STATE-WIDE SEQUENCE FRAMEWORK OF MIXED CARBONATE-SILICICLASTIC RAMP RESERVOIRS: MISSISSIPPIAN BIG LIME, WEST VIRGINIA, USA

by

THOMAS C. WYNN

Dissertation submitted to the Faculty of the Virginia Polytechnic Institute and State University in partial fulfillment of the requirements for the degree of

DOCTOR OF PHILOSOPHY

in

GEOLOGICAL SCIENCES

J. Fred Read (chair)
Richard Bambach
Cahit Coruh
John Dennison
Kenneth Eriksson
Richard Law

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State-Wide Sequence Framework of Mixed Carbonate-Siliciclastic Ramp Reservoirs: Mississippian Big Lime, West Virginia, USA

by

Thomas C. Wynn

J. F. Read, Chairman

Department of Geosciences, Virginia Tech

ABSTRACT

Well-cuttings data and wireline logs in conjunction with limited core and outcrop data are used to generate a regional, three dimensional, high resolution sequence framework for Upper Mississippian (Chesterian), Big Lime carbonates, West Virginia, U.S.A.. The analysis was done using the washed coarse fraction (1-2 mm) of cuttings for each sample interval, which were separated into Dunham rock types, counted to determine relative abundance and the data presented as percent lithology plotted against depth for each well. Digitized wireline logs and the cuttings-percent logs were slipped to take into account drilling lag and lithologic columns produced from the combined data. Sequence stratigraphic cross sections through the basin and into the outcrop belt, with a resolution of 10 feet were then produced. Sequence stratigraphic time slices were generated as isopachs maps of the sequences, and of lowstand-transgressive, and highstand tracts with major facies shown. This data was then used to document the stratigraphic response of the foreland basin to tectonics and, with isotope data from the slope section, evaluates evidence for glacio-eustasy during the transition into ice-house times. The major mappable sequences are fourth-order sequences, a few meters to over 90 meters (300 feet) thick. They consist of updip red beds and eolianites, lagoonal muddy carbonates, ooid grainstone and skeletal grainstone-packstone shoal complexes, deeper ramp and slope wackestone-mudstone, and laminated argillaceous lime mudstone. Maximum flooding surfaces on the ramp slope occur at the base of deeper water facies that overlie lowstand-to transgressive siliciclastic or carbonate complexes, whereas on the ramp, maximum flooding
surfaces cap near-shore shale or lime mudstone beneath widespread grainstones. The highstand systems tracts contain significant grainstone units, interlayered with extensive lagoonal lime mudstones.

In spite of differential subsidence rates across the foreland, fourth-order eustatic sea level changes documented by isotopic signals in basinal facies, controlled regional sequence development. Thrust-load induced differential subsidence of fault-blocks of the foreland basement controlled the rapid basinward thickening of the depositional wedge while subtle structures such as arches at high angles as well as parallel to the margin, affected thicknesses and facies development.
ACKNOWLEDGMENTS

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CHAPTER 1: INTRODUCTION

This dissertation evaluates the use of well-cuttings data and wireline logs in conjunction with limited core and outcrop data, to generate a regional, three dimensional high resolution sequence framework for Upper Mississippian (Chesterian), Big Lime carbonates, West Virginia, U.S.A. Wireline logs or seismic can not by themselves be used to determine lithologies in wells in carbonate dominated sections. However, well-cuttings analyzed using modern carbonate facies concepts, with the data in a GIS framework can provide 1) high resolution sub-seismic sequence stratigraphy, at reservoir scale 2) time slice maps showing facies distribution within systems tracts, above the resolution of the biostratigraphy and 3) high resolution time slice maps that allow differentiation between effects of tectonic subsidence and eustasy.

Chapter 2 is a methods paper that describes how well-cuttings and wireline logs can be used to generate 3-D, high resolution sequence stratigraphies of carbonate ramps in the subsurface even where there is only limited core and outcrop data. Well-cuttings classified according to Dunham (1962), and wireline logs were used in a GIS framework to determine the vertical stacking of lithologies. From this, sequence picks were made along the ramp margin beneath lowstand sandstones and calcareous siltstones. On the ramp, sequence boundaries are overlain by thin transgressive siliciclastics and eolianites. Only a few sequence boundaries are calichified. Cross sections compiled from the well data then allow regional sequences to be mapped.

Chapter 3 shows how well-cuttings data from 190 wells were used to generate the high resolution 3-D framework for the Big Lime reservoirs throughout the state. Regional cross-sections were made from the wells that had been analyzed and bounding surfaces and sequences picked. Using regional markers the sequences were traced throughout the cross-sections. Isopachs and isolith maps generated for sequences show dominant facies of system tracts for each sequence. The GIS time slice maps provide the first view of potential Big Lime reservoir trends and associated facies throughout the whole state at the 4th order sequence scale, documenting the stratigraphic response of the foreland basin to tectonics and, with isotope data from the slope section, evaluates evidence for glacio-eustasy during the transition into ice-house times.
CHAPTER 2: REGIONAL HIGH RESOLUTION SEQUENCE ANALYSIS VIA WELL-CUTTINGS, MISSISSIPPIAN BIG LIME, WEST VIRGINIA, USA.

ABSTRACT

Well-cuttings analysis largely pre-dated modern carbonate facies analysis, sequence stratigraphy, reflection seismic and advanced down-hole logging techniques. These higher tech methods have resulted in well-cuttings being neglected as an important data source for high resolution subsurface analysis. However, binocular analysis of well indurated Paleozoic well-cuttings from relatively shallow wells (0-4000 ft.), can provide detailed vertical facies successions when tied to wireline logs, and can be used to generate high resolution sequence stratigraphic frameworks and time slice maps for the subsurface, at a higher resolution than is available from other methods alone. The analysis was done using the washed coarse fraction (1-2 mm) of the cuttings for each sample interval, which were separated into Dunham rock types, counted to determine relative abundance, and the data as percent lithology plotted against depth for each well. Digitized wireline logs and the cuttings-percent logs were slipped (typically 10 ft or so) to take into account drilling lag and lithologic columns produced from the combined data. Distinctive gamma ray marker horizons were used to tie the sections together. Sequence stratigraphic cross sections through the basin and into the outcrop belt, with a resolution of 10 feet were then produced. Maps of sequence stratigraphic time slices were generated showing isopachs of the sequences, and distribution of lowstand-transgressive, and highstand tracts with major facies shown. This allowed us to track the complex influence of tectonics and eustasy and their effects on the stacking patterns of reservoirs.

INTRODUCTION

This article illustrates how well-cuttings and wireline logs can be used to generate 3-D, high resolution sequence stratigraphies of carbonate ramps in the subsurface even where there is only limited core and outcrop data. We selected the Mississippian Big Lime (Greenbrier Group) throughout West Virginia as a test case (Fig. 1). Examples of well-cuttings studies include Mazzullo and Reid (1989) and Coffey and Read (2002), but these are at lower resolution than the present
Figure 1. Geologic location map of West Virginia study area, showing wells used and distribution of exposed Mississippian Greenbrier carbonate rocks (gray shading) in the Appalachian Basin, eastern U.S.A. Isopach contours (non-palinspastic, in feet) show total thickness of Greenbrier carbonates, which thicken into the Appalachian foredeep to the southeast (modified from Pryor and Sable 1974; MacQuown and Pear 1980; Yeilding and Dennison 1986; Dever 1995; Sable and Dever 1990, Dever et al. 1990). Well (●), core (★), and outcrop (■), locations used in study on West Virginia county base.
study. Well-cuttings integrated with wireline logs provide the means to generate lithologic data at a resolution needed for detailed sequence analysis for regional subsurface studies in carbonate prone areas given the general scarcity of continuous core. Well-cuttings in mature areas such as the Appalachian Basin, where seismic data are limited, may provide the major data set for sequence stratigraphic analysis.

Well-cuttings analysis largely pre-dated sequence stratigraphy, reflection seismic and advanced down-hole logging techniques (Vail and Mitchum 1977; Rider 1996). These “high tech” methods have resulted in well-cuttings becoming of secondary importance as a reliable data source for subsurface analysis. Thus, binocular analysis of Paleozoic well-cuttings from relatively shallow wells (0-4000 ft.) using modern carbonate facies concepts, provides detailed vertical facies successions in the wells when tied to wireline logs. The study demonstrates the power of the cuttings-based, sequence analysis for the subsurface to achieve a detailed picture of the 3-D facies make-up at a higher resolution than is available from other methods alone.

PROBLEMS INVOLVED WITH USING WELL-CUTTINGS

Problems due to Mixing of Interbedded Lithologies

Interbedding can give the appearance of mixing in well cuttings and can be a major problem if not recognized. There are two types of interbedding (Fig. 2) that cause problems when working with well cuttings; thinly interbedded and larger scale interbedding of two or more lithologies within the sample interval. Thinning interbedded lithologies usually occur as pairs of lithologies (i.e. shale interbedded with limestone) in subequal amounts in the sample interval, with individual beds being below the resolution of the logging tool which would then give an average value for the thinly interbedded units. This made it difficult to distinguish the interbedding from mixing of well cuttings. Where one lithology had far greater abundance than the others then it was assigned as the dominant lithology for the interval, but when they were subequal, both were considered to be the dominant lithology for the interval.

Apparent mixing can also result from two or more lithologic units stacked within the sample
Figure 2. Examples of mixing of cuttings due to small scale interbedding and lithologic succession in which rock units are smaller than the sample interval.
interval. Thus for example, in a typical sample interval of ten feet containing three lithologies each approximately 3 feet in thickness, it was difficult to relocate the lithologic units in the section unless one had a distinctive wireline log signal (eg. shale, evaporites). This was a problem where the percentage of each lithology was similar, making it difficult to assign a dominant lithology to the interval. The lithologies were assigned to their correct depth where one or more had a distinctive wireline log signature (gamma ray or bulk density) (Fig. 2), for example shale or shaley limestone (gamma ray), porous limestone or anhydrite (density). When these could not be differentiated in the logs, there was no way to determine the actual succession of lithologic units in the sample interval.

Sample Interval Induced Mixing

The number and location of the sampled intervals of well cuttings can also cause mixing of lithologies within the interval. There is minimum mixing due to sampling spacing, where the sample interval boundaries roughly coincided with the major lithologic unit boundaries (Fig. 3). However, if the sample intervals were larger than the lithologic unit boundaries, then this resulted in mixing of lithologies in the interval. Again, relocation of the cuttings to the actual depth occurrence required a distinctive wireline log signature for one or more of the lithologies, otherwise it could not be done.

Drilling-Induced Mixing

Mixing of well-cuttings and contamination from upper beds during drilling is in most cases due to improper mud viscosity (Hills 1949). Low viscosity drilling fluids do not allow a good “mud cake” to form on the drill hole wall, thus allowing contamination from beds higher up in the drill hole. Deposition of “mud cake” on the wall of the drill hole will limit contamination from upper beds. Improper viscosity drilling mud allows well-cuttings from different sample intervals to mix together, apparently distorting the primary depositional stacking in the well, but with proper drilling mud viscosity, well-cuttings may be held in suspension even when drilling stops (Hills 1949; Rider 1996).
Boundaries of lithologic units almost coincide with sampling interval.

Boundaries of lithologic units occur within sampling interval.

**Figure 3.** Examples of mixing of cuttings due to sampling interval.
SEQUENCE ANALYSIS OF WEST VIRGINIA USING WELL-CUTTINGS

Geologic Setting

The study area is located within the Appalachian foreland (Fig. 1). It overlies the 1 Ga. Grenville basement, and the margin underwent Neo-Proterozoic rifting, ending at about 570 ma to form the passive margin. Renewed Middle Cambrian rifting in-board from the margin formed the Rome Trough, a failed rift. The area underwent fault reactivation during collisional events starting in the Middle Ordovician and ending with continent-continent collision in the Pennsylvanian-Permian (Ettensohn 1994; Shumaker and Wilson 1996). Antecedent faults subdivide the region into a series of rigid blocks, arches and domes which have moved relative to each other (MacQuown and Pear 1983; Boswell and Donaldson 1988; Henika 1994; Shumaker and Wilson 1996; Yang 1998), and some of the faults underwent slight inversion in the Late Paleozoic (Shumaker and Wilson 1996).

The study area consists of little deformed, flat lying strata beneath the Appalachian Plateau. Outcropping Mississippian rocks are restricted to the leading edge of the over thrust belt in the east and to the distal western edge of the basin bordering the Cincinnati Arch (Al-Tawil and Read in press; Al-Tawil et al. in press). Between these outcrop belts, the subsurface Mississippian rocks are overlain by 0 to 4000 feet of Pennsylvanian rocks. The Mississippian interval is penetrated by numerous wells, 190 of which were used for this study, along with two cores and seven outcrop sections (Al-Tawil et al. in press) (Fig. 1).

Regional Stratigraphic Framework and Facies

Regional Mississippian stratigraphy and biostratigraphy of the Mississippian in Virginia and West Virginia are given in Reger (1926), Butts (1940, 1941), Wells (1950), Flowers (1956), de Witt and McGrew (1979), Rice et al. (1979), and Maples and Waters (1987). The rocks in the study area overlie the Price/Borden (Kinderhookian-Mid Osagean) in the northeast (Branson 1912; Butts 1940; Butts 1941; Bartlett 1974; Bjerstedt and Kammer 1988; Pashin and Ettensohn 1995) and the Fort Payne-Salem (Late Osage-Early Meramecian) (Bjerstedt and Kammer 1988; Sable and Dever 1990; Khetani and Read 2002). The Greenbrier Group (Hillsdale to Alderson interval called the Big Lime by drillers) and the Lillydale Shale and Glenray and Reynolds Limestones (drillers Little
Lime) of the Lower Bluefield Formation (basal Mauch Chunk Group) make up the study interval (0 to 3000 feet thick). The study interval is overlain by siliciclastics of the Upper Mississippian and the Pennsylvanian.

Regionally, the Greenbrier units thicken to over 2000 feet to the southeast into the foredeep, where they are dominated by thick units of dark skeletal wackestone-mudstone and basin and/or slope laminated argillaceous carbonate mudstones, and thin quartz sandstone and skeletal and/or ooid grainstone. Al Tawil et al (in press) provided the first detailed sequence stratigraphic framework for the region based on the eastern outcrop belt and limited subsurface data. The Mississippian facies of the eastern Appalachian Basin (Leonard 1968; Carney and Smosna 1989) are schematically shown on an idealized ramp model (Fig. 4) and resemble facies described elsewhere by Ettensohn et al. (1984), Smith and Read (1999, 2001), Al-Tawil et al. (in press) and Al Tawil and Read (in press). Facies consist of updip red beds and quartz peloid eolianites, lagoonal muddy carbonates, ooid grainstone and skeletal grainstone-packstone shoal complexes, deeper ramp and slope dark gray wackestone-mudstone, and dark gray laminated argillaceous lime mudstone. Facies descriptions and their environments are summarized in Table 1.

Data Collection and Analysis

Data were collected from 190 wells with cuttings, two cores (cuttings and cores provided by West Virginia Geological Survey, Morgantown, WV), and eight outcrop sections from previous studies (Wray 1980; Yeilding 1984; Yeilding and Dennison 1986; Al-Tawil 1998; Al-Tawil et al. in press). The coarse fraction (1-2 mm) of the cuttings for each sample interval (typically 10 feet) were washed, acid etched (2.5% HCL) and if dolomitic were stained with Alizarin Red S, and examined under a binocular microscope. For each sample interval, the lithofacies present were classified using Dunham (1962), and counted to determine relative abundance of lithologic types. Percentages were recorded on data sheets, entered into Microsoft®Excel and saved as comma delineated files. For each well the comma delineated files were imported into Rockware’s Log Plot2001 and plotted against depth to form a “percent lithology” log. Wireline logs were digitized initially using Digirule®FastLog and later the NeuraScanner™ with automated Neuralog™ software.
Figure 4. Schematic facies profile for the Mississippian of the Appalachian Basin. Actual facies distributions are more complex, in that skeletal grainstone/packstone and ooid grainstone facies are not only developed on the ramp margin, but also in local areas far back in the ramp interior.
and exported as LAS files. These LAS files were imported into Rockware® Log Plot2001 and plotted alongside the cuttings-percent logs, after slipping the logs typically 10 ft or so for best match to take into account drilling lag. The combination of cuttings-percent logs and gamma ray logs were used to produce lithologic columns with a resolution of 10 feet showing dominant lithology and gamma ray response. Well data were internally calibrated against geophysical logs if available, with gamma ray and bulk density being the most useful. The integration of gamma ray and bulk density logs with the cuttings data helped remove errors due to drilling lag depth, but they also were used to identify facies below detection with cuttings alone. The gamma ray and bulk density logs when combined with well-cuttings data made it possible to identify siliciclastic units only a few feet thick. The five to six distinctive gamma ray marker horizons are associated with several regional siliciclastic units, which are mostly transgressive shales. The well to well correlations were constrained by the correlated gamma ray markers.

Sequences in the wells were recognized on the basis of major landward and basinward shifts in diagnostic facies belts such as eolianites, red beds, siliciclastic sands/calcareous siltstones, and lime grainstones. Sequence boundaries in outcrop are disconformities on shallowing-upward carbonate units. In the wells, the sequence boundaries were placed beneath lowstand or transgressive siliciclastic red bed, sandstone and shale that rest on the underlying highstand carbonate (Fig. 4, 5, and 6). Along the ramp margin and slope, sequence boundaries were placed beneath the shallowest water facies within the dominantly deep-water successions (Fig. 4, 5, and 6). In wireline logs, many sequence boundaries (and some parasequence boundaries) are associated with high gamma ray response due to the presence of terrigeneous mudrocks, shale, or argillaceous dolomites veneering the surfaces (Fig. 4, 5, and 6). Because it was not clear using the subsurface data whether siliciclastic units were late highstand, lowstand or early transgressive, they were arbitrarily referred to as “lowstand to transgressive” units.

Gamma ray logs typically show five to six distinctive marker horizons that were used along with the lithologic columns to produce three dip-oriented and two strike-oriented sequence stratigraphic cross sections (Fig. 1) through the basin and into the outcrop belt reference sections.
<table>
<thead>
<tr>
<th>Lithofacies</th>
<th>Description</th>
<th>Biota</th>
<th>Depositional Environment</th>
</tr>
</thead>
<tbody>
<tr>
<td>Red Beds</td>
<td>Red, maroon and green mottled, mudrocks and siltstones, massive, to laminated, rare current- and wave ripples on siltstone interbeds, rare mudcracks</td>
<td>Root/ burrow traces in paleosols</td>
<td>Subaerial to Marginal Marine</td>
</tr>
<tr>
<td>Gray Shale</td>
<td>Dark gray to olive green, clay and silt, Poorly fissile to massive</td>
<td>Poorly fossiliferous to very fossiliferous. Mollusks, ostracods, some echinoderms brachiopods and bryozoa; biota sparse and restricted in updip shale</td>
<td>Lagoonal</td>
</tr>
<tr>
<td>Quartz sandstone and Calcareous siltstones</td>
<td>White to light gray, Well-sorted fine to medium-grained shaly sandstones and siltstones of quartz, lesser carbonate grains, Cross-bedded to structureless; flaser, lenticular and wavy bedded locally</td>
<td>Rare</td>
<td>Shoreline Clastic Complex and Barrier Siltstones</td>
</tr>
<tr>
<td>Anhydrite</td>
<td>White-glassy, bedded, sometimes sandy with dolomite</td>
<td>None</td>
<td>Sabkha</td>
</tr>
<tr>
<td>Caliche</td>
<td>Yellow to brown, cryptocrystalline and fibrous calcite crusts and fracture fills; patches of caliche-coated peloids and pisolites. Variably silicified.</td>
<td>None</td>
<td>Subaerial</td>
</tr>
<tr>
<td>Quartz Peloidal Grainstone</td>
<td>Light to dark gray. Rounded and abraded peloids and some ooids, abraded skeletal fragments, and subangular very fine-to-fine quartz up to 50%</td>
<td>Abridged, rounded skeletal fragments. No in-situ biota.</td>
<td>Coastal Eolianite, Minor Marine Sand Sheets</td>
</tr>
<tr>
<td>Dolomite</td>
<td>Yellowish-tan. Poorly fossiliferous to unfossiliferous. Fine grained dolomite crystals; may include quartz silt and clay</td>
<td>None to sparse, small mollusks, small crinoids, and ostracods</td>
<td>Tidal Flat</td>
</tr>
<tr>
<td>Fine-Grained Lime Wackestone/ Mudstone</td>
<td>Light gray to creamy white, un fossiliferous to moderately fossiliferous fine wackestone and mudstone and pellet packstone. Skeletal debris fine grained; may contain quartz silt and clay. Locally cherty.</td>
<td>None to sparse, may have mollusks, small crinoid columnals, ostracods, small oncolites, rare corals, and brachiopods.</td>
<td>Low Energy Lagoon</td>
</tr>
<tr>
<td>Peloid and Ooid Grainstone</td>
<td>Light gray to white Well sorted, rounded, medium to coarse grainstone of sand-size ooids, peloids, lesser skeletal fragments and minor intraclasts.</td>
<td>Crinoid, brachiopod, bryozoan and mollusk fragments, forams</td>
<td>High Energy Shoal</td>
</tr>
<tr>
<td>Skeletal Grainstone/ Packstone</td>
<td>Light to medium gray. Variably fragmented echinoderms, brachiopod, bryozoa, mollusks and rare ooids. Mud-free grainstones to grain-rich packstones and minor wackestones, some with argillaceous seams.</td>
<td>Abundant echinoderms, brachiopods and bryozoa; and lesser mollusks.</td>
<td>Ramp Margin and Lagoonal Skeletal Sheets and Shoals</td>
</tr>
<tr>
<td>Argillaceous Skeletal Wackestone</td>
<td>Medium Gray to Dark Gray Echinoderm, brachiopod, bryozoa, and lesser mollusks. Include wackestone and lesser packstone; abundant lime mud; terrigenous clay disseminated or in seams and stringers</td>
<td>Abundant echinoderm, common brachiopods and bryozoa, rare mollusk</td>
<td>Deep Subtidal</td>
</tr>
<tr>
<td>Laminated Shaly Lime Mudstone</td>
<td>Typically dark gray Carbonate and silicilastic clay and silt</td>
<td>None to sparse small skeletal fragments</td>
<td>Ramp slope-Basin</td>
</tr>
</tbody>
</table>
Figure 5. Example of how raw cuttings data are converted to dominant lithology and sequences are picked.
Figure 6. Example of cross-section produced from well-cuttings.
Using the five cross-sections as a starting framework, the sequence picks were extended to nearby wells. For each sequence, the following data was recorded in Microsoft® Excel: county, permit number, latitude and longitude, sequence number, sequence thickness, lowstand-transgressive tract thickness, lowstand-transgressive tract dominant facies, the highstand tract thickness, dominant highstand facies, aggregate grainstone thickness in the sequence (to generate isolith maps), dominant marine grainstone type (skeletal, ooid or peloid), sandstone isolith, caliche (present or absent), and production type (oil, gas, or oil and gas). The sequence data for each well were then imported into ESRI® ArcviewGIS and plotted as point themes. Lowstand-transgressive and highstand dominant facies, isopachs and isolith maps were produced in ESRI® ArcviewGIS for each sequence using the data from the point themes. The isopachs and isolith maps were produced using Golden Software® Surfer8, edited by hand and then imported into ESRI® ArcviewGIS.

These data were compiled into regional cross-sections showing the vertical and lateral distribution of facies (Fig. 6). Sequence boundaries and maximum flooding surfaces were traced from section to section, guided by distinctive log markers and/or biostratigraphy. This generated a high resolution sequence stratigraphic framework (Fig. 6). With the well data in GIS, the succession throughout the region of interest was then time-sliced into sequences and systems tracts (Fig. 7). Isopach maps were constructed for individual sequences, or for a) lowstand to transgressive b) highstand units (Fig. 7). In addition dominant facies maps could then be rapidly made for each system tract to quickly define major potential reservoir trends, and aggregate grainstone maps (isolith maps) (Fig. 7) were generated to show regional location of grainstone facies (reservoirs). The maps and cross-sections (Fig. 6 and 7) illustrate the power of GIS and that sequence analysis of well-cuttings is capable of generating a high resolution sequence stratigraphic framework.

CONCLUSIONS

This study, used as a test case the Mississippian Greenbrier (Big Lime) in the West Virginia Appalachian Basin, illustrates how well-cuttings integrated with wireline logs can be used to generate 3-D, high resolution sequence stratigraphies of carbonates in the subsurface.

Well-cuttings classified according to Dunham (1962), and wireline logs were used in a GIS
Figure 7. Examples of isopach, lowstand-transgressive, highstand and isolith maps produced for each sequence.
framework to determine the vertical stacking of lithologies, and sequence picks were made. Regional cross-sections were made using regional markers and sequences were traced throughout the cross-sections. Isopachs and isolith maps generated for sequences show dominant facies of system tracts for each sequence. The GIS time slice maps provide the first view of potential Big Lime reservoir trends and associated facies throughout West Virginia at the 4th order sequence scale.

Where seismic data are limited, well-cuttings and wireline logs provide the major data set for sequence stratigraphic analysis. In areas where seismic data are available, well-cuttings can provide the necessary lithologic data for intelligent seismic interpretation, where core coverage are insufficient to provide regional coverage. Three-dimensional mapping of the sequence stratigraphic time slices and the resulting isopach maps of 3rd and 4th order sequences can clarify subtle differential subsidence patterns and identify relatively subdued regional structures that are spatially too complex to be evaluated by 2-D cross-sections.
REFERENCES


CHAPTER 3: SEQUENCE DEVELOPMENT ON A FORELAND CARBONATE RAMP, MISSISSIPPIAN APPALACHIAN BASIN, WEST VIRGINIA, USA.

ABSTRACT

This paper evaluates the use of well-cuttings data and wireline logs in conjunction with limited core and outcrop data, to generate a regional, 3 dimensional high resolution sequence framework for Upper Mississippian (Chesterian), Greenbrier Group carbonates, West Virginia, U.S.A. These data are then used to document the stratigraphic response of the Appalachian foreland basin to tectonics and, with isotope data from the slope section, evaluates evidence for glacio-eustasy during the transition into ice-house times. The major mappable sequences are fourth-order sequences, a few meters to over 90 meters (300 feet) thick. They consist of updip red beds and eolianites, lagoonal muddy carbonates, ooid grainstone and skeletal grainstone-packstone shoal complexes, deeper ramp and slope wackestone-mudstone, and laminated argillaceous lime mudstone. The sequences are bounded along the ramp margin by lowstand sandstones and calcareous siltstones. On the ramp, sequence boundaries are overlain by thin transgressive siliciclastics and eolianites. Only a few sequence boundaries are calichified. Maximum flooding surfaces on the ramp slope occur at the base of deeper water facies that overlie lowstand- to transgressive siliciclastic or carbonate complexes, whereas on the ramp, maximum flooding surfaces cap near-shore shale or lime mudstone beneath widespread grainstones. The highstand system tracts contain significant grainstone units, interlayered with extensive lagoonal lime mudstones.

Fourth-order sequences are the dominant mappable unit. They are bundled into weak composite sequences composed of three to four 4th order sequences, which are bounded updip by extensive red beds.

In spite of differential subsidence rates across the foreland basin, third- and fourth-order eustatic sea level changes documented by isotopic signals in basinal facies, controlled regional sequence development. Thrust-load induced differential subsidence of fault-blocks of the foreland basement controlled the rapid basinward thickening of the depositional wedge while subtle structures
such as arches trending at high angles as well as parallel to the margin, affected thicknesses and facies development.

INTRODUCTION

The Mississippian Greenbrier Group, West Virginia, is a 30 to 500 meters (100 to 1600 feet) thick succession of mixed carbonate-siliciclastic sedimentary rock that formed on the Appalachian foreland. It provides an outcrop and subsurface analog to better understand Mississippian three-dimensional (3-D) facies distribution, reservoir stacking patterns, and the stratigraphic signature of global greenhouse to icehouse conditions within an active foreland setting. Prior to Al-Tawil (1998), Al-Tawil et al. (in press) and this study, the Greenbrier Group was lumped into large litho-stratigraphic units that were heterogeneous (Wells 1950; Flowers 1956; Leonard 1968; Wray 1980; Yeilding 1984; Yeilding and Dennison 1986; Carney and Smosna 1989; Kelleher 1990; Kelleher and Smosna 1993). This rock-stratigraphic approach did not allow the facies distribution, stacking patterns, reservoir trends, and effects of tectonics to be elucidated because there was no obvious way to time slice the intervals. High resolution sequence analysis using well-cuttings from 190 wells throughout the state allowed the interval to be subdivided into 3rd and 4th order depositional sequences and systems tracts to be traced throughout the subsurface. This allowed time slice maps to be made that showed the complex regional facies geometries while several regional cross-sections define the stacking of facies and reservoirs.

Well-cuttings analysis largely pre-dated sequence stratigraphy, reflection seismic and advanced down-hole logging techniques. These “high tech” methods have resulted in well-cuttings becoming of secondary importance as a reliable data source for subsurface analysis. However, binocular analysis of well-indurated Paleozoic well-cuttings from relatively shallow wells (0-4000 ft.) within the framework of modern carbonate facies analysis, provided detailed vertical facies successions in the wells when tied to wireline logs. The study demonstrates the power of the cuttings-based high resolution sequence analysis for the subsurface to achieve a detailed picture of the 3-D facies and make up at a higher resolution than is available from other methods.
GEOLOGIC SETTING

Regional Structural Framework

The study area is located within the Appalachian foreland (Fig. 1), which developed on the 1.0 Ga. Grenville orogenic belt, that was the site of a failed rift (Middle Cambrian Rome Trough). The area underwent fault reactivation during late Paleozoic collisional events (Ettensohn 1994; Shumaker and Wilson 1996). These antecedent faults subdivided the region into a series of rigid blocks, arches and domes which have moved relative to each other (MacQuown and Pear 1983; Boswell and Donaldson 1988; Henika 1994; Shumaker and Wilson 1996; Yang 1998). Some of the faults underwent slight inversion in the Late Paleozoic (Shumaker and Wilson 1996).

The Appalachian Basin is an elongate foreland basin bordered on the east by overthrust Precambrian and Cambrian metasedimentary rocks of the Blue Ridge Province, and on the west by the Cincinnati Arch (Colton 1970; De Witt and McGrew 1979)(Fig. 1). In the western part of the basin, the Paleozoic rocks are undeformed to gently folded (Appalachian Plateau), whereas the eastern portion of the basin lies in the folded and thrusted Valley-and-Ridge Province. The bulk of the study area lies beneath the little deformed Appalachian Plateau. Outcropping Mississippian rocks are restricted to the leading edge of the over thrust belt in the east and to the distal western edge of the basin bordering the Cincinnati Arch (Al-Tawil and Read in press; Al-Tawil et al. in press). Between these outcrop belts, the subsurface Mississippian rocks underlie 0 to 1200 meters (4000 feet) of Pennsylvanian sediments. The Mississippian interval is penetrated by numerous wells, 190 of which were used for this study along with two cores and seven outcrop sections (Fig. 2).

The Appalachian foreland basin was tectonically active during deposition of the Mississippian carbonates which thicken into the foredeep in southwestern Virginia and Tennessee (Fig. 1). The depocenter of the Late Mississippian carbonate ramp is located in the overthrust belt in the Greendale Syncline, southwestern Virginia, where Mississippian units thicken to over 600 meters (2000 feet) of basinal, slope, and ramp-margin facies (Cooper 1948; Bartlett and Webb 1971; Al-Tawil 1998). The Cincinnati Arch on the western edge of the basin subsided very little during Chesterian time,
Figure 1. Geologic location map of study area, showing distribution of exposed Mississippian Greenbrier carbonate rocks (gray shading) in the Appalachian Basin, eastern U.S.A. Isopach contours (non-palinspastic, in feet) show total thickness of Greenbrier carbonates, which thicken into the Appalachian foredeep to the southeast (modified from Pryor and Sable, 1974; MacQuown and Pear, 1980; Yeilding and Dennison, 1986; Dever, 1986, 1995; Sable and Dever, 1990, Dever et al., 1990). Orogenic highlands bounded the basin to the southeast.
Figure 2. Well (●), core (★), outcrop (■), and C-O isotope sample (▲) locations used in study on West Virginia county base.
while the Illinois Basin to the west and the Appalachian Basin to the east subsided more rapidly (Donaldson 1974; Pryor and Sable 1974; De Witt and McGrew 1979; Whitehead 1984; Sable and Dever 1990; Dever 1995). The West Virginia Dome was a positive structure active during Early Chesterian time just north of the “38th parallel” lineament zone (Yielding and Dennison 1986). A northwest-southeast trending arch (Burning Springs-Mann anticline) lies at right angles to the regional structure, contiguous with the “Virginia Promontory” a long-lived Paleozoic high (Thomas 1977; Read 1980; Read 1989). Shumaker et al. (1986) showed this structure to be a down-to-the-east, basement normal fault (Burning Springs fault) of Cambrian age. The arch is offset by the southern portion of the 38th parallel lineament zone to form a shallow-rooted fold (Mann anticline) that detaches in the Devonian shale (Perry 1980).

Several faults are associated with regional “hinges” and “hinge zones” (Fig. 3) (Donaldson 1974; Kelleher 1990; Kelleher and Smosna 1993; Shumaker and Wilson 1996; Yang 1998). These include Western-Margin faults and Eastern-Margin faults of the Rome Trough, as well as several zones parallel and to the east of these. The Summersville fault, a normal fault that dips to the southeast (Yang 1998), the Beckley fault, another normal fault, and the Boissevain fault, a normal fault that was re-activated and inverted by Alleghanian thrusting (pers. comm. Bill Henika, 2003). The small east-west trending hinge that occurs along the 122 meter (400 foot) Greenbrier isopach (Fig. 3) seems to be associated with the southern portion of the 38th parallel lineament zone (Fig 4).

**Regional Stratigraphic Framework**

The rocks in the study area overlie the Price/Borden (Kinderhookian-Mid Osagean) in the northeast (Branson 1912; Butts 1940; Butts 1941; Bartlett 1974; Bjerstedt and Kammer 1988; Pashin and Ettensohn 1995) and the Fort Payne-Salem (Late Osage-Early Meramecian) (Bjerstedt and Kammer 1988; Sable and Dever 1990; Khetani and Read 2002). Regional Mississippian stratigraphy and biostratigraphy of the Mississippian in Virginia and West Virginia (Figs. 5, 6) are given in Butts (1922, 1940, 1941), Reger (1926), Wells (1950), Flowers (1956), de Witt and McGrew (1979), and Maples and Waters (1987). The Greenbrier Group (Hillsdale to Alderson interval) and the Lillydale Shale and Glenray Limestone of the Lower Bluefield Formation (Mauch Chunk Group)
Figure 3. Map of total Greenbrier isopach (Big Lime) on a structural base map.
Figure 4. Detailed basement structure map of West Virginia.
Figure 6. Schematic lithostratigraphy and sequence stratigraphy showing the major Mississippian stratigraphic units in the area. The study interval overlies Early Mississippian siliciclastics of the Price/Borden formations, and is overlain by and interfingers with the late Mississippian Mauch Chunk siliciclastics.
make up the study interval (0 to 900 meters thick) (Fig. 2). It unconformably overlies the Price-Maccrady siliciclastic units (Figs. 5, 6). The Late Meramecian Little Valley-Hillsdale Limestone, the basal unit of the study interval is a very cherty, pelletal/peloidal and skeletal packstone/wackestone with the distinctive corals *Syringopora* and *Acrocyathus proliferus* (*Lithostrotionella*) (Wells 1950; DiRenzo 1986). The overlying Chesterian Denmar Formation, an oolitic unit, contains numerous exposure surfaces in outcrop, and has the crinoid *Platycrinites penicillus* and the earliest occurrence of the zonal conodonts *G. bilineatus* and *C. charactus* (Butts 1922; Butts 1940; Butts 1941; Wells 1950; McFarlan and Walker 1956; Dever et al. 1990). It is overlain by the Taggard red beds, a regional marker consisting of red shale, siltstone and minor carbonates.

The oolitic and skeletal limestone-rich Gasper Formation contains the first appearance of the crinoid index fossil, *Talarocrinus* (Butts 1940; Butts 1941; Wells 1950). The interval is subdivided into the Pickaway and Union limestone, Greenville Shale, and Alderson Formation (Reger 1926). This is overlain by Lillydale Shale and the Glenray Limestone (Reger 1926), consisting of ooid and skeletal limestone and containing the zonal conodonts *G. bilineatus* and *Kladognathus* (Rexroad and Clarke 1960). The succession is capped by Mauch Chunk Group red sandstone, siltstone and mudrock (McCulloch 1957; Manspeizer 1958; Thomas 1959; Swires 1972; Brezinski 1989; Maynard 1999).

Regionally, the Greenbrier units thicken to over 600 meter (2000 feet) to the southeast into the foredeep, where they are dominated by thick units of dark skeletal wackestone/mudstone and basin/slope laminated argillaceous carbonate mudstones, and thin quartz sandstone and skeletal/oolid grainstone. In the subsurface, the Hillsdale to Alderson interval is the drillers’ “Big Lime” and the Glenray Limestone is the “Little Lime” (Fig. 6).

**METHODS**

Data were collected from 190 wells with cuttings, two cores (cuttings and cores provided by West Virginia Geological Survey, Morgantown, WV), and eight outcrop sections from previous studies (Wray 1980; Yeilding 1984; Al-Tawil 1998). The coarse fraction (1-2 mm) of the cuttings
for each sample interval (typically 10 feet) was washed, acid etched (2.5% HCL) and if needed stained (Alizarin Red S), and examined under a binocular microscope. For each sample interval, the lithofacies present were separated into lithologic types using the carbonate classification of Dunham (1962), and counted to determine relative abundance. Percentages were recorded on a data sheet, and then entered into Microsoft®Excel and saved as comma delineated files. For each well the comma delineated files were imported into Rockware’s Log Plot2001 and plotted against depth to form a “percent lithology” log. Wireline logs were digitized initially using Digirule®FastLog and later the NeuraScanner™ with automated Neuralog™ software and exported as LAS files. These LAS files gamma ray were imported into Rockware®Log Plot2001 and plotted alongside the cuttings-percent logs, after slipping the logs typically 10 ft or so for best match to take into account drilling lag. The combination of cuttings-percent logs and gamma ray logs were used to produce lithologic columns showing dominant lithology and gamma ray response.

Gamma ray logs typically show 5-6 distinctive marker horizons that were used along with the lithologic columns to produce three dip oriented and two strike oriented sequence stratigraphic cross sections through the basin and into the outcrop belt reference sections, with a resolution of 3 meters (10 feet). Using the five cross-sections (Figs. 7a-d) as a starting framework, the sequence picks were extended to nearby wells. For each sequence, the following data were recorded in Microsoft®Excel: county, permit number, latitude and longitude, sequence number, sequence thickness, lowstand-transgressive tract thickness, lowstand-transgressive tract dominant facies, the highstand tract thickness, dominant highstand facies, aggregate grainstone thickness in the sequence (to generate isolith maps), dominant marine grainstone type (skeletal, ooid or peloid), sandstone isolith, caliche (present or absent), and production type (oil, gas, or oil and none). The sequence data for each well were then imported into ESRI®ArcviewGIS and plotted as point themes. Lowstand-transgressive and highstand dominant facies maps were produced in ESRI®ArcviewGIS for each sequence using the data from the point themes. Isopach and isolith maps were produced using Golden Software®Surfer8 and then edited by hand and imported into ESRI®ArcviewGIS.

Basinal muddy carbonates from the Appalachian Basin ramp slope (Fig. 2) were sampled at
Figure 7a. High resolution sequence stratigraphic cross-section (A-A') from updip to downdip in West Virginia based on subsurface wells analyzed using cuttings and wireline logs. Upper units were hung from the Lillydale Shale marker as in Figure 6, but the lower sequences were hung from the base of the upper Taggard equivalent (quartz peloid eolianites/calcareous siltstones at base of sequence C6).
Figure 7b. High resolution sequence stratigraphic cross-section (B-B’) from updip to downdip in West Virginia based on subsurface wells analyzed using cuttings and wireline logs. Upper units were hung from the Lillydale Shale marker as in Figure 6, but the lower sequences were hung from the base of the upper Taggard equivalent (quartz peloid eolianites/calcareous siltstones at base of sequence C6).
Figure 7c. High resolution sequence stratigraphic cross-section (C-C’) from updip to downdip in West Virginia based on subsurface wells analyzed using cuttings and wireline logs. Upper units were hung from the Lillydale Shale marker as in Figure 6.
Figure 7d. High resolution sequence stratigraphic strike cross-sections (D-D and E-E’) in West Virginia based on subsurface wells analyzed using cuttings and wireline logs. Upper units were hung from the Lillydale Shale marker as in Figure 6.
1.5 meter (5 feet) intervals for C-O isotope analysis, in order to construct C-O isotope curves for the Chesterian Appalachian Basin carbonate ramp. The diagenetic resetting of these basinal muddy carbonates was relatively small (Niemann and Read 1988; Nelson and Read 1990). The Appalachian Basin ramp slope section C-O isotope curves were compared to the curves of the mid-contontinental United States and Europe (Bruckschen et al. 1999; Mii et al. 1999) where the isotope curves were largely based on brachiopod data.

GREENBRIER GROUP FACIES

The facies developed in the eastern Appalachian Basin (Leonard 1968; Carney and Smosna 1989) are schematically shown on an idealized ramp model (Fig. 8) and resemble facies described elsewhere by Ettensohn et al. (1984), Smith and Read (1999, 2001), and Al-Tawil et al. (in press). Facies descriptions and their environments are summarized in Table 1. To facilitate the understanding of the complex facies distribution found in the Greenbrier Group two broadly applicable models are presented, a Lowstand-Transgressive and a Highstand system model (Fig. 9).

**Facies Model for Lowstand-Transgressive System**

The Lowstand-Transgressive system (Fig. 9) is composed of a shoreline complex, lagoonal shales, barrier siltstones, eolianites, red beds and lowstand carbonate facies. The shoreline clastic complex include the most westward facies and is predominantly calcareous siltstones, marine sands and minor red beds. In front of the shoreline clastic complex facies are the lagoonal and eolian facies. The lagoonal facies are green/gray shales and siltstones. The eolian facies are mostly quartz peloid grainstone and are located in the northeast. The most eastward facies seen in the study area are the red bed and barrier siltstone facies. The red bed facies are composed of red shales, silts and sands from the Taggard input system in the northeast. These extend from the northeast southward along the West Virginia and Virginia border. Southwest of the red bed facies are the barrier siltstone facies. The barrier siltstone facies are composed of calcareous siltstones and minor marine sands and are located in front of the lagoonal facies. The barrier siltstone facies are most likely sourced from the Taggard, Greenbrier and Mercer input systems to the northeast and east (Fig. 10). Basinward
Figure 8. Schematic facies profile for the Mississippian of the Appalachian Basin. Actual facies distributions are more complex, in that skeletal grainstone/packstone and ooid grainstone facies are not only developed on the ramp margin, but also in local areas far back in the ramp interior.
<table>
<thead>
<tr>
<th>Lithofacies</th>
<th>Description</th>
<th>Biota</th>
<th>Depositional Environment</th>
</tr>
</thead>
<tbody>
<tr>
<td>Red Beds</td>
<td>Red, maroon and green mottled, mudrocks and siltstones, massive, to laminated, rare current- and wave ripples on siltstone interbeds, rare mudcracks</td>
<td>Root/ burrow traces in paleosols</td>
<td>Subaerial to Marginal Marine</td>
</tr>
<tr>
<td>Gray Shale</td>
<td>Dark gray to olive green, clay and silt, Poorly fissile to massive</td>
<td>Poorly fossiliferous to very fossiliferous. Mollusks, ostracods, some echinoderms brachiopods and bryozoa; biota sparse and restricted in updip shale</td>
<td>Lagoonal</td>
</tr>
<tr>
<td>Quartz sandstone and Calcareous siltstones</td>
<td>White to light gray, Well-sorted fine to medium -grained shaly sandstones and siltstones of quartz, lesser carbonate grains, Cross-beded to structureless; flaser, lenticular and wavy bedded locally</td>
<td>Rare</td>
<td>Shoreline Clastic Complex and Barrier Siltstones</td>
</tr>
<tr>
<td>Anhydrite</td>
<td>white-glassy, bedded, sometimes sandy with dolomite</td>
<td>None</td>
<td>Sabkha</td>
</tr>
<tr>
<td>Caliche</td>
<td>Yellow to brown, cryptocrystalline and fibrous calcite crusts and fracture fills; patches of caliche-coated peloids and pisoliths. Variably silicified.</td>
<td>None</td>
<td>Subaerial</td>
</tr>
<tr>
<td>Quartz Peloidal Grainstone</td>
<td>Light to dark gray. Rounded and abraded peloids and some ooids, abraded skeletal fragments, and subangular very fine-to-fine quartz up to 50%.</td>
<td>Abraded, rounded skeletal fragments. No in-situ biota.</td>
<td>Coastal Eolianite, Minor Marine Sand Sheets</td>
</tr>
<tr>
<td>Dolomite</td>
<td>Yellowish-tan. Poorly fossiliferous to unfossiliferous. Fine grained dolomite crystals; may include quartz silt and clay</td>
<td>None to sparse, small mollusks, small crinoids, and ostracods</td>
<td>Tidal Flat</td>
</tr>
<tr>
<td>Fine-Grained Lime Wackestone/ Mudstone</td>
<td>Light gray to creamy white, un fossiliferous to moderately fossiliferous fine wackestone and mudstone and pellet packstone. Skeletal debris fine grained; may contain quartz silt and clay. Locally cherty.</td>
<td>None to sparse, may have mollusks, small crinoid columnals, ostracods, small oncolites, rare corals, and brachiopods.</td>
<td>Low Energy Lagoon</td>
</tr>
<tr>
<td>Peloid and Ooid Grainstone</td>
<td>Light gray to white Well sorted, rounded, medium to coarse grainstone of sand-size ooids, peloids, lesser skeletal fragments and minor intraclasts.</td>
<td>Crinoid, brachiopod, bryozoan and mollusk fragments, forams</td>
<td>High Energy Shoal</td>
</tr>
<tr>
<td>Skeletal Grainstone/ Packstone</td>
<td>Light to medium gray. Variably fragmented echinoderms, brachiopod, bryozoa, mollusks and rare ooids. Mud-free grainstones to grain-rich packstones and minor wackestones, some with argillaceous seams.</td>
<td>Abundant echinoderms, brachiopods and bryozoa; and lesser mollusks.</td>
<td>Ramp Margin and Lagoonal Skeletal Sheets and Shoals</td>
</tr>
<tr>
<td>Argillaceous Skeletal Wackestone</td>
<td>Medium Gray to Dark Gray Echinoderm, brachiopod, bryozoa, and lesser mollusks. Include wackestone and lesser packstone; abundant lime mud; terrigenous clay disseminated in or in seams and stringers</td>
<td>Abundant echinoderm, common brachiopods and bryozoa, rare mollusk</td>
<td>Deep Subtidal</td>
</tr>
<tr>
<td>Laminated Shaly Lime Mudstone</td>
<td>Typically dark gray. Carbonate and silicilastic clay and silt</td>
<td>None to sparse small skeletal fragments</td>
<td>Ramp slope-Basin</td>
</tr>
</tbody>
</table>
Figure 9. Generalized facies models for the Greenbrier Group.
Figure 10. Greenbrier Group (Sequences M0-C10) sandstone isolith map (values in feet) is based on approximately 200 data points including wells that penetrated the Greenbrier Group and measured sections. Major input locations are marked with arrows.
and southeast of the barrier siltstone facies and outside of the study area are the lowstand-transgressive carbonates. These are made up of ooid, peloid, and skeletal grainstones/packstones and deeper water wackestone/mudstones.

**Facies Model for Highstand System**

The Highstand system (Fig. 9) is composed of lagoonal muddy carbonates, one to three grainstone fairways, deeper water wackestone/mudstone, and basinal carbonates. The lagoonal muddy carbonates form in low energy lagoonal settings on the ramp interior and between grainstone shoals on the ramp. Lack of open marine fossils, and association with shallow marine facies and exposure surfaces in updip areas indicate a relatively restricted, possibly periodically hypersaline setting. In some places in western West Virginia patchy dolomites and anhydrite occur. The northwest-to-southeast trending grainstone fairways are from a few kilometers to 40 km (25 miles) wide. The inner and middle grainstone fairways are composed of ooid and peloid grainstones with minor skeletal grainstone. The ramp margin grainstone fairway is composed of mostly ooid grainstone and skeletal grainstone, which occur on the ramp margin. The deeper water wackestone/mudstone facies are composed of argillaceous packstone/wackestone. They formed below fair-weather wave-base and above storm wave-base on the open ramp in water depths of 20 to 60 m by comparison with the Persian Gulf (Purser and Seibold 1973). This facies locally contains very small bryozoan reefs. The basinal carbonate facies occurs rarely in the study area, but is a major component of sections seen in the Greendale syncline to the southeast, in Virginia. The deep-water ramp slope-to-basin floor facies are composed of shaly lime mudstones which lack shallow water sedimentary structures and biota. Fine carbonate sediment was brought onto the slope perhaps by storms and tidal currents and mixed with land-derived terrigenous silts and clays.

**SEQUENCE STRATIGRAPHY**

Sequences were recognized on the basis of major landward and basinward shifts in diagnostic facies belts such as eolianites, red beds, siliciclastic sands/calcareous siltstones, and lime grainstones. Sequence boundaries in outcrop are disconformities on shallowing-upward carbonate units. In the wells, the sequence boundaries were placed below lowstand or transgressive siliciclastic red beds, sandstones and shale that rest on the underlying highstand carbonates. Along the ramp margin and
slopes, sequence boundaries were placed beneath the shallowest water facies within the dominantly deep-water successions. In wireline logs, many sequence boundaries (and some parasequence boundaries) are associated with high gamma ray response due to presence of terrigeneous mudrocks, shale, or argillaceous dolomites veneering the surfaces. Because it was not clear using the subsurface data whether siliciclastic units were late highstand, lowstand or early transgressive, they have been arbitrarily referred to as lowstand to transgressive units. Isopach maps, dominant facies maps for lowstand-transgressive and highstand systems tracts, and isolith maps of aggregated dominant grainstone are shown in figures 11 to 15.

**Sequences M0**-M4 **(Hillsdale Limestone)**

Sequences M0 to M4 (Hillsdale Limestone) rest on the Maccrady red beds and extend up to the base of sequence C1 (Fig. 6). Sequences M0 to M4 contain the distinctive Late Meramecian corals *Acrocyathus proliferus* (“Lithostrotionella”) (Reger 1926; Butts 1940; Butts 1941; Cooper 1948), and the *T. varians*-Synclydognathus (“Apatagnathus”) and Synclydognathus-Cavusgnathus zonal conodonts (Huggins 1983; Chaplin 1984; Stamm 1997). They correspond to Mamet foraminiferal zones 13 to 15 (Mamet and Skipp 1970; Baxter and Brenckle 1982; Ross and Ross 1987) (Fig. 5).

**Isopachs.**——

Total isopachs for sequences M0 to M4 (Hillsdale Limestone) interval range from zero in the west to over 91 meters (300 feet) in the southeast (Fig. 11a). The interval thins over the Burning-Springs anticline, and the Mann anticline in the southeast.

Sequences M0 to M2 (Fig. 11b) successively step further west but do not extend northeast beyond the Mann anticline, producing a broad depositional thickening with localized, narrow depositional thickenings defined by isopachs greater than 18 meters (60 feet) in sequence M1 and M2. Sequences M3 and M4 extend no farther west than earlier sequences, but they do extend northeastward, beyond the Mann anticline. Sequences M3 to M4 also have localized, narrow depositional thickenings defined by isopachs greater than 18 meters (60 feet) in the southeast. Notably, sequence M4 extends 40 km (25 miles) further west than it does south of the Mann anticline. All sequences thicken toward the southeast, ranging from 0 to 24 meters (80 feet), except M0 and
Figure 11a. Sequences M0 to M4 total isopach map with structural base on West Virginia county base.
Figure 11b. Sequences M0 to M4 isopach maps with basement structures on West Virginia county base.
M2 which only reach 18 meters (60 feet). The sequences on the inner ramp range from 0 to 15 meters (50 feet) thick, whereas the mid-ramp and ramp margin units thicken from 15 to 24 meters (50 to 80 feet) (Fig. 11b).

Lowstand to Transgressive Systems Tracts.—

M0 to M4 sequence boundaries are placed beneath widespread siliciclastic-prone units (sandstone, calcareous siltstone, shale and red beds), that range from a few meters in the west to over 12 meters (40 feet) thick in the east and northeast, and beneath eolianites (quartz peloidal grainstone/packstone) that range from a few meters in the west to 6 meters (20 feet) thick in the east (Fig. 11c). The lowstand-transgressive tracts of sequences M0 to M4 extend progressively further onto the foreland; M0 just extends into the study area, whereas M4 covers almost twenty-five percent of the state (Fig. 11b and c). Sequences M0 to M3 lowstand-transgressive systems are composed of marine calcareous siltstones, patchy marine sands, shales and eolianites. Patchy red beds are present in sequences M0 and M2, but they become more extensive in the southeast in sequence M3 (Fig. 11c). The siliciclastics in sequences M0 to M3 appear to be fed from the Mercer input system in the southeast (Fig. 10 and 11c). Sequence M4 lowstand-transgressive system is made up of extensive barrier siltstones and eolianites which in the southeast are backed by lagoonal shales with patchy marine sands (Fig. 11c). Sequence M4 siliciclastics are fed by the Mercer, and Greenbrier input systems (Fig. 10 and 11c).

Highstand Systems Tracts.—

The highstand tract of sequence M0 barely extends into the study area, whereas the younger units (M1 to M4) extend progressively further onto the foreland, M4 highstand covering almost thirty percent of the study area (Fig 11 b and d). Highstands of sequences M0 to M3 thicken from a few meters in the west to over 12 meters (40 feet) in the southeast, except sequence M3 which continues to thicken to over 24 meters (80 feet) in the southeast.

Sequences M0 to M3 consist of ramp interior facies (muddy lagoonal carbonates) with patchy dolomite; a small area of anhydrite deposition occurs in the northwest in sequence M1 (Fig. 11d). The minor grainstone shoals evident on sequences M1 to M3 isolith maps form localized dip
Figure 11c. Sequences M0 to M4 lowstand-transgressive facies maps showing dominant facies present with lowstand-transgressive tract isopachs on West Virginia county base.
Figure 11d. Sequences M0 to M4 highstand facies maps showing dominant facies present with highstand tract isopachs on West Virginia county base.
oriented units (Fig. 11 d and e). The sequence M4 highstand system thickens from a few meters in the west to over 24 meters (80 feet) in the southeast. The M4 highstand facies are mostly muddy lagoonal carbonates, although the M4 highstand has fairly continuous ramp margin grainstone fairways, with some dip oriented shoals in the southeast. The fairways are composed mostly of ooid and peloid grainstones to the southwest, with skeletal grainstone in the southeast on the ramp margin. The ramp margin fairways of sequences M3 to M4 are interrupted by embayments of muddy lagoonal carbonates just northeast of the Mann anticline and in the southeast (Fig. 11 d and e).

Sequences M0 to M4 (Hillsdale) Grainstone Isoliths.—

The total grainstone isolith map (Fig. 11f) compiled from individual grainstone thicknesses in sequences M0 to M4 (Hillsdale Limestone), shows that aggregate grainstone thicknesses in the M0 to M4 interval range from a few meters to over 24 meters (80 feet) in the southeast. The largest aggregate thicknesses of grainstones are located on the ramp margin (Fig. 11f) and they form dip-oriented shoals flanking the Mann anticline (Fig. 11f). The isolith map shows skeletal grainstone are dominant along the outer ramp fairway, passing updip into ooid grainstones and then into peloid grainstones in the west. The ramp margin fairway of the total aggregated thicknesses of grainstones are interrupted by embayments of muddy lagoonal carbonates just northeast of the Mann anticline and in the southeast (Fig. 11f).

Sequences C1 – C5 (Denmar and Lower Taggard)

The C1 to C5 interval rests on the Hillsdale Limestone. Sequences C1 to C4 are the Denmar Limestone and sequence C5 is the Lower Taggard red beds (Fig. 6). The interval extends up to the base of C6 (Upper Taggard red beds). This interval contains the first appearance of the G. bilineatus-C. charactus zonal conodonts (Huggins 1983; Chaplin 1984; Stamm 1997) and spans Mamet foraminiferal zones 14 to 16 (Mamet and Skipp 1970; Baxter and Brenckle 1982; Ross and Ross 1987), indicating an Early Chesterian age as defined by Maples and Waters (1987). Sequence C1 has the first appearance of the crinoid Platycrinites penicillus which has its last occurrence in sequence C5 (Reger 1926; Butts 1940; Butts 1941; Cooper 1948; Bartlett and Webb 1971) (Fig. 5).
Figure 11e. Sequences M1 to M4 grainstone isolith maps showing dominant facies present on West Virginia county base.
Figure 11f. Sequences M0 to M4 (Hillsdale Limestone) total grainstone isolith maps compiled from individual grainstone thicknesses in sequences M0 to M4 showing dominant grainstone facies present on West Virginia county base.
The Meramecian and Chesterian boundary as defined by Maples and Waters (1987) is at the base of sequence C1 (not at the base of C5, Lower Taggard red beds as was previously done). Thus prior to Maples and Waters (1987), sequences C1 to C4 were considered Late Meramecian rather than Early Chesterian in age, which likely caused some confusion for intercontinental correlations.

Isopachs.—

Total isopachs for C1 to C5 interval range from zero in the west to over 107 meters (350 feet) in the southeast (Fig. 12a). There is a pronounced increase in thickness southeast of the 61 meter (200 feet) isopach increasing to over 107 meters (350 feet) in 40 km (25 miles) (an increase of ~2 m/km). This unit thins over the Burning-Springs and Mann anticlines (Fig. 12a).

Each sequence steps further west and northeast, and thickens toward the southeast (Fig. 12b). Sequences C1 to C5 range from 0 to 36 meters (120 feet), except C3 which is less than 21 meters (70 feet). The inner ramp thicknesses range from 0 to 15 meters (50 feet) and the mid-ramp and ramp margin units thicken from 15 to 36 meters (50 to 120 feet). All the sequences are affected by the Burning-Springs anticline, with C2 and C5 showing the greatest thinning changes over the structure. Sequence C4 thins to six meters (20 feet) or less over a positive feature in the northeast, and sequences C2, C3 and C5 thin to six meters (20 feet) or less over a positive feature in the southwest. In the northwest sequence C3 thickens locally up-dip to over 9 meters (30 feet) (Fig. 12b). Sequences C1 and C4 have narrow areas of depositional thickenings defined by isopachs greater than 24 meters (80 feet) in the southeast, while sequences C2, C3 and C5 have broader areas of depositional thickenings defined by isopachs greater than 24 meters (80 feet) (C3 is defined by isopachs greater than 60 feet) in the southeast.

Lowstand to Transgressive Systems Tracts.—

C1 to C5 sequence boundaries were placed beneath widespread siliciclastic-prone units of sandstone, calcareous siltstone, shale and red beds that are a few meters in the west to over 18 meters (60 feet) in the east and northeast (Fig. 12c). Sequence boundaries were also placed beneath eolianites (quartz peloidal grainstone/packstone) that are a few meters in the west to over 12 meters (40 feet) in the northeast (Fig. 10b). Sequences C1 and C2 lowstands-transgressive units are
Figure 12a. Sequences C1 to C5 total isopach map with structural base on West Virginia county base.
Figure 12b. Sequences C1 to C5 isopach maps with basement structures on West Virginia county base.
Figure 12c. Sequences C1 to C5 lowstand-transgressive facies maps showing dominant facies present with lowstand-transgressive tract isopachs on West Virginia county base.
composed of marine calcareous siltstones, patchy marine sands, red beds, eolianites, and shales. The siliciclastics in sequences C1 and C2 appear to be fed from the Mercer input system in the southeast; while the Taggard input system appears to have fed the siliciclastics of sequence C2 (Fig. 10 and 12b). The lowstand-transgressive systems of sequences C3 to C5 are made up of well developed barrier siltstones backed by lagoonal shales in the southwest and eolianites in the northeast (Fig. 12b). The eolianites are patchy in sequence C3 but become more extensive by sequence C5 time in the northeast while continuing to be patchy in the southeast. A shoreline clastic complex occurs in the west in sequences C3 to C5 (Fig 12b). Patchy red beds are present in sequences C3 and C4, but red beds become more extensive in the northeast in sequence C5 (Fig. 12b). Patchy areas of caliche are found in sequence C3 in the southeast and become more extensive in sequences C4 and C5. Sequence C3 to C5 siliciclastics are fed by the Taggard, Mercer, and Greenbrier input systems and thicken to over 12 meters (40 feet) in the northeast and southeast (Fig. 10 and 12b).

**Highstand Systems Tracts.**—

Highstand tracts of sequences C1 and C2 thicken from a few meters in the west; the sequence C1 highstand is over 18 meters (60 feet) in the southeast, but sequence C2 thickens to over 37 meters (120 feet) in the southeast. Sequences C1 to C2 are mostly composed of muddy lagoonal carbonates with discontinuous ramp margin grainstone fairways with incipient development of a second grainstone fairway shoreward of the ramp margin fairway (Fig 12d). Muddy lagoonal carbonates also occur on the C1 and C2 ramp margins where grainstones are absent. Not all the grainstone shoals parallel the margin; some form dip-oriented shoals as in sequences C1 and C2 in the southeast (fig. 12d and e). The fairways are mostly ooid and peloid grainstone, with patchy skeletal grainstone on the ramp margin in the southeast. The ramp margin fairways are interrupted by embayments of muddy lagoonal carbonates. In sequences C1 and C2, dolomite has a patchy distribution in the west, and C2 has a few local dolomites in the southeast (Fig. 12d).

Sequences C3 to C5 highstand systems thicken from a few meters in the west, to over 30 meters (100 feet) in the southeast, except for sequence C3 which is only 12 meters (40 feet) thick. The westernmost facies are patchy dolomites and anhydrites in C3 and C4 highstands; anhydrites
Figure 12d. Sequences C1 to C5 highstand facies maps showing dominant facies present with highstand tract isopachs on West Virginia county base.
Figure 12e. Sequences C1 to C5 grainstone isolith maps showing dominant facies present on West Virginia county base.
are absent from sequence C5. The dominant highstand facies in sequences C3 to C5 are mostly composed of muddy lagoonal carbonates (Fig. 12d). The ramp margin grainstone fairways are discontinuous with incipient development of two more grainstone fairways shoreward of the ramp margin fairway. Sequences C4 and C5 have dip oriented shoals in the southeast. The fairways are composed mostly of ooid and peloid grainstones, with patchy skeletal grainstone in the southeast on the ramp margin, except sequence C5 which has a fairly continuous skeletal ramp margin. Muddy lagoonal carbonates occupy the ramp margin where grainstones are absent (Fig. 12d). Sequence C5 has a deeper water wackestone in the southeast, seaward of the ramp margin skeletal facies. It also has a 16 km (10 miles) wide zone during the highstand of non-deposition extending from the northwest to the northeast (Fig. 12d).

**Sequences C1 to C5 (Denmar) Grainstone Isoliths.**—

A total grainstone isolith map (Fig. 12f) compiled from individual grainstone thicknesses in C1 to C5 (Denmar), shows total that grainstone isoliths range from a few meters to over 30 meters (100 feet) in the southeast. The thickest grainstones are located on the ramp margin, with the exception of an area in the northeast with aggregate thicknesses over 12 meters (40 feet) (Fig. 12f). The grainstones locally extend northward, halfway across West Virginia, but they also occur locally in the northeast. The grainstone shoals flanking the Burning-Springs and Mann anticlines are dip-oriented shoals. Skeletal grainstone facies are dominant on the outer ramp margin passing updip into ooid grainstones and then into peloid grainstones in the west.

**Sequences C6 to C10 (Gasper)**

Sequences C6 to C10 make up the Gasper Limestone, which rests on sequence C5 (Lower Taggard red beds). Sequence C6 includes the Upper Taggard red beds and the Lower Pickaway Limestone. Sequence C7 is the Upper Pickaway. Sequences C8 and C9 make-up the Lower and Upper Union Limestone, and sequence C10 comprises the Alderson Limestone. Sequences C6 to C8 are in the *G. bilineatus-C. charactus* conodont zone (Huggins 1983; Chaplin 1984; Stamm 1997) while sequences C9 to C10 are in the *G. bilineatus-C. altus* conodont zone (Collinson et al. 1971; Huggins 1983; Chaplin 1984; Stamm 1997), and spans Mamet foraminiferal zones 16 to 18.
Figure 12f. Sequences C1 to C5 total grainstone isolith maps compiled from individual grainstone thicknesses in sequences C1 to C5 showing dominant grainstone facies present on West Virginia county base.
(Mamet and Skipp 1970; Baxter and Brenckle 1982; Ross and Ross 1987). Sequence C6 has the first occurrence of the crinoid *Talarocrinus* (Reger 1926; Butts 1940; Butts 1941; Cooper 1948; Bartlett and Webb 1971) (Fig. 5).

Isopachs.

Total isopachs for the interval made up of sequences C6 to C10 range over much of the ramp from less than 15 meters (50 feet) in the northwest, thickening to 61 meters (200 feet) to the southeast (Fig. 13a). The interval thickens rapidly southeast of the 61 meter (200 feet) isopach, reaching over 168 meters (550 feet) in 40 km (25 miles) (an increase of 4m/km). Individual sequences range from a few meters to 12 meters (40 feet) thick on the inner ramp, thickening southeast onto the slope from 12 to over 30 meters (40 to over 100 feet) (Fig. 13b). Sequences C7 and C8 show a rapid basinward thickening with a relatively steep gradient of 2 m/km. Sequence C7, C9 and C10 have narrow areas of depositional thickenings defined by isopachs greater than 24 meters (80 feet) in the southeast, while sequences C6 and C8 have broader areas of depositional thickenings defined by isopachs greater than 24 meters (80 feet) in the southeast (Fig. 13b).

The Burning-Springs anticline affects most sequences except C7 and C8. Most sequences thin over the Mann anticline, except sequences C7 and C9, which instead thicken over this structure (Fig. 13b). Sequences C6 to C10 show major thinning in a belt parallel to but 40 km (25 miles) northeast the Burning-Springs anticline (Fig 13b). These thinned areas are extensive in sequence C6 but become patchy in sequence C8 but are again extensive in sequences C9 and C10. There is also an area of thinning that is located on top of the Burning-Springs anticline in sequences C9 and C10 (Fig 13b). Positive areas also occur southwest of the Burning-Springs anticline in sequences C6 to C8. They increase in extent from sequence C6 to 8, decrease and disappear in C9, only to reappear in C10 in the far southwest (Fig. 13b).

Lowstand to Transgressive Systems Tracts.

Sequences C6 to C8 have extensive lowstand to transgressive marine calcareous siltstones and lagoonal shales (Fig 13c). The lagoonal shales are more extensive in sequence C8, with the marine calcareous siltstones becoming patchy except in the northeast and southeast of the state.
**Figure 13a.** Sequences C6 to C10 total isopach map with structural base on West Virginia county base.
Figure 13b. Sequences C6 to C10 isopach maps with basement structures on West Virginia county base.
Figure 13c. Sequences C6 to C10 lowstand-transgressive facies maps showing dominant facies present with lowstand-transgressive tract isopachs on West Virginia county base.
Eolianites are extensive in sequence C6 but become progressively patchier until they occupy only small areas in sequence C8. Sequence C6 has extensive red beds in the northeast but these backstep toward the east in sequence C7 and are confined to small areas by sequence C8 (Fig 13c). A well developed shoreline clastic complex exists in sequences C6 and C7 but is not evident in sequence C8, where the sands are very patchy. Sequences C6 to C8 siliciclastics are fed by the Taggard, Greenbrier, Pre-Sharon/Perry, and Pre-Middlesboro input systems (Fig 10 and 13c). Caliche is not evident in the interval containing the sequence boundary in sequence C6, but caliche is more common in sequences C7 and C8 (Fig 13c).

Sequences C9 and C10 are dominated by extensive marine calcareous siltstone interspersed with elongate belts of shale, except in C9 where the shale is patchier. Red beds and sands in sequences C9 and C10 also are patchy, with small areas of eolianite. Caliches decrease from C9 to C10 (Fig 13c). The Taggard and Mercer input systems may feed the siliciclastics (Fig 10 and 13c).

**Highstand Systems Tracts.**

The highstand tract of sequence C6 is composed of muddy lagoonal carbonates, with patchy dolomites in the west (Fig 13d). This sequence lacks a grainstone-dominated barrier. The C6 grainstone isolith map (Fig. 13e) indicates that the grainstone shoals are discontinuous, dip-oriented units. There also is an area of non-deposition or erosion in the northeast (Fig 13d).

Highstand tracts of sequences C7 to C10 are composed of muddy lagoonal carbonates and minor dolomites (C7) with patchy grainstone shoals on the shallow ramp (Fig 13d). Sequences C8 to C10 show a progressive increase in small skeletal grainstone shoals on the shallow ramp. Sequences C7 to C10 have a well developed ramp margin grainstone that is extended updip via lobe–like, dip-oriented shoals that merge with the middle grainstone fairway (Fig. 9, 13d and e). Sequences C7 to C9 have broad embayments in front of the ramp margin grainstones filled with deeper water wackestone. Sequence C8 also contains a small area of basinal carbonates extending in from the southeast (Fig 13d). By sequence C10, these embayments had retreated to the southeast almost from the study area.
Figure 13d. Sequences C6 to C10 highstand facies maps showing dominant facies present with highstand tract isopachs on West Virginia county base.
Figure 13e. Sequences C6 to C10 grainstone isolith maps showing dominant facies present on West Virginia county base.
Sequences C6 to C10 Grainstone Isoliths.—

The total grainstone isolith map for the combined C6 to C10 interval (Fig. 13f), shows total grainstone aggregate thicknesses from a few meters, increasing to over 79 meters (260 feet) to the southeast. The thickest grainstones are along the ooid grainstone-dominated ramp margin, where they are from 12 meters (40 feet) to over 79 meters (260 feet) in aggregate thickness (Fig 13f). The ramp margin fairway and middle ramp fairway had merged into a single broad ramp margin fairway that is 96 km (60 miles) across at its widest. On the inner ramp, the aggregate grainstone thicknesses suggest a patchy distribution of shoals ranging from a few meters to 12 meters (40 feet) thick (Fig 13f).

Sequences C11 to C12

Sequences C11 (Glenray Limestone) to C12 (Webster Springs Sandstone and Reynolds Limestone) rest on the Alderson Limestone (C10). Sequence C11 contains the first occurrence of the *G. bilineatus-K. mehli* conodont zone (Rexroad and Clarke 1960; Collinson et al. 1971), and the Mamet foraminiferal zones 16 to 18 (Mamet and Skipp 1970; Baxter and Brenckle 1982; Ross and Ross 1987) (Fig. 5).

C11 Isopachs Sequence.—

Sequence C11 ranges from zero to 12 meters (40 feet) thick in the northeast and zero to 24 meters (80 feet) thick in the southwest (Fig. 14a). Sequence C11 thins over the Burning-Springs anticline, while west of the Mann anticline sequence C11 has a broader depositional thick area in the southeast, defined by isopachs greater than 24 meters (80 feet) (Fig. 14a). West of the 12 meter (40 feet) isopach there are positive features, defined by the less than 6 meters (20 feet) isopach, that parallel the margin. In the northwest there is an elongate positive feature, defined by the less than 6 meters (20 feet) isopach, extending from the west 80 km (50 miles) toward the east (Fig 14a).

Lowstand to Transgressive System Tract C11.—

Sequence C11 lowstand-transgressive tract has extensive red beds fed by the Taggard input system in the northeast that pass into a broad marine shale belt further to the south and west. Four wells in the northeast show an elongate sand bodies that also may be fed by the Taggard input.
Figure 13f. Sequences C6 to C10 total grainstone isolith maps compiled from individual grainstone thicknesses in sequences C6 to C10 showing dominant grainstone facies present on West Virginia county base.
Figure 14a. Sequence C11 isopach map with basement structures on West Virginia county base.
system (Fig. 10 and 14b). In the southeast, local sands mark inputs that fed the barrier siltstone system in the far southeast. These siliciclastics were fed by the Greenbrier input system (Fig 10 and 14b). Farther toward the west there is a shoreline clastic complex that feeds a southwest trending marine calcareous siltstone system backed by shales (Fig 14b). Sequence C11 has minor patchy eolianites that are localized in the east and locally developed caliches (Fig 14b).

Highstand System Tract Sequence C11.—

In the southern half of West Virginia, the sequence C11 highstand tract is dominated by thick fossiliferous shales and thin interbedded limestone, with a lobe of shale extending from the center of the state to the northeast (Fig 14c). The northern half of sequence C11 is dominated by muddy carbonates and patchy skeletal shoals with minor ooid-peloid shoals (Fig 14c).

Sequence C11 Grainstone Isolith.—

The sequence C11 grainstone isolith map (Fig 14d) shows that grainstone facies are present in the sequence, although in the southeast these are a minor component (less than thirty percent) of the entire shale dominated succession. Where developed, the ramp margin carbonates are mainly skeletal grainstone backed by ooid grainstone. Sequence C11 has a wide ramp margin fairway, which appears to have formed by backstepping of the ramp margin, which then merges with the middle ramp fairway.

Lowstand to Transgressive System Tract C12.—

Sequence C12 rests on C11 and caps the study interval. Only the C12 lowstand-transgressive tract is shown (Fig. 15), because the sections are clastic dominated and the carbonates present in this sequence are below the resolution of the cuttings analysis. Sequence C12 has an extensive, lowstand red bed system in the east flanked by sands in the south and shales to the west; these are fed by the Taggard input system in the northeast (Fig. 10 and 15). In the far southwest, a barrier siltstone complex is backed by lagoonal shales. This is bordered to the west by a well developed shoreline clastic complex, fed by the Pre-Sharon/Perry, and Pre-Middlesboro input systems (Fig. 10 and 15).
Figure 14b. Sequence C11 lowstand-transgressive facies map showing dominant facies present with lowstand-transgressive tract isopachs on West Virginia county base.
Figure 14c. Sequence C11 highstand facies map showing dominant facies present with highstand tract isopachs on West Virginia county base.
Figure 14d. Sequence C11 grainstone isolith map showing dominant facies present on West Virginia county base.
Figure 15. Sequence C12 lowstand-transgressive facies map showing dominant facies present with lowstand-transgressive tract isopachs on West Virginia county base.
DISCUSSION

_Tectonics_

Ettensohn (1994) has suggested that the region was tectonically quiescent during deposition of the study interval, and in a state of elevational equilibrium between the filled foreland basin and the beveled Devonian orogenic highlands. Using the model of Quinlan and Beaumont (1984) and Beaumont et al. (1988), Ettensohn (1994) provided a flexural tectonic model for the Mississippian rocks of the Appalachian Basin. Widespread deepening followed by progradation of Early Mississippian (Kinderhookian to Early Meramecian) siliciclastics were suggested to have formed during loading, then load-relaxation and eastward bulge migration following the last Acadian orogenic event which extended into the Early Mississippian. It was proposed that the foreland basin filled with terrigenous sediment from the eroding orogenic highlands, leading to a state of near-elevational equilibrium, at which time the widespread Mississippian shallow water carbonates were deposited.

The Ettensohn (1994) model fails to account for the many-fold thickening of the sequences and the deep water facies, which would have required much more differential subsidence in the southeast. Seismites in the Late Meramecian (Greb and Dever 2002) and in sequence C7 near Lewisburg, West Virginia (A. Al-Tawil and J. Read, pers. commun., 2002), and the “stair-step” thickening of units on dip-oriented cross-sections of these Late Meramecian to Chesterian carbonates are indicate penecontemporaneous tectonism and faulting. Given the widely quoted 1 to 10 m.y. time constants of lithospheric relaxation (Quinlan and Beaumont 1984), this early Chesterian subsidence occurred too long (some 10 to 15 m.y.) after the inferred earliest Mississippian loading event to be related to relaxation (Al-Tawil and Read in press).

Slingerland and Beaumont (1989), Yang (1998), and Al-Tawil et al. (in press) suggest that the rapid eastward thickening of the study interval across the basin hinge, and the presence of deep water facies in the proximal foreland basin, mark reinitiation of compression and thrust-loading of the foreland basin (Slingerland and Beaumont 1989; Yang 1998; Al-Tawil and Read in press). This caused increased differential subsidence of the foreland following the relatively stable condition in
the Late Osagean-Early Meramecian, development of southeast paleoslopes and caused movement of many of the foreland fault blocks and structures (MacQuown and Pear 1983; Shumaker et al. 1986; Yielding and Dennison 1986; Dever et al. 1990; Shumaker and Wilson 1996; Yang 1998; Al-Tawil et al. in press). The highly faulted foreland slab appears to have been very important in influencing how it responded to loading and unloading (cf. Dorobek 1995). With such a highly fractured basement, the foreland would not have reacted as a simple flexing beam. Instead, it likely responded to flexing by differential movement along basement faults; such a complexly fractured foreland would also have inhibited migration of any peripheral bulge during loading and unloading, for which there appears to be little firm stratigraphic evidence (Shumaker and Wilson 1996). The Late Devonian and Early Mississippian highlands east and northeast of the study area were the source for the Taggard, Greenbrier and Mercer siliciclastic input systems.

Flexural downwarping to the southeast, and to a lesser extent, movement along the Burning-Springs and Mann anticlines, controlled the paleo-slope direction. This in turn controlled the paleo-drainage from the northwest (Pre-Middlesboro and Pre-Sharon/Perry input systems). Uplift along the Burning-Springs and Mann anticlines divided the Pre-Middlesboro and Pre-Sharon/Perry input systems into two depositional sub-basins.

Load-induced differential subsidence (25 cm/k.y.) of the proximal foreland caused downwarping in the southeast, while a much lower subsidence rate (1 to 3 cm/k.y.) occurred on the distal foreland. This differential subsidence caused rapid thickening across the ramp margin hinge (Beckley Fault zone), as well as at various places on the distal foreland. It thus was an important control on the accommodation created during deposition of each sequence. Uplift in the western portion of the study area along the eastern and western margin faults of the Rome Trough and Summersville fault zone may have been related to load-induced flexure of the foreland. The load induced subsidence in the proximal foreland appears to have been associated at least locally with downwarping along the Beckley fault zone, thus producing a relatively positive feature updip on which the high energy grainstone shoals were localized. Movement along the Mann anticline and un-mapped structures in the southeast may have produced the narrow areas of depositional thickening.
oriented at high angles to the margin, and have localized muddy embayments extending updip into the grainstone fairways of the ramp margin.

**Climate and Eustasy**

Semi-arid climate existed from the late Meramecian (sequences M0 to M4) to the early Chesterian (sequences C1 to C8) but the climate became progressively more humid during deposition of sequence C9 onward (Al-Tawil et al. in press). These climatic changes are borne out by paleogeographic reconstructions (De Witt and McGrew 1979; Scotese and McKerrow 1990) which show the North American plate gradually rotating counterclockwise and moving northward from the Pangean arid sub-equatorial belt for the Meramecian to Chesterian (latitudes of 25 degrees or less), with the study area reaching close to the equator by Pennsylvanian time. The semi-arid climate of sequences M0 to C8 could also be in part due to a rain shadow on the western side of the Mississippian highlands, but from sequence C8 onward the influence of this rain shadow appears to have decreased, possibly due to cooling (cf. Johnson and Beaumont 1995).

Smith and Read (1999, 2000, and 2001), Al-Tawil and Read (in press) and Al-Tawil et al. (in press) showed that several of the major sea level events in the Illinois and Appalachian Basins are evident on the 3rd order sea-level curve of Ross and Ross (1987), suggesting a eustatic cause at least for some of the 3rd order sequences. With regard to the 4th order sequences, Smith and Read (1999 and 2001), Al-Tawil and Read (in press), and Al-Tawil et. Al. (in press) showed that the 4th order sequences can be correlated for the most part between the Appalachian and Illinois basins, suggesting eustatic origin. Both the eastern United States and British (Wright and Vanstone 2001; Barnett et al. 2002) sections show an increase in magnitude of sea-level change from the late Meramecian to the Chesterian, which would imply the increase was due to a global cause, namely increased ice-buildup in Gondwana (Smith and Read 2000; Wright and Vanstone 2001; Barnett et al. 2002; Al-Tawil and Read in press).

The oxygen stable isotope data from the mid-continental United States and Europe for the Meramecian-Chesterian show several major climate events, marked by increased δ¹⁸O values interpreted as cooling or ice build up events, whereas decreased δ¹⁸O values marked warming and/or ice melting events (Bruckschen et al. 1999; Mii et al. 1999). Such a mid-Meramecian cooling
event (Bruckschen et al. 1999; Mii et al. 1999) likely marks the sea level fall at the base of sequence M0. This was followed by warming event in the early Chesterian (Bruckschen et al. 1999; Mii et al. 1999), that promoted transgression and deposition of the Meramecian sequences M0 to M4 and Chesterian sequences C1 to 4. This was followed by a major global cooling event and a significant positive oxygen isotope excursion (Bruckschen et al. 1999; Mii et al. 1999) that likely corresponds to the widespread red bed deposition of sequences C5 and 6 (Taggard red beds); this appears to corresponds to the major sea level fall at the Asbian/Brigantian boundary in Britain (Bruckschen et al. 1999; Mii et al. 1999). Subsequent warming and a negative oxygen isotope excursion probably caused the long-term transgression and deposition of sequences C7 through 10. The subsequent cooling trend (Bruckschen et al. 1999; Mii et al. 1999) and falling sea levels were associated with progradation of Mauch Chunk siliciclastic-prone sequences (C11 and younger units).

The deeper water lime muds in the Appalachians appear to preserve a global isotope record and thus can be used to evaluate eustasy. The downdip deeper water lime mudstones were not significantly reset isotopically by meteoric aquifer systems, which did not extend into the more rapidly subsiding proximal foreland (Niemann and Read 1988; Nelson and Read 1990). Their relatively pristine isotope signature is supported by the fact that the overlapping carbon and oxygen compositions of the downdip muds with mid-continent brachiopod data (Mii et al. 1999) and have the heaviest values of any calcite phase on the Appalachian ramp (Niemann and Read 1988; Nelson and Read 1990). Preservation of the original oxygen isotope values in the Appalachian deep water lime probably was due to early cementation under the influence of marine pore fluids, during early compaction and pressure-solution, and their consequent low porosities, low water content and low permeabilities protected them from significant burial fluids that strongly modified the updip carbonates. This low porosity and permeability of the downdip mudstones with their low water-rock ratios probably allowed their isotope compositions to be rock buffered for carbon isotopes, while their low water content and/or low permeabilities limited resetting of the oxygen isotope values in the calcites. The Saltville thrust sheet containing the deeper water units also appears to have been relatively high in the thrust stack and may have not been buried as deeply as other thrust sheets (Harris et al. 1978).
The Appalachian Basin isotope curve (Fig. 16) for sequences C2 and C3 shows relative heavy δ^{18}O values near the inferred sequence boundaries (cooling or ice buildup) that become up to 2 ‰ lighter as the platform floods, and then they become heavier toward the top of the sequence; superimposed on this overall trend are three higher frequency excursions possibly at the parasequence scale, evident in outcrops along the margin (Al-Tawil unpublished data). The δ^{18}O values for sequence C4 show a weak short lived positive excursion near the sequence boundary that becomes more negative with flooding (warming or ice sheet decay), and then becomes up to 2 ‰ heavier toward the top of the sequence (cooling or ice buildup); superimposed on this are five higher frequency excursions. Carbon isotopes covary with oxygen, suggesting increased productivity during colder phases (Holser 1997). No data is available for sequence C5, due to poor outcrop exposure.

The isotope data of Mii et al. (1999) and Bruckschen et al. (1999) suggest a cooling event at the Asbian-Brigantian boundary that likely corresponds to deposition of the red beds of sequences C5 and 6 (Taggard red beds) (Fig. 5). Subsequent warming evident on the Mii et al. (1999) and Bruckschen et al. (1999) isotope curves as a negative δ^{18}O excursion probably caused the long-term transgression and deposition of sequences C7 through 10 and that of the Brigantian of western Europe (Bruckschen et al. 1999; Mii et al. 1999). For this interval, the C-O isotope curves for the Appalachian deep water section show increased δ^{18}O values at the C5 sequence boundary, but no data are available for the sequence C5 and lower C6 intervals (Fig. 16). The Appalachian isotope curve (Fig. 16) for sequences C7 to C10 shows relative heavy δ^{18}O values (cooling or ice buildup) near the inferred sequence boundaries. There is almost a 2 ‰ positive excursion at the C7 lowstand; two data points at the C8 sequence boundary suggest over 1 ‰ positive excursion here and C9 and C10 sequence boundaries are also 1 to 1.5 ‰ positive excursions (Fig. 16). The oxygen isotope signal becomes lighter with flooding, with negative excursions of over 2 ‰ in C7, 1.5 ‰ in C8, 1 ‰ in C9 and C10 (Fig. 16). They then become heavier toward the top of the sequences. The isotope signals show several superimposed higher frequency excursions suggesting parasequence scale excursions.

The δ^{18}O and δ^{13}C of the Appalachian deep water section (Fig. 2) do not show the same
Figure 16. $\delta^{18}$O and $\delta^{13}$C data from the basinal Greendale section. Sequence boundary placement is based on the stratigraphy of Al-Tawil et al. (in press).
strong 3rd order trends common to those of the mid-continental United States and western Europe, although this may be because samples for sequences C5 to C6 and the upper portion of C10 to C11 were not able to be sampled at the section measured. The dominant signal in the Appalachian deep water section stable isotope data are 4th order cycles, that also dominate the shallow platform stratigraphy (Fig. 16). Six out of nine of the sequences show a positive excursion of 1 to 2 per mil in $\delta^{18}O$ near 4th order sequence boundaries. A portion of this signal could reflect ice volume changes, although the presumed sea level changes of less than 50 m in the late Meramecian-early Chesterian would only account for 0.5 ‰, whereas the larger late Chesterian events might have generated 1 ‰ change (Smith and Read 2000). In addition, cooling of shallow platform waters (from which the lime muds were derived) by 5°C or so could have generated 1 to 1.5 ‰ oxygen isotope shifts, that along with the ice volume effect could account for the magnitude of the excursions (Fig. 16).

The $\delta^{18}O$ signal is not just due to temperature change in the water column due to shallowing, because the signatures become heavier with shallowing, which would be opposite to that expected with the move from deeper cooler to shallower warmer waters. The oxygen isotope data suggest relatively rapid warming and flooding of the platform following lowstand. These flooding events appear to be at a high frequency at the scale of parasequences seen along the ramp margin. Rapid deglaciation and warming also characterized the Pleistocene climate cycles, with cooling, glaciation and sea level fall being much slower events (Hays et al. 1976). This trend of rapid flooding following the lowstand is also seen in Pleistocene sediments of the Persian Gulf, where shallow water lowstand deposits are overlain by deep basinal marls (Uchupi et al. 1999).

The Appalachian Basin carbon isotope data co-vary with the oxygen isotope data. If the isotope signatures were related to meteoric diagenesis then the carbon and oxygen isotopes should be lighter beneath sequence boundaries, whereas they are heavy. The approximately 1 per mil increase in $\delta^{13}C$ near sequence boundaries for sequences C2 to C4 and C7 to C10 could be due to increased productivity during glacial stages due to increased ocean circulation. With global warming, and subsequent flooding of the platform, and decreased thermal gradients, the decreased oceanic circulation reduced productivity, resulting in a negative carbon isotope excursion. Part of the carbon
isotope signal also may relate to input of heavy carbon from widespread limestone dissolution during lowstands and carbonate production during highstands (Mii et al. 1999). The 4th order carbon excursions are opposite to the long-term positive carbon isotope excursion common in the geological record, which are due to long term sequestering of isotopically light organic carbon (Holser 1997). However, the positive $\delta^{13}C$ excursions toward the sequence boundaries could be due to effects of shallowing, which would bring the sea floor into more positive $\delta^{13}C$ waters.

The sequence stratigraphic and isotopic evidence points to eustasy as the dominant factor in sequence development in the Appalachian Basin, even though the basin was also tectonically active. Sea level rise augmented tectonically induced subsidence by sediment- and water-loading, thus increasing accommodation space during transgression and highstand. Increased accommodation allowed for the preservation of more complete sequences in which two or more parasequences (evident in outcrop) were developed. Sea level falls were amplified by removal of the water load from the platform, thus causing slight rebound of the platform, exaggerating the effects of sea level fall, and promoting progradation and large scale basinward shifts in facies, and even exposure. It is unlikely that erosion was an important factor in changing the sediment load, given the sparse evidence of erosional incision. The lowstand of each sequence returned to a similar area on the ramp slope. This was because the low gradient (5 cm/km) of the ramp top promoted large scale basinward shifts in shoreline position. However, once the lowstand shoreline migrated onto the ramp slope (gradients of 20 cm/km) then it required much larger sea level fall to shift the shoreline further downslope.

Sequence Interpretations

The interval M0 to M4 (Hillsdale Limestone) has a duration between 4 to 12 m.y. (Sando 1985; Ross and Ross 1987; Claoue-Long et al. 1995; Roberts et al. 1995), which suggest a range of 1.6 m.y. ± 0.8 m.y. for each sequence. Thus, they are 3rd order sequences (Weber et al. 1995).

Sequences M0 to M4 correlate with the St. Louis interval of the Mississippi Valley (Collinson et al. 1971; Huggins 1983). Based on biostratigraphic evidence (Fig. 5), the interval M0 to M4 correlates with the Lower Asbian Lower Urswick Limestone of northwest England (Horbury 1989; Horbury and Adams 1989), Lower Asbian limestones of North Wales (Somerville 1979) and Lower
Asbian Cecilstown Member of the Ballyclough Limestone Formation of County Cork, Ireland (Gallagher and Somerville 1997). Based on the dominant 3rd order sequences, lack of evidence of disconformable 4th order sequences in the Hillsdale Limestone, and dominance of lagoonal, shallow water facies in the interval M0 to M3 with the increase of grainstone shoals in sequence M4, the interval M0 to M4 likely formed during a time of transition from low amplitude (greenhouse) to moderate amplitude sea-level conditions. The underlying Fort Payne-Salem-Warsaw interval of eastern Kentucky was interpreted as greenhouse by Khetani and Read (2002) as was the Holkerian of the British Isles (Wright and Vanstone 2001). Wright and Vanstone (2001) and Barnett et al. (2002) also believe that the Early Asbian was a time of transition from low amplitude (greenhouse) to moderate amplitude sea-level conditions.

The climate was semiarid during the deposition of sequences M0 to M4 (indicated by red beds, coastal eolianites, and anhydrite). Eolianites are more widespread in sequence M4, indicating a more arid climate than in previous sequences.

The area covered by the interval M0 to M4 (Hillsdale Limestone) is far less extensive in West Virginia than in Kentucky immediately to the southwest, where the St. Louis limestone extends further to the northeast (Sable and Dever 1990; Al-Tawil 1998; Al-Tawil and Read in press), suggesting that tectonics may have controlled the extent of sequences M0 to M4 in West Virginia. Movement adjacent to the Mann anticline, possibly along an un-mapped fault to the southeast at the edge of the study area, and regional flexure along the Beckley fault zone produced the broad, elongate depositional thickening that filled from the southeast in sequences M1 and M2 (Fig. 11b). The localized, narrow depositional thickens in sequences M1 and M2 may have been produced by movement along the Mann anticline to the northeast and possibly un-mapped faults to the southwest, producing localized areas of increased subsidence (Fig. 11b). Regional flexure along the Beckley fault zone produced a more elongate basin in sequences M3 and M4, but the localized, narrow depositional thick in sequences M3 and M4 are likely related to movement along the Mann anticline to the northeast and possibly un-mapped faults to the southwest, producing localized areas of increased subsidence (Fig. 11b). The westward extension of the basin in sequence M4
could be due to movement of the Burning-Springs anticline or an un-mapped structure to the northeast of the Burning-Springs anticline (Fig. 11b). Grainstone shoals were localized on the high formed by the Mann anticline creating dip-oriented shoals evident on the grainstone isolith map of sequence M4 (Fig. 11a, d and e). These shoals were produced because of the Mann anticline’s almost perpendicular orientation to the ramp margin. The presence of embayments filled with muddy lagoonal carbonates that interrupt the ramp margin shoals may be related to movement along the Mann anticline producing areas on its flanks that are bathymetrically deeper than surrounding areas (Fig. 11a, d, e). Tectonic deformation in northeast Kentucky at this time is indicated by seismites in the St. Louis Limestone (Greb and Dever 2002).

The sparse radiometric data, and the difficulties in intercontinental correlation between Mississippian brachiopod, conodont, and foraminiferal zones, makes estimation of the duration of the C1 to C5 interval difficult. Al-Tawil et al. (in press) and Smith and Read (1999, 2001) divided the C1 to C11 interval which has a duration of 2 to 7 m.y. (Sando 1985; Ross and Ross 1987; Claoue-Long et al. 1995; Roberts et al. 1995), into two 3rd order sequences (C1 to C5, and C6 to C10). These two 3rd order sequences thus each had an average duration of 1 to 3.5 m.y. Given this duration for the 3rd order sequences, then the individual 4th order sequences had likely durations of 200 to 700 k.y.

The C1 to C5 interval correlates with the Ste. Genevieve Limestone of the Illinois Basin and eastern Kentucky (Collinson et al. 1971; Collinson et al. 1979; Smith and Read 1999; Al-Tawil and Read in press; Al-Tawil et al. in press). Biostratigraphic evidence (Fig. 5) suggests that interval C1 to C5 correlates with the Late Asbian Upper Urswick Limestone of northwest England (Horbury 1989; Horbury and Adams 1989), the Loggershead Limestone of North Wales (Somerville and Strank 1984), and Dromdowney Member of the Ballyclough Formation of County Cork, Ireland (Gallagher and Somerville 1997). These Late Asbian units have widespread grainstone units, that probably formed under moderate-amplitude eustasy, marking the onset of glaciation in western Europe and eastern North America (Horbury 1989; Smith and Read 2000; Wright and Vanstone 2001; Barnett et al. 2002; Al-Tawil et al. in press).
The semiarid climate during the deposition of sequences C1 to C2 continued from the Meramecian (sequences M0 to M4). Sequences C3 to C4 during lowstand and transgression may have been associated with a more arid climate, with widespread eolianites which continued into the highstand with deposition of anhydrite updp. Conditions returned to a more seasonal semiarid climate in sequence C5, with extensive deposition of red beds.

Sequences C1 to C2 show a similar magnitude of relative sea-level change with each lowstand-transgressive and highstand system tract having a similar aerial extent. In contrast, although the lowstand-transgressive tracts of sequences C3 to C5 occupy similar areas on the ramp, each highstand systems tract steps progressively farther back on the ramp, indicating an increase in magnitude of relative sea-level rise through time.

Movement on the Burning-Springs anticline may have caused thinning of sequences C2 and C5 over the structure. The thinning over the positive areas in the southwest seen in sequences C2, C3, and C5 may be related to uplift on the eastern margin faults of the Rome Trough. The positive area in the northeastern part of sequence C4 may be related to movement of fault blocks in the vicinity of the West Virginia Dome. The narrow depositional thickenings of sequences C1, C3, and C4 in the southeast may be related to movement of the Mann anticline and other un-mapped faults in the southeast, producing localized areas of increased subsidence. The broader depositional thickenings of C2 and C5 also may be related to movement of the Burning-Springs and Mann anticlines. The depositional thickenings in the northeast of sequence C3 may be related to movement of the Burning-Springs anticline, the western margin faults and interior faults of the Rome trough.

The formation of a more landward grainstone fairway (middle grainstone fairway of Figure 9) may indicate uplift on the eastern margin faults of the Rome Trough and Summersville Fault, thus producing a positive feature on which the grainstone shoal could localize, but could also just be due to increased flooding relative to underlying units. Uplift on these structures may have been related to load-induced flexure of the foreland, that increased subsidence southeast of the 61 meter (200 feet) isopach (Fig. 12a) by downwarping along the Beckley fault zone.

Grainstone shoals were localized on highs formed by the Burning-Springs and Mann
anticlines, creating dip-oriented shoals in sequences C1, C2, and C5 (Fig. 12d and e). This was influenced by the almost perpendicular orientation of the Burning-Springs and Mann anticlines to the ramp margin. The presence of muddy carbonate embayments that interrupt the ramp margin shoals in sequences C1 to C4 may be related to movement along the Mann anticline producing areas on its flanks that are bathymetrically deeper than surrounding areas (Fig. 12d, e, f). The embayment of muddy lagoonal carbonates in sequence C5 in the southeast may be due to the exposure of the area in the previous lowstand and transgression thus producing a lagoon, but it could also be related to clastic input stopping the formation of ooid shoals and the conditions of the area was too restricted for the formation of skeletal banks.

The C6 to C10 interval correlates with the Paoli/Renault, Beaver Bend, Reelsville, Beech Creek, and Haney of the Illinois Basin and eastern Kentucky (Collinson et al. 1971; Collinson et al. 1979; Smith and Read 1999; Al-Tawil and Read in press; Al-Tawil et al. in press). Biostratigraphic evidence (Fig. 5) suggests that interval C6 to C10 correlates with the Brigantian Gleaston Formation of northwest England (Horbury 1989; Horbury and Adams 1989), the Cefn Mawr Limestone of North Wales (Somerville and Strank 1984), and the Liscarroll Limestone Formation of County Cork, Ireland (Gallagher and Somerville 1997).

The seasonal semiarid climate during the deposition of sequence C5 continued into sequence C6 with the deposition of eolianites and the extensive red beds of the upper Taggard. Caliches are extensive in sequences C7 and C8, suggesting a more seasonal semi-arid climate. The increase of siliciclastics, in sequences C9 to C10, could be due to the more humid climate that may have caused the decrease of caliche. The siliciclastics could have formed a veneer on disconformities thus protecting them from caliche formation during the late highstand-lowstand, and then the siliciclastics would have been reworked and/or eroded during the following transgression (Al-Tawil and Read in press). The change from semi-arid to more humid climate along with rising sea-level would have converted the previously ooid-dominated ramp to a skeletal-dominated system.

The progressive backstepping in the highstand systems tracts could be attributed to the long term rise in sea-level, but it fails to explain why the lowstand-transgressive tracts returned to
a similar location on the ramp. It probably requires that the magnitude of sea-level rise increased with each sequence (C6 to C10), with each highstand systems tract progressively stepping farther back on the ramp while lowstand-transgressive tracts returned to similar areas on the ramp margin (cf. Smith and Read 2001).

Uplift along an un-mapped structure parallel to but 40 km (25 miles) northeast the Burning-Springs anticline in sequences C6, C7, C9 and C10 was most likely responsible for the extensive areas of thinning, defined by isopachs less than 6 meters (20 feet). The thinning over the positive areas in the southwest seen in sequences C6 to C8, and C10 may be related to uplift on the western and eastern margin faults of the Rome Trough accompanied by movement along other un-mapped faults near the Burning-Springs anticline (Fig. 13b). Movement on the Burning-Springs anticline may have caused thinning of sequences C9 and C10 over the structure. The narrow depositional thickenings of sequences C7, C9, and C10 in the southeast may be related to movement of the Mann anticline and other un-mapped faults in the southeast, producing localized areas of increased subsidence.

The formation of a third grainstone fairway (inner grainstone fairway of Figure 9) may indicate uplift on the western margin faults of the Rome Trough and Summersville Fault, thus producing a positive feature on which the grainstone shoal could localize, but could also be due to increased flooding relative to underlying units. Uplift on these structures may have been related to load-induced flexure of the foreland, that increased subsidence southeast of the 76 meter (250 feet) isopach (Fig. 13a) by downwarping along the Beckley fault zone.

Grainstone shoals were localized on highs formed by the Burning-Springs and Mann anticlines, creating dip-oriented shoals in sequences C6, C7, and C10 (Fig. 13d and e). This was influenced by the almost perpendicular orientation of the Burning-Springs and Mann anticlines to the ramp margin. The presence of muddy carbonate embayments that interrupt the ramp margin shoals in sequences C6 to C10 may be related to movement along the Mann anticline producing areas on its flanks that are bathymetrically deeper than surrounding areas (Fig. 13d, e, f). The embayment of muddy lagoonal carbonates in sequence C6 in the southeast may be due to the
exposure of the area in the previous lowstand and transgression, thus producing a lagoon, but it could also be related to siliciclastic input stopping the formation of ooid shoals and the conditions of the area was to restricted for the formation of skeletal banks.

The C11 interval correlates with the Glen Dean Limestone of the Illinois Basin and eastern Kentucky whereas C12 is equivalent to the Vienna Limestone, again suggesting a eustatic cause (Rexroad and Clarke 1960; Collinson et al. 1971; Collinson et al. 1979; Smith and Read 1999; Al-Tawil and Read in press; Al-Tawil et al. in press). However, no isotope data are available for C11 and C12 for the Appalachian Basin.

Climate through this interval continued to become more humid, with marine shales and sands and/or silts becoming the dominant facies in the lowstand-transgressive tracts in sequences C11 and C12 as a result of increased influx of fine terrigenous sediments (Fig. 14b and 15). The marine shales in the highstand tract of C11 are due to this humid climate that converted the skeletal-dominated ramp margin of sequence C10 into a fine grained siliciclastic-dominated margin in sequence C11.

The broad depositional thickening of sequence C11 in the southeast, defined by isopachs greater than 24 meters (80 feet) in figure 14a, may be related to movement of the Mann anticline and un-mapped faults, and increased subsidence southeast of the 24 meter (80 feet) isopach (Fig. 14a) due to downwarping along the Beckley fault zone. The thinning over the positive areas west of the 12 meter (40 feet) isopach that parallel the margin in the southwest in sequence C11 may be related to uplift on the eastern margin faults of the Rome Trough. The elongate positive feature in the northwest could be related to an un-mapped structure that extends from the west eastward.

*Comparison of Mississippian Ooid Shoals to the Holocene*

Mississippian ooid shoals show similar width scales (10 to 20 km) to Holocene ooid shoals of the Bahamas (8 to 16 km) (Purdy 1963; Harris 1979). However in Holocene marine belt sands of Joulters Cay, there is only a narrow zone (1 to 3 km) of active ooid formation (Harris 1979), the remainder of the width of the shoals being inactive, stabilized ooid packstone, formed by micritization.
of ooids, mixing with other grains, and infiltration of mud (Harris 1984). In contrast, on oolitic tidal bars, active oolite formation occurs in a much broader zone (24 km wide) (Handford 1988). These tidal bars are a more comparable setting given evidence for the tidally influenced Mississippian setting and tidal bar geometries in the subsurface (Kelleher and Smosna 1993). Thus the Mississippian oolites may be related to tidal bar origin, rather than progradation. Mississippian ooid shoals show similar length scales (20 to 161 km) to Holocene ooid shoals of the Bahamas (20 to 112 km) (Harris 1994). Thicknesses of Mississippian shoals and Holocene Bahamian shoals of Joulters Cays are also very similar, ranging from less than 3 meters to over 6 meters (10 feet to over 20 feet) thick (Major et al. 1996) and are comparable with an origin related to 4th order glacio-eustatic cycles.

The active Bahamian shoals are covered by less than a meter to 10 meters (33 feet) of water (Harris 1984), which likely was similar to conditions on the Mississippian ramp. The Bahamian platform only had these water depths during the last 10m (33 feet) of flooding of the 100m (328 feet) magnitude 100k.y. cycle (ice-house conditions). Given the lower magnitude for the Early Chesterian the Mississippian ramp would have flooded to 10m (33 feet) water depths during the upper one third of sea level rise although just how much time was involved is difficult to ascertain. It is possible that some of the shoals localized on subtle depositional topography formed above grainstone units within the previous cycle, or they could have localized over subtle tectonic highs. Documenting these causes was not possible due to the resolution of the data set.

*Importance of Well-cuttings in Generating Regional 3-D Framework*

This study illustrates how well-cuttings and wireline logs can be used to generate 3-D, high resolution sequence stratigraphies of carbonate ramps in the subsurface even where there is only limited core and outcrop data. Well-cuttings integrated with wireline logs provide the means to generate lithologic data at a resolution needed for detailed sequence analysis for regional subsurface studies in carbonate prone areas given the general scarcity of continuous core. Well-cuttings in areas like the Appalachian Basin, where seismic data is limited, may provide the major data set for sequence stratigraphic analysis.
Three-dimensional mapping of the sequence stratigraphic time slices and the resulting isopach maps of 3rd and 4th order sequences have clarified the subtle differential subsidence patterns in this foreland basin setting that are spatially too complex to evaluate by 2-D cross-sections. It also has shown how the subtle regional thickness variations may relate to relatively subdued regional structures. The approach has allowed construction of isopach and facies maps of systems tracts and isolith maps that show aggregate thickness and trends of potential reservoir facies and provide insight into the influence of tectonics and eustasy on the facies distribution.

CONCLUSIONS

This study demonstrates that well-cuttings integrated with wireline logs and limited core and outcrop data can be used to generate 3-D high resolution sequence stratigraphy of a carbonate prone mixed carbonate-siliciclastic ramp succession in the subsurface. It used as a test case, the Mississippian Meramecian to Chesterian Greenbrier succession in the West Virginia Appalachian Basin, which formed on a tropical ramp on a differentially subsiding foreland during initiation of Southern Hemisphere glaciation of Gondwana.

Well-cuttings were used to determine the detailed vertical stacking of lithologic units in the wells, which were then subjected to sequence stratigraphic analysis. Regional cross-sections were made using regional markers to map the sequences. The cross-sections then were used to make sequence picks in nearby wells.

The Greenbrier succession is made up of fourth-order depositional sequences of glacio-eustatic origin, which is supported by regional interbasinal correlation (Al-Tawil 1998; Smith and Read 1999; Smith and Read 2001; Al-Tawil and Read in press; Al-Tawil et al. in press) and the stable isotope record of the Appalachian deep water section. Fourth-order sequences are up to tens of meters (30 feet) to a hundred meters (300 feet) thick. Lowstand-transgressive tracts are composed of a siliciclastic shoreline complex, in lagoonal shale and eolian sands, bordered by red beds seaward and barrier siliciclastic sourced from highlands, along the ramp margin. Local transgressive banks developed along the ramp margin, but on the shallow ramp, transgressive carbonates are relatively
rare. Highstand ooid- and skeletal grainstone formed along the ramp margin, and as isolated shoals along regional trends of the ramp interior, along with extensive lagoonal lime wackestone-mudstone. On the deeper ramp and ramp slope skeletal wackestone graded seaward into laminated, argillaceous lime mudstone.

Differential subsidence of the foredeep, probably marked early onset of Late Paleozoic thrust-loading. This generated a basinward thickening wedge of sediments 0 to 900 meters (3000 feet) thick. A fault influenced hinge localized the ramp margin grainstone trend, while other faults localized less extensive interior grainstone shoals. Positive structures perpendicular to the ramp trend caused thinning of sequences, and produced localized embayments in the ramp margin grainstone belt as well as dip-oriented grainstone trends.
REFERENCES


HARRIS, P.M., 1979, Facies anatomy and diagenesis of a Bahamian ooid shoal: Sedimenta, no.7, University of Miami, 163 p.

HARRIS, P.M., 1984, Cores from a Modern Carbonate Sand Body; The Joulters Ooid Shoal, Great Bahama Bank, in Harris, P.M., ed., Carbonate Sands - A Core Workshop: SEPM Core Workshop No. 5: Tulsa, Oklahoma, Society of Economic Paleontologists and Mineralogists, p. 429-462.


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STAMM, R.G., 1997, Late Mississippian conodont biostratigraphy of the Appalachian Basin; preliminary correlations to the Eastern Interior Basin and eustatic curves: Geological Society of America, South-Central Section 31st annual meeting and Rocky Mountain Section 50th annual meeting, El Paso, Texas.


APPENDIX A: Identifying Well-Cuttings

1. Sieve sample to collect cuttings that are between 2-5 millimeters in size. Try to get approximately 20-30 cuttings and place them in a labeled (Sample ID and depth) petri-dish, also label a plastic beaker.

2. Etch sample with 2.5% HCL for a few seconds making sure to wet entire sample. You may have to spray water on sample to reduce the fizzing.

3. Examine sample under microscope making sure not to have light source to bright. (This can cause headaches)

4. Use Dunham’s classification to ID cuttings and record on worksheet. Record starting and ending depth of each sample only for the first sample on each sheet used for well. For every other sample on sheet only record end depth, also if sample intervals have a skip (i.e 2000-2010 sample interval followed by 2020-2025 sample interval) make-up a sample interval to cover the missing interval (i.e. 2010-2020). Make sure to place each cuttings type identified/recorded in its own separate pile/grouping in the petri-dish.

5. Make sure to stack the petri-dishes in stacks of no more than 12 petri-dishes.

6. When you are done with the cuttings samples, wash them from the Petri dish into the labeled plastic beakers and pour off any excess water. Wash the sample a couple of times again while in beaker and pour off excess water each time, this washes away the HCL.

7. Dry cuttings in an oven set below 90° C till dry or leave cuttings in beakers to open air dry.

8. After cuttings are dried return cuttings to envelopes and place back into boxes, making sure to place the envelopes in appropriate boxes (i.e. appropriate footage that is labeled on box).

9. Wash plastic beakers in alcohol to remove ink and rinse in water and air dry beakers.
### APPENDIX B: How to identify lithologies in well-cuttings

<table>
<thead>
<tr>
<th>Lithology</th>
<th>How to ID Cuttings</th>
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<tbody>
<tr>
<td><strong>Red Sandstone/Siltstone</strong></td>
<td>Red color, grain-size, side of metal needle makes a scratching sound and will leave behind metal flakes.</td>
</tr>
<tr>
<td><strong>Non-Red Quartz Sandstone</strong></td>
<td>Grainsize, side of metal needle makes a scratching sound and will leave behind metal flakes.</td>
</tr>
<tr>
<td><strong>Non-fossiliferous Gray Shale/Siltstone</strong></td>
<td>Silt/Clay size terrigenous grains (quartz, micas, feldspars, etc.), May or may not fizz in 1.5% HCL but sometimes does due to calcareous cement, does not have marine fossils in it but may have plant fragments</td>
</tr>
<tr>
<td><strong>Fossiliferous Gray Shale/Siltstone</strong></td>
<td>Silt/Clay size terrigenous grains (quartz, micas, feldspars, etc.), May or may not fizz in 1.5% HCL but sometimes does due to calcareous cement, has marine fossils in it but may have plant fragments</td>
</tr>
<tr>
<td><strong>Quartz-Calcisiltite</strong></td>
<td>Reacts with HCL (may be slow if small amounts of carbonate present) contains ~50% quartz silt and 50% carbonate silt composed of pellets, and peloids.</td>
</tr>
<tr>
<td><strong>Sandy Lime</strong></td>
<td>Reacts with HCL but contains sand size quartz(carbonate is dominant) contains ~50% quartz sand and 50% carbonate</td>
</tr>
<tr>
<td><strong>Quartz Peloidal Grainstone</strong></td>
<td>Mix of peloidal and quartz grains, side of metal needle makes a scratching sound due to quartz grains and will leave behind metal flakes.</td>
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<tr>
<td><strong>Caliche</strong></td>
<td>Resinous brown color, forming bands or coatings on grains</td>
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<tr>
<td><strong>Fine Dolomite</strong></td>
<td>Fine grained, won't fizz with 1.5% HC or fizzes very slowly, yellowish tan color</td>
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<tr>
<td><strong>Burial Dolomite</strong></td>
<td>Coarse grained, won't fizz with 1.5% HCL or fizzes very slowly, some crystals are rhomb shaped, milky color</td>
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<tr>
<td><strong>Fenestral Lime Mudstone</strong></td>
<td>Clear to cloudy calcite-filled blebs in lime mudstone</td>
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<tr>
<td><strong>Light Wackestone/Mudstone</strong></td>
<td>Light color, fossils float in a lime mud</td>
</tr>
<tr>
<td><strong>Peloidal Grainstone/Packstone</strong></td>
<td>Sand-size carbonate grains that are internally structureless.</td>
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<tr>
<td><strong>Oolite</strong></td>
<td>Good radial/concentrically coated spherical grains, sometimes only a few coatings are present so don't confuse with Peloidal G/P.</td>
</tr>
<tr>
<td><strong>Skeletal Grainstone/Packstone</strong></td>
<td>Composed of skeletal material, grain-supported (crinoidal material gives marble-like look due to the cleavage of the crinoid particles)</td>
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<tr>
<td><strong>Dark Wackestone/Mudstone</strong></td>
<td>Dark color, fossils float in a lime mud</td>
</tr>
<tr>
<td><strong>Anhydrite</strong></td>
<td>Doesn't fizz with HCL and is white-glassy, good right angle cleavage, softer than quartz but same hardness as calcite</td>
</tr>
</tbody>
</table>

1. Siderite has a yellowish-brown stain, and reacts slowly with 10%HCL.
2. Glauconite green grains.
3. Plant material is dark colored.
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APPENDIX C: Example of well-cuttings data sheet.
VITA

Thomas C. Wynn

Thomas C. Wynn was born on November 15, 1962 in Mebane, North Carolina. He graduated from Eastern Alamance High School in May of 1981 and entered the University of North Carolina at Chapel Hill in the fall of that year. In May 1987 he was awarded a Bachelor of Arts with a major in Biology from the University of North Carolina at Chapel Hill. In August of 1994 he entered Old Dominion University as a Masters student, and was awarded a Master of Science in Geology in May of 1998. He started a doctoral program at Virginia Tech in January of 1998 and completed his dissertation in November of 2003.