35 m), trending northwest-southeast, parallels to the edge of exposed land in southwestern Virginia (Smosna and Patchen, 1978). To the west, its top grades into the Wills Creek and into the "upper Keefer" Sandstone to the southwest.

Supersequence S-3: Upper Silurian- Wills Creek-Tonoloway-"Lower Keyser" Interval

Supersequence S-2 consists of the Wills Creek, Tonoloway and Lower Keyser formations (upper Upper Silurian, Ludlovian to Pridolian; Fig. 9); age constraints are based on brachiopods, corals and conodonts (Head, 1974; Helfrich, 1975; Cook, 1981; Sartain, 1981; Patchen et al., 1985).

Facies are summarized in Appendix A. The basal unit in supersequence S-3 is the Wills Creek Formation (Gathright et al., 2003) which overlies the Keefer-Bloomsburg-Williamsport siliciclastics. The Wills Creek is largely confined to western Virginia, and is typically 0 to 120 m thick, and consists of thin bedded, fine sandy limestone, shale and mudstone with locally common brachiopods and ostracods (Butts, 1940; Gathright et al., 2003). Well-data from eastern West Virginia (Smosna et al., 1977) indicates that the Wills Creek is a single 3rd-order sequence, with a basal sequence boundary on the underlying thin Williamsport Sandstone or Bloomsburg redbeds. The Wills Creek Formation consists of a lower transgressive interval of interbedded limestone and shale (increasing gamma ray); pebbly sandstone and sandy limestone occur in the lower 15 m of the equivalent Hancock Dolomite, western southwest Virginia (Butts, 1940). The transgressive Wills Creek beds grade up into highstand limestone and shaly limestone (relatively low gamma ray), and then up into dolomite and sandstone.

The overlying uppermost Wills Creek beds above the Wills Creek sandy interval appear to be early transgressive deposits of the Upper Silurian Tonoloway Limestone. The Tonoloway Limestone, which is 0 to 110 m in Virginia, is developed in the northwest part of the Valley and Ridge, and is described by Bell and Smosna (1999). This carbonate-prone unit overlies the siliciclastic Wills Creek Formation or in the east, the Bloomsburg redbeds. Data in Bell and Smosna (1999) suggest that the Tonoloway Limestone is a 3rd-order sequence (1 to 3 m.y. duration) composed of two minor sequences. The lower minor sequence in the Tonoloway Limestone consists of transgressive peritidal sabkha and lagoonal carbonate facies, overlain by highstand thrombolitic units capped by peritidal sabkha facies. Lowstand to transgressive siliciclastic-prone sabkha carbonates and siliciclastics occur at the base of the Tonoloway upper minor sequence, and are overlain by highstand deep ramp, shallow ramp, and then peritidal sabkha and lagoonal carbonates. The deep ramp carbonates in the southwest pass updip to the northwest into shallow ramp and sabkha facies, possibly associated with an east west trending fault system.

The Lower Keyser carbonates make up the top of the upper Silurian supersequence, and are disconformable on underlying units only along the depositional basin margin in southwest Virginia. The Lower Keyser mainly consists of deeper ramp limestones grading downdip into shaly cherty carbonates. Its upper boundary is a karstic unconformity in the midcontinent, between Silurian and Devonian strata.

The Wills Creek-Tonoloway sequence formed during a period of very arid climate, indicated by leached evaporites in outcrop, and subsurface evaporites and salt in the Appalachian and Michigan basins (Smosna et al., 1977). This low rainfall setting coupled with long-term sea level rise, may have decreased siliciclastic influx into the basin, favoring carbonate

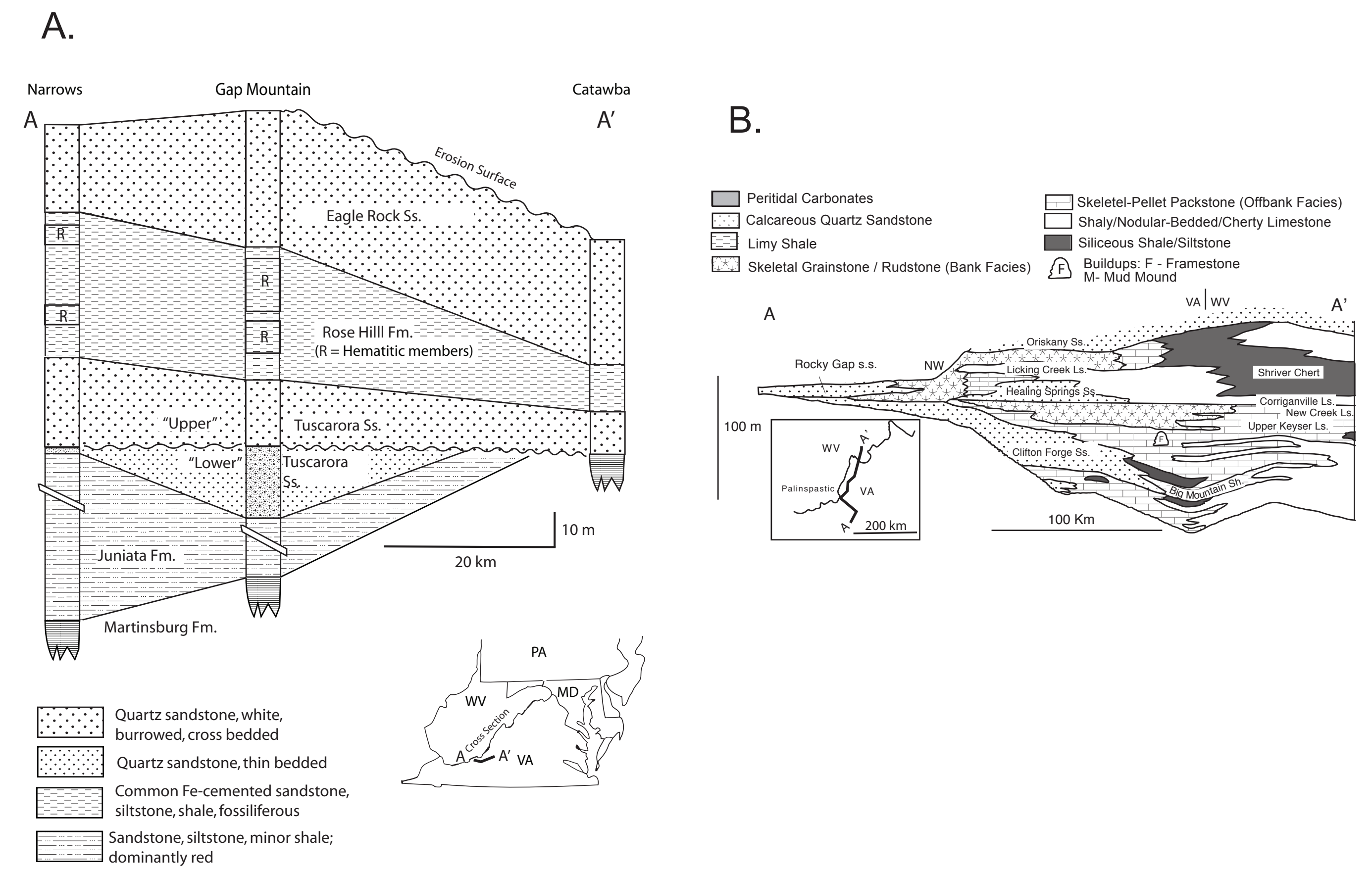


***Figure 11.** Regional cross sections of Silurian-Early Devonian supersequences. A. Cross section of Silurian Tuscarora-Rose Hill-Keefer formations (modified from Lampiris, 1975; Dorsch et al., 1994). B. Cross section of Late Silurian-Early Devonian Helderberg Group carbonates comprised of Lower and Upper Keyser and Helderberg Limestone formations (modified from Dorobek and Read, 1986).*

deposition. Bell and Smosna (1999) suggest that the Tonoloway ramp had very low slope (few cms/km), but that downwarping during middle Tonoloway deposition promoted deep ramp sedimentation until the basin shallowed, with salt deposition in northwest West Virginia (Smosna et al., 1977). Subsequently, climate appears to have become more humid and waters more open marine during Lower Keyser time.

Supersequence D-1: Upper Keyser/Helderberg-Oriskany Interval

The Early Devonian "Upper Keyser"-Helderberg Group in Virginia (Fig. 12) has been described by Head (1974) and Dorobek and Read (1986 and references therein). It overlies the Upper Silurian "Lower Keyser" Formation (Fig. 11B). It is a mixed carbonate siliciclastic succession 0 to 100 m thick; the isopachs increase to the northwest and the northeast, with the greatest thicknesses in the northern part of Virginia (Fig. 10B). The basin was closed to the southwest where the unit is absent. The Devonian Helderberg-Oriskany spans 8–10 m.y. and contains two 3rd-order sequences. The Clifton Forge Sandstone/Big Mountain Shale (uppermost Silurian) form the base of Supersequence D-1. Sequence 1 consists of the Clifton Forge sandstone/shale and "Upper Keyser" Limestone, and New Creek Limestone-Elbow Ridge Sandstone. Helderberg sequence 2 includes the Healing Spring Sandstone, Corriganville Limestone, the Licking Creek Limestone, Mandata Shale/Shriver Chert; it is capped by the Ridgeley/Oriskany Sandstone. Facies are summarized in Appendix A.

The sandstones (Clifton Forge, Elbow Ridge/Healing Springs, and Rocky Gap sandstones) are locally developed, late highstand to early transgressive bodies, which are localized and occur adjacent to the southwestern and eastern depositional basin margins (Fig. 11B; Dorobek and Read, 1986). They range from many tens to 150 kms in width and are a few meters to 40 m thick. The sandstones include nearshore, tidal sandstones to offshore bar sandstones and interbedded local bafflestones. Input sources were inherited from the Silurian (Dennison, 1970; Smosna and Patchen, 1978).

Pelmatozoan fringing banks 3 to 12 m thick and local mudmounds formed offshore from sand bodies and along the regional eastern shoreline. Further offshore, stromatoporoid-coral reefs developed during periods of shoaling (Smosna and Warshauer, 1979; Stock and Holmes, 1986). Deep ramp argillaceous carbonates (15 to 25 m water depths) formed seaward of the banks, while near or below storm wave base, fossiliferous limy shales, 0 to 12 m thick were deposited. Deeper water facies (25 to 50 m water depths) include nodular, shaly or cherty storm-bedded, spicule-bearing limestone up to 20 m thick with diverse assemblages and common hardgrounds.

The Oriskany or Ridgeley Sandstones generally conformably overlie the Helderberg carbonates or, where these are absent, unconformably overlie Keefer Sandstone; they have been described by Diecchio (1985a, b) and Seilacher

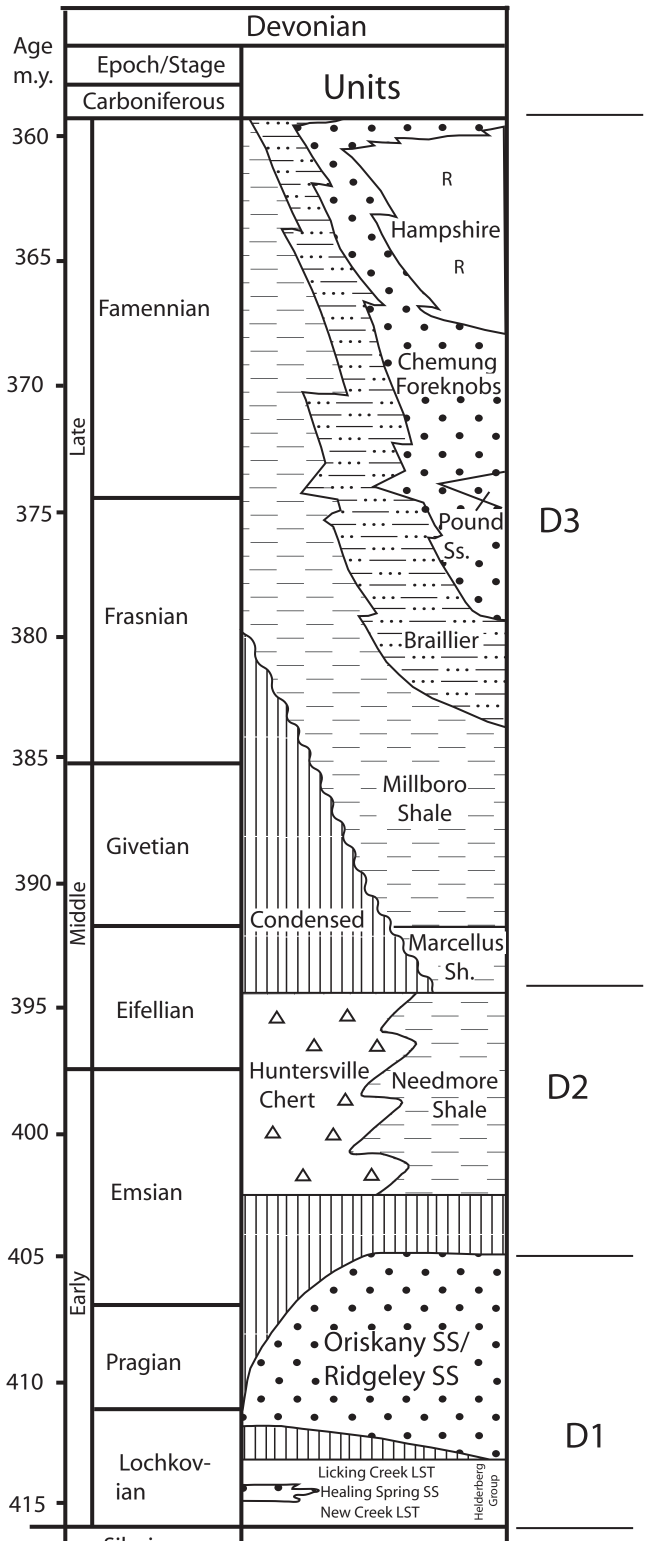


Figure 12. *Chronostratigraphic chart of Devonian supersequences (based on Cook, 1981; Sartain, 1981; Patchen et al., 1985). Redbeds (R), sandstone (dots), mixed siliciclastics (dot-dash), shale (dashes), and chert (open triangles).*

(1968). They are thin or absent in southwestern Virginia; here, local sandstones have been included in the Wildcat Sandstone (Dennison and Head, 1975; Jones, 1982). The Oriskany Sandstone thickens to over 60 m in northwestern Virginia and has its depocenter in western Maryland (Patchen, 1968). It is a calcareous quartz arenite that is locally silica-cemented, with heavy shelled brachiopods and some snails. In outcrop, fossils are leached to form a fossil-moldic quartz arenite. The Oriskany/Ridgeley sandstone progressively overlies older units of Silurian and late Ordovician age to the east. Its eastern limit coincides with the outcrop belt, and beyond this, it has been eroded following Alleghanian deformation. The Oriskany Sandstone further north (in Pennsylvania) is a transgressive marine sand, deposited in upper shoreface (above fair-weather wave-base) to lower shoreface (near shallow storm wave base; Seilacher, 1968; Abplanalp and Lehmann, 2004). The Oriskany/Ridgeley sandstones are overlain with regional unconformity by Needmore Shale or Huntersville Chert in Virginia.

The Devonian Helderberg-Oriskany succession resulted from eustatic fluctuations, coupled with relatively low subsidence rates and tectonic quiescence. Slopes on the younger, Helderberg ramp were 10 to 15 cm/km, and subsidence rates continued to be low (1 to 2 cm/k.y.) and compatible with this tectonically quiescent time. Siliciclastics entered the basin from a few point sources draining the accretionary prism, generating lobe-shaped late highstand to early transgressive sands. Low sediment influx from the accretionary prism during highstands allowed the carbonates to form a fringing-bank complex and localized near-shore sand bodies just offshore from the tectonic highlands. Maximum deepening is marked by the Shriver Chert. The disconformity on the Oriskany Sandstone marks a major sea-level drop, the Wallbridge discontinuity (Wheeler, 1963; Dennison and Head, 1975).

DEVONIAN-LOWER MISSISSIPPIAN ACADIAN FORELAND BASIN SUPERSEQUENCES

Supersequence D-2: Middle Devonian Huntersville Chert to Mahantango Formation

The Middle Devonian supersequence (D-2) in Virginia (Fig. 12) is approximately 16 m.y. duration and consists, in ascending order, of: 1) transgressive systems tracts of chert, shale and sandstone (e.g., Huntersville Chert/Needmore Shale and Bobs Ridge Sandstone); 2) highstand systems tracts of Mahantango Formation and in southwestern Virginia, lower Millboro Shale (Figs. 13, 14; Patchen et al., 1985). The supersequence is up to 400 m or more (Sevon, 1985).

The basal unit, the Huntersville Chert and equivalents are commonly 10 to 30 m thick and generally are unconformable on the Oriskany/Ridgely Sandstone; they may unconformably overlie units down to the Martinsburg Formation in the east. In southwestern Virginia, the interval consists of sandstone and sandy fossiliferous cherty argillaceous limestone (mapped as Onondaga Limestone by Butts, 1940) that contains diverse assemblages including brachiopods, mollusks. They grade westward and northward into deeper water chert (Huntersville Chert). It grades into the fossiliferous Needmore Shale north of the New River, and throughout northern Virginia. Locally, such as near Wytheville, the cherts are unconformably overlain by chert-lithoclast, very shelly sandstone and dark gray silicified sandstone (Bob's Ridge Sandstone; Dennison, per. comm.). This sandstone is up to 10 m thick and contains spiriferids, and chert clasts. It is overlain by the thin Tioga bentonite. In northern

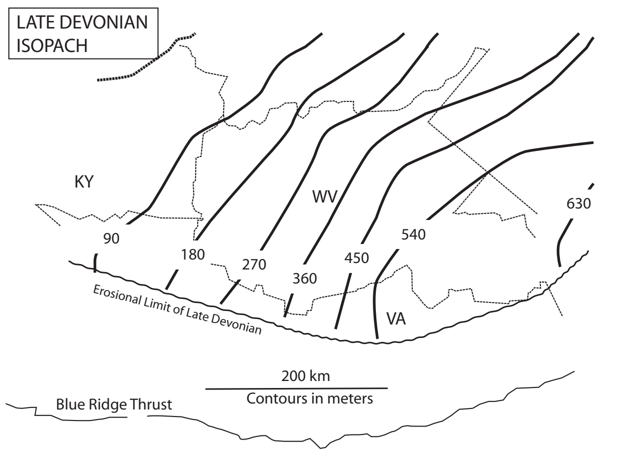


Figure 13. *Isopach map (palinspastic base in fold thrust belt) of Late Devonian supersequence (modified from Ayrton, 1963 and Sevon, 1985).*

Virginia, the units are overlain by the Mahantango Formation, consisting of fossiliferous sandstone and interbedded shale; the top of the formation becomes unconformable to the west (Ettensohn,1994; Ryder et al., 2009). The Huntersville Chert (equivalent to the Onandaga Limestone outside the state) is the transgressive tract of the supersequence, with local emergence (tectonic?) marked by the Bob's Ridge Sandstone.

Supersequence D-3: Late Devonian Millboro Shale to Hampshire Formations

The Late Devonian supersequence is approximately 25 m.y. duration (House and Gradstein, 2004); the formations are diachronous and are mainly Frasnian to Famennian in age although the Millboro Shale extends into the Middle Devonian in southwestern Virginia (Fig. 12; Dennison, 1985). Biostratigraphic correlation in Virginia is based primarily on brachiopods (Brame, 2001). Facies are summarized in Appendix A.

The Millboro to Hampshire interval records initial flooding and progressive upward-shallowing as a result of westward progradation of the Augusta lobe located in central Virginia (Dennison, 1985). In Virginia, the southwestward-thinning wedge of strata represents an oblique cross section through this progradational deltaic succession (Fig. 14). The Millboro Formation is a pro-deltaic black shale that was deposited under anaerobic or dysarobic conditions resulting from stratification of warm equatorial waters (Dennison, 1970, 1996; Ettensohn, 1985). Overlying strata of the Brallier Formation consist of turbidite-slope grey or black shale-mudstone with intercalated graded or massive siltstone/mudstone (Lundegard et al., 1985; Hasson and Dennison, 1988). The Foreknobs/Chemung Formation is made up of parasequences of storm-shelf, interbedded grey or black shale, mudstone up into siltstone with sublitharenite (lithics of quartz-muscovite schist and polycrystalline quartz with vermiform chlorite inclusions) and capping conglomerate lags. Conglomerate lags are developed within proximity to the lobes whereas bioclastic lags are developed in more distal settings. The Pound Sandstone and conglomerate is a significant regressive unit at the Frasnian-Famennian boundary. Interbedded bioturbated mudstones are fair-weather deposits (Randall, 1983; McClung, 1983). The Brallier and Foreknobs/Chemung formations contain 11 progradational/retrogradational sedimentation cycles in which siltstone-sandstone and/or conglomerate intervals are the most proximal facies (Filer 2002). Intercalated red beds may

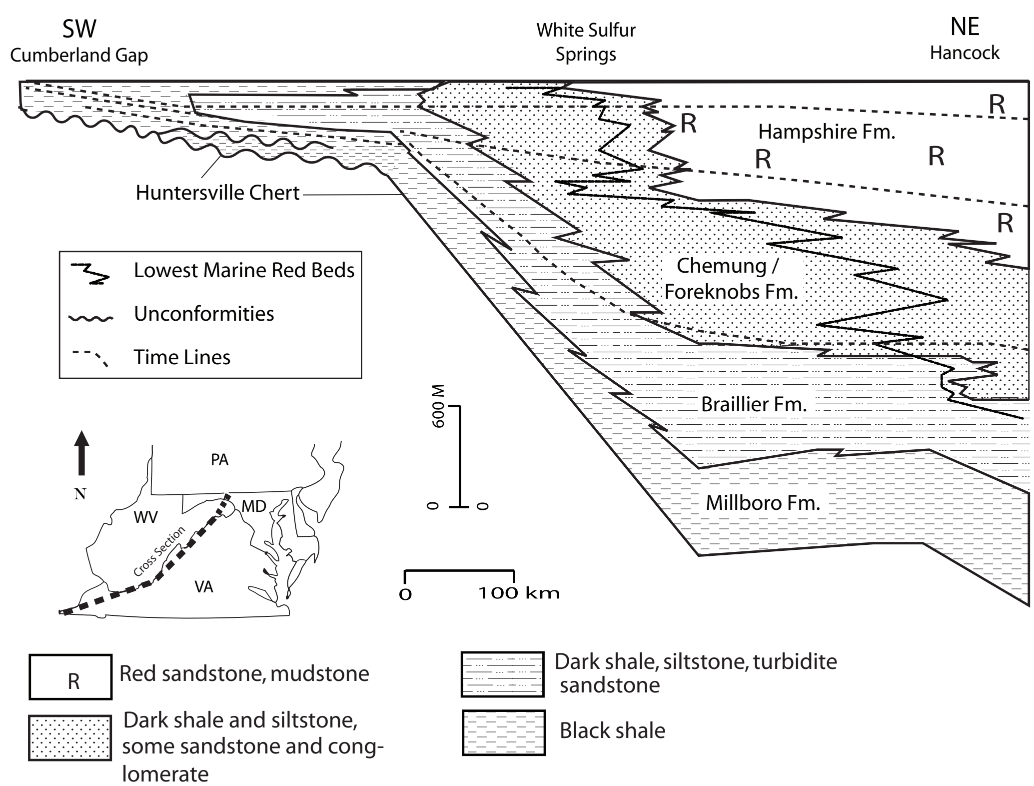


Figure 14. *Regional cross section of Middle to Late Devonian supersequence (modified from Dennison, 1985).*

mark sealevel lowstands

The upper part of the supersequence in northern Virginia consists of red beds of the Hampshire Formation (Fig. 14). They consist of non-marine, braided and meandering channel and sheetflood sandstones and overbank red mudstones, some with calcareous paleosols (McClung, 1983). Upward in the Hampshire, plant debris becomes more abundant (Brezinski et al., 2009). The Late Devonian cycles are attributed to high-frequency, 4th-order (100-150 k.y.), eustatic sea-level changes related to a late Frasnian-Famennian glaciation (van Tassell, 1994; Filer, 2002).

Supersequence M-1: Mississippian Cloyd-Sunbury-Price-Maccrady Formations

The Fort Payne Chert is thin and poorly developed in Virginia, but it thickens markedly to the west into Kentucky into a mixed carbonate-siliciclastic supersequence; because of its poor development in Virginia we have included this unit into our supersequence M-1. The limited development of the Fort Payne-Salem-Warsaw supersequence in Virginia suggests that there is a significant unconformity on top of the Maccrady, unless this unit extends up into the Meramecian. Early Carboniferous supersequence M-1 (Fig. 15) has been described by Kreisa and Bambach (1973), Bartlett (1974), Whitehead (1984), Bjerstedt and Kammer (1988) and Matchen and Kammer (1994). Supersequence M-1 consists of the latest Devonian Cloyd Conglomerate Member, the Devonian to earliest Mississippian Sunbury Shale-Price Formation (Kinderhookian to early Osagean) and the Maccrady Formation redbeds (Osagean; Fig. 17A). The succession is up to 450 m in the east, thinning westward to less than 100 m in far southwest Virginia and to 60 m in the north (Fig. 16A). At least five sequences occur in supersequence M1. Facies are summarized in Appendix A.

The Cloyd Conglomerate is a 0 to 48 m thick, fluvial-deltaic conglomeratic unit at the base of the supersequence in southwest Virginia. Channels trend northwest and are unconformably incised into underlying units of the Chemung-Hampshire formations. It wedges out into fossiliferous sandstone and siltstone to the south and west, and fines upwards into slightly fossiliferous, estuarine and tidal flat facies (T1 transgression, Bjerstedt and Kammer, 1988).

The Cloyd Conglomerate is overlain by the generally poorly exposed Sunbury Shale Member and equivalents; this shale unit contains the Devonian-Mississippian boundary, and the T2 transgression of Bjerstedt and Kammer (1988). In the west, it is a dark gray to black shale with thin tabular sands. The base of the sequence in Virginia is the Ceres sandstone member (cf. Bartlett, 1974).

The overlying Price deltaic succession (Fig. 15) is a broadly, upward-coarsening unit containing the T3 and T4 transgressions with the base of the sequences marked by the Broadford and Hayters Gap Sandstone members (cf. Bartlett, 1974). Fan turbidites pass upwards into storm-deposited fine-grained sandstone and shale. In the west, a bioturbated silty shale with phosphate pebbles marks the T3 transgression, possibly corresponding to glauconitic units in the east. Above this, distal to proximal lower shoreface tempestites coarsen upwards into upper shoreface sandstone, and then into local

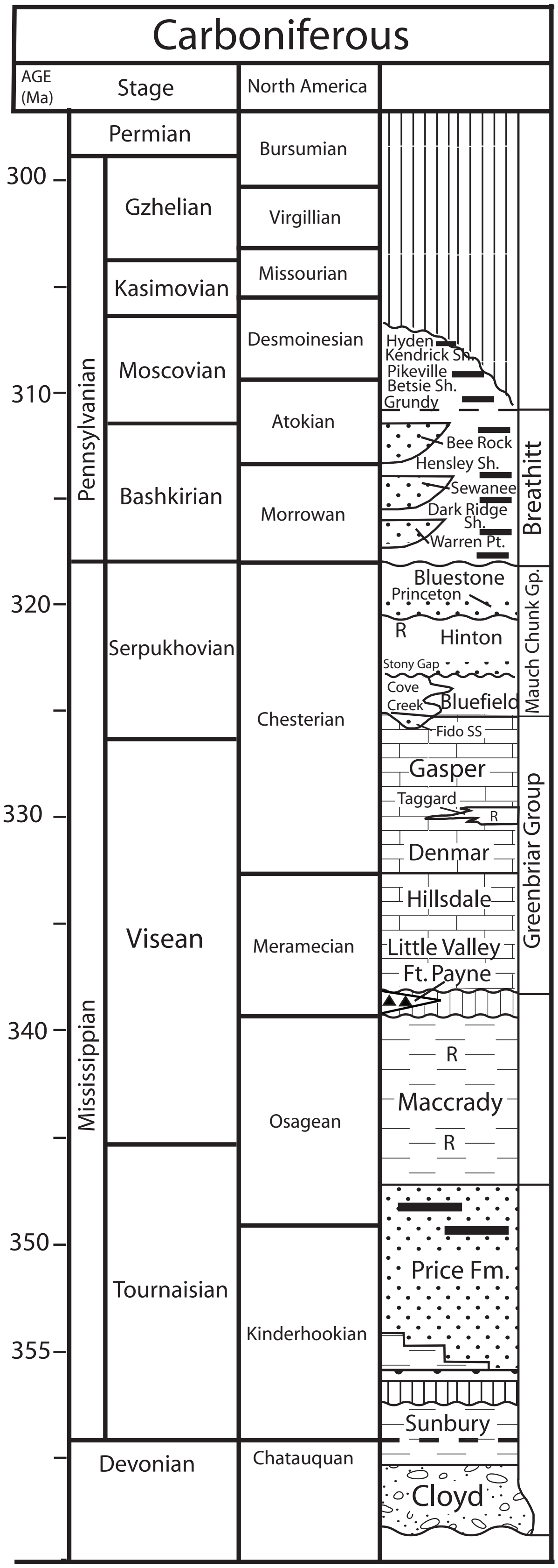


__Figure 15.__ Chronostratigraphic chart of Mississippian supersequence and lower part of Pennsylvanian supersequence preserved in southwestern Virginia (based on Huggins, 1983; references in Al-Tawil et al., 2003; Patchen et al., 1985). Redbeds (R), sandstone (dots), mixed siliciclastics (blank) and coals (heavy black line), shale-siltstone (dashes), limestone (brick pattern).

back-bar and bay, finer sandstone-shale. Distributary mouth-bar sands form coarsening upward units, overlain by local coals and non-marine, subaerial delta plain units with local, distributary channel-fill sandstones (Lick Run member, Bartlett, 1974). Price sandstones are sublitharenites with lithics of muscovite- and quartz-muscovite schist.

Alluvial coastal plain deposits of the Maccrady Formation interfinger with and overlie the Price Sandstone (Fig. 15). The Maccrady beds are dominantly non-marine red mudstones with channel-fill sandstone, crevasse-splay deposits and calcretes. At the base of the Maccrady Formation in southwestern Virginia, are some thin limestones (T5 transgression), as well as significant evaporites especially in the Saltville area (Dennison and Wheeler, 1975, p. 86).

The Cloyd Conglomerate was deposited within channel complexes draining the uplifted Late Devonian highlands during lowstand. These were reworked in shoreline settings adjacent to the river mouths. Initial transgression reworked the top of the Cloyd Conglomerate to develop mega wave-ripples followed by estuarine to tidal flat sedimentation. Shales of the Sunbury Member developed on the drowned shelf in 150 m of water (Bjerstedt and Kammer, 1988), followed by progradation of the Price deltaic succession. Climate was humid, as evidenced by coals and logs in channel sandstones, but became arid during deposition of the Maccrady redbeds and evaporites (Bartlett, 1997). This drying may have set the stage for the subsequent widespread carbonate sedimentation of the Greenbrier Group.

MISSISSIPPIAN-PENNSYLVANIAN ALLEGHANIAN FORELAND BASIN SUPERSEQUENCES

Supersequence M-2: Late Mississippian Greenbrier - Mauch Chunk Interval

The Greenbrier carbonates form the transgressive limb of the Late Mississippian supersequence, whereas the Mauch Chunk siliciclastics are the highstand tract (Fig. 15). Outcrops of the Greenbrier Group carbonates are mainly in southwestern Virginia. The Greenbrier Group ranges in thickness from 100 m in the west, where a superb section is exposed just over the Virginia State Line on Pine Mountain Coalfields Expressway, to 800 m in the foredeep to the east (Greendale syncline; Fig. 16A). The Greenbrier carbonates unconformably overlie the Price-Maccrady redbeds. They include the Little Valley and the Hillsdale formations (Meramecian), and the Denmar and Gasper formations (Chesterian). Studies include Bartlett and Webb (1971), Blancher (1974), and Huggins (1983) on conodonts. Al-Tawil et al. (2003), although focused on eastern West Virginia sections, also include the foredeep section in the Greendale syncline. Al-Tawil and Read (2003) describe the succession in Kentucky, including the Pine Mountain belt outcrops adjacent to the Virginia state line. Limited data in Virginia preclude the construction of detailed isopach maps for Virginia, although these are available for the bordering states of West Virginia and Kentucky (Al-Tawil and Read, 2003; Al-Tawil et al., 2003;

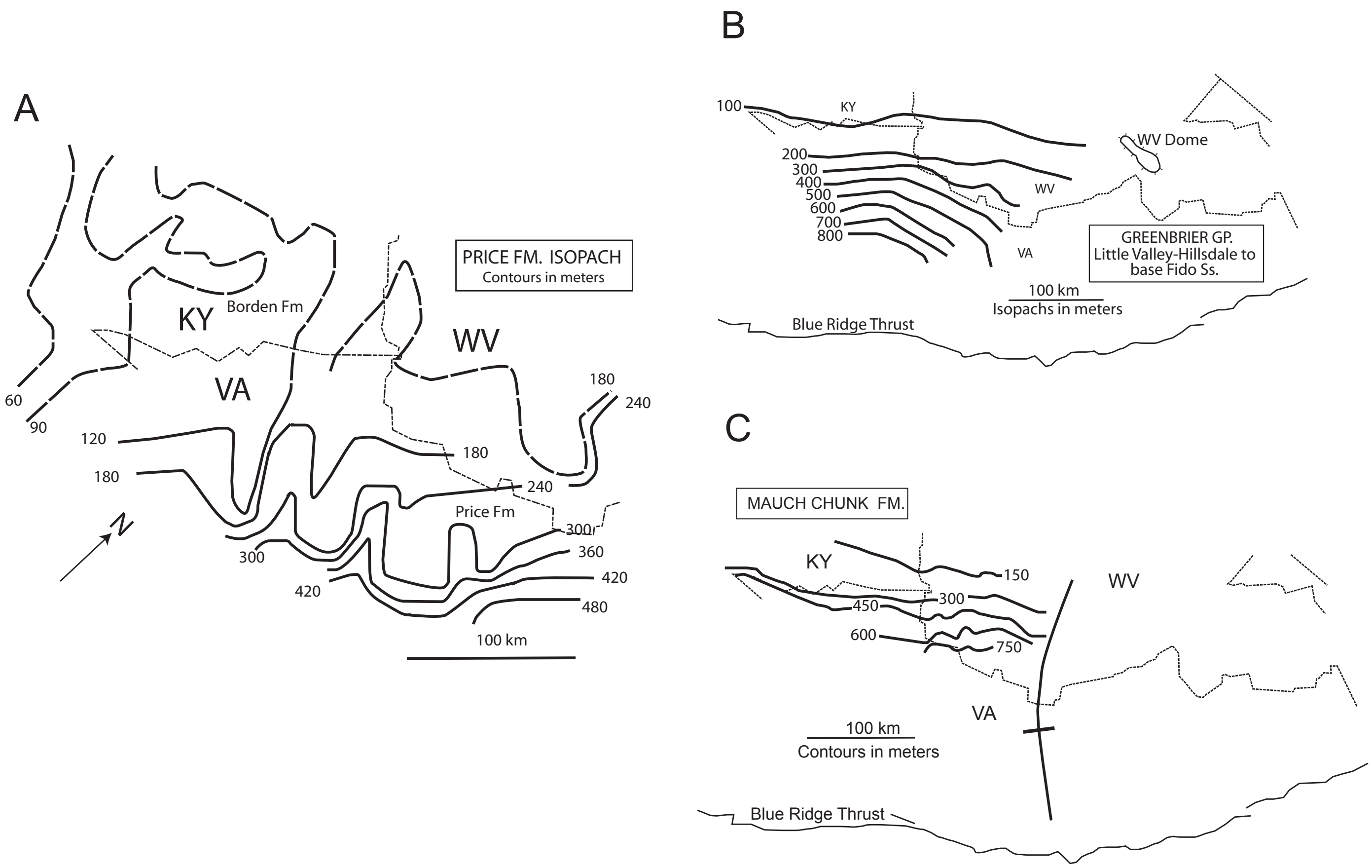


Figure 16. *Isopach maps (palinspastic base) of Mississippian supersequences. A. Lower Mississippian Price-Maccrady formations, southwestern Virginia (modified from Bartlett, 1974; Whitehead, 1984; solid isopachs from Bartlett, 1974, for Price Formation only, dashed isopachs are for the equivalent Borden Fm in Kentucky-West Virginia. B. Greenbrier Group (modified from Al-Tawil and Read, 2003; Al-Tawil et al., 2003; Wynn and Read 2008). C. Mauch Chunk Group (modified from Arkle, 1974).*

Wynn and Read, 2008). Facies are summarized in Appendix A.

The basal Greenbrier units are Upper Meramecian, Little Valley-Hillsdale carbonates (0 to 270 m thick) that are equivalent to the St. Louis Limestone to the west (Huggins, 1983). They form a composite sequence of five to perhaps six sequences (Fig. 17B). In the Greendale syncline foredeep sections, the Little Valley Formation appears to contain four sequences 30 to 40 m thick, consisting of lowstand sandstone and rare thin oolite, overlain by deeper, skeletal wackestone and fossiliferous limy mudrocks, capped by thin oolite and barren mudrock; evaporite pseudomorphs occur at the top of the Little Valley Formation. An 8 m thick black, very fossiliferous nautiloid-bearing shale in the middle of the Little Valley-Hillsdale succession may mark the maximum flooding of the composite sequence. The Hillsdale Limestone consists of at least two 30 to 40 m thick sequences of lowstand siltstone/mudrock to transgressive thin oolite, overlain by highstand skeletal wackestone (commonly onkoidal), capped by oolite and lime mudstone. On the shallow ramp to the west, the equivalent facies are very thin units of minor skeletal grainstone, oolite and peritidal dolomitic lime mudstone with the corals Syringopora and Acrocyathus.

The Chesterian Denmar composite sequence (up to 270 m thick and equivalent to the Ste. Genevieve Formation), contains four 4th-order sequences (Fig. 17B). Sequences on the ramp slope (Greendale syncline) are thick (~30 m) and composed of lowstand sandstone/ siltstone, overlain by transgressive skeletal wackestone to grainstone, with thick highstand deep-water laminated silty mudstones overlain by skeletal wackestone. Updip on the ramp, these sequences are only a few meters thick, contain the crinoid Platycrinities penicillus, and have abundant oolite, lagoonal muddy limestone, minor eolianite and paleosols.

The next Chesterian composite sequence (up to 240 m thick) includes two 4th-order sequences with basal regressive redbeds on the shallow ramp (Taggard redbeds of eastern West Virginia), overlain by four 4th-order carbonate-prone sequences 5 to 15 m thick (Gasper Formation) with the crinoid Talarocrinus (Fig. 17B). These tend to become less oolitic and more skeletal upward in the composite sequence, marking overall backstepping of the shallow water facies on the ramp. In the foredeep, the sequences are thick (20 to 70 m) and consist of lowstand to transgressive clastics or shallow water carbonates, near the base and the upper part of the composite sequence, and highstand deep water laminated silty mudstone overlain by skeletal wackestone. The uppermost Gasper 4th-order sequence lacks deeper water mudstone, marking an overall upward shallowing.

The Mauch Chunk Group is relatively conformable on the Greenbrier carbonates, although the basal incised Fido redbeds suggest some erosion. The Mauch Chunk interval is a westward-tapering wedge with a maximum thickness of over 900 m (Englund and Thomas, 1990). It consists of, in ascending order, the Bluefield, Hinton, Princeton and Bluestone formations (Figs. 15, 16C, 17C). Important lithostratigraphic units include the Stony Gap Sandstone Member (base of the Hinton Formation), the Little Stone Gap Limestone Member (middle of the Hinton Formation) and the Pride Shale (base of the Bluestone Formation). With the exception of the Stony Gap Sandstone (a quartz arenite), the other Mauch Chunk sandstones are sublitharenites (lithics of muscovite- and quartz-muscovite schist). The top of the Mauch Chunk Group is the Mississippian-Pennsylvanian unconformity. The Chesterian-age Mauch Chunk Group is approximately 9 Ma (Miller and Eriksson, 2000; Maynard et al., 2006).

Maximum flooding of the Middle and Upper Mississippian supersequence is in the Cove Creek Limestone/Lillydale Shale (Pencil Cave Shale of drillers) at the base of the Bluefield Formation. Three composite 3rd-order sequences, each consisting of stacked 4th-order sequences of roughly 400 k.y. duration, comprise the Mauch Chunk Group (Miller and Eriksson, 2000; Maynard et al., 2006).

The lowermost 3rd-order composite sequence (Bluefield Formation) contains four sequences; the lower two sequences are dominated by marine limestones and grey and black shales. In the foredeep, they are the Cove Creek Limestone, which has a basal red sandstone (Fido Sandstone) incised into the underlying Gasper marine carbonates. Here, two Cove Creek sequences are dominated by deep water, laminated silty lime mudstones with a thin lowstand oolite between them. On the shallow ramp (Pine Mountain), the equivalent units are basal gray shale overlain by open shelf limestone, with locally developed paleosols on sequence boundaries. The upper two sequences are dominated by coastal plain red beds and tidal deltaic sandstones and on the foreland (northwestern) side of the basin, by incised-valley sandstones (Fig. 17C; Maynard et al., 2006).

The second 3rd-order composite sequence (Hinton Formation) begins with the regionally extensive Stony Gap Sandstone Member of incised-valley origin and extends to the base of the Princeton Formation. Coastal plain red beds dominate this composite sequence, consisting of up to twelve 4th-order sequences, the bases of which are defined by minor incised valley, lenticular sandstone bodies (e.g., Hackett, Neal, Tallery, Falls Mills units). Within the composite sequence, 3rd-order sequences are stacked into a transgressive and highstand systems tract, in which the Little Stony Gap Limestone marks the maximum flood.

The base of the uppermost composite sequence (Bluestone Formation) is defined by the major incised valley fill of the Princeton Formation that displays an upward transition from braided-alluvial to estuarine facies containing rhythmically bedded tidal creek deposits. The overlying Pride Shale Member (Fig. 17C) developed in a prodeltaic setting dominated by tides (Appendix A; Miller and Eriksson, 1997). The remainder of the Bluestone Formation is dominated by red beds and is interpreted as the upper part of a transgressive systems tract. Highstand deposits of this sequence are truncated by the Mississippian-Pennsylvanian systemic unconformity.

The rapid thickening of the Greenbrier carbonates into the foredeep that developed above the earlier Price deltas, indicate that tectonically induced subsidence had started to increase relative to sedimentation, perhaps in response to reactivated, incipient thrust loading. Basin infilling was coincident with northward paleolatitudinal drift through subequatorial latitudes (Cecil, 1990). Arid climate prevailed during Greenbrier deposition, indicated by widespread oolitic carbonates on the ramp, local sulfate evaporites in core, and early dolomitization. Vertisols with well-developed slickenslides and peds, present throughout the upper Bluefield, Hinton and upper Bluestone formations (Miller and Eriksson, 1997) imply seasonal rainfall and overall semi-arid conditions (Cecil et al., 1985; Cecil,

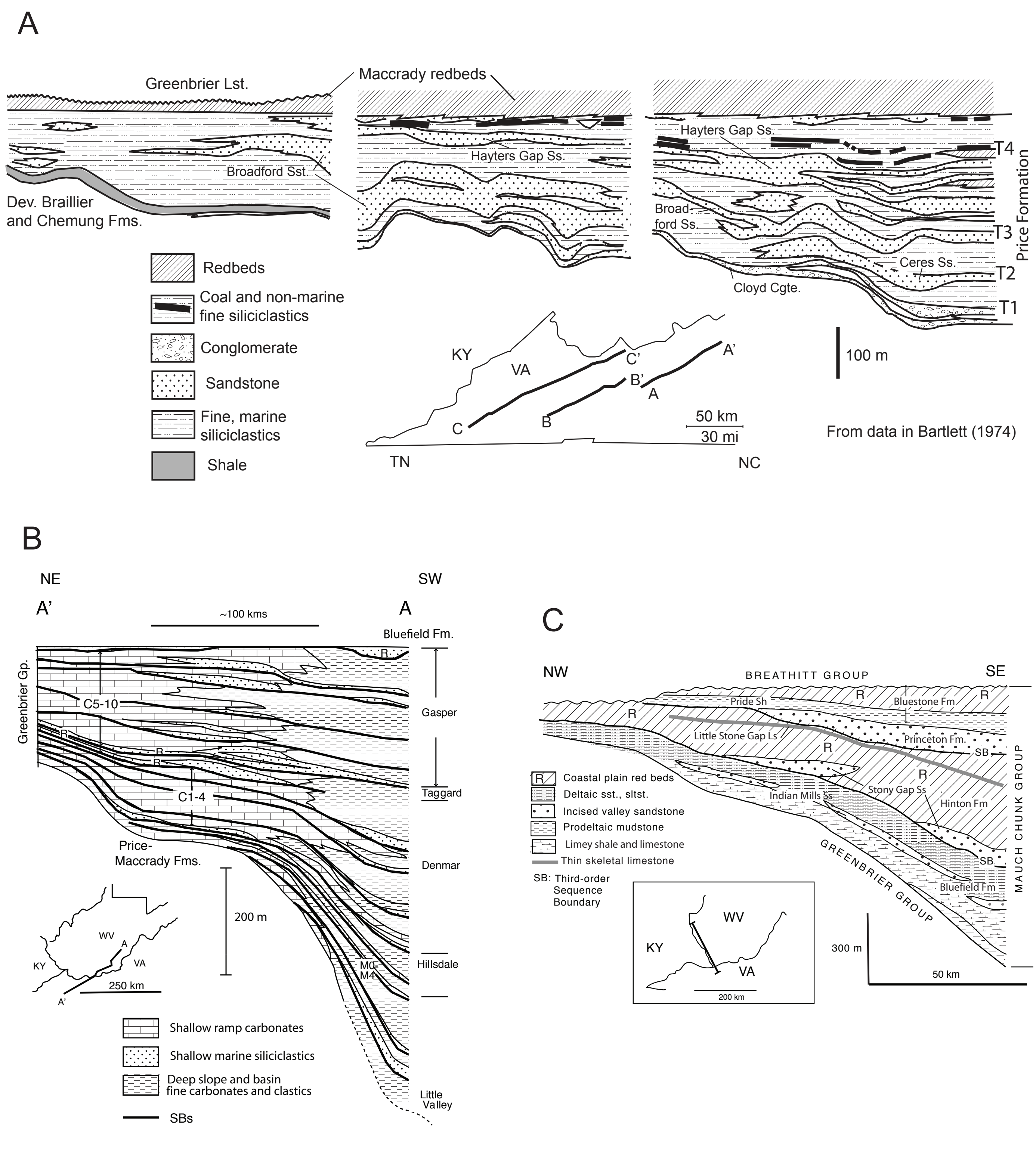


***Figure 17.** Regional cross sections of Mississippian supersequences. A. Lower Mississippian Price-Maccrady formations (drawn mainly from data in Bartlett 1974, and from Kreisa and Bambach, 1973; Bjerstedt, 1986; Bjerstedt and Kammer 1988;). B. Greenbrier Limestone (modified from Al-Tawil et al., 2003; Huggins, 1983). C. Mauch Chunk Group (modified from Miller and Eriksson, 2000).*

1990; Caudill et al., 1996). Local caliche layers in the Hinton Formation indicate that evaporation exceeded rainfall during part of the Upper Mississippian. Evidence of seasonal rainfall provided by paleosols is consistent with the inferred monsoonal control on annual cycles in the Pride Shale.

The sequence stratigraphic hierarchy recognized in the Upper Mississippian supersequence reflects two orders of relative sea-level change in the prevailing ice-house world in which much of Gondwana was covered by a continental ice sheet (Frakes et al., 1992). The dominant 400 k.y. sequences in the Mississippian carbonates and overlying siliciclastics result from waxing and waning of ice sheets under the influence of long-term eccentricity forcing. The 3rd-order composite sequences may be tectonoeustatic in origin (Miller and Eriksson, 1997)

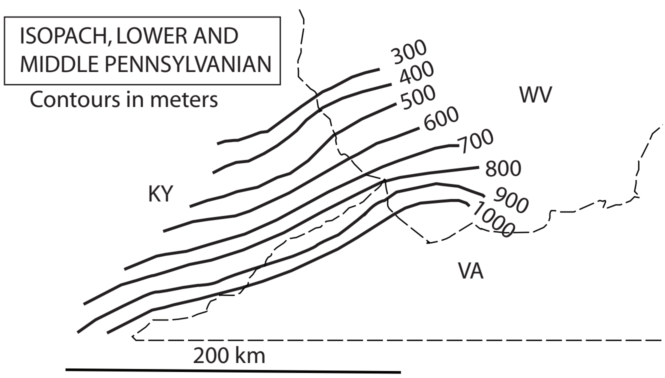


Figure 18. Isopach map (non-palinspastic base) of early and middle Pennsylvanian strata preserved in southwestern Virginia (modified from Arkle, 1974).

or, perhaps reflect long-term (~2.5 Ma) glacioeustatic changes related to very long-term orbital eccentricity (cf. Matthews and Frolich, 2002).

Supersequence P-1: Early to Middle Pennsylvanian Breathitt Group

The Breathitt Group is a westward-tapering wedge of coal-bearing strata that is well developed in Kentucky (Fig. 15). Equivalents of only the lower and middle Breathitt Group are preserved in southwestern Virginia in stratigraphic thicknesses of up to 1,700 m (Fig. 18; Arkle, 1974; Nolde, 1994). The Breathitt Group nomenclature developed for Kentucky and extended into Virginia by Chesnut (1994) and Greb et al. (2004) is used here. In ascending order, the coal-bearing formations making up the lower Breathitt Group in Virginia are the Pocahontas, Bottom Creek and Alvy Creek formations in the southeast and correlative Warren Point (lower Middleboro of Virginia), Sewanee (upper Middleboro) and Bee Rock quartz arenite units to the northwest (Fig. 19); the Dark Ridge and Hensley Shale Members of Chesnut (1994) separate these formations. Lower Pennsylvanian strata are of Bashkirian age (~312-318 Ma; Davydov et al., 2004).

The Middle Pennsylvanian record in Virginia consists of the coal- bearing Grundy (Norton), Pikeville (Wise) and Hyden (Harlan) formations (Fig. 19) with the equivalent of the Corbin Formation quartz arenite belt of Kentucky preserved only in Lee County as the Nease Sandstone (Nolde, 1994). An unnamed shale member separates the Alvy Creek and Grundy formations; the three Middle Pennsylvanian Formations are separated by the Betsie and Kendrick Shale members. The Middle Pennsylvanian strata in Virginia are of lower Moscovian age (~308-312 Ma; Davydov et al., 2004).

The Warren Point, Sewanee, Bee Rock and Corbin (Appendix A; Fig. 19) are tabular, quartz arenite bodies with channelized scour and fining-upward channel fills, unidirectional southwesterly paleocurrents, and ubiquitous basal quartz-pebble conglomeritic lags (e.g., Rice and Schwietering, 1988; Chesnut, 1988; Wizevich, 1992; Archer and Greb, 1995). In contrast to the quartz arenites, correlative sandstones to the southeast are locally derived, lenticular, sublitharenite bodies with abundant schist grains and local perthitic felspar grains (Reed et al., 2005). These sublitharenites occur as channelized scour and upward-fining successions with trough cross beds indicating flow towards the west and northwest.

Whereas there has been a general consensus that the sublitharenites were derived from a fold- and -thrust terrane to the southeast (Rice and Schwietering, 1988; Greb and Chestnut, 1996; Korus, 2002), the origin of the quartz arenites has been more controversial. A barrier beach model for these arenites was proposed originally by Ferm and Cavaroc (1969), Ferm (1974), and Horne et al. (1974) who interpreted Upper Mississippian, Lower Pennsylvanian and Upper Pennsylvanian strata as facies equivalents without regard to the Mississippian-Pennsylvanian systemic unconformity (Fig. 15). More recently, the quartz arenite belts have been interpreted as braided-alluvial deposits of major trunk river systems flowing to the southwest down the axis of the foreland basin. Detrital zircon ages are consistent with recycling of zircons from older passive-margin and older foreland-basin sandstones, whereas the Archean age zircons are consistent with the inferred longitudinal river system draining the craton.

Sequence stratigraphic analysis reveals that Lower Pennsylvanian strata in Virginia consist of three 3rd-order (~2.5 Ma) composite sequences that are comprised of stacked 4th-order (~400 k.y.) sequences (Bodek, 2006). Facies are summarized in Appendix A. Both 3rd- and 4th-order sequences are bounded by erosional surfaces (sequence boundaries) that developed during lowstands of sea-level. The overlying tabular to lenticular sandstones are mostly of braided-alluvial origin and were deposited in transverse and longitudinal incised valleys during rise in base level. Lenticular sandstones at the base of 4th-order sequences pass laterally into interfluves. The fluvial sandstones typically grade upwards into estuarine tidal rhythmites that are gradational into coal seams. These upward-fining successions are interpreted as lowstand-transgressive systems tract. Thin shales cap the transgressive systems tracts and mark maximum floods. Overlying highstand deposits are overall upward-coarsening and consist of stacked parasequences of deltaic origin. Coal seams located above parasequences reflect peat marsh development with limited siliciclastic influx

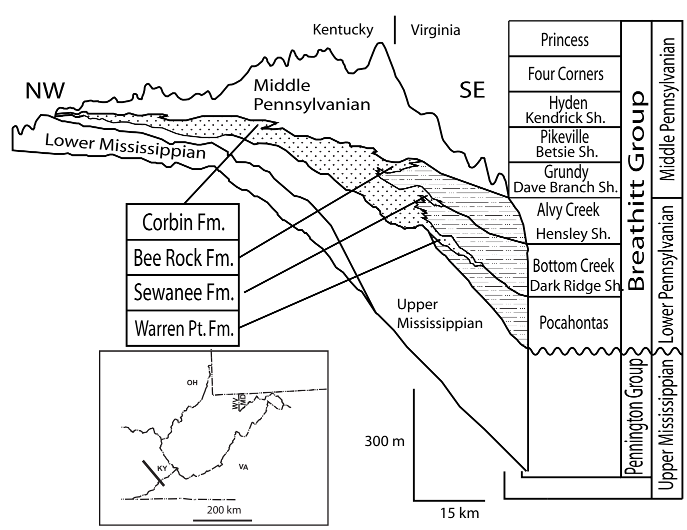


Figure 19. Regional cross section of Pennsylvanian supersequences (modified from Englund and Thomas, 1990; Chestnut, 1994).

following channel avulsion and delta-lobe switching (Bodek, 2006).

Detailed sequence stratigraphic analysis has not been completed for Middle Pennsylvanian strata in Virginia. However, the unnamed shale member above the Alvy Creek Formation, along with the Betsie and the Kendrick Shale members all are interpreted as 3rd-order flooding events (Chesnut, 1994). The Corbin quartz arenite in eastern Kentucky is a trunk river deposit to the tributary alluvial and deltaic facies of the Grundy Formation. Trunk river equivalents are lacking for overlying formations.

In contrast to the Upper Mississippian stratigraphic record, red beds are absent from Lower and Middle Pennsylvanian strata leading Cecil et al. (1985) and Cecil (1990) to infer an everwet climate, in which the coal seams (histosols; Mack et al., 1993) characteristically overlie highly-leached underclays (seat earths). In common with the Upper Mississippian record, the stratigraphic hierarchy recognized in Lower Pennsylvanian strata also reflects two orders to relative sea-level change in an ice-house world. The 4th-order sequences (~400 k.y.) record waxing and waning of the Gondwana ice sheet related to orbital eccentricity. The 3rd-order sequences may be tectonoeustatic (Miller and Eriksson, 1997) or reflect long-term (~2.5 Ma) glacioeustatic changes related to long-term orbital eccentricity (cf., Matthews and Frolich, 2002).

Detrital Zircon Geochronology

In recent years, U-Pb geochronology of individual zircon grains in modern sediments and in sedimentary rocks has become a widely used technique for identifying sediment provenance ages, location of provenances, maximum depositional ages and tectonic evolution of orogenic belts (e.g., Cawood and Nemchin, 2001; Eriksson et al., 2004; Thomas et al., 2004; Dickinson and Gehrels, 2009; Park et al., 2010). Modern techniques of U-Pb geochronology using laser ablation inductively coupled plasma mass spectrometry (LA-ICP-MS) now permit rapid determination of ages of multiple grains (e.g., Black et al., 2004; Gehrels et al., 2006). In the past decade, a number of papers have been published on detrital zircon geochronology of sandstones in the Central Appalachians of Virginia and adjacent eastern West Virginia (Eriksson et al., 2004; Becker et al., 2005; Park et al., 2010). Data for 24 samples from these studies and for two unpublished samples are summarized below and inferred provenances are identified on Figure 20 (after Park et al., 2010). Provenance terrains older than 950 Ma, with the exception of the Trans-Amazonian of South America, are located west of the Appalachian orogenic belt whereas Pan-African and Paleozoic provenances are located to the east. Sample numbering follows the supersequence terminology (O-1, etc) used in this chapter.

Three samples are available for the rift-passive margin supersequences. Detrital zircons from a late Neoproterozoic, rift-related Unicoi Formation sample (base of Ꞓ-1) define a unimodal age spectrum with the majority of grains of Grenvillian age (950-1250 Ma) and with a few isolated grains reflecting a Granite-Rhyolite source with ages of 1300-1500 Ma (Fig. 20; Eriksson et al., 2004). A sample from the early Cambrian Erwin Formation (upper part of Ꞓ-1) contains grains of Grenvillian, Granite-Rhyolite and Yavapai-Mazatzal (1600-1800 Ma) provenance (Fig. 20; Eriksson et al., 2004). Unpublished data for a sample of the late Cambrian Copper Ridge sandstone (Ꞓ-4) are dominated by zircons of Granite-Rhyolite and Yavapai-Mazatzal provenance with subordinate Grenvillian and Archean modes and no concordant zircons younger than 1000 Ma (Fig. 20).

Data from Taconian foreland basin supersequences are from five samples (Fig. 20). Detrital zircons from a feldspathic arenite and a quartz arenite clast in the Fincastle conglomerate of the Bays Formation (O-2) are predominantly of Grenvillian and less common Granite-Rhyolite provenance. A sandstone sample from the Bays Formation (also in O-2) has a comparable age spectrum to the clasts but with a few Yavapai-Mazatzal zircons and one grain as young as Pan-African (Fig. 20; Eriksson et al., 2004). Samples from the Fincastle Member of the Martinsburg Formation and the Oswego Formation (O-3) have similar age spectra to the Bays Formation except that the Oswego Formation contains a number of Superior and older Archean zircons (2500-3300 Ma) as well as Paleozoic zircons of comparable depositional age to the host sandstone, and attributed to syndepositional volcanism (Fig. 20; Park et al., 2010).

Data from post-Taconian foreland basin supersequences are from samples of the upper Tuscarora, Rose Hill, Eagle Rock-Keefer formations (S-1; Fig. 20) and the Oriskany sandstone (S-3; Fig. 20). All five samples have similar age spectra with peaks coinciding with Grenvillian and Granite-Rhyolite provenances. Rare Superior-aged zircons are present in four of the samples (Fig. 20; Eriksson et al., 2004; Park et al., 2010).

Data from the Devonian-Lower Mississippian Acadian foreland basin supersequences are from samples of the Foreknobs, Chemung and Hampshire formations (D-3; Fig. 20), the Cloyd Conglomerate and a sandstone from the Price Formation (M-1; Fig. 20). Whereas zircons of Grenvillian and Granite-Rhyolite provenance predominate and three of the five samples contain zircons of Superior provenance, four of the five samples contain zircons of Paleozoic and less common Pan-African provenance. In addition, the Foreknobs and Chemung samples contain a few zircons of Penokean and Trans-Amazonian provenance (Fig. 20; Eriksson et al., 2004; Park et al., 2010).

Samples from the Mississippian-Pennsylvanian Alleghanian foreland basin sequences are from the Upper Mississippian Hinton, Princeton and Bluestone formations (M-2; Fig. 20) and the Lower Pennsylvanian Pocahontas, Warren Point, lower and upper Raleigh (Sewanee equivalent in WV), and Bottom Creek formations (P-1; Fig. 20). Mississippian samples are dominated by zircons of Grenvillian and Granite-Rhyolite provenance and the proportion of zircons of Yavapai-Mazatzal, Superior, Paleozoic and Pan-African provenance decreases up-section (Fig. 20; Park et al., 2010). Zircons from Pennsylvanian quartz arenite bodies (Warren Point and upper Sewanee) display a wide spread in ages indicating Grenvillian, Granite-Rhyolite, Yavapai-Mazatzal and Penokean provenances as well as a strong (>10%) Superior and older Archean contributions (Fig. 20; Eriksson et al., 2004; Becker et al., 2005). Sublitharenite bodies (Pocahontas, Bottom Creek and lower Sewanee) are dominated by zircons of Grenvillian and Granite-Rhyolite provenance but lack Archean zircons (Fig. 20; Eriksson et al., 2004; Becker et al., 2005). Minor Paleozoic and Pan-African provenances are indicated for one of the samples (P-1; Fig. 20).

Discussion

Figure 21 summarizes the chronostratigraphy, accommodation history, paleolatitudes and paleoclimate, and the main tectonic events that affected the Appalachian Basin within Virginia. It also provides a comparison between the northern and southwestern Virginia stratigraphic units and nomenclature.

Tectonic Evolution

The geological evolution of the central Appalachians commenced ca. 750 Ma ago with rifting of the Neoproterozoic supercontinent Rodinia and, more specifically, of Grenville basement in the eastern USA (Fig. 5B). The first rifting event at ca. 750 Ma (U-Pb zircon, Aleinikoff et al., 1995) did not proceed to continental separation, but it initiated deposition of Neoproterozoic, shallow- to deep-water Lynchburg siliciclastics on attenuated crust (Wehr, 1983; Wehr and Glover, 1985).

The second rifting event, dated at 600-550 Ma (U-Pb zircon, Aleinikoff et al., 1995; Sinha et al., 1988), is represented by siliciclastic sedimentary rocks of the Swift Run Group west of the Blue Ridge, the Lynchburg Group east of the Blue Ridge (Chapter 5, this volume), basalts, subordinate rhyolites and sedimentary rocks of the Catoctin Formation (Chapter 5, this volume), the lower half of the Unicoi Formation and the Weverton Formation at the base of the Chilhowee Group. Detrital zircon ages (Fig. 20, Є-1) indicate a mostly local provenance consisting of Grenvillian basement for the Unicoi felspathic sandstones. Quartz arenites of the upper Unicoi Formation demarcate the onset of breakup and thermal subsidence of the Laurentian margin at the beginning of the Paleozoic (Simpson and Eriksson, 1989). The overlying ca. 3.5 km-thick passive-margin deposits of supersequences Є-1 to Є- 4 and O -1 record decreasing rates of thermal subsidence over a time period of ~100 million years (Bond et al., 1984). Detrital zircon ages from lower and upper Cambrian sandstones (Fig. 20, Є-1 and Є-4) indicate a change in provenance through time from predominantly Grenvillian to predominantly Granite-Rhyolite sources. This change is attributed to the progressive onlap onto the Grenvillian basement during marine transgression of Laurentia. The Middle Cambrian Rome Trough rifting event of West Virginia and Kentucky is evidenced in Virginia by increased subsidence in the southern depocener (Fig. 1).

Passive-margin sedimentation was terminated with the onset of the Taconian Orogeny (Fig. 5B; Rodgers, 1971; Lash, 1988; Hatcher, 1989). Hinterland expressions of this orogeny are the Chopawamsic and Milton terranes (Chapter 5, this volume). In the Valley and Ridge Province of Virginia, the transition from passive-margin to foreland-basin sedimentation is demarcated by the middle Ordovician Knox unconformity that developed

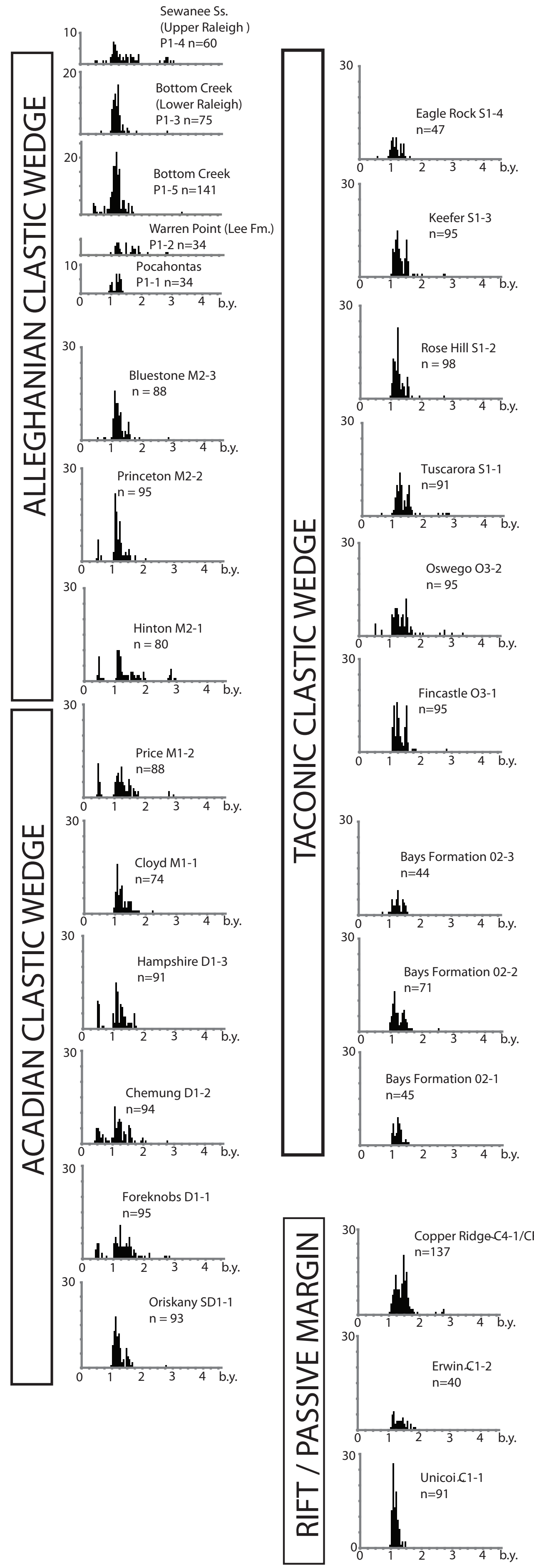


__Figure 20.__ Histograms of U-Pb ages for detrital-zircon plots from Late Neoproterozoic to Pennsylvanian sandstones of the Central Appalachian Basin from Virginia and easternmost West Virginia. All ages are less than 10% discordant based on the agreement between 207Pb/206Pb* and 206Pb*/238U ages (asterisk indicates radiogenic Pb, i.e. total minus common Pb). Ages older than 800Ma calculated from 207Pb*/206Pb*ratios and ages less than 800 Ma based on 206Pb*/238U ratios. Bin sizes are 50 million years. Data from: Eriksson et al. (2004); Becker et al. (2005); Park et al. (2010).*

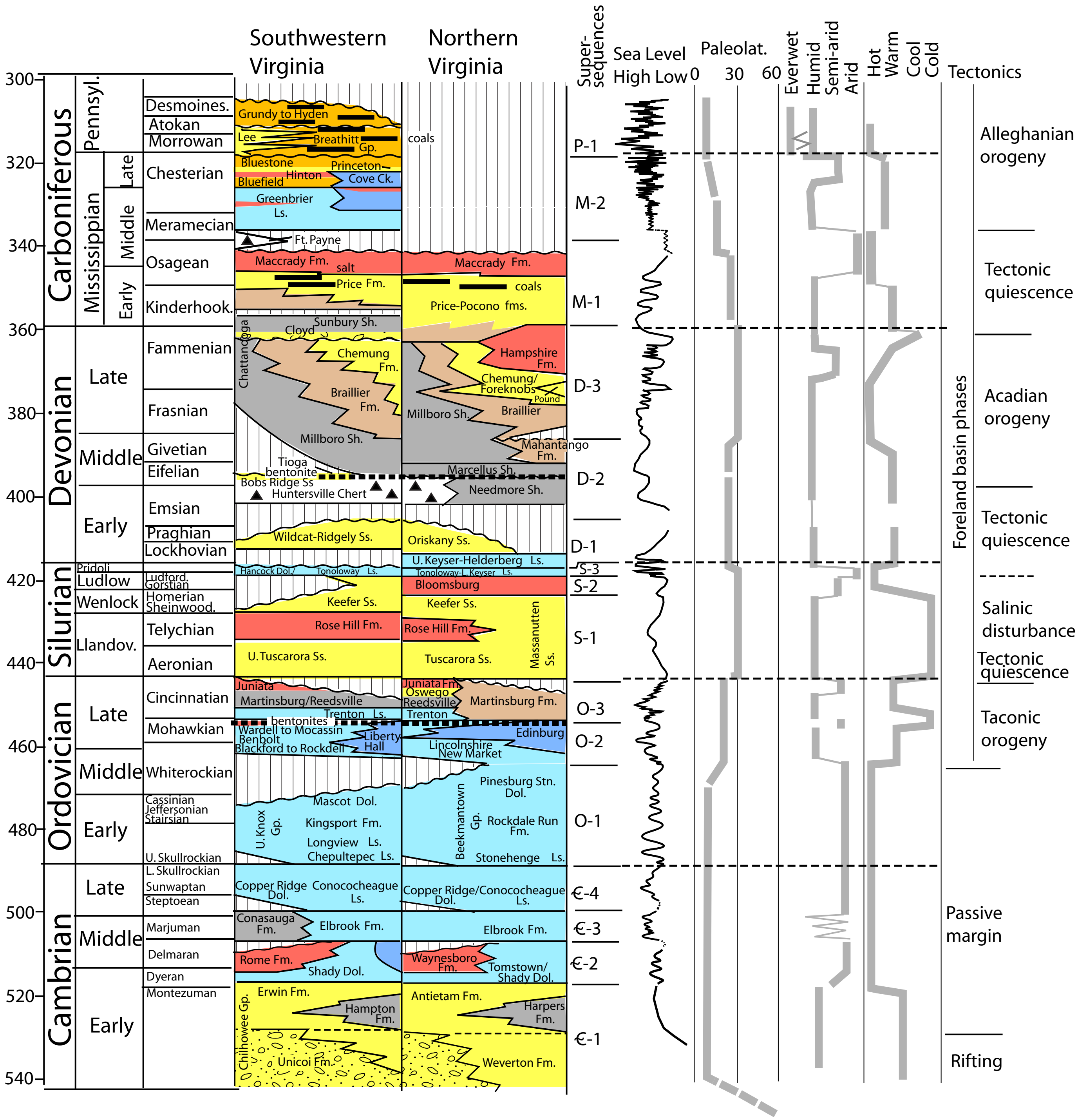


Figure 21. *Summary chronostratigraphic diagram showing stratigraphic units and nomenclature in northern and southwestern Virginia, the accommodation (relative sea level) history, paleolatitude, paleorainfall (in terms of everwet to arid), paleotemperature (hot to cold) and and tectonics. Paleolatitudes and paleoclimate from Scotese 2001; other references given in discussion in text. Red-beds (red), conglomerate (yellow, open ellipses and dots), sandstone (yellow), mixed siliciclastics (orange), coal (heavy black lines), prodelta or turbidite clastics (pale brown), shale (gray), chert (solid triangles), carbonates (light blue), offshelf carbonates (dark blue), volcanic ash (heavy dashed lines), unconformities (vertical lines).*

in response to peripheral bulge formation landward of the downwarping shelf during thrust loading (Mussman and Read, 1986; Ettensohn, 1994). Shallow ramp carbonates formed on the foreland (western) side of the basin (Read, 1989b). The initial bulge in southwestern Virginia, the Tazewell Arch (Read, 1980), is recorded by thinning of the Ordovician O-2 supersequence (so-called "Middle Ordovician" Limestones). A subsequent bulge (the Shenandoah Axis; Rader and Perry, 1976a) developed in northern Virginia during continued foundering of the carbonate ramp during deposition of the Late Ordovician Martinsburg Formation (supersequence O-3; Fig. 5B). This bulge separated the deep basin in northern Virginia from a shallower basin to the west (Woodward, 1951; Diecchio, 1993). The initial orogenic event (Blountian phase) occurred earlier in the south and progressed northward during the classic Taconic phase (Rodgers, 1971) as reflected in the development of the southern depocenter prior to the northern depocenter. These are the Blountian and Taconic Tectophases of Ettensohn (1994). The top of the Taconic clastic wedge is defined by an unconformity in part related to isostatic rebound following the cessation of thrusting and erosion of the thrust load (Dorsch and Driese, 1995).

Petrography and detrital zircon geochronology of Taconian-age sandstones, together with clast composition data from the Fincastle Conglomerate, imply that foreland-basin sandstones and conglomerates were derived from recycling of Neoproterozoic rift and Cambro-Ordovician passive-margin sedimentary rocks. Sedimentary clasts and rock fragments in the Taconian conglomerate can be matched with stratigraphic units in the passive-margin succession. Mussman and Read (1986) argued that the passive-margin successions were uplifted in an accretionary prism associated with eastward subduction during the early convergent history of the Appalachian orogen. Rare gneiss clasts probably were derived from thrust slices of Grenville basement caught up in the accretionary prism. Quartz-chlorite clasts as well as the low-grade metamorphic rock fragments in these sandstones could also have been derived from relatively shallow levels of the accretionary prism. Detrital zircon age distribution patterns of Taconian sandstones (Fig. 20, sequences O-2, O-3) are remarkably similar and are consistent with uplift and recycling from the passive-margin succession in that they are dominated by Grenvillian and Granite-Rhyolite provenance ages (Eriksson et al., 2004). Older zircons, including those of Archean age in the Oswego Formation, were also most likely recycled from passive-margin strata (Park et al., 2010); the presence of Archean-age zircons in samples from supersequence Ꞓ-4 lends support to this contention.

The overlying post-Taconian upper Tuscarora, Rose Hill and Eagle Rock formations (supersequence S-1) reflect a phase of waning tectonism during the convergent history of the orogen (Castle, 2001). Sandstones of the upper Tuscarora and Eagle Rock formations are anomalous in composition for a foreland basin, in consisting predominantly of monocrystalline with subordinate polycrystalline quartz (Hayes, 1974). Dorsch et al. (1994) attributed the maturity of these sandstones to recycling of older passive-margin sandstones coupled with intense weathering and marine reworking of less mature sediment. Ettensohn (1994) suggested there may have been weak tectonism during the Silurian (mainly Ludlow) when Baltica or separate Avalonia terranes underwent convergence with northeastern Laurentia (Salinic disturbance, Rodgers, 1971). Two Salinic events (the first and second Salinic tectophases), were postulated by Ettensohn (1994), one associated with the Tuscarora-Rose Hill succession (supersequence S-1), and the second with the Keefer-McKenzie-Bloomsburg succession (S-2). Evidence for these events includes angular relations and presumed flexural sequences related to thrust loading. The Bloomsburg redbeds would appear to be coeval with the main Salinic event but they also correlate with a major eustatic fall (cf. Johnson, 1996). Separating the effects of eustasy from such weak tectonic effects is difficult. Detrital zircon ages of the post-Taconian (S-1 and S-2) are consistent with recycling of Cambrian passive-margin and/or Taconian sandstones (Eriksson et al., 2004; Park et al., 2010). The absence of Pan-African age zircons in post-Taconian sandstones mitigates against accretion of Avalonia in the Silurian.

Ettensohn (1994) ascribed the three Devonian supersequences (roughly corresponding to D-1 to D-3 herein) to his first, second and third Acadian tectophases. Major convergence associated with the Late Devonian-Early Mississippian Acadian Orogeny caused subsidence of the foreland, drowning of the Early Devonian carbonate ramp, and uplift of the hinterland to the southeast. This was followed by southwestward delta progradation and deposition of supersequences D-3 and M-1, the latter occurring during Ettensohn's (1994) fourth Acadian tectophase. The predominance of metamorphic rock fragments and mica coupled with the paucity of extrabasinal sedimentary rock fragments in Acadian sandstones (as opposed to Taconian sandstones), suggests unroofing of deeper crustal levels with time. Acadian orogenesis is attributed to accretion of the late Neoproterozoic Carolina terrane onto Laurentia (Wortman et al., 2000; Hatcher, 1989) (Chapter 5, this volume). Detrital zircon data (Fig. 20, D-2 and M-1) are consistent with recycling from older stratigraphic units and the presence of Pan-African age zircons supports the accretion of the Carolina arc during the Acadian (Park et al., 2010). This is supported by the presence of the few Trans-Amazonian age zircons in two of the samples. The strong Paleozoic signal in most of the samples (Fig. 20) can be related to uplift and erosion of Taconian-age plutons in the hinterland (Chapter 5, this volume).

The Alleghanian orogeny in the Late Mississippian-Pennsylvanian of Virginia was related to collision of Africa with Laurentia coupled with far field effects associated with collision of South America and southern Laurentia (Hatcher, 1989; Ettensohn 1994). Initial onset of thrust loading is reflected in the abrupt thickening of the Mississippian Greenbrier carbonates (TST of supersequence M-2) into the foredeep in southwestern Virginia. Synorogenic, mostly siliciclastic sedimentary rocks preserved in the Valley and Ridge and Appalachian Plateau provinces make up the highstand of supersequence M-2 and supersequence P-1. Detrital zircon data from upper Mississippian sandstones (Fig. 20, M-2) are consistent with recycling of older stratigraphic units (Eriksson et al., 2004) and a decrease over time in the contribution of Paleozoic plutons and the Carolina terrane. Data from Pennsylvanian sandstones (Fig. 20, P-1) also support recycling of older stratigraphic units. The presence of Archean zircons in quartz arenite bodies reflects their deposition in longitudinal braided rivers that were sourced, in part, by the

Superior provenance to the north (Eriksson et al., 2004). Their absence in sublitharenite bodies is because these were transverse river systems eroding highlands to the southeast; the presence of Pan-African and Paleozoic zircons in one of these sandstones is consistent with a provenance to the southeast. The Alleghanian orogeny culminated in massive overthrusting of Grenville basement from the southeast to produce the outcropping Blue Ridge Province (Hatcher, 1972; Cook et al., 1979).

EUSTATIC CONTROLS ON SEQUENCE HIERARCHY

Correlation of the relative sea-level changes in the Appalachian Paleozoic succession is hampered by biostratigraphic resolution of the fossils, and scarcity of other time controls such as dated ash beds, magnetostratigraphy and isotope stratigraphy in the successions in Virginia. However, there is strong evidence that eustatic sea-level changes exerted a major control on sequence deposition. For example, in the passive margin succession, many of the Cambrian sea-level cycles in the Appalachians can be matched with those of the Great Basin in the western U.S. (Osleger and Read, 1993; Read, 1989a, b). Similarly, the Early Ordovician sea-level cycles recognized in the Appalachians are recognizable elsewhere (Read, 1989b; Read and Goldhammer, 1988; Goldhammer et al., 1993; Schutter, 1992; Taylor et al., 1992).

Even in the tectonically active foreland basin, many of the relative sea-level cycles appear to have a global cause, although tectonics may have slightly modified the timing of unconformity development over bulges and on the proximal foreland where uplift during rebound may have occurred (cf. Haynes et al., 2005). Middle and Late Ordovician sequences in the Appalachian Basin have been shown to relate to eustatic sea-level changes (Holland and Patzkowsky, 1996; Pope and Read, 1997; Dennison, 1976). Silurian and Devonian sea-level cycles have long been considered to be the result of glacio-eustasy (Johnson, 1996; Dennison and Head, 1975; Diecchio and Broderson, 1994; Filer, 2002). Mississippian 4th-order sequences have been traced from the Appalachian Basin into the Illinois Basin, which, along with their ~400 k.y. estimated periodicities and association with Gondwanan glaciation, is indicative of a glacio-eustatic driver (Miller and Eriksson, 2000; Smith and Read, 2000; Al-Tawil et al., 2003). Similarly, the Mississippian composite sequences can be tied to the Mississippian sea-level curve of Ross and Ross (1987). Pennsylvanian sequences throughout much of North America have been shown to have formed under high amplitude glacio-eustasy, driven by Gondwanan glaciation (Crowell, 1999; Heckel, 1985), in which 3rd order sea-level cycles possibly related to 2.5 m.y. orbital eccentricity, have superimposed 4th-order (400 k.y.) orbital eccentricity cycles.

The average duration of the 2nd-order supersequences in the Paleozoic of the Virginia Appalachians is ~14 m.y., similar to astronomically driven 2nd-order sequences (Al-Husseini and Matthews, 2010). These are bounded by regional unconformities. Tectonics also played a significant role in development of the supersequence-bounding unconformities (Ettensohn, 1994). It is likely that uplift in proximal foreland basin areas at the end of each tectophase, or during subsequent bulge uplift amplified eustatic sea-level falls, to form the major Appalachian unconformities that bound many of the supersequences in the foreland basin succession. On the distal foreland, bulge uplift during initial downwarping of the foredeep may have influenced unconformity formation.

PALEOCLIMATE CONTROLS ON DEPOSITION

Paleoclimate history and paleolatitudinal position of Virginia, based on Scotese (2001) and other sources, are summarized in Figure 21. Glacigene deposits between 750 and 570 m.y. attest to a cold climate along eastern North America in the Late Proterozoic; glaciation may have lingered up to 520 m.y. (until Early Cambrian in Africa, summarized in Crowell, 1999). The position of Laurentia in the latest Proterozoic (~570 m.y.) at the time of rifting is contentious. It may have been located at high latitudes, based on paleomagnetic evidence from the Catoctin volcanics, with Virginia near the South Pole (Meert et al., 1994). However, it also has been suggested that Laurentia lay at low latitudes at this time (Meert and Torsvik, 2004).

For the bulk of Cambrian-Early Ordovician time, the passive margin was in low latitudes (Meert and Torsvik, 2004), implying that in the latest Neoproterozoic, Laurentia had either migrated north from high latitudes to low latitudes, or was located in low latitudes, in which case the paleoclimate indicators are the product of warming of the globe from the late Proterozoic into the Cambrian. This low latitude setting favored the development of a microbial reefal rim along the Laurentian margin. The Cambrian-Early Ordovician earth had a global greenhouse climate with little, if any, polar ice. This resulted in precessionally driven sea-level changes that were small, forming the classic meter-scale cycles that dominate the passive margin succession (Koerschner and Read, 1989; Simpson and Eriksson, 1990). The local climate was hot and relatively dry, indicated by the abundant microbial laminites and silicified evaporite nodules in the carbonates. Periodic, short lived humid climate phases may have punctuated this hot dry climate during Milankovitch-driven high stands in the Middle Cambrian intrashelf basin, when large volumes of marine shale were deposited in the southern Appalachians, under the influence of paleotrade wind-driven longshore currents (Markello and Read, 1981; Dalrymple et al., 1985; Read, 1989a, b).

The climate became increasingly humid during much of the Middle to Late Ordovician, indicated by decrease in ooids, microbial laminites and evaporite pseudomorphs, and presence of paleosinkholes and solution pits on exposed carbonates, meteoric cementation in regional aquifer systems, and stratification of basin waters beneath less saline surface waters (Witzke, 1990; Grover and Read, 1978, 1983; Pope and Read, 1998). This increased humidity could have been due to southward migration of Laurentia and elevation of the Taconic tectonic highlands, increasing the monsoonal rainfall. Global cooling following explosive volcanism (Kolata et al., 1996) initiated Gondwana glaciation and may have triggered widespread deposition of cooler water carbonates at the base of Late Ordovician supersequence O-3. Glaciations may have been initiated during the Caradoc (early Late Ordovician); these resulted in moderate amplitude Milankovitch-driven sea-level changes that influenced deposition of the distinctive later Ordovician parasequences in Kentucky (Pope and Read, 1998). These glaciations peaked on Gondwana at the end of the Ordovician with a short-lived, glacial episode approximately 2 m.y. duration (Brenchley et al., 1994). This caused a significant sea-level fall and siliciclastic progradation onto the distal foreland, terminating Ordovician deposition. Periodic small glaciations on Gondwana continued into the Early Silurian (Caputo, 1998). Continued cool, humid conditions favored widespread siliciclastic deposition in Virginia, terminating with deposition of the Bloomsburg redbeds under drier climate. In the later Silurian (Wills Creek-Tonoloway interval), gradual warming and drying shut down siliciclastic influx and led to widespread deposition of cyclic peritidal carbonates and evaporites, including halites in the depocenter in West Virginia. The abundant meter-scale cycles in the Tonoloway (Bell and Smosna, 1999) are suggestive of relatively ice free, greenhouse global climates at this time.

Late Silurian-Early Devonian Keyser-Helderberg mixed carbonate-siliciclastic units were deposited in more humid, warm subtropical settings, which favored influx of quartz sands and muds from riverine point sources during lowered sea levels, and promoted later meteoric cementation of the carbonates by regional aquifers (Dorobek, 1987). Late Devonian climate was warm temperate to tropical humid, with widespread development of salinity stratified basins, in which a less saline surface water mass rested on a normal marine bottom water mass resulting in anoxic deeper water facies of the Millboro Formation (Algeo et al., 1995). Frasnian-Famennian glaciation is well documented on Gondwana, and waxing and waning of these ice sheets likely formed the strong 100 to 400 k.y. eccentricity signal evident in the Late Devonian siliciclastics (Filer, 2002; van Tassell, 1994). The climate became drier during Hampshire deposition, but then cooled and became wetter in the upper Hampshire (Brezinski et al., 2009). The lower Hampshire redbeds formed in drier conditions, but the upper Hampshire Formation records cooling toward the Devonian-Mississippian boundary (Brezinski et al., 2009). Humid conditions continued into the early Mississippian during Price delta deposition.

Climate became arid in the Early Meramecian resulting in deposition of the Maccrady redbeds and salt. Renewed flooding of the region in the late Meramecian brought more humid conditions (plant detritus in the Little Valley units), but climate then became drier during the early Chesterian when the oolitic Greenbrier ramp developed; widespread eolianites at this time attest to aridity during lowered sea level stages (Smith et al., 2001). Conditions then became more humid into the later Chesterian (upper Greenbrier-Bluefield formations) which shows a decrease in oolitic facies, and massive influx of shale in the Bluefield Formation. Climate again became drier during Mauch Chunk highstand deposition, with widespread development of redbeds and vertisols indicative of semi-arid conditions (Cecil, 1990; Miller and Eriksson, 2000)

Ice sheets were initiated on Gondwana during the Early Chesterian Greenbrier deposition, and resulted in 400 k.y. sea level changes of a few tens of meters, that generated the dominant 4th-order sequences in the Greenbrier-Mauch Chunk succession. Magnitude of sea level changes may have increased into the later Mississippian perhaps almost reaching 100 m (Miller and Eriksson, 2000; Smith and Read, 2000). Third-order composite sequences typical of the Mississippian (and the Pennsylvanian), may relate to the long-term (~2.5 Ma) glacioeustatic changes related to orbital eccentricity (cf., Matthews and Frolich, 2002).

By Pennsylvanian time, the area had migrated into the equatorial belt, and the climate in the Appalachians had become dominantly humid once again. The humid phases commonly occurred during glacio-eustatic lowstands, alternating with seasonally dry-subhumid phases during late highstands and early transgression (Cecil et al., 2003; Miller and Eriksson, 1999). These drier phases related to increased seasonality of rainfall in low latitudes due to the mega-monsoon effect. The alternating climates were driven by Milankovitch eccentricity forcing, typically at ~400 k.y. periodicities. Pennsylvanian sea-level changes driven by waxing and waning of Gondwana ice sheets probably were large, from 60 to 100 m magnitude.

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APPENDIX A SUMMARY OF FACIES IN SUPERSEQUENCES

Supersequence Ꞓ-1 Chilhowee Group Siliciclastics

TST: 1. braided-alluvial succession of channelized conglomerates and feldspathic sandstones; 2. shallow-marine parasequences of mudstone with interbedded sandstone, 3. quartz arenite, capped by large-scale symmetrical and asymmetrical ripples (on drowning surfaces). Typify Unicoi-Weverton ;

HST: 1. deep shelf carbonaceous mudstone (Hampton-Harpers); 2. outer- to inner-shelf, massive, bioturbated or laminated mudstone, lesser rippled siltstone and sandstone 3. inner-shelf to shoreface quartz arenites with abundant Skolithus; typify Erwin-Antietam Formations.

Supersequence Ꞓ-2: Upper Erwin-Shady-Rome formations

Transgressive inner shelf quartz arenites (Upper Erwin Fm).

Ramp phase: Shady Dolomite: 1. deep ramp/slope nodular argillaceous lime mudstones, with lime conglomerates in upper part (Patterson Member). 2. carbonate mudmounds with stromatactis and archaecyathids; 3. (Austinville Member) -- white, massive oolitic dolomite and cyclic dolomite with lowstand quartzose units.

Rimmed shelf phase: 1. downslope periplatform apron of thin bedded, black shaly fine grained carbonate rhythmites, carbonate sands, turbidites and megabreccias. 2. reefal rim - dolomitized grainstone, with archeocyathid and Epiphyton-Renalcis microbial reefs; 3. (Upper Shady Dolomite) peritidal cyclic massive/bedded subtidal dolomites with microbial laminite caps, and lesser, interbedded red clastics; 4. siliciclastic inner shelf (Rome Formation) - meter-scale parasequences of lower intertidal and shallow subtidal red and gray shale and sandstone, and local coarse channel-fills, laminated fine shale and sandstone, tidal–supratidal redbeds and fenestral dolomite and silicified evaporite beds

Supersequence -Ꞓ3: Conasauga-Elbrook Interval

1. Intrashelf basin fill; 1 to 4 m, upward shallowing parasequences of green-grey shale to thin, flat-laminated to hummocky laminated calcareous siltstones capped by thin, skeletal-oolitic packstone and flat-pebble conglomerate storm beds; glauconite and hardgrounds common near tops. 2. deep ramp ribbon carbonates (argillaceous carbonate mudstone, minor wackestone, and thin storm lags (Maryville and Maynardville Limestone); 3. cross-bedded, dolomitized ooid grainstones interbedded with thin units of microbial laminate or intraclast rudstone border intrashelf basin (Honaker Formation); 4. cyclic peritidal platform interior; meter-scale parasequences of intraclastic lags (local), microbial thrombolite heads or thin bedded pellet mudstone, capped by microbial laminites and minor shale. Subtidal facies commonly limestone, laminite caps typically dolomitized (Elbrook Formation). In Middle Limestone (LST) parasequences of thin black shale to calcisiltite to flat pebble-oolitic rudstones capped by microbial mounds (2 m relief).

Supersequence Ꞓ-4: Conococheague-Copper Ridge formations

Pertidal platform with northern and eastern facies dominated by cyclic limestone (Conococheague Limestone), western facies dominated by cyclic dolomite (Copper Ridge Dolomite). TST cycles thicker, and more subtidal facies; HST cycles thin and more restricted facies. Facies: 1. shallow subtidal dark ooid intraclast grainstone sheets (1 cm to 2 m thick), cross bedded and rippled, and thin lags of flat pebble conglomerate; grainstones may fill and fine up within rare channels; commonly transgressive; 3. digitate bioherms and thrombolites (10 cm to 3 m thick) locally capped by silicified digitate stromatolite veneer 4. thinly bedded, mechanically laminated, rippled and scoured pellet packstone/mudstone, arranged in centimeter fining-upward layers; may have skeletal intraclast lags in lower part; 5. intertidal light colored, thick-laminite to microbial, crinkled, small stromatolites (rare) and stratiform microbial thin-laminite sheets, mudcracked with silicified evaporite nodules; quartz sandstone laminae and beds common in laminite caps in the lower and upper parts of supersequence; 5. subaerial breccias (locally common near sequence boundaries.

Supersequence O -1:Chepultepec-Kingsport-Mascot(south-western Va) and Stonehenge, Rockdale Run, Pinesburg Station formations (northern Va)

Limestone-dominated subtidal parasequences are typical of TSTs and northern and eastern belts and are best developed low in the Chepultepec-Stonehenge interval; completely dolomitized restricted parasequences typical of western belts and HSTs. Facies include 1. storm-influenced deeper ramp, thin bedded wackestone-mudstone with storm-deposited, fining upward layers, and laminated and cross laminated skeletal-pellet grainstone; 2. pellet grainstone and intraclast conglomerates; 3. thrombolite and stromatolite mounds (biostromes and mound complexes) locally capped by cross bedded skeletal sands; 4. restricted shallow subtidal massive dolomudstone; 5. intertidal thick-laminite to microbial laminite with silicified evaporite nodules, layers and beds, and thin shale drapes. Facies are sandy near sequence boundaries.

Supersequence O-2: Middle Ordovician Limestone

Facies include 1. slope/basin thin, evenly bedded black slope/basin limestone and shale, and black graptolitic shale (Liberty Hall Rich Valley Formation and Paperille Shale); 2. lower slope nodular, dark gray lime mudstone/ minor wackestone (Lantz Mills formation); 3. deep ramp shaly nodular skeletal wackestone/mudstone and minor packstone (Ben Bolt Formation); 4. transgressive carbonate banks (tens of kilometers across) on mid-ramp, and narrower (few kilometers wide) but thicker buildups (up to 250 m thick) on deeper slope; banks have muddy cores and crinoid-rich grainy flanks (Effna, Rockdell); highstand banks are skeletal grainstone sheets (Wardell/ Wassum); 5. subtidal quiet lagoonal cherty, pellet mudstone and cherty pellet grainstone sandflat facies and channel-fill grainy carbonates (Elway/Witten); 6. intertidal fenestral and laminated lime mudstone, some with microkarst tops (New

Market, Mosheim, Five Oaks, Gratton); 7. detrital dolomite breccia, dolomudstone and shale (Blackford Fm.). Siliciclastic proximal basin margin 1. deep water channelized turbidite fan deposits composed of interbedded shale, lithic arenite turbidites and channel conglomerates (Knobs Formation and Fincastle Conglomerate); conglomerate clasts include gray and red feldspathic and quartz arenite, siltstone, mudstone, dolomite, limestone, chert and quartz-chlorite rock and rare gneissic clasts; lithic arenites contain recycled sedimentary and quartz-mica schist grains; 2. Bays deltaic/marine, lithic and felspathic sandstone, cross-bedded and planar bedded locally, burrowed, interlayered with mudrock and thin sandstone layers 3. Bowen and Moccasin red mudrock, siltstone and sandstone and rare quartz conglomerate sheets.

Supersequence O-3: Martinsburg to basal Tuscaroora Interval

Lower Martinsburg Formation 1. basinal black limestone and shale (Lower Martinsburg Formation, N. Virginia and Reedsville Shale; 2. thick, siliciclastic turbidite apron facies - coarsening-up units of green to gray immature sandstone, siltstone and shale; 3. interbedded fossiliferous siltstone, shale and limestone, and then immature sandstone with Orthorhynchula and Lingula in upper Reedsville; 4. inner shelf; cross-bedded sandstone (green-gray felspathic litharenite to sublitharenite); locally with polymictic conglomerate (clasts of quartzite, volcanic and sedimentary rock fragments, including oolitic dolomite, slate and chert; Oswego).

Southwest Virginia: Martinsburg 1. storm influenced deep ramp, shale with fining-up units of calcarenite/calcisiltite (locally quartz-silty) 2. interbedded shale and storm beds of shelly packstone, quartz silty calcarenite-calcisiltite, siltstone and sandstone; 3. whole fossil packstone proximal ramp (developed as transgressive unit at base of Martinsburg); 4. siliciclastic lower-to middle-shoreface trough-cross bedded sandstone, highly burrowed in lower parts and wave-rippled in upper parts; capped by foreshore laminated and local tabular cross-laminated sandstone (Martinsburg-Juniata transition); 5A. siliciclastic marine channeled sandflat- and inlet-fills of red and lesser green cross bedded sublitharenite/quartz arenite with mudstone ripups, and mixed flat, interbedded mudstone and arenite to wacke with Skolithos burrows, capped by 6. intertidal/supratidal burrowed mudstone facies (some Lingula, abundant trace fossils, Skolithos, Cruziana and Rusophycus)(Juniata Fm.). Juniata Fm. grades laterally into Sequatchie Fm. 5B. storm influenced neashore ramp; thin bedded, gray and red mudstone and gray, laminated and rippled, silty peloid limestone, calcareous siltstone (erosional bases and intraclast lags), and very fine sandstone plus skeletal packstone, diverse biotas; overlain by 6. peritidal unfossiliferous mudcracked or burrowed mudstone, and thin-bedded, plane- and ripple-laminated, very fine sandstone-siltstone, intraclastic lags.

Supersequence S-1: Tuscarora-Rose Hill-Keefer/Eagle Rock Interval

1. Incised valley fill conglomerate (basal unconformity), 2. upper shoreface quartz arenites, massive with rare cross beds, ripples, Arthrophycus, burrows and mudcracks (Keefer/Eagle Rock), 3. storm influenced lower shoreface, cross bedded and rare hummocky cross stratified arenites with thin mud drapes, Skolithos, Monocraterion, Arthrophycus and less common Arenicolites, Ruzophycus, Cruziana, Planolites, Phycodes, 4. more distal, shallow-marine deposits (Dorsch and Driese, 1995) of heterolithic facies made up of hematite-cemented quartz-rich sandstones with rare hummocky cross stratification, interbedded shales and siltstones, and fossiliferous shales. Trace fossils include Skolithos, Planolites, Monocraterion and less common Chondrites, Ruzophycus and Cruziana; bioturbation is commonly intense, (Rose Hill Formation).

Supersequence S-2: Rochester-McKenzie-Bloomsburg Interval

1. Subtidal shale and thin limestone interbeds, some with ostracods (KcKenzie Formation), 2. thin trangressive marine shale (Rochester), 3. sandy shoreline facies Williamsport) grading laterally into non-marine red mud rocks and red sandstone, and grey-green sandstone and shale (Bloomsburg redbeds).

Supersequence S-3: Wills Creek, Tonoloway-Lower Keyser Interval

Mixed siliciclastics-carbonates: Wills Creek thin-bedded, fine grained sandy limestone, calcareous shale and mudstone, local pebbly sandstone (TST) and limestone and shaly limestone, dolomite and sandstone (HST).

Carbonates: Tonoloway- carbonate and mixed carbonate-siliciclastic sabkha carbonates 1. nodular argillaceous wackestone-mudstone, and deep ramp, storm deposited skeletal packstone and argillaceous laminated carbonates, 2 off-shoal stromatoporoid reefs along with packstone and wackestone, 3. shallow ramp crinoidal grainstone , 4. microbially laminated carbonates and leached evaporites, lagoonal peritidal thin bedded carbonates.

Supersequence D-1: Helderberg-Oriskany Interval

Helderberg Group; 1. deep ramp, cherty argillaceous laminated carbonates or limy shale with increasing storm beds of skeletal packstone updip; brachiopods, horn corals, spicules and hargrounds common; 2. nodular argillaceous wackestone-mudstone or argillaceous pellet packstone; 3. downslope stromatoporoid reefs along with skeletal packstone and wackestone; 4. shallow ramp crinoidal grainstone sheets and local mudmounds; 5. lagoonal peritidal thin bedded muddy carbonates and sabkha microbially laminated carbonates and evaporites. In Helderberg, basin margin sandstones (Clifton Forge, Elbow Ridge/Healing Springs, and Rocky Gap sandstones) include 1. offshore, fine argillaceous sandstone, laminated, hummocky cross-laminated, highly burrowed; 2. offshore-bar, tabular to trough-cross bedded fine to medium-grained sandstone interbedded with local bafflestones; 3. nearshore, cross-bedded, medium to coarse sandstones with tidal structures.

Oriskany or Ridgeley Sandstone; calcareous quartzarenite, with heavy shelled brachiopods and some snails; in outcrop, fossils leached to form fossil-moldic quartzenarite.

Supersequence D-2: Huntersville Chert to Mahantango formations

Southwest Virginia: 1. black, nodular chert, argillaceous and phosphatic at base, relatively pure spicular chert upsection (Huntersville Chert); to north grades into Needmore Shale; dark green, diverse biota (brachiopods, trilobites, bryozoa, coral and mollusks); 3. sandstone (locally fossiliferous) and sandy fossiliferous cherty argillaceous limestone (Onondaga Limestone, Butts, 1940) with brachiopods, mollusks.

Supersequence D-3: Millboro to Hampshire formations

1. Pro-deltaic black shale (anaerobic or dysarobic) (Millboro Formation); 2. turbidite-slope grey or organic-rich, black shale-mudstone with intercalated graded beds of siltstone/mudstone or massive siltstone (Brallier Formation); 3. storm-shelf, interbedded grey or black shale-mudstone and siltstone, local sandstone and conglomerate (Foreknobs/Chemung Formation). Storm-beds 5 to 200 cm thick; basal lags of bioclasts and/or quartz-pebble conglomerates to hummocky cross-stratification to wave-rippled top; conglomerate lags proximal to lobes; bioclastic lags more distal; intensely bioturbated mudstone intervals, fair-weather deposits; 4. red, channel cross-bedded sandstones and overbank mudstones (Hampshire Formation); low-sinuosity, braided and less common high-sinuosity meandering river as well as sheetflood deposits developed; intercalated pedogenic red mudstones.

Supersequence M-1: Mississippian Cloyd-Sunbury-Price-Maccrady Interval

Basal unit; 1. fluvial-deltaic conglomeratic incised channel-belt (Cloyd Conglomerate) wedges out into fossiliferous sandstone and siltstone to the south and west, up into 2. slightly fossiliferous, estuarine and tidal flat facies

Progradational deltaic succession: 1. Basinal deep water, dark gray to black shale with thin tabular sands (Sunbury Shale Mbr.); 2. fan turbidites 3. storm-deposited fine-grained sandstone and shale (distal to proximal lower shoreface tempestites) 4. massive upper shoreface sandstone; locally horizontally laminated or cross-bedded, and back-bar and bay, finer grained sandstone-shale; and distributary mouth-bar sands (coarsening upward units of fine to medium grained sandstone with mud chip lags; 6. non-marine, subaerial delta plain facies of sandstone, siltstone and silty shale with local, distributary channel-fill sandstones and offlapping coals; 7. non-marine coastal alluvial plain, red mudstones with channel-fill sandstone, crevasse-splay deposits and calcretes; some thin limestones and significant evaporites at base of Maccrady Formation in southwest Virginia.

Supersequence M-2: Mississippian Greebrier Limestone-Mauch Chunk Interval

Greenbrier Limestone: Little Valley Formation-Hillsdale Interval: 1. very fossiliferous nautiloid-bearing shale (mfs?); 2. deeper, open marine skeletal wackestone and fossiliferous limy mudrock; may be oncoidal 3. thin skeletal grainstone and oolite, thin sandstone and unfossiliferous mudrock with evaporite pseudomorphs. Denmar-Gasper interval: 1. basinal laminated, silty mudstone and fine silty packstone, 2. skeletal wackestone 3. coarse skeletal grainstone/packstone 4. ooid grainstone 5. lagoonal carbonate mudstone, variably dolomitic; 7. lowstand-transgressive siliciclastics, local incised valley fills (eolian, sandflat-mudflat, lagoonal shale).

Overlain by progradational highstand (Mauch Chunk Gp.): 1. deeper water limestone and gray and black shale (Cove Creek) 2. open shelf skeletal packstone/wackestone and gray shale, 3. prodeltaic shale/siltstone with tidal rhythmites; tidal signals are diurnal, fortnightly neap-spring, and monthly; annual monsoon-modulated tidal cycles average 10 cm in thickness and contain up to 17 neap-spring cycles (Pride Shale); 4. tidal deltaic sandstone 5. coastal plain redbeds (Hinton), 6. incised valley fill sandstones and conglomerate lags (Fido, Indian Mills, Stony Gap, Princeton sandstones).

Supersequence P-1: Early to Middle Pennsylvanian Breathitt Group

Breathitt Group: 1. thin marine shales 2. deltaic sandstones, interbedded sandstone and shale, coals 3. estuarine tidal rhythmites 4. tabular to lenticular sandstones, braided alluvial, and incised valley fills (lowstand) .

Warren Point, Sewanee, Bee Rock and Corbin; tabular, quartz arenite bodies; channelized scour and fining-upward channel fills, unidirectional southwesterly paleocurrents, ubiquitous basal quartz-pebble conglomeritic lags, in-situ tree roots, abundant plant debris, associated peat deposits and a paucity of shelly body fossils.

Correlative sandstones to southeast are lenticular, lithic arenite bodies that contain abundant schist grains, have channelized scour and upward-fining successions, locally derived sandstone and coal-clasts; trough cross beds indicate flow towards west and northwest.

REFERENCES

Abplanalp, J. and Lehmann, D., 2004, High resolution stratigraphy and parasequences of the Devonian, Oriskany Sandstone, Mapleton, Pennsylvania: Geological Society of America Abstracts with Programs, v. 36, no. 2, p. 88.

Algeo, T. J., Berner, R. A., Maynard, J. B., and Scheckler, S. E., 1995, Late Devonian oceanic anoxic events and biotic crises; "rooted" in the evolution of vascular land plants?: GSA Today, v. 5, p. 45, 64-66.

Al-Husseini, M. I., and Matthews, R. K., 2010, Calibrating Mid-Permian to Early Triassic Khuff sequences with orbital clocks: GeoArabia, v. 15, no. 3, p. 171-206.

Al-Tawil, A., and Read, J. F., 2003, Late Mississippian (Chesterian) glacio-eustatic sequence development on an active distal foreland ramp, Kentucky. Carboniferous and Permian Carbonate Platforms: SEPM Special Publication, p. 35-55.

Al-Tawil, A., Wynn, T. C., and Read, J. F., 2003, Sequence response of a foreland carbonate ramp during Mississippian glacio-eustasy, Appalachian Basin, eastern U.S.A.: Carboniferous and Permian Carbonate Platforms, SEPM Special Publication, p. 11-34.

Aleinikoff, J. N., Zartman, R. E., Walter, M., Rankin, D. W., Lyttle, P. T., and Burton, W. C., 1995, U-Pb ages of metarhyolites of the Catoctin and Mount Rogers formations, central and southern Appalachians: evidence for two pulses of Iapetan rifting: American Journal of Science, v. 295, p. 428-454.

Archer, A. W., and Greb, S. F., 1995, An Amazon-scale drainage system in the Early Pennsylvanian of central North America: Journal of Geology, v. 103, p. 611-628.

Arkle, T., Jr., 1974, Stratigraphy of the Pennsylvanian and Permian systems of the central Appalachians, in Briggs, G., ed., Carboniferous of the southeastern United States: Geological Society of America Special Paper 148, p. 5-30.

Ayrton, W. G., 1963, Isopach and lithofacies map of the Upper Devonian of northeastern United States: in Shepps, V. C., ed., Symposium on Middle and Upper Devonian stratigraphy of Pennsylvania and adjacent states: Pennsylvania Geological Survey, General Geology Report, G39, p. 3-6.

Bambach, R. K., 1987, The Ordovician-Silurian unconformity in western Virginia and adjacent West Virginia, in Shumaker, R.S., compiler, Appalachian Basin Industrial Associates, Fall 1987, v. 13, p. 2-14, Morgantown, West Virginia.

Barnaby, R., and Read, J. F., 1990, Evolution of Early to Middle Cambrian carbonate continental shelf, Virginia Appalachians: Geological Society of America Bulletin, v. 102, p. 391-404.

Bartlett, C. S., 1974, Anatomy of the Lower Mississippian delta in southwestern Virginia [Ph.D. dissertation]: Knoxville, University of Tennessee, 374 p.

Bartlett, C. S., 1997, The rock salt deposit at Saltville, Virginia, in Bartlett, C. S., Jr., and Whisonant, R. C., eds., A geological sample of Saltville and vicinity: Southern Appalachian Geological Association Guidebook, p. 1-10.

Bartlett, C. S. Jr., and Webb, H. W., 1971, Geology of the Bristol and Wallace quadrangles, Virginia: Virginia Division of Mineral Resources, Report of Investigations 25, 93 p.

Becker, T.P, Thomas, W.A., Samson, S. D., and Gehrels, G. E., 2005, Detrital zircon evidence of Laurentian crustal dominance in the lower Pennsylvanian deposits of the Alleghanian clastic wedge in eastern North America: Sedimentary Geology, 182, 59-86.

Bell, S. C., and Smosna, R., 1999, Regional facies analysis and carbonate ramp development in the Tonoloway Limestone (Upper Silurian; Central Appalachians): Southeastern Geology, v. 38, n. 4, p. 259-278.

Betzner, J., and Read, J. F., 2007, Evidence for Greenhouse global climate: Early Cambrian Shady Dolomite, Virginia: Geological Society of America Southeastern Section 56th Annual Meeting, Abstracts with Programs, Paper No. 23-6.

Bjerstedt, T. W., 1986, Regional stratigraphy and sedimentology of the lower Mississippian Rockwell Formation and Purslane Sandstone based on the new Sideling Hill road cut, Maryland; Southeastern Geology, v. 27, p. 69-94.

Bjerstedt, T. W., and Kammer, T. W., 1988, Genetic stratigraphy and depositional systems of the Upper Devonian-Lower Mississippian Price-Rockwell deltaic complex in the Central Appalachians, U.S.A.: Sedimentary Geology, v. 54, p. 265-301.

Black, L. P., Kamo, S. L., Allen, C. M., Davis, D. W., Aleinikoff, J. N., Valley, J. W., Mundil, R. et al., 2004, Improved 206Pb/238U microprobe geochronology by the monitoring of trace-element-related matrix effect: SHRIMP, ID-TIMS, ELA-ICP-MS and oxygen isotope documentation for a series of zircon standards: Chemical Geology, v. 205, p. 115-140.

Blancher, D. W., 1974, Sediments and depositional environments of the Maccrady Formation and the Greenbrier Group (Mississippian) of the Hurricane Ridge syncline of southwestern Virginia and West Virginia [unpublished Ph.D. dissertation]: Blacksburg, Virginia Polytechnic Institute and State University, 229 p.

Bodek, R. J., 2006, Sequence stratigraphic architecture of early Pennsylvanian, coal-bearing strata of the Cumberland Block: A case study from Dickenson, County, Virginia [MS Thesis]: Blacksburg, Virginia Polytechnic Institute and State University, 55 p.

Bond, G. C., Nickeson, P. A., and Kominz, M. A., 1984, Breakup of a supercontinent between 625 and 555 Ma: New evidence and implications for continental histories: Earth and Planetary Science Letters, v. 70, p. 325-345.

Bova, J. A., and Read, J. F., 1987, Incipiently drowned facies within a cyclic peritidal ramp sequence, Early Ordovician Chepultepac interval, Virginia Appalachians: Geological Society of America Bulletin, v. 98, p. 728-738.

Brame, R. I., 2001, Revision of the Upper Devonian in the Central Appalachian Basin [PhD dissertation]: Blacksburg, Virginia Polytechnic Institute and State University, 168 p.

Brenchley, P. J., Marshall, J. D., Carden, G. A. F., Roberston, D. B .R., Meidla, T., Hints, L., and Anderson, T. F., 1994, Bathymetric and isotopic evidence for short-lived Late Ordovician glaciation in a greenhouse period: Geology, v. 22, p. 295-298.

Brezinski, D. K., Cecil, C. B., Skema, V. W., and Kertis, C. A., 2009, Evidence for long-term climate change in Upper Devonian strata of the central Appalachians:

Palaeogeography, Palaeoclimatology, Palaeoecology, v. 284, p. 315-325.

Butts, C., 1940, Geology of the Appalachian Valley and Ridge in Virginia, Part I – Geologic text and illustrations: Virginia Geological Survey Bulletin 52, University of Virginia, 568 p.

Butts, C., 1941, Geology of the Appalachian Valley and Ridge in Virginia, Part II –Fossil Plates and Explanations: Virginia Geological Survey Bulletin 52, University of Virginia, 271 p.

Byrd, W. J., 1973, Petrology of the Cambrian Shady Dolomite in North Carolina, northeast Tennessee, and southwest Virginia [Unpublished Ph.D. Dissertation]: University of North Carolina, Chapel Hill, 152 p.

Caputo, M., 1998, Ordovician-Silurian glaciations and global sea-level changes, in Landing, E., and Johnson, M. E., eds., Silurian Cycles, Linkages of Dynamic Stratigraphy with Atmospheric, Oceanic and Tectonic Changes: James Hall Centennial Volume, New York State Museum Bulletin 491, p. 15-25.

Castle, J. W., 1998, Regional sedimentology and stratal surfaces of a Lower Silurian clastic wedge in the Appalachian foreland basin: Journal of Sedimentary Research, v. 68, p. 1201-1211.

Castle, J. W., 2001, Appalachian basin stratigraphic response to convergent-margin structural evolution: Basin Research, v. 13, p. 397-418.

Caudill, M. R., Driese, S. G., and Mora, C. I., 1996, Preservation of a paleo-vertisol and an estimate of Late Mississippian paleoprecipitation: Journal of Sedimentary Research, v. A66, p. 58-70.

Cawood, P.A. and Nemchin, A.A., 2001, Paleogeographic development of the eastern Laurentian margin: constraints from U-Pb dating of detrital zircons from the northern Appalachians. Bulletin Geological Society of America, v. 113, p. 1234-1246.

Cecil, C. B., 1990, Paleoclimate controls on stratigraphic repetition of chemical and siliciclastic rocks: Geology, v. 18, p. 533-536.

Cecil, C. B., Dulong, F. T., West, R. D., Stamm, R., Wardlaw, B., and Edgar, N. T., 2003, Climate controls on the stratigraphy of a Middle Pennsylvanian cyclothem in North America, in Cecil, C. B. and Edgar, N. T., eds., Climate controls on stratigraphy, SEPM (Society for Sedimentary Geology) Special Publication, no. 77, p. 151-180.

Cecil, C. B., Stanton, R. W., Neuzil, S. G., Dulong, F. T., Ruppert, L. F. and Pierce, B. J., 1985, Paleoclimate controls on Late Paleozoic sedimentation and peat formation in the Central Appalachian Basin: International Journal of Coal Geology, v. 5, p. 195-230.

Chesnut, D. R., Jr., 1988, Stratigraphic Analysis of the Carboniferous Rocks of the Central Appalachian Basin [Ph.D. dissertation]: Lexington, University of Kentucky, 297 p.

Chesnut, D. R., Jr., 1994, Eustatic and tectonic control of deposition of the Lower and Middle Pennsylvanian strata of the central Appalachian basin, in Dennison, J. M., and Ettensohn, F. R., eds., Tectonic and eustatic controls on sedimentary cycles: SEPM Concepts in Sedimentology and Paleontology, v. 4, p. 51-64.

Colton, G. W., 1970, The Appalachian Basin – Its depositional sequences and their geologic relationships, in Fisher, G. W., Pettijohn, F .J., and Reed, J. C., Jr., eds., Studies of Appalachian Geology: Central and Southern: New York, Interscience Publishers, p. 5-47.

Cook, E. G., 1981, Conodont biostratigraphy and paleoecology of the Lower Devonian Helderberg Group of Virginia [unpublished M.S. Thesis]: Virginia Polytechnic Institute and State University, Blacksburg, 152 p.

Cook, F. A., Albaugh, D. S., Brown, L. D., Kaufman, S., Oliver, J. E., and Hatcher, R. D., 1979, Thin skinned tectonics in the crystalline southern Appalachians - COCORP Seismic-reflection profiling of the Blue Ridge and Piedmont: Geology, v. 7, p. 563-567.

Cooper, B. N., 1944a, Industrial limestones and dolomites in Virginia: New River-Roanoke District: Virginia Geological Survey Bulletin 62, 98 p.

Cooper, B. N., 1944b, Geology and mineral resources of the Burkes Garden Quadrangle, Virginia: Virginia Geological Survey Bulletin 60, 299 p.

Cooper, B. N., 1945, Industrial limestones and dolomites in Virginia: Clinch Valley district: Virginia Geological Survey Bulletin 66, 259 p.

Cooper, B. N., 1960, The geology of the region between Roanoke and Winchester in the Appalachian Valley of western Virginia: Johns Hopkins University Studies in Geology, no. 18, Guidebook 2, 84 p.

Cooper, B. N., and Cooper, G. A., 1946, Lower Middle Ordovician stratigraphy of the Shenandoah Valley, Virginia: Geological Society of America Bulletin, v. 57, p. 35-114.

Cooper, B. N. and Prouty, C. E., 1943, Stratigraphy of the Lower Middle Ordovician rocks of Tazewell County, Virginia: Geological Society of America Bulletin, v. 54, p. 819-886.

Crowell, J. C., 1999, Pre-Mesozoic ice ages: their bearing on understanding the climate system: Geological Society of America Memoir 192, 106 p.

Currier, L. W., 1935, Zinc and lead region of southwestern Virginia: Virginia Geological Survey Bulletin 43, 122 p.

Dalrymple, R. W., Narbonne, G. M., and Smith, L., 1985, Eolian action and the distribution of Cambrian shales in North America: Geology, v. 13, p. 607-610.

Davydov, V., Wardlaw, B. R., and Gradstein, F. M., 2004, The Carboniferous Period, in Gradstein, F., Ogg, J., and Smith, A., eds., A Geologic Time Scale: Cambridge University Press, p. 222-248.

Demicco, R. V., 1985, Patterns of platform and off-platform carbonates of the Upper Cambrian of western Maryland: Sedimentology, v. 32, p. 1-22.

Dennison, J. M., 1970, Silurian stratigraphy and sedimentary tectonics of southern West Virginia and adjacent Virginia, in Silurian stratigraphy, central Appalachian Basin: Appalachian Geological Society, Field Conference Guidebook, p. 2-33.

Dennison, J. M., 1976, Appalachian Queenston Delta related to eustatic sea-level drop accompanying Late Ordovician glaciation centered in Africa, in Basset, M. G., ed., The Ordovician system: Proceedings of the Palaeontological Society of Birmingham, September 1974, University of

Wales Press and Natural Museum of Wales, Cardiff, p. 107-120.

Dennison, J. M., 1985, Catskill delta shallow marine strata, in Woodrow, D. L., and Sevon, W. D., eds., The Catskill Delta: Geological Society of America Special Paper, v. 201, p. 91-106.

Dennison, J. M., 1996. Geologic field guide to Devonian hydrocarbon stratigraphy of southeastern West Virginia and adjacent Virginia: Appalachian Geological Society, Charleston, WV, 160 p.

Dennison, J. M., and Head, J. W., 1975, Sea level variations interpreted from the Appalachian basin Silurian and Devonian: American Journal of Science, v. 275, p. 1089-1120.

Dennison, J. M., and Wheeler, W. H., 1975, Stratigraphy of Precambrian through Cretaceous strata of probable fluvial origin in southeastern United States and their potential as uranium host rocks: Southeastern Geology, Special Publication, no. 5, 210 p.

Dennison, J. M., Bambach, R. K., Dorobek, S. L., Filer, J. K., and Shel, J. A., 1992, Silurian and Devonian unconformities in southwestern Virginia, in Dennison, J. M. and Stewart, K. G., eds., Geologic Field Guides to North Carolina and Vicinity: Department of Geology, University of North Carolina, Chapel Hill, Guidebook 1, p. 79-105.

Derby, J. R., 1965, Paleontology and stratigraphy of the Nolichucky Formation in southwest Virginia and northeast Tennessee [Ph.D. thesis]: Virginia Polytechnic Institute and State University, Blacksburg, 468 p.

Dickinson, W.R. and Gehrels, G.E., 2009. Use of U-Pb ages of detrital zircons to infer maximum depositional ages of strat: a test against a Colorado Plateau Mesozoic data base: Earth Planetary Science Letters, v. 288, p. 115-125.

Diecchio, R. J., 1985a, Post-Martinsburg Ordovician stratigraphy of Virginia and West Virginia: Virginia Division of Mineral Resources Publication 57, 77 p.

Diecchio, R. J., 1985b, Regional controls of gas accumulation in Oriskany Sandstone, central Appalachian basin: American Association of Petroleum Geologists Bulletin, v. 69, p. 722-732.

Diecchio, R.J., 1993, Stratigraphic interpretation of the Ordovician of the Appalachian Basin and implications for Taconian flexural modeling: Tectonics, v. 12, p. 1410-1419.

Diecchio, R J., and Broderson, B. T., 1994, Recognition of regional (eustatic?) and local (tectonic?) relative sea-level events in outcrop and gamma-ray logs, Ordovician, West Virginia, in Dennison, J. M., and Ettensohn, F. R., eds, Tectonic and Eustatic Controls on Sedimentary Cycles: SEPM Concepts in Sedimentology and Paleontology #4, p. 171-180.

Diecchio, R. J., and Dennison, J. M., 1996, Silurian stratigraphy of central and northern Virginia and adjacent West Virginia, in Broadhead, T. W., ed., Sedimentary environments of Silurian Taconia - fieldtrips to the Appalachians and southern craton of eastern North America: University of Tennessee, Studies in Geology, v. 26, p. 107-127.

Dorobek, S. L., 1986, Petrography, geochemistry and origin of burial diagenetic facies, Siluro-Devonian Helderberg carbonate rocks, Central Appalachians: American Association of Petroleum Geologists Bulletin 71, p. 492-514.

Dorobek, S. L., and Read, J. F., 1986, Sedimentology and evolution of the Siluro-Devonian Helderberg Group, central Appalachians: Journal of Sedimentary Petrology, v. 56, p. 601-613.

Dorsch, J., and Driese, S. G., 1995, The Taconic foredeep as sediment sink and sediment exporter: Implications for the origin of the white quartzarenite blanket (Upper Ordovician-Lower Silurian) of the central and southern Appalachians: American Journal of Science, v. 295, p. 201-243.

Dorsch, J., Bambach, R. K., and Driese, S. G., 1994, Basin rebound hypothesis for the "Tuscarora Unconformity" in southwestern Virginia and its bearing on the nature of the Taconic orogeny: American Journal of Science, v. 294, p. 237-255.

Edmundson, R. S., 1958, Industrial limestones and dolomites in Virginia: James River district west of the Blue Ridge: Virginia Division of Mineral Resources Bulletin, v. 73, 137 p.

Englund, K. J., and Thomas R. E., 1990, Late Paleozoic depositional trends in the central Appalachian basin: United States Geological Survey Bulletin, v. 1839-F, p. 1-19.

Eriksson, K. A., Campbell, I. H., Palin, J. M., Allen, C. M., and Bock, B., 2004, Evidence for multiple recycling in Neoproterozoic through Pennsylvanian sandstones of the central Appalachian basin: Journal of Geology, v. 112, p. 261-276.

Erwin, P. N., Jr., 1981, Stratigraphy, depositional environments and dolomitization of the Maryville and Upper Honaker formations (Cambrian), Tennessee, and Virginia [Master's thesis]: Duke University, 232 p.

Ettensohn, F. R., 1985, 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, v. 201, p. 65-77.

Ettensohn, F. R., 1994, Tectonic control on formation and cyclicity of major Appalachian unconformities and associated stratigraphic sequences, in Dennison, J. M., and Ettensohn, F. R., eds., Tectonic and Eustatic Controls on Sedimentary Cycles: Concepts in Sedimentology and Paleontology, no. 4, p. 217-242.

Ferm, J. C., 1974, Carboniferous environmental models in eastern United States and their significance, in Briggs, G., ed., Carboniferous of the Southeastern United States: Geological Society of America Special Paper, v. 148, p. 79-95.

Ferm, J. C., and Cavaroc, V. V., 1969, A Field Guide to Allegheny Deltaic Deposits in the Upper Ohio Valley: Pittsburgh Geological Society, Pittsburgh, 21 p.

Filer, J. K., 2002, Late Frasnian sedimentation cycles in the Appalachian basin – possible evidence for high frequency eustatic sea-level changes: Sedimentary Geology, v. 154, p. 31-52.

Frakes, L. A., Francis, J. E., and Syktus, J. I., 1992, Climate modes of the Phanerozoic: the history of the Earth's climate over the past 600 million years: Cambridge University Press, 274 p.

Gao, Z. and Eriksson, K. A., 1991, Internal - tide deposits in an Ordovician submarine channel: previously unrecognized facies: Geology, v. 19, p. 734-737.

Gathright, T. M., Jr., 1976, Geology of the Shenandoah National Park, Virginia: Virginia Division of Mineral Resources, Bulletin 86, 93 p.

Gathright, T. M. II, Rader, E. K., Taylor, A. R., and Whitlock, W. W., 2003, Mapped units of the Appalachian plateaus and Valley and Ridge, in Geologic Map of Virginia – Expanded Explanation, Commonwealth of Virginia Department of Mines, Minerals, and Energy, Division of Mineral Resources, Charlottesville, Virginia, p. 1-24.

Gehrels, G.E., Valencia, A. and Pullen, A., 2006, Detrital zircon geochronology by laser ablation multicollector ICPMS at the Arizona LaserChron Center, in Olszewski, T. (ed.) Geochronology: emerging opportunities: Paleontological Society Paper 12, 67-76.

Goldhammer, R. K., Lehmann, P. J., and Dunn, P. A., 1993, The origin of high-frequency platform carbonate cycles and third-order sequences (Lower Ordovician El Paso Gp, West Texas); constraints from outcrop data and stratigraphic modeling: Journal of Sedimentary Petrology, v. 63, no. 3, p. 318-359.

Gradstein, F. M., Ogg, J. G., and Smith, A. G., 2004, A Geologic Time Scale. Cambridge University Press, 589 p

Greb, S. F., and Chesnut, D. R., Jr., 1996, Lower and Lower Middle Pennsylvanian fluvial to estuarine deposition, central Appalachian basin: Effects of eustasy, tectonics, and climate: Geological Society of America Bulletin, v. 108, p. 303-317.

Greb, S. F., Chesnut, D. R., Jr., and Eble, C. F., 2004, Temporal changes in coal-bearing depositional sequences (Lower and Middle Pennsylvanian) of the central Appalachian basin, USA, in Pashin, J. C., and Gastaldo, R. A., eds., Sequence Stratigraphy, Paleoclimate, and Tectonics of Coal-Bearing Strata: American Association of Petroleum Geologists, Studies in Geology, v. 51, p. 89-120.

Grover, G. A., and Read, J. F., 1978, Fenestral and vadose diagenetic fabrics, Middle Ordovician limestones, Virginia: Journal of Sedimentary Petrology, v. 48, p. 453-473.

Grover, G. A., and Read, J. F., 1983, Paleoaquifer and deep burial related cements defined by regional cathodoluminescence patterns, Middle Ordovician carbonates, Virginia: American Association of Petroleum Geologists Bulletin 67, p. 1275-1303.

Harris, L. D., 1964, Facies relations of exposed Rome Formation and Conasauga Group of northeastern Kentucky and Virginia: U.S. Geological Survey Professional Paper 501–B, p. B25-B29.

Hasson, K. O. and Dennison, J. M., 1988, Devonian shale lithostratigraphy, central Appalachians, USA, in McMillan, N .J., Embry, A. F. and Glass D. J., eds., Devonian of the World: Canadian Society of Petroleum Geologists, Proceedings of the Second International Symposium on the Devonian System, Memoir 14, p. 157-178, Calgary, Canada.

Hatcher, R. D., Jr., 1972, Development model for the southern Appalachians: Geological Society of America Bulletin, v. 83, p. 2735 - 2760.

Hatcher, R. D., Jr., 1989, Tectonic synthesis of the U.S. Appalachians, in Hatcher, R. D., Jr.; Thomas, W. A.; and Viele, G. W., eds. The Appalachian-Ouachita Orogen in the United States, Boulder Colorado, Geological Society of America Bulletin, The Geology of North America, F-2, p. 511-535.

Hayes, A. W., 1974, Origin of the Tuscarora Formation (Lower Silurian), southwestern Virginia [Ph.D. dissertation]: Blacksburg, Virginia Polytechnic Institute and State University, 162 p.

Haynes, J. T., and Goggin, K. E., 1993, Field guide to the Ordovician Walker Mountain Sandstone Member: proposed type section and other exposures: Virginia Minerals, v. 39, no.4, p. 25-37.

Haynes, J. T., Melson, W. G., and Goggin, K., 2005, Petrology and significance of quartz arenites and conglomerates in the Ordovician Blount Molasse from Virginia to Alabama, Southern Appalachians: Geological Society of America, Northeastern Section, 40th Annual Meeting, Saratoga Springs, N.Y., Abstracts with Programs, v. 37, no. 1, p. 18.

Head, J.W., 1974, Correlation and paleogeography of upper part of Helderberg Group (Lower Devonian) of central Appalachians: American Association of Petroleum Geologists Bulletin, v. 58, p. 247-259.

Heckel, P., 1985, Recent interpretations of Late Paleozoic cyclothems, in Proceedings of the Third Annual Field Conference, Mid-Continent Section, SEPM, Lawrence, Kansas, p. 1-22.

Helfrich, C. T., 1975, Silurian conodonts from Wills Mountain Anticline, Virginia, West Virginia, Maryland: Geological Society of America Special Paper 161, 82 p.

Hergenroder, J. D., 1966, The Bays Formations (Middle Ordovician) and related rocks of the southern Appalachians [Ph.D. thesis]: Virginia Polytechnic Institute and State University, Blacksburg, 325 p.

Heyman, L., 1970, Petrology of the basal Middle Ordovician Blackford Formation of the type belt, Russell County, Virginia [unpublished Ph.D. Dissertation]: Virginia Polytechnic Institute and State University, Blacksburg, Virginia, 272 p.

Holland, S. M., and Patzkowsky, M. E., 1996, Sequence stratigraphy and long-term paleoceanographic change in the Middle and Upper Ordovician of the eastern United States, in Witzke, B. J., Ludvigson, G. A., and Day, J., eds., Paleozoic Sequence Stratigraphy: Views from the North American Craton: Geological Society of America Special Paper 306, p. 117-129.

Horne, J. C., Ferm, J. C., and Swinchatt, J. P., 1974, Depositional model for the Mississippian – Pennsylvanian boundary in northeastern Kentucky, in Briggs, G., ed., Carboniferous of the Southeastern United States: Geological Society of America Bulletin Special Paper, v. 148, p. 97-114.

House, M. R. and Gradstein, F. M., 2004, The Devonian Period, in Gradstein, F., Ogg, J., and Smith, A., eds., A Geologic Time Scale: Cambridge University Press, p. 202-221.

Huggins, M. J., 1983, Meramecian conodonts and biostratigraphy of the (Upper Mississippian) Greenbrier Limestone (Hurricane Ridge and Greendale synclines), southwestern Virginia and southern West Virginia [unpublished Ph.D.

dissertation]: Blacksburg, Virginia Polytechnic Institute and State University, 305 p.

Jannsens, A., 1973, Stratigraphy of the Cambrian and Lower Ordovician rocks in Ohio: Ohio Geological Survey Bulletin 64, 197 p.

Johnson, M. E., 1996, Stable cratonic sequences and a standard for Silurian eustasy, in Witzke, B. J., Ludvigson, G. A., and Day, J. E., (eds.), Paleozoic Sequence Stratigraphy, Views from the North American Craton. Geological Society of America Special Paper 306, p. 203-211.

Jones, S. E., 1982, Stratigraphy and petrology of the Wildcat Valley Sandstone (Lower Devonian), southwestern Virginia [Master's thesis]: Chapel Hill, University of North Carolina, 127 p.

Karpa, J. B., 1974, The middle Ordovician Fincastle Conglomerate north of Roanoke, Virginia and its implications for Blue Ridge tectonism [M.S. thesis]: Blacksburg, Virginia Polytechnic Institute and State University, 104 p.

Koerschner, W. F. III, and Read, J. F., 1989, Field and modelling studies of Cambrian carbonate cycles, Virginia Appalachians: Journal of Sedimentary Petrology, v. 59, p. 654-687.

Kolata, D. R., Huff, W. D., and Bergstrom, S. M., 1996, Ordovician K-bentonites of eastern North America: Geological Society of America Special Paper 313, 84 p.

Korus, J. T., 2002, The Lower Pennsylvanian New River Formation: A non-marine record of glacioeustasy in a foreland basin [M.S. thesis]: Blacksburg, Virginia Polytechnic Institute and State University, 55 p.

Kreisa, R.D., 1980, The Martinsburg Formation (Middle and Upper Ordovician) and related facies, southwestern Virginia [unpublished Ph.D. Dissertation]: Blacksburg, Virginia, Virginia Polytechnic Institute and State University, 358 p.

Kreisa, R. D., 1981, Storm-generated sedimentary structures in subtidal marine facies with examples from the Middle and Upper Ordovician of southwestern Virginia: Journal of Sedimentary Petrology, v. 51, p. 823-848.

Kreisa, R. D., and Bambach, R. K., 1973, Environments of deposition of the Price Formation (Lower Mississippian) in its type area, southwestern Virginia, in Glover, L., and Ribbe, P. H., eds., The Byron N. Cooper Volume, American Journal of Science, p. 326-342.

Lampiris, N., 1975, Stratigraphy of the clastic Silurian rocks of central western Virginia and adjacent West Virginia [Ph.D. dissertation]: Virginia Polytechnic Institute and State University, Blacksburg, 205 p.

Lash, G. G., 1988, Middle and Late Ordovician shelf activation and foredeep evolution, central Appalachian orogen, in Keith, B. D., ed., The Trenton Group (Upper Ordovician Series) of eastern North America: American Association of Petroleum Geologists Studies in Geology 29, p. 37-53.

Laurence, R. A., and Palmer, A. R., 1963, Age of the Murray shale and Hesse quartzite on Chilhowee Mountain, Blount County, Tennessee: United States Geological Survey Professor Paper 475-C, p. C53-C54.

Lundegard, P. D., Samuels, N. D., and Pryor, W. A., 1985, Upper Devonian turbidite sequence, central and southern Appalachian basin: Contrasts with submarine fan deposits, in Woodrow, D. L., and Sevon, W. D., eds., The Catskill Delta: Geological Society of America Special Paper, v. 201, p. 107-121.

Mack, G. H., James, W. C., and Monger, H. C., 1993, Classification of paleosols: Geological Society of America Bulletin, v. 105, p. 129-136.

Markello, J. R., and Read, J. F., 1981, Carbonate ramp to deeper shale shelf transitions of an Upper Cambrian intrashelf basin, Nolichucky Formation, Virginia, U.S.A: Sedimentology, v. 28, p. 573-597.

Markello, J. R., and Read, J. F., 1982, Upper Cambrian intrashelf basin, Nolichucky Formation, southwest Virginia Appalachians: American Association of Petroleum Geologists Bulletin, v. 6, p. 860-878.

Markello, J. R., Tillman, C. G., and Read, J. F., 1979, Lithofacies and biostratigraphy of Cambrian and Ordovician platform and basin facies carbonates and clastics, southwestern Virginia: Geological Society of America Southeastern Section Guides to Field Trips 1-3, p. 42-86.

Matchen, D. L., and Kammer, T. W., 1994, Sequence stratigraphy of the Lower Mississippian Price and Borden formations in southern West Virginia and eastern Kentucky: Southeastern Geology, v. 34, p. 25-41.

Matthews, R. K., and Frohlich, C., 2002, Maximum flooding surfaces and sequence boundary comparisons between orbital forcing in the Cretaceous and Jurassic (65-190 Ma): GeoArabia, v. 7, p. 503 - 539.

Maynard, J. P., Eriksson, K. A., and Law, R. D., 2006, The Upper Mississippian Bluefield Formation in the central Appalachian basin: A hierarchical sequence stratigraphic record of a greenhouse to icehouse transition: Sedimentary Geology, v. 192, p. 99-122.

McBride, E. F., 1962, Flysch and associated beds of the Martinsburg Formation (Ordovician), central Appalachians: Journal of Sedimentary Petrology, v. 32, p. 39-91.

McClung, W. S., 1983, Marine offshore to alluvial plain transitions within the "Chemung"- Hampshire interval (Upper Devonian) of the southern central Appalachians [MS Thesis]: Blacksburg, Virginia Polytechnic Institute and State University, 257 p.

Meert, J. G. and Torsvik, T. H., 2004, Paleomagnetic constraints on Neoproterozoic 'Snowball Earth' continental reconstructions: Jenkins, G. S., McMenamin, M. A. S., McKay, C. P., and Sohl, L., eds., AGU Monograph, v. 146, p. 5-11.

Meert, J. G., van der Voo, R. and Payne, T., 1994, Paleomagnetism of the Catoctin volcanic province: a new Vendian-Cambrian apparent polar wander path for North America: Journal of Geophysical Research, v. 99, p. 4625-4641.

Meyer, S. C., Textoris, D. A., and Dennison, J. M., 1992, Lithofacies of the Silurian Keefer Sandstone, east-central Appalachian basin, USA: Sedimentary Geology, v. 76, p. 187-206.

Miller, D. J., and Eriksson, K. A., 1997, Late Mississippian prodeltaic rhythmites in the Appalachian basin: a hierarchical record of tidal and climatic periodicities: Journal of Sedimentary Research, v. 67, p. 653-660.

Miller, D. J., and Eriksson, K. A., 1999, Linked sequence development and global climate change: The Upper Mississippian record in the Appalachian Basin: Geology,

v. 27, p. 35-38.
Miller, D. J., and Eriksson, K. A., 2000, Sequence stratigraphy of Upper Mississippian strata in the central Appalachians: A record of glacioeustasy and tectonoeustasy in a foreland basin setting: American Association of Petroleum Geologists Bulletin, v. 84, p. 210-233.
Montañez, I. P., 1989, Regional dolomitization of Early Ordovician, Upper Knox Group, Appalachians [PhD dissertation]: Virginia Polytechnic Institute and State University, Blacksburg, 322 p.
Montañez, I. P., and Read, J. F., 1992, Eustatic control on early dolomitization of cyclic peritidal carbonates: evidence from the Early Ordovician Upper Knox Group, Appalachians: Geological Society of America Bulletin, v. 104, p. 872-886.
Mussman, W., and Read, J. F., 1986, Sedimentology of a passive to convergent margin conformity: Middle Ordovician Knox unconformity, Virginia Appalachians: Geological Society of America Bulletin, v. 97, p. 282-295.
Mussman, W. J., Montañez, I. P., and Read, J. F., 1988., Ordovician Knox unconformity, in James, N. P. and Choquette, P. W., eds., Paleokarst: Springer-Verlag, New York, p. 211-228.
Nolde, J. E., 1994, Devonian to Pennsylvanian stratigraphy and coal beds of the Appalachian plateau province, in Geology and Mineral Resources of the Southwest Virginia Coalfield: Virginia Division of Mineral Resources Publication 131, p. 1-85.
Oder, C. R. L., 1934, Preliminary subdivision of the Knox dolomite in east Tennessee: Journal of Geology, v. 42, p. 469-497.
Osleger, D., and Read, J. F., 1993, Comparative analysis of methods used to define eustatic variations in outcrop: Late Cambrian interbasinal sequence development: American Journal of Science, v. 293, p. 157-216.
Palmer, A. R., 1981, On the correlatibility of Grand Cycle tops, in Taylor, M. E., ed., Second International Symposium on the Cambrian System: U. S. Geological Survey Open-File Report 81-743, p. 156-157.
Palmer, A. R., and James, N. P., 1979, The Hawke Bay event: a circum-Iapetus regression near the Lower to Middle Cambrian boundary, in Wones, D. R., ed., Proceedings, Caledonides in the U.S.A.: Virginia Polytechnic Institute and State University Memoir 2, Blacksburg, p. 15-18.
Park, H., Barbeau, D. L., Jr., Rickenbaker-Krug, D. and Gehrels, G. E., 2010, Application of foreland basin detrital-zircon geochronology to the reconstruction of the Southern and central Appalachian orogen: Journal of Geology, v. 118, p. 23-24.
Patchen, D. G., 1968, Oriskany Sandstone-Huntersville Chert gas production in the eastern half of West Virginia: West Virginia Geological and Economic Survey Circular 9, 30 p.
Patchen, D. G., Avery, K. L., and Erwin, R. B., 1984, Correlation of stratigraphic units in North America—Southern Appalachian region correlation chart: American Association of Petroleum Geologists, 1 sheet.
Patchen, D. G., Avery, K. L., and Erwin, R. B., 1985, Correlation of stratigraphic units in North America—Northern Appalachian region correlation chart: American Association of Petroleum Geologists, 1 sheet.
Pfeil, R. W., and Read, J. F., 1980, Cambrian carbonate platform margin facies, Shady Dolomite, southwestern Virginia, U.S.A: Journal of Sedimentary Petrology, v. 50, p. 91-116.
Pope, M. C., and Read, J. F., 1997, High-resolution surface and subsurface sequence stratigraphy of Late Middle to Late Ordovician (Late Mohawkian-Cincinnatian) foreland basin rocks, Kentucky and Virginia: American Association of Petroleum Geologists Bulletin, v. 81, no. 11, p. 1866-1893.
Pope, M. C., and Read, J. F., 1998, Sequences and meter scale cycles of Ordovician rocks in Kentucky and Virginia: Implications for Ordovician climate and eustatic fluctuations in the central Appalachian Basin, U.S.A.: Palaeoclimatology, Palaeogeography, and Palaeoecology, v. 138, p. 27-42.
Rader, E. K., and Biggs, T.H., 1976, Geology of the Strasburg and Toms Brook quadrangles, Virginia: Virginia Division of Mineral Resources Report of Investigations 45, 104 p.
Rader, E. K., and Henika, W. S., 1978, Ordovician shelf-to-basin transition, Shenandoah Valley, Virginia: Virginia Division of Mineral Resources, Publication no. 7, Contributions to Virginia Geology: III, p. 51-65.
Rader, E. K., and Perry, W. J. Jr., 1976a, Stratigraphy as a key to the arch-related origin of the Little North Mountain structural front, Virginia and West Virginia (Abstract): American Association of Petroleum Geologists Bulletin, v. 60, p. 1623.
Rader, E. K., and Perry, W. J., Jr., 1976b, Reinterpretation of the geology of Brocks Gap, Rockingham County, Virginia: Virginia Division of Mineral Resources, Virginia Minerals, v. 22, p. 37-45.
Rader, E. K., and Read, J. F., 1989, Early Paleozoic continental shelf to basin transition, northern Virginia: I.G.C. Guidebook for Field Trip T-221, 9 p.
Randall, A. H., 1983, Storm-dominated shelf sedimentation marginal to a delta: an example from the Devonian of southwestern Virginia and eastern West Virginia [MS Thesis]: Blacksburg, Virginia Polytechnic Institute and State University, 175 p.
Rankin, D. W., Drake, A. A., Jr. and seven others, 1989, Pre-orogenic terranes, in Hatcher, R. D., Jr., Thomas, W. A., and Viele, G. W., eds. The Appalachian-Ouachita Orogen in the United States, Boulder, Colorado: Geological Society of America, The Geology of North America, F-2, p. 7–100.
Raymond, L. A., Webb, F., and Moore, D., 1979, Submarine fan facies of an Ordovician foreland basin, Abingdon/Lodi area, southwestern Virginia (abs.): Geological Society of America Abstracts with Programs, v. 11, p. 209.
Read, J. F., 1980, Carbonate ramp to basin transitions, and foreland basin evolution, Middle Ordovician, Virginia Appalachians: American Association of Petroleum Geologists Bulletin, v. 64, p. 1575-1612.
Read, J. F., 1982, Geometry, facies and development of Middle Ordovician carbonate buildups, Virginia Appalachians: American Association of Petroleum Geologists Bulletin, v. 66, p. 189-209.
Read, J. F., 1989a, Evolution of Cambro-Ordovician passive margin, U. S. Appalachians. Decade of North American Geology Synthesis: Volume Appalachian-Ouachitas, p. 42-57.

Read, J. F., 1989b, Controls on the evolution of the Cambro-Ordovician passive margin, U.S. Appalachians, in Crevello, P. D., Read, J. F., Wilson, J. L. and Sarg, J. F., eds., Controls on Carbonate Platform and Basin Development: SEPM Special Publication 44, p. 146-165.

Read, J. F., and Goldhammer, R. K., 1988, Use of Fischer plots to define 3rd order sea level curves, Ordovician, Appalachians: Geology, v. 16, p. 895-899.

Read, J. F., and Grover, G. A., 1977, Scalloped and planar erosion surfaces, Middle Ordovician limestones, Virginia: Analogues of Holocene exposed karst or tidal rock platforms: Journal of Sedimentary Petrology, v. 47, p. 956-972.

Reed, J. S., Eriksson, K. A. and Kowalewski, M., 2005, Controls on Carboniferous sandstone diagenesis, Central Appalachian basin: a qualitative and quantitative assessment: Sedimentary Geology, v. 176, p. 225-246.

Rice, C. L., and Schwietering, J. F., 1988, Fluvial deposition in the central Appalachians during the early Pennsylvanian: United States Geological Survey Bulletin, v. 1839-B, p. 1-9.

Rodgers, J., 1971, The Taconic orogeny: Geological Society of America Bulletin, v. 82, p. 1141-1178.

Ross, C. A., and Ross, J. R. P., 1987, Late Paleozoic sea levels and depositional sequences, in Ross, C. A., and Haman, D., eds., Timing and depositional history of eustatic sequences : constraints on seismic stratigraphy: Washington, D.C., Cushman Foundation for Foraminiferal Research, p. 137-149.

Ross, R. J., Adler, F. J. and twenty six others, 1982, The Ordovician system in the United States, correlation chart and explanatory notes: International Union of Geological Sciences, Publication 12, 73 p.

Rowley, D. B., and Kidd, W. S. F., 1981, Stratigraphic relationships and detrital compositions of the medial Ordovician flysch of New England: Implications for the tectonic evolution of the Taconic orogeny: Journal of Geology, v. 89, p. 199-218.

Ruppel, S. C., and Walker, K. R., 1984, Petrology and depositional history of a Middle Ordovician carbonate platform: Chickamauga Group, northeastern Tennessee: Geological Society of America Bulletin, v. 95, p. 568-593.

Ryder, R. T., Crangle, R. D., Fr., Treppe, M. H., Swezey, C.S., Lentz, E. E., Rowan, E. L., and Hope, R. S., 2009, Geologic cross section D-D' through the Appalachian Basin from the Findlay Arch, Sandusky County, Ohio, to the Valley and Ridge Province, Hardy County, West Virginia; Scientific Investigations Map 3067, United States Geological Survey, 52 p.

Samman, N. F., 1975, Sedimentation and stratigraphy of the Rome formation in east Tennessee [Unpublished Ph.D. Dissertation]: University of Tennessee, Knoxville, 337 p.

Sartain, R. R., 1981, Stratigraphy and conodont paleontology of Late Silurian-Early Devonian strata of western Virginia [M.S. Thesis]: Virginia Polytechnic Institute and State University, Blacksburg, 142 p.

Schutter, S. R., 1992, Ordovician hydrocarbon distribution in North America and its relationship to eustatic cycles, in Webby, B. D. and Laurie, J. R., (eds.), Global Perspectives on Ordovician Geology, p. 421-432.

Schwab, F. L., 1970, Origin of the Antietam Formation (Late Precambrian?-Lower Cambrian), central Virginia: Journal of Sedimentary Petrology, v. 40, p. 354-366.

Schwab, F. L., 1971, Harpers Formation, central Virginia: A sedimentary model: Journal of Sedimentary Petrology, v. 41, p. 139-149.

Schwab, F. L., 1972, The Chilhowee Group and late Precambrian, early Paleozoic sedimentary framework in the central and southern Appalachians, in Lessing, P., Hayhurst, R. I., Barlow, J. A., and Woodford, L. D., eds., Appalachian Structures: Origin, Evolution and Possible Potential for Exploration Frontiers, A Seminar: West Virginia University and West Virginia Geologic and Economic Survey, Morgantown, p. 59-101.

Scotese, C. R., 2001, Atlas of Earth History, Volume 1, Paleogeography, PALEOMAP Project, Arlington, Texas, 52 pp.

Seilacher, A., 1968, Origin and diagenesis of the Oriskany Sandstone (Lower Devonian, Appalachians) as reflected in its shell fossils, in Muller, G. and Friedman, G. M., eds., Recent developments in carbonate sedimentology in central Europe: New York, Springer-Verlag, p. 175-185.

Sevon, W. D., 1985, Nonmarine facies of the Middle and Late Devonian Catskill coastal alluvial plain, in Woodrow, D. L., and Sevon, W. D., eds., The Catskill Delta: Geological Society of America Special Paper, v. 201, p. 79-90.

Shanmugam, G. and Lash, G. G., 1982, Analogous tectonic evolution of Ordovician foredeeps, southern and central Appalachians: Geology, v. 10, p. 562-566.

Shergold, J. H., and Cooper, R. A., 2004, The Cambrian Period, in Gradstein, F. M., Ogg, J. G., and Smith, A., eds., A Geologic Time Scale: Cambridge University Press, p. 147-164.

Simpson, E., and Eriksson, K. A., 1989, Rift to incipient passive-margin sedimentation within the Lower Cambrian Unicoi Formation, southern and central Virginia: Geological Society of America Bulletin, v. 101, p. 42-54.

Simpson, E. L., and Eriksson, K. A., 1990, Early Cambrian progradational and transgressive sedimentation patterns in Virginia: an example of the early history of a passive margin: Journal of Sedimentary Petrology, v. 60, p. 84-100.

Simpson, E. L., and Sundberg, F. A., 1987, Early Cambrian age from synrift deposits of the Chilhowee Group of southwestern Virginia: Geology, v. 12, p. 123-126.

Sinha, A. K., Badger, R. L., Hund, E, Hogan, J. P., and Guy, R. E., 1988, Late Precambrian and Paleozoic accretionary history of the Central and Southern Appalachians; characterization through magmatic and deformational events, in Bartholomew, M. J., Hyndman, D. W., Mogk, D. W., and Mason, R. (eds.) Proceedings of the International Conference on Basement Tectonics, v. 8, p.738-739.

Smith, L. B. and Read, J. F., 2000, Rapid onset of Late Paleozoic glaciation: evidence from Upper Mississippian strata of the Midcontinent, United States: Geology, v. 28, p. 279-282.

Smith, L. B., Al-Tawil, A., and Read, J. F., 2001, Sequence stratigraphic setting of Mississippian eolianites, Appalachian and Illinois Basins, in Abegg, R., ed., Modern and Ancient Carbonate Eolianites: Sedimentology, Sequence Stratigraphy and Diagenesis: SEPM Special Publication 71, p. 167-181.

Smosna, R. A., and Patchen, D. G., 1978, Silurian evolution of the central Appalachian Basin: American Association of Petroleum Geologists Bulletin, v. 62, p. 2308-2328.

Smosna, R., and Warshauer, S. M., 1979, A very early Devonian patch reef and its ecological setting: Journal of Paleontology, v. 53, p. 142-152

Smosna, R. A., Patchen, D. G., Warshauer, S. M., and Perry, W. J., 1977, Relationships between depositional environments, Tonoloway Limestone, and distribution of evaporites in the Salina Formation, West Virginia, in Fischer, J. A., ed., Reefs and Evaporites: American Association of Petroleum Geologists Studies in Geology No. 5, p. 125-143.

Southworth, S. and Aleinikoff, J. N., 2007, Time versus tectonic significance; the disconformity at the base of the Chilhowee Group and the Neoproterozoic-Cambrian boundary: Abstracts with Programs - Geological Society of America, Southeastern Section, 56th Annual Meeting, Savannah, Georgia, United States, v. 39, no. 2, p. 9.

Stock, C. W. and Holmes, A. E., 1986, Upper Silurian/Lower Devonian Stromatoporoidea from the Keyser formation at Mustoe, Highland County, west-central Virginia: Journal of Paleontology, v. 60, p. 555-580.

Taylor, J. F., Repetski, J. E., and Orndorff, R. C., 1992, The Stonehenge transgression: a rapid submergence of the central Appalachian platform in the Early Ordovician, in Webby, B. D. and Laurie, J. R., (eds.), Global Perspectives on Ordovician Geology, p. 409-418.

Thomas, W. A., 1977, Evolution of Appalachian-Ouachita salients and recesses from reentrants and promontories in the continental margin: American Journal of Science, v. 277, p. 1233-1278.

Thomas, W. A., 1983, Continental margins, orogenic belts, and intracratonic structures. Geology, v. 11, p. 270-272.

Thomas, W. A., 2006, Tectonic inheritance at a continental margin: GSA Today, v. 16, p. 4-11.

Thomas, W. A., Becker, T. P., Samson, S. D. and Hamilton, M. A., 2004, Detrital zircon evidence of a recycled orogenic foreland provenance for Alleghanian clastic-wedge sandstones: Journal of Geology, v. 112, p. 23-37.

Van Tassell, J., 1994, Evidence for orbitally-driven sedimentary cycles in the Devonian Catskill delta complex, in Dennison, J. M. and Ettensohn, F. R., Tectonic and Eustatic Controls on Sedimentary Cycles, SEPM Concepts in Sedimentology and Paleontology, v. 4, p. 121-131.

Walcott, C. D., 1891, Correlation papers, Cambrian: United States Geological Survey Bulletin 81, 447 p.

Webb, E. J., 1980, Cambrian sedimentation and structural evolution of the Rome trough in Kentucky [Ph.D. thesis]: University of Cincinnati, 98 p.

Weber, L. J., Sarg, J. F., and Wright, F. M., 1995, Sequence stratigraphy and reservoir delineation of the Middle Pennsylvanian (Desmoinesian), Paradox Basin and Aneth Field, Southwestern USA, in Read, J. F., Kerans, C., Weber, L. J., Sarg, J. F., and Wright, F. M., eds., Milankovitch Sea-Level Changes, Cycles, and Reservoirs on Carbonate Platforms in Greenhouse and Ice-House Worlds: SEPM Short Course No. 35, Part 3, p. 1-81.

Wehr, F., 1983, Geology of the Lynchburg Group in the Culpeper and Rockfish River areas, Virginia [Unpublished Ph.D. dissertation]: Virginia Polytechnic Institute and State University, 245 p.

Wehr, F., and Glover, L., III, 1985, Stratigraphy and tectonics of the Virginia-North Carolina Blue Ridge: evolution of a late Proterozoic-early Paleozoic hinge zone: Geological Society of America Bulletin, v. 96, p. 285-295.

Wheeler, H. E., 1963, Post-Sauk and pre-Absaroka Paleozoic stratigraphic patterns in North America: American Association of Petroleum Geologists Bulletin, v. 47, p. 1497-1526.

Whitehead, N. H., III, 1984, Paleogeography and depositional environments of the Lower Mississippian of the east-central United States, in Nassichuk, W. W., ed., Paleogeography and Paleotectonics: Neuvième Congrès International de Stratigraphie et de Géologie due Carbonifère 1979, v. 3, part 2, p. 280-290.

Willoughby, R., 1977, Paleontology and stratigraphy of the Shady Formation near Austinville, Virginia [M.S. Thesis]: Virginia Polytechnic Institute and State University, Blacksburg, Virginia, 189 p.

Witzke, B. J., 1990, Paleoclimatic constraints for Paleozoic paleolatitudes of Laurentia and Euramerica, in McKerrow, W. S., and Scotese, C. R., (eds.), Palaeozoic Palaeogeography and Biogeography: Geological Society of London Memoir 12, p. 57-74.

Wizevich, M. C., 1992, Sedimentology of Pennsylvanian quartzose sandstones of the Lee Formation, central Appalachian basin: fluvial interpretation based on lateral profile analysis: Sedimentary Geology, v. 78, p. 1-47.

Woodward, H. P., 1951, Ordovician System of West Virginia: West Virginia Geological Survey, v. 21, 627 p.

Wortman, G. L., Samson, S. D., and Hibbard, J. P., 2000, Precise U-Pb zircon constraints on the earliest magmatic history of the Carolina terrane: Journal of Geology, v. 108, p. 321-338.

Wynn, T. C. and Read, J. F., 2008, Sequence development on a foreland carbonate ramp, Mississippian Appalachian Basin, West Virginia, U.S.A.: Sedimentary Geology, v. 55, p. 357-394.