I, _____ Patrick Ian McLaughlin ________________________________, hereby submit this work as part of the requirements for the degree of:

Doctor of Philosophy

in:

Geology

It is entitled:

Cratonic sequence stratigraphy: advances from analysis of mixed carbonate-siliciclastic successions

This work and its defense approved by:

Chair: __Dr. Carlton Brett__________

__Dr. Warren Huff__________

__Dr. Arnold Miller__________

__Dr. J. Barry Maynard__________

__Dr. Mark Wilson__________
CRATONIC SEQUENCE STRATIGRAPHY:
ADVANCES FROM ANALYSIS OF
MIXED CARBONATE-SILICICLASTIC SUCESSIONS

A dissertation submitted to the

Graduate School
University of Cincinnati

in partial fulfillment of the requirements for the degree of

DOCTOR OF PHILOSOPHY

Department of Geology
McMicken College of Arts and Sciences

Submitted May 16, 2006

By

Patrick Ian McLaughlin

A.A.S., Illinois Central College, 1997
B.S., Illinois State University, 1999
M.S., University of Cincinnati, 2002

Committee Chair:
Dr. Carlton E. Brett
The following series of papers addresses the need for a sequence stratigraphic model specifically designed for cratonic mixed carbonate-siliciclastic successions. Case studies are provided primarily from the mixed carbonate-siliciclastic strata of the Lexington Limestone of Kentucky. These studies explore the sub-regional and regional distribution of decameter-scale couplets composed of clean skeletal grainstone and argillaceous limestones interbedded with shales. Analysis of six couplets that make up the Lexington Limestone along a basin profile reveals that both parts of decameter-scale couplets are widely traceable, though each undergoes a gradual lateral facies change. Subsequent investigation reveals that these six couplets are regionally traceable along strike, showing particularly good similarity to age equivalent strata in New York. More detailed studies are also provided that focus on different aspects of the couplets to help reinforce their sequence stratigraphic significance. The uppermost skeletal grainstone unit of the Lexington Limestone and basal portion of the overlying Kope Formation are analyzed in great detail, incorporating stratigraphic correlation of individual beds between closely spaced exposures, sedimentology of condensed beds and discontinuity surfaces, and faunal and taphonomic gradient analysis of limestones within this interval. The data generated suggest that this grainstone-rich succession represents a deepening-upward succession formed during sea level rise, though with slightly varying degrees of influx of argillaceous sediments (lowstand, early transgression, and late transgression, respectively). Additional case-studies focus on the contact at the base of the grainstone-dominated half of the couplet. This contact, contrary to previous studies, is almost always sharp and erosional. In fact, detailed analysis reveals two closely spaced erosion surfaces, one at the contact of the two halves of the couplet (forced
regression surface), typically overlain by argillaceous calcarenite (falling stage systems tract), and one slightly higher (sequence boundary) overlain by more massive grainstones (lowstand, etc.). Combination of the case-study data with the literature of discontinuity surfaces and condensed beds allows for the formation of a general sequence stratigraphic framework for foreland basins. Finally, a unified model for foreland basin sequence stratigraphy is presented by integrating knowledge of the carbonate margin of middle Paleozoic foreland basins with well-established models concerned with the siliciclastic margin.
# TABLE OF CONTENTS

<table>
<thead>
<tr>
<th>Section</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>ACKNOWLEDGEMENTS</td>
<td>3</td>
</tr>
<tr>
<td>PREFACE</td>
<td>5</td>
</tr>
<tr>
<td><strong>CHAPTER I:</strong> HIGH-RESOLUTION SEQUENCE STRATIGRAPHY OF A MIXED</td>
<td>8</td>
</tr>
<tr>
<td>CARBONATE-SILICICLASTIC CRATONIC RAMP (UPPER ORDOVICIAN; KENTUCKY-OHIO, USA): INSIGHTS INTO THE RELATIVE INFLUENCE OF EUSTASY AND TECTONICS THROUGH ANALYSIS OF FACIES GRADIENTS</td>
<td></td>
</tr>
<tr>
<td><strong>CHAPTER II:</strong> COMPARATIVE SEQUENCE STRATIGRAPHY OF TWO CLASSIC</td>
<td>69</td>
</tr>
<tr>
<td>UPPER ORDOVICIAN SUCCESSIONS: TRENTON SHELF (NEW YORK-ONTARIO) AND LEXINGTON PLATFORM (KENTUCKY-OHIO): IMPLICATIONS FOR EUSTASY AND LOCAL TECTONISM IN EASTERN LAURENTIA</td>
<td></td>
</tr>
<tr>
<td><strong>CHAPTER III:</strong> SIGNATURES OF SEA LEVEL RISE IN MIXED CARBONATE-</td>
<td>144</td>
</tr>
<tr>
<td>SILICICLASTIC FORELAND BASIN SUCCESSIONS</td>
<td></td>
</tr>
<tr>
<td><strong>CHAPTER IV:</strong> FORCED REGRESSION IN FORELAND BASIN SETTINGS:</td>
<td>224</td>
</tr>
<tr>
<td>SYNCHRONIZED RESPONSES FROM THE CARBONATE- AND SILICICLASTIC-DOMINATED MARGINS</td>
<td></td>
</tr>
<tr>
<td><strong>CHAPTER V:</strong> HIERARCHY OF SEDIMENTARY DISCONTINUITY SURFACES</td>
<td>295</td>
</tr>
<tr>
<td>DERIVED FROM ANALYSIS OF MIDDLE PALEozoic FORELAND BASINS OF EASTERN NORTH AMERICA</td>
<td></td>
</tr>
<tr>
<td><strong>CHAPTER VI:</strong> A UNIFIED SEQUENCE STRATIGRAPHIC MODEL FOR FORELAND</td>
<td>350</td>
</tr>
<tr>
<td>BASINS: ADVANCES FROM ANALYSIS OF MIXED CARBONATE-SILICICLASTIC</td>
<td></td>
</tr>
<tr>
<td>SUCCESSIONS</td>
<td></td>
</tr>
<tr>
<td>BIOGRAPHY</td>
<td>401</td>
</tr>
</tbody>
</table>
ACKNOWLEDGEMENTS

The various chapters that follow each contain their own acknowledgements section to recognize those geologists and support agencies that contributed to each of the respective studies. However, this acknowledgements section is dedicated to family and friends that have supported me in my academic journey.

My years within the Department of Geology at the University of Cincinnati have been highly enjoyable. The faculty and staff of the Department of Geology at the University of Cincinnati are witty, intelligent, and kind. In particular, I would like to recognize my advisor Dr. Carlton Brett who has motivated and supported me as mentor, colleague, and friend throughout my years as a graduate student. His patience with my constant debate is greatly appreciated. I would also like to thank the members of my committee: Dr. Arnold Miller, Dr. Warren Huff, and Dr. Barry Maynard as well as my outside committee member Dr. Mark Wilson at the College of Wooster for their support and encouragement. A special thanks to Dr. Thomas Algeo who, as former director of graduate studies, aided me in my application for and ultimate receipt of a University Distinguished Graduate Fellowship, the funding from which allowed me to pursue my research interests full time for three amazing years. I would also like to acknowledge fellow graduate students Alex Bartholomew, Sean Cornell, and Mike DeSantis who are not only good friends, but talented geologists in their own rights from whom I have learned much.

This dissertation is dedicated to my family who has supported me in good times and in bad. Thank you to my grandparents Shirley and John for always keeping a smile on my face and teaching me the value of hardwork, to my mother Glenda and father Patrick for your love and guidance, and to my brother Tim for eternal friendship. I owe the largest debt of gratitude to my wife Susie who has not only acted as loving and patient spouse, but also editor, sounding board,
field assistant, and collaborator in my geological endeavors. Finally, my dogs Roscoe, Stanley, and Wilson were always by my side, be it in the field or at the computer, throughout the course of my studies to give comfort, entertainment, and often a much needed break from the rigors of research. Thanks guys!
PREFACE

Sequence stratigraphy is one of the most unifying models in geology. It has become an essential starting point for studies in sedimentary geology, paleontology, and tectonics because it predicts the location and characteristics of unconformities within the rock record (i.e. sequence boundary, maximum flooding surface), and provides a theoretical framework of relative sea level fluctuation to explain their genesis. Yet, sequence stratigraphic models designed specifically to address the depositional dynamics of epicontinental seas are at best partially developed. A surprising insight, as these deposits are the focus of the vast majority of paleontological and tectonic studies and serve as the foundation for training students in sedimentary geology, especially in the USA. The studies that make up the following chapters seek to address this issue by stepwise development of a sequence stratigraphic model applicable to cratonic mixed carbonate-siliciclastic successions.

The primary goal of the following chapters is to characterize the regularly repeating stratigraphic patterns in the Upper Ordovician of Kentucky (Fig. 1, 2) and to test for the processes responsible for their genesis, resulting in a revised sequence stratigraphy model for this interval. Chapter 1 focuses specifically on stratigraphic patterns in the mixed carbonate-siliciclastic Upper Ordovician Lexington Limestone in Kentucky and Ohio. This study establishes a base line for further studies through division of this interval into a series of facies that represent deposition under a spectrum of environmental energy and sediment regimes. Many facies are bound stratigraphically by distinctive marker horizons such as hardgrounds or erosion surfaces that enabled regional tracing of stratal packages along an onshore-offshore gradient. Ultimately, a series of six depositional sequences were recognized within the Lexington Limestone. Building off of the framework established in Chapter 1, Chapter 2 focuses
on comparison of the Lexington Limestone succession in central Kentucky with the age equivalent Trenton Limestone in its type area in central New York. These two successions show striking similarity in their vertical facies succession and are matched sequence-for-sequence. Notable differences between the two intervals included: A) the Lexington Limestone overall was deposited in a slightly shallower environment, B) K-bentonites are more abundant in the Trenton Limestone, and C) seismites are surprisingly more common in the Lexington Limestone, despite its greater distance from the orogenic front. Having established the regional nature of these depositional sequences from the Upper Ordovician more detailed studies follow focusing on narrow stratigraphic intervals to more fully characterize the architecture of systems tracts (Chapters 3 and 4). Chapter 3 focuses primarily on the characteristics of sea level rise revealed through semi-quantitative analysis of sedimentological, taphonomic, and paleoecological characteristics of the uppermost member of the Lexington Limestone and overlying basal Kope Formation. A general comparison of this interval with other grainstone-dominated packages within mixed carbonate-siliciclastic strata reveals great similarity. Chapter 4 follows by examining the characteristics of sea level fall. In this chapter, the sedimentology and paleoecology of slightly younger Upper Ordovician siliciclastic-dominated depositional sequences (Kope and Fairview formations) are compared against a series of slightly older mixed carbonate-siliciclastic depositional sequences from the upper Lexington Limestone. Though these two intervals show moderate disparity lithologically, they contain very similar facies changes indicative of sea level fall. Discussions in both chapters 3 and 4 seeks to relate the similarity of Upper Ordovician depositional sequences to younger strata. Chapter 5 explores foreland basin sequence stratigraphy in relation to the distribution of discontinuity surfaces. It is shown that foreland basins contain a series of discontinuity surfaces and condensed beds that
represent a general temporal hierarchy. Timing of sediment starvation is tied largely to periods of rapid sea level rise. The final chapter provides a general sequence stratigraphic model for foreland basin successions through synthesis of recent siliciclastic sequence stratigraphic concepts with those established in the earlier chapters for mixed carbonate-siliciclastic successions.

The primary hypothetical differences between patterns proposed here and those of “traditional” sequence stratigraphy are: a) the basic building blocks of depositional sequences in mixed carbonate-siliciclastic successions are small-scale symmetrical cycles rather than asymmetrical shallowing-upward cycles (parasequences; “symmetrical” is used here to convey that these cycles contain a record of both sea level rise and fall, in contrast to the traditional interpretation, and is not meant to be taken in the most literal sense of mirror image symmetry, see Figure 4), b) the motif of depositional sequences varies little from sequence to sequence (see Fig. 1, 2), c) each of the three main portions of a depositional sequence (recording shallowing/deepening of relative sea level; i.e. systems tracts) are dominated by distinct lithologies (e.g. skeletal grainstone dominates the transgressive systems tract), which is a response to changes in siliciclastic sediment input, d) both the forced regression surface and the maximum starvation surface represent significant diastems and correspond to the maximum rate of sea level fluctuation (Fig. 6), e) the period of lowest relative sea level (LST) is commonly represented in cratonic successions, rather than typically absent as previously proposed by Holland (1993), and f) though mixed carbonate-siliciclastic successions bear many similarities to pure siliciclastic successions, mixed carbonate-siliciclastic successions should be considered to represent a more complete record of relative sea level fluctuation and, therefore, contain stratigraphic units not present in pure siliciclastic systems.
In conclusion, this study presents a new model of mixed carbonate-siliciclastic cratonic sequence stratigraphy through integrated qualitative and semi-quantitative analysis of a classic North American stratigraphic interval combined with comparative analysis of other well-studied mixed carbonate-siliciclastic intervals from North America. It is the intent of this study to add to the understanding of the cratonic marine sedimentary record as the model predicts the timing of changes in sedimentation rate, early diagenetic processes, and faunal community composition in response to changes in relative sea level.
CHAPTER 1

High-resolution sequence stratigraphy of a mixed carbonate-siliciclastic, cratonic ramp (Upper Ordovician; Kentucky–Ohio, USA): insights into the relative influence of eustasy and tectonics through analysis of facies gradients

Patrick I. McLaughlin, Carlton E. Brett, Susannah L. Taha McLaughlin, and Sean R. Cornell

H.N. Fisk Laboratory for Sedimentary Geology, Department of Geology, University of Cincinnati, Cincinnati, Ohio 45221, U.S.A.

Abstract

Detailed facies analysis and event stratigraphy of an Upper Ordovician (Rocklandian–Edenian) cratonic ramp succession in eastern North America yields insights into eustatically driven sequence architecture and localized tectonic instability. Seven, predominantly subtidal, mixed carbonate-siliciclastic depositional sequences (3rd-order) are identified and correlated across the length of a 275 km ramp to basin profile. Within the larger depositional sequences (3rd-order) at least two smaller orders (4th- and 5th-) of cyclicity are recognizable. Three systems tracts occur within each sequence (transgressive, TST; highstand, HST; regressive, RST) and are considered in terms of their component parasequences (5th-order). Generally, TSTs are composed of skeletal grainstone-rudstone facies, HSTs are dominated by shaly nodular wacke-packstone facies, and RSTs are mostly calcarenite facies. Systems tracts, sequence boundaries and their correlative conformities, maximum flooding surfaces, and forced regression surfaces were traced from shallow shelf to basinal settings. This high-resolution framework also provides insight into the timing of tectonic fluctuations on this cratonic ramp during the Taconic Orogeny and documents the relative influence of tectonism on lateral facies distributions and eustatically-derived cyclicity.

Keywords: Upper Ordovician, mixed carbonate-siliciclastic sequence stratigraphy, far-field tectonics, seismites, K-bentonites, Trenton, correlative conformity

1. Introduction

Many studies in sedimentary geology draw their data from easily accessible cratonic successions. A great number of these studies require an accurate stratigraphy within which to
study regional variations in depositional environment, biota, and climate during a discrete period of time. Most modern studies draw upon the principles of sequence stratigraphy to provide an integrative technique for forming and testing correlation based hypotheses, capable of yielding highly robust chronostratigraphic frameworks. However, the applicability of sequence stratigraphy to cratonic mixed carbonate-siliciclastic successions, as opposed to the purely siliciclastic passive margin successions from which its principles were developed (Vail, 1987; Van Wagoner et al., 1988), is still being refined.

This study describes seven depositional sequences assigned to the Upper Ordovician Lexington and Kope Formations of central Kentucky and southwestern Ohio. These strata are predominantly subtidal and formed on a tectonically active ramp. Regional facies analysis has provided new insights into the expression of mixed carbonate-siliciclastic depositional sequences in epicontinental settings, including details of the down-ramp expressions of systems tracts and their bounding surfaces, as well as the relative influence of eustasy and tectonics on the formation of cyclic bedding.

1.1 Study interval

The Upper Ordovician (Rocklandian–Edenian) Lexington and Kope formations are exposed in hundreds of closely spaced roadcuts on the Jessamine Dome (northern Cincinnati Arch) in north central Kentucky and western Ohio. These rocks are richly fossiliferous and record a broad spectrum of carbonate litho- and biofacies. We have conducted a detailed analysis through the 155–200 m thick upper Mohawkian–lower Cincinnatian (Rocklandian–Edenian) age succession along a 275 km south-to-north transect from central Kentucky into western Ohio (Figs. 1a, b).
Numerous studies have focused on the Upper Ordovician (Rocklandian–Edenian) strata of the Jessamine Dome, resulting in a plethora of stratigraphic terms. Lithostratigraphers repeatedly referred to these strata as a facies mosaic (Black et al., 1965; Cressman, 1973; Borella and Osborne, 1978; Weir et al., 1984; Ettensohn, 1992). However, recent detailed stratigraphic study indicates that nearly all decameter- to meter-scale cycles are actually traceable throughout much of the study area. Resolving this complex stratigraphy was greatly aided by close examination of a series of new road cuts from the northern Frankfort region to the northern Swallowfield region (Fig. 1a), which affords almost continuous exposure from shore face to outer shelf settings. Large-scale, low-resolution sequence stratigraphic analysis of these strata was published by Holland (1993), Holland and Patzkowsky (1996), and Pope and Read (1997a). Using these studies as a base line we have built a high-resolution sequence stratigraphic framework, incorporating repeated comparison of facies, parasequence stacking patterns, and event beds between nearby outcrops/cores, throughout the entire study interval. The resulting sequence stratigraphic interpretation varies in many subtle, but fundamental aspects from those of previous authors, including: the motif of depositional sequences, the expression of systems tracts, the sedimentary record of sea level rise and subsequent siliciclastic sediment starvation, and recognition of multiple regional erosion surfaces.
Fig. 1. (A) Location map detailing positions of outcrops and cores used in this study from northern Kentucky and southern Ohio. White centered triangles and squares represent cores and outcrops (respectively) directly used in the A-A’ cross-section (black dashed line). Gray triangles and squares represent supporting data. Note that the study area is divided into five depositional regions denoted by gray dashed lines. Light gray lines mark out county boundaries. (B) Regional basement structure map for the study area (modified from Drahovzal et al., 1992). Note the position of the Grenville Front (heavy gray line with tick marks). Remaining faults are reactivated normal faults occupying two dominant trends. The northeast-southwest faults parallel the Grenville Front, but are probably associated with (Proterozoic) Hadrynian rifting. East-west trending faults are associated with the 38th parallel lineament. Black and gray dashed lines represent cross section and depositional regions from figure 1A for reference.
1.2 Paleogeography

The paleogeographic evolution of eastern Laurentia (Fig. 2) during the Late Ordovician has been debated for some time (see introduction in Kolata et al., 2001; and Ettensohn et al., 2002 for review). The current consensus suggests that the Taconic Orogeny, which initiated in the late Middle Ordovician, was the result of a series of collisions of an island arc with the southern margin of Laurentia (Fig. 2; Stanley and Ratcliff, 1985; Rowley and Kidd, 1991) resulting in at least two tectophases (Ettensohn, 1992). During Turinian time a shallow carbonate platform with little topography covered most of Laurentia (Cressman and Noger, 1976) and the Laurentian sea floor remained relatively flat through Rocklandian time (Cressman, 1973; Keith, 1988; Kolata et al., 2001; Ettensohn et al., 2002; Brett et al., this volume) as confirmed by recognition of the Guttenberg C$^{13}$ isotopic excursion found in mid-Rocklandian age strata of similar facies across most of eastern North America (Bergstrom, 2001). Through the Kirkfieldian and Shermanian stages, carbonate platforms developed along basement highs confining a linear bathymetric low (Sebree Trough), which stretched from modern day western Tennessee to western Pennsylvania where it connected with the Taconic foreland basin (Kolata et al., 2001; Ettensohn et al., 2002). Subsequently, the carbonate platforms were partially smothered in the Edenian by a major pulse of siliciclastic sediment influx (Ettensohn et al., 2002; Brett et al., this volume). These recent models depict the Sebree Trough as an open corridor with little to no deposition and possibly erosion for much of its duration until it was filled in the latest Shermanian or Edenian (Hohman, 1998; Kolata et al., 2001, Ettensohn et al., 2002). However, results of this study suggest that deposition occurred within the Sebree Trough throughout the Rocklandian–Edenian (Fig. 3).
Fig. 2. (A) General Late Ordovician paleogeography for Laurentia (modified from Scotese and McKerrow, 1990; Witzke, 1990) and regional paleogeography for the tristate region of Kentucky, Ohio, and Indiana (modified from Mitchell and Bergström, 1991). Black box outlines study area, black dotted line represents A-A' cross section.
1.3. Tectonic setting

A complex set of faults and other basement structures has been detected in the subsurface of the mid-continent, which had a strong influence on deposition of later Paleozoic sediments (Fig. 1b; Black and Haney, 1975; Black, 1986; Drahovzal et al., 1992; Gao et al., 2000). In particular, the Kentucky River fault zone extends across the southern portion of the study area (east-west trending high angle faults) and the Grenville Front lies just east of the study area (north to south trending low angle faults; Fig. 1b). Several authors have implicated many of these features as primary controls on deposition of Upper Ordovician strata in the study area (Cressman, 1973; Borella and Osborne, 1978; Weir et al., 1984; Wickstrom et al., 1992; Pope and Read, 1997; Ettensohn et al., 2002).
Fig. 3 Composite biostratigraphic, lithostratigraphic, and sequence correlation diagram along regional south-to-north A-A' cross-section (Fig. 1). The cross-section is approximately perpendicular to depositional strike as defined by Cressman (1973). Swallowfield section duplicated at page margin for clarity. With the exception of the two northernmost sections, representative stratigraphic columns are composites of measured core and nearby outcrop exposures. Note presence of tidal flat facies in lower half of the column in the south in the Danville and Frankfort regions in the middle of the column. Also note thinning of units near Frankfort and thinning of units in the northern sections.
2. Facies and relative bathymetry of upper Mohawkian–lower Cincinnatian strata of the Lexington Platform and Sebree Trough.

Facies provide critical information for interpreting the pattern and extent of vertical and lateral changes along shelf–basin gradients. Facies analysis for any one locality enables recognition of the bounding surfaces of systems tracts and the relative magnitude of bathymetric variation they represent within depositional sequences. A number of previous studies have characterized facies of the Upper Ordovician in the study area (e.g., Cressman, 1973; Weir et al., 1984; Holland, 1993; Holland and Patzkowsky, 1998; Pope and Read, 1997a). We have incorporated salient features of these previous facies subdivisions. However, because of their importance in making interpretations of relative onshore–offshore position as well as paleoenvironments, especially bathymetry, we reconsider and further subdivide the facies of the Rocklandian–Edenian age strata in the Lexington Platform–Sebree Trough (Table 1). We have emphasized aspects of taphonomy and paleoecology of fossils that provide insights into water depth, substrate, and depositional rates. In the following sections we will briefly discuss aspects of litho-, tapho-, and biofacies and present an inferred depositional environment. Facies are described in order from proximal (near-shore) to distal (basinal).

2.1 Green mudstone facies

The green mudstone facies is composed of light greenish-gray desiccation cracked dolomitic shales and thin argillaceous dolostones. Commonly the green mudstone facies is massively bedded with alternating brownish-gray and light-green laminations. Fossils are scarce, but may include large leperditian ostracodes, bryozoans, gastropods and small pterioid bivalves.
<table>
<thead>
<tr>
<th>Facies</th>
<th>Lithology</th>
<th>Fauna/Taphonomy</th>
<th>Inferred Environment</th>
</tr>
</thead>
<tbody>
<tr>
<td>1) Green mudstone</td>
<td>Light greenish-gray, desiccation-cracked, dolomitic shales and thin, argillaceous dolostones</td>
<td>Sparse; rare leperditian ostracodes, bryozoan fragments, gastropods: (BA-1)*</td>
<td>Supra- to intertidal flats</td>
</tr>
<tr>
<td>2) fenestral micrite</td>
<td>Pale gray to pinkish brown (dove) fenestral micrites; minor greenish shale</td>
<td>Tetradid corals, cyrtodont bivalves gastropods, Bathycorax (trilobites) vertical burrows (Phytopora): (BA-2)</td>
<td>Lower intertidal to shallow subtidal, lagoonal</td>
</tr>
<tr>
<td>3) micritic wackestone</td>
<td>Medium to dark gray, burrow-mottled thick-bedded to massive, cherty wackestone and packstone</td>
<td>Tetradid corals, stromatoporoids, crinoid fragments; nautiloids, strophomenid brachiopods solenoporid algae: (BA-2)</td>
<td>Shallow protected low energy shell, &quot;lagoon&quot;; sorted, fine to medium grainstone (calcareite) grades into facies 1, 5, 6 Absolute depth: &lt;10 m</td>
</tr>
<tr>
<td>4) carbonaceous shale</td>
<td>Black, carbonaceous, fissile shale (lenses in crinoidal grainstones)</td>
<td>Barren except for carbonaceous dencrystal algae; rare ostracodes, inarticulate brachiopods: (BA 2)</td>
<td>Shallow stagnant lagoon; estuary or interdune ponds Absolute depth &lt;10 m (probably &lt;5 m)</td>
</tr>
<tr>
<td>5) calcarenite</td>
<td>Buff-orange weathering, well sorted, fine-to medium-grained grainstone (calcareite), thin planar, trough, and tabular (including &quot;herringbone&quot;) cross-stratification; minor channeling; may appear &quot;pinstriped&quot; in weathered exposures; commonly deformed; may show light gray cherty bands (Tanglewood facies, in part)</td>
<td>Mainly fragmentary, comminuted skeletal fragments, including crinoid ossicles; patches of ramose bryozoans, Rafinesquina and other strophomenids; Skolithos present (BA-2 to 3)</td>
<td>Shallow, high energy skeletal shoals; normal wave base to average storm wave base Absolute depth: 10-20 m</td>
</tr>
<tr>
<td>6) grainstone-rudstone</td>
<td>Light pinkish gray, coarse skeletal grainstone to rudstone; medium-bedded to massive; minor interbeds of Facies 4, cm-scale gray, fossiliferous shales minor channeling and cross-bedding; (Tanglewood facies, in part)</td>
<td>Fragmentary to well preserved bryozoans, brachiopods (especially Rafinesquina); crinoid pluricolumnals; large fragments of Isotelus; gastropods (Stamping Ground-Strodes Creek variants yield solenoporid algae and stromatoporoids) (BA-3)</td>
<td>Shallow, siliciclastic-starved, moderate to high energy; average storm wave base and euphotic zone Absolute depth: ~10-30 m</td>
</tr>
<tr>
<td>7) shaly nodular wacke-packstone</td>
<td>Medium to dark gray, nodular, thin- and wavy-bedded wacke- to packstones, minor grainstone, and alternating thin (1-5 cm) partings of calcareous shale.</td>
<td>Articulated to fragmentary brachiopods: Elytopsida, Rafinesquina, Strophomena, Heberella, and Platyphora; ramose, massive, and encrusting bryozoans abundant; fragments of trilobites; Chondrites and Planolites common; solenoporid algae and stromatoporoids in some intervals: (BA 3-4)</td>
<td>Muddy, common storm wave-winnowed shelf dysphotic zone; relatively slow burial, time-averaging, common reworking Absolute depth: ~20-40 m</td>
</tr>
<tr>
<td>8) shale and limestone</td>
<td>Medium to pale olive gray sparsely fossiliferous shales and mudstones, non-calcareous, interbedded with skeletal pack- or grainstone beds 1 to 40 cm thick and thin planar to hummicky laminated calcisiltites some with gutter casts; small ellipsoidal carbonate concretions may be present (Kope facies)</td>
<td>Mudstones carry small brachiopods mainly Ommatella, and Seymouritella; small crinoids; Ectenicrinus, Cincinnaticrinus, Locrinus and Mereocrinus graptolites common in some beds trilobites include Isotelus, Cryponithus and Flexicalymene; ramose bryozoans limestones composed of brachiopods and bryozoans; typically fragmented and abraded: (BA-4 to 5)</td>
<td>Deep to moderate muddy soft substrates alternating with shell hash gravels formed by storm reworking; abundant gradient current deposits; moderate to rapid rates of deposition; dysoxic to fully oxic; dysphotic Absolute depth: ~30-80 m</td>
</tr>
<tr>
<td>9) rhythmite</td>
<td>Dark gray, laminated shales, calcisiltites, and argillaceous limestones</td>
<td>Sparse fossils include small dalmanelid brachiopods, inarticulates (Mesoboles), bivalves, sponges, nautiloids, rare trilobites (Cryptolithus, Triarthrus), graptolites, and small Chondrites: (BA 5-6)</td>
<td>Deep shelf to ramp, storm wave base, distal gradient currents; dysoxic Absolute depth: 80-100 m</td>
</tr>
<tr>
<td>10) laminated shale</td>
<td>Dark olive to brownish gray, laminated to massive, slightly calcareous, clay shale with rare interbedded tabular calcilutites</td>
<td>Mostly barren; some bedding planes with abundant graptolites; small inarticulates (Mesoboles, Lingula) rare Triarthrus trilobites; traces rare, thin pyritic burrows: (BA-6)</td>
<td>Basinal, dysoxic to anoxic, muddy substrates, aphytic to dysphotic Absolute depth: &gt;100 m ?</td>
</tr>
</tbody>
</table>

*BA = benthic assemblage; a standardized onshore-offshore succession of faunas  
** water depth estimates are only approximations; these represent our best estimates based upon combined data of photic, wavebase, storm, and paleontological (e.g. algal) indicators

Table 1. Facies and inferred depositional environments of the (Upper Ordovician) Turonian–Edenian age strata of Kentucky and Ohio, emphasizing lithology, sedimentary structures, taphonomy of skeletal material, and representative fauna.
*Interpretation:* These dolomitic to shaly beds are interpreted as the shallowest facies in the study interval. The presence of desiccation cracks, nearly planar stromatolitic lamination, and possible teepee structures all point to a high intertidal to supratidal environments. Lack or rarity of evaporites suggests a humid tidal flat rather than sabkha conditions (Pope and Read, 1998).

2.2 *Fenestral micrite facies*

The fenestral micrite facies consists of pale gray to pinkish-brown micrites. Bedding is typically medium to thick with very minor greenish-gray shale partings separating beds. These beds characteristically contain fenestrae (sparry clots and birdseye structure), pellets, and may possess vertical burrows (*Phytopsis*), which crosscut horizontal laminations. Whole fossils are rare, but thin bioclastic stringers present within this facies often include a combination of gastropods, cyrtodont bivalves, ostracodes, and less common *Tetradium* corals and stromatoporoid sponges. Desiccation features are rare.

*Interpretation:* The carbonate muds composing these fenestral micrites are thought to have accumulated in intertidal to shallow shelf-lagoonal conditions. The extremely uniform microcrystalline grain size suggests that this facies represents very low energy inner-shelf conditions that were conducive to accumulation of pelletal lime muds. Thin bioclastic stringers likely represent the signature of intermittent high energy events such as storms.

2.3 *Micritic wackestone*

The micritic wackestone facies is composed of massive, medium dark-gray micritic limestones, which display a range of ichnofabrics from distinct burrowed galleries through disrupted fabrics (Fig. 4a). Fossils often include the tabulate corals *Tetradium* (typically as...
fragments, but locally as biostromes of intact coralla), and *Foerstphyllum*, as well as *Stromatocerium* and *Labechia* stromatoporoids. Ostracodes, gastropods, and strophomenid brachiopods may also be abundant; crinoid debris is present in some lenses.

*Interpretation:* This burrow-mottled limestone is thought to represent shallow subtidal (lagoonal to shallow protected shelf) environments probably less than 10 m deep. As these carbonates contain numerous stenohaline corals, brachiopods, crinoids, and cephalopods, it is evident that they were deposited under normal marine salinities. However, the low diversity of faunas also suggests somewhat restricted inner shelf conditions.

2.4 *Carbonaceous shale facies*

Black, organic rich, fissile, carbonaceous shales are found as thin lenses (<10 cm) and partings associated with fenestral micrite and calcarenite facies. These shales may contain carbonaceous remains including dasycladacean green algae (Marashi, 1974; Traub, 1982) and brachiopods.

*Interpretation:* These black shales have sometimes been construed as deep-water facies and their carbonaceous remains identified as graptolites (Ettensohn, 1992). However, their association with other shallow water, low energy facies and the presence of dasycladacean algae indicates that they instead were formed in very shallow water within the photic zone, representing a lagoonal environment.

2.5 *Calcarenite facies*

The calcarenite facies consists of fine to medium, sand sized, well-sorted skeletal fragments (Fig. 4b). Skeletal material is largely composed highly abraded and fragmented robust bryozoans
Fig. 4. Outcrop photographs of various lithofacies. Note that A-F represents a shallow-deep succession. (A). Very shallow shaly nodular facies. Note wavy nodular Tetradium and ostracode wackestone (lagoonal?; Faulconer Bed, Danville region, hammer for scale). (B). Calcarenite facies. Note herringbone cross-bedding and minor flaser-bedding in fine-grained calcarenite (Tanglewood Member, Frankfort region; hammer for scale). (C). Transitional calcarenite to shallow shaly nodular packstone facies. Note wavy nodular to cross-bedded packstone to grainstone with overturned, fragmented stromatoporoid (Stamping Ground Member, Frankfort region; 1 cm tick marks on measuring stick) (D) Shallow shaly nodular wackestone-packstone facies. Note stromatoporoid bearing shaly nodular wackestones and packstones with widely spaced light gray grainstones (Stamping Ground Member, northern Frankfort region; black bar = 30 cm). (E) Distal shaly nodular wackestone-packstone facies. Note shaly nodular wackestones and widely spaced, rhythmically bedded light gray tabular pack-grainstones (Bromley shale, Swallowfield region, black bar = 30 cm). (F) Rhythmite facies. Note the very uniform bedding of calcisiltite and shale. (Bromley shale, Swallowfield region, black bar = 30 cm)
and brachiopods. Calcarenite facies may also contain up to 50 percent quartz silt, however values around 10 percent are much more typical. Planar to trough cross-bedding is common and locally, sections have herringbone cross stratification (Fig. 4b). Some bedding planes show limonitic staining, which accents their appearance in weathered outcrops.

**Interpretation:** The calcarenite facies represents one of the most common shallow-water environments preserved in the study interval. The highly reworked character of the skeletal grains suggests a high-energy environment associated with shoals (James and Kendall, 1992). The presence of thin, wavy laminae with mud drapes and herringbone cross-bedding indicates tidally influenced deposition (Hrabar et al., 1971) and depths within fair weather wave base of probably 10 m or less.

### 2.6 Coarse skeletal grainstone-rudstone facies

The skeletal grainstone-rudstone facies is one of the most widespread facies in the study area. It is composed of well- to poorly-sorted, medium- to thick-bedded skeletal grainstones and rudstones, which may locally grade into packstone. Grainstone-rudstone facies grades up dip into calcarenite facies, but persists as grainstone-rudstone facies well into the basin where it typically pinches out into a singular, pyrite- and phosphate-rich horizon. Features associated with grainstone-rudstone facies typically show unidirectional trends regionally and are consistently described below from one extreme to the other. Sedimentary structures range from herringbone cross bedding, small-scale (tidal?) channels, and lenticular—amalgamated bedding to megaripples, hummocky cross stratification, planar lamination, small gutter casts, and tabular bedding. Likewise, fossil associations range from robust branching and dome-shaped bryozoans, brachiopods, gastropods, red algae, and stromatoporoids, to delicate branching bryozoans,
crinoids, thin-shelled brachiopods, and trilobites. Faunal gradient analysis performed for many of these taxa suggests a relatively wide range of bathymetric values (Holland et al., 2001). Similarly, the taphonomic character of the skeletal grains ranges from highly abraded, to pristine, to corroded suggests a wide range of physical and chemical conditions. Fossil preservation typically varies little locally within individual beds as compared with differences observed regionally within a single bed or between successive beds.

**Interpretation:** The great variation in fossil preservation, sedimentary structures, and faunal assemblages observed at the regional scale suggests that the grainstone-rudstone facies spans a range of depths. The shallow end of the gradient intertongues with the calcarenite facies and is represented by bryozoan (massive and domal) and gastropod grainstone-rudstone, which may also include a component of robust brachiopods. Green and red algae and micritized crinoid grains within this end of the grainstone-rudstone facies spectrum indicate deposition within part of the photic zone. Cyclocrinitid green algae are similar to modern dasycladacean green algae, which occur only at depths of about 10–15 m, with an absolute maximum in clear water of 30 m (Brett et al., 1993). Hence, shallow grainstone-rudstone facies, although slightly below fair-weather wave base, based on sedimentary structures, probably record depths no greater than about 15 to 20 meters. On the opposite end of the spectrum, crinoids, delicate branching bryozoans, and thin-shelled brachiopods comprise a low diversity faunal assemblage suggesting low and energy and perhaps, near dysoxic conditions. The persistence of the grainstone-rudstone facies into depths near or below storm wave base, and the typically well preserved state of fossils at that end of the facies spectrum suggests that these deposits are accumulating by processes other than intense storm winnowing alone. Though the taphonomic grade of the fossils is somewhat broad in any given outcrop sample and suggestive of time averaging, evidence of
extensive abrasion is generally lacking except in the shallowest end of the spectrum. Commonly shells in the distal grainstone-rudstone facies are whole, displaying only minor fragmentation and occasionally mineralization and corrosion, suggesting relatively quiet conditions, reinforcing interpretations of sediment starvation as the primary mechanism in shell bed accumulation (Kidwell 1991; Brett and Baird, 1993). Sediment starvation during sea level rise would not only allow for formation of these relatively pure shell beds, but also agrees with trends of increasingly better preservation and more dysaerobic-adapted faunal assemblages upward through thick intervals of this facies.

2.8 Shaly nodular wackestone-packstone facies

Similar to the grainstone-rudstone facies, the shaly nodular wackestone-packstone facies is widely distributed throughout the study interval. Generally, this facies consists of medium to dark gray, nodular wackestones and packstones with thin (1–5 cm) alternations of calcareous shale. In weathered outcrops this shaly nodular facies typically has a rhythmic appearance from regular spacing of decimeter-scale wavy bedded limestone with alternating shale partings (Fig. 4e). Decimeter scale beds also may be bundled in groups of 15 to 20 between thicker or more persistent pack- to grainstone beds.

This facies is commonly very fossiliferous and trends in fossil content, distribution, and preservation closely parallel those trends described for the grainstone-rudstone facies. Additionally in this case, highly abraded skeletal constituents and more common robust forms typically coincide with a decrease in shale content (Fig. 4c). Again, similar to the grainstone-rudstone facies the shaly nodular wackestone-packstone facies grades into calcarenite facies up
dip, however down dip it grades into rhythmite facies (Fig. 4f). The trace fossils *Chondrites* and *Planolites* are common throughout much of the range of this facies.

**Interpretation:** Similar to the coarse grainstone-rudstone facies, this facies spans a gradient from shallow moderate-energy environments (Figs. 4c,d), where it contains more robust fossil forms and intertongues with calcarenite facies (Fig. 4b) to relatively low-energy environments (Fig. 4e) where it contains more gracile fossil forms and intertongues with tabular interbedded calcisiltites and shales (rhythmite facies; Fig. 4f; see below). The wide variation in characteristics and fossil associations of this facies suggests deposition between fair weather and storm wave base.

The abundance of burrows suggests that bioturbation played a major role in the fragmentation and minor reorientation of small skeletal grains. However, the occurrence of aligned fossils, and overturned stromatoporoids, suggests intermittent strong turbulence from storm waves and currents. Occasional mud layers occur within stromatoporoid coenostea suggesting episodic input of fine-grained sediment, probably following storms in more proximal areas. Solenoporid red algae are occasionally found near the shallower end of this facies, indicating deposition within the euphotic zone.

### 2.9 Shale and limestone facies

The shale and limestone facies is composed of interbedded gray shale and thin- to medium-bedded wacke-packstone, which is characteristic of the Kope Formation (Edenian) in the Cincinnati region. The siliciclastic portion is characterized by medium to pale olive gray, silty claystones and mudstones that are typically soft and non-calcareous. These beds are sparsely
fossiliferous and are interbedded with thin calcisiltites, wackestones and typically more tabular packstones.

Faunas of the siliciclastic mudstones tend to be typified by small, thin shelled brachiopods, such as *Onniella*, and *Sowerbyella*, a few trilobites, including *Flexicalymene* and *Isotelus*, small crinoids with relatively long stems and slender small crowns, e.g., *Ectenocrinus*, *Cincinnaticrinus*, and *Iocrinus*. Graptolites are present on some bedding planes. Fossils in the shales are commonly articulated. The wacke-packstones are composed predominantly of bryozoans, and fragmentary brachiopods.

**Interpretation:** This facies is considered bathymetrically equivalent to the much of the shaly nodular wackestone-packstone facies, but contains a much greater proportion of mud. Relatively rapid deposition of muds may have led to both sparse distribution of faunas as well as the excellent preservation of fossils seen in some beds (i.e. *Triarthrus* molt clusters; Kohrs, 2002). Thin lenticular wacke-packstone beds are believed to represent storm winnowing events during periods of siliciclastic input in contrast to associated thicker grainstone-rudstone beds, which represent winnowing during periods of sediment starvation.

### 2.11 Rhythmite facies

This facies is very common, especially in the northern portion of the study area. It consists of dark gray, slightly organic-rich, sub-laminated shales, with interspersed thin planar to hummocky cross-stratified calcisiltites (Fig. 4f). More shale-rich variants of this facies contain argillaceous concretions. Bedding is generally very regular with alternating decimeter-thick bands of shale and calcisiltite. Rhythmite facies grades laterally into dark brownish gray, laminated shale facies down ramp and into distal shaly nodular facies up ramp.
Shales are typically sparsely fossiliferous, but may contain well preserved bivalves and thin shell hash layers with abundant dalmanellid brachiopods. Low diversity associations of graptolites, inarticulate brachiopods, sponges, ostracodes, and a few species of trilobites, notably *Isotelus, Cryptolithus*, and, rarely, *Triarthrus*, dominate fossil assemblages near the deeper range of this facies. Small *Chondrites* traces may be present locally. Most fossil material is well preserved, but disarticulated. However, some horizons contain very well preserved complete fossils.

**Interpretation:** Lamination of sediments (a near absence of burrowing), and the sparse, relatively small fauna indicate low energy, dysoxic conditions, below the effects of all but the deepest storm waves. The presence of thin, sorted layers of fossil debris suggests occasional distal storm wave/current effects. Thin calcisiltites are interpreted as a distal expression of storm-generated gradient currents, which imported carbonate silt from shallower shelf areas. The presence of small carbonate concretions indicates the development of alkaline conditions intermittently within the sediment column, probably associated with the development of a zone of sulfate reduction within the upper sediments, and implies periods of low net sedimentation.

2.12 *Laminated shale facies*

This facies ranges from brownish-gray shales to alternating olive gray and brownish gray shale with rare to abundant calcilutite lamina. It contains a limited fauna including the inarticulate brachiopods *Lingula* and *Leptobolus*, graptolites, and the trilobite *Triarthrus*. Graptolite rhabdosomes commonly show some degree of fragmentation and current alignment. There is little to no bioturbation and concretions are absent.
Fig. 5. Idealized stratigraphic column of a single depositional sequence and component cycles. Lithologic descriptions given at right. Varying scales of cyclicity with bounding surfaces provided at left (symbols correspond with figure 3). 5th order cycles alternate in shading to clarify boundaries. RST especially representative of sequences M6A-C1. No scale intended.
**Interpretation:** The dark, organic rich muds accumulated in anoxic to dysoxic deep waters below storm wave base. Minor turbidity or gradient currents occasionally disturbed the bottom, but most benthic organism remains accumulated in proximity to living sites. Water depths of more than 100 m are commonly invoked and some authors suggest depths of several hundred to thousands of meters (Joy et al., 2000). However, interfingering with gray mudstones and the affinities of faunas of brownish gray shales with those of rhythmite facies favors the shallower depth estimates.

3. **Sequence architecture**

Recognition of the facies spectrum described above was essential to establishing a sequence stratigraphic framework for the study interval. In subtidal sequences, sharp facies dislocations/offsets are the best indicators of systems tract bounding surfaces (e.g. maximum flooding surfaces). In general, systems tracts are dominated by a discrete group of facies. The vertical distribution of facies may appear complex, but we suggest this is the result of the interference and reinforcement of three readily identifiable scales of cyclicity (Fig. 5). These interpretations are reinforced by event bed correlation, which has formed an independent basis for testing the proposed sequence stratigraphic framework. We present a highly detailed 30 km cross-section from outcrops of the Frankfort to Swallowfield regions to serve as an example of the following concepts (Fig. 6).

3.1 **General sequence surfaces and systems tracts**

*Sequence boundary (SB):* marks lowest relative sea level and are expressed as a subaerial unconformity or sharp facies dislocation (i.e. evidence for disjunctive juxtaposition of shallower
Fig. 6. Detailed 30 km south to north cross-section of the M6B sequence from the northern Frankfort region to the mid Swallowfield region along the A-A' profile. Measured sections taken from large newly created roadcuts representing nearly continuous exposure representing a profile from inner to outer ramp conditions. Note the decreasing proportion of calcarenite to shaly nodular wacke-packstone facies in the down ramp direction (from left to right). Lithologic key equivalent to figure 5, symbols same as figure 3. Note event bed correlation including stromatoporoid and rhodolite epiboles, K-bentonites, chert horizons, and hardgrounds in addition to major facies offsets. 3rd-order systems tracts delineated by black lines, 4th-order systems tracts additionally include gray lines (see fig. 5 for details of nested scales of cyclicity). Black and white bars on left in meters.
over deeper water facies; type 2 sequence boundary of Van Wagoner et al., 1990; Figs. 5, 7a). In these epicontinental successions where it appears sea level did not fall below a shelf-slope break, the sequence boundary is inferred to represent only a small hiatus in deposition. The sequence boundary progresses from a subaerial unconformity in the shallowest areas, to a sharp planar marine erosion surface where the effects of constant wave agitation have beveled the sea floor, to an increasingly irregular contact below fair weather wave base where the sequence boundary becomes a correlative conformity (Van Wagoner et al., 1990). Correlative conformities may be difficult to recognize in mid-shelf areas where calcarenite facies of the underlying RST (or ramp margin wedge) are overlain by calcarenite to grainstone-rudstone facies of the TST. Correlative conformities are more easily recognized in the outer shelf where clean skeletal grainstone-rudstone facies rest upon muddy, shaly nodular wackestone-packstone facies of the RST.

*Transgressive systems tract (TST):* is dominated by grainstone-rudstone facies, which is thought to occur primarily as a result of sediment starvation due to relative sea level rise and trapping of siliciclastic sediments in estuaries near their source areas. Unlike purely siliciclastic successions, TSTs in mixed carbonate-siliciclastic successions are recognizable across a spectrum of environments from tidal flats to basinal settings (Figs. 5, 7a). In peritidal environments TSTs typically occur as fenestral micrite facies, which can be traced down depositional dip into calcarenite and finally grainstone-rudstone facies. Portions of this same facies succession can be observed upward within the TST suggesting deepening upward (retrograding) patterns. Within monotonous grainstone-rudstone successions, the deepening upward pattern is revealed by analysis of sedimentary structures and faunal trends from more robust to more fragile forms. In outer shelf to basin margin areas the pack-grainstones of the TST stand in sharp contrast to the surrounding rhythmite and laminated shale facies. Within basinal
Fig. 7. Photographs of diagnostic surfaces and event beds used in correlation. (A) Sequence boundary. Note the abrupt planar truncation of the underlying rhythmite facies (Brannon Member) by the overlying coarse skeletal grainstones of the TST (Sulphur Well Member). This facies dislocation is diagnostic of many sequence boundaries within the study interval (black bar = 1 m; eastern Frankfort region). (B) Flooding surface. These surfaces are often heavily stained and mineralized (limonite from weathering of pyrite) and mark an abrupt shift from relatively shallower clean skeletal grainstone-rudstone facies (Strodes Creek Member) to deeper water rhythmite facies (Greendale Member) associated with maximum flooding (arrow points to steel tape on flooding surface; Swallowfield region). (C) Forced regression surface (base of Point Pleasant Formation). Note channelized erosion surface (lower arrow) marking an abrupt shift from relatively deep-water rhythmite facies (Bromley shale) to shallow water calcarenite facies. Upper arrow marks C1 sequence boundary (black bar = 1 m; eastern Swallowfield region). (D) Seismite (soft sediment deformed zone, Brannon Member, Danville region). These zones of typically widespread soft-sediment deformation form excellent markers in regional correlation. (E) Hardground (Point Pleasant Formation, Swallowfield region), note the crinoid holdfasts and borings on this lumpy bryozoan encrusted bedding plane (pencil for scale). (F) Epiboles of Solonopora and Labechia stromatoporoids (Strodes Creek Member, Swallowfield region).
settings TSTs occur as thin laminae of pyrite- and phosphate-rich shell hash consisting of dalmenellid brachiopods, and in the most distal environments of the Sebree Trough are reduced to laminae rich in the inarticulate brachiopod *Leptobolus*, ostracods, and conodont elements.

*Maximum flooding surface (MFS):* or more appropriately maximum flooding zone, marks an abrupt shift to more siliciclastic-rich, finer-grained, deeper-water facies, recording rapid rise of relative sea level. In some cases the MFS occurs as a thin interval (decimeters) of corroded hardgrounds with pyrite and phosphate coatings (Figs. 5, 7b). Platters (bored and encrusted hardground clasts) are also common of these intervals suggesting that these surfaces also record periods of minor submarine erosion. These mineralized surfaces are typically overlain by rhythmite facies. Intervals surrounding maximum flooding surfaces may also contain condensed beds with a high density of pelagic elements (e.g. conodonts) and K-bentonites.

*Highstand systems tract (HST):* is characterized by a progradational succession typically dominated by shaly nodular wacke-packstone and rhythmite facies (Figs. 4a-f). These mud-rich strata suggest an increase in fine-grained siliciclastic input over that of the TST in response to slowing and initial drop of sea level, allowing progradation of previously sequestered sediments. The basal HST typically contains multiple obrution deposits (smothered fossil assemblages), recording rapid accumulation of mud layers. Fauna within HSTs typically grade from gracile to more robust forms upward, suggesting increasingly higher energy environments associated with lowering of relative sea level (Holland et al., 2001). A given horizon within the HST typically will grade from laminated shale facies in the most distal settings up ramp into rhythmite, shaly nodular wackestone-packstone, calcarenite, and finally micritic wackestone facies. Segments of this same progression can be observed in vertical succession within the HST indicating upward shallowing, following Walther’s Law. Unlike highstands of carbonate-dominated ramps
(Handford and Loucks, 1993), HSTs of mixed carbonate-siliciclastic ramps are rich in siliciclastic mud and stand in sharp contrast to the clean carbonates of the TST.

**Forced regression surface (FRS):** is an erosion surface marked by a sharp facies dislocation (Plint and Nummedal, 2000), typically showing an abrupt shift from rhythmite or shaly nodular facies into calcarenite facies. The FRS represents a period of rapid sea level lowering that separates the HST from overlying RST (Figs. 5, 7c). The FRS is often cryptic in the basin, displaying the most dramatic erosion in mid to outer ramp settings where it typically forms channels oriented in the down ramp direction. The FRS truncates up to a few meters of the underlying HST in inner-to mid-ramp areas, becoming conformable down ramp. The FRS typically represents a significant erosion surface in epicontinental sequences where sea level does not fall below the shelf-slope break. In cases of the latter it typically merges with the overlying sequence boundary.

**Regressive systems tract (RST):** is dominated by calcarenite facies and exhibits a generally progradational (shallowing upward) pattern bounded by the underlying FRS and the overlying SB (Figs. 5, 7c). The RST usually forms a progradational wedge, draped by the overlying TST. Strata of the RST may be lithologically similar to sediments of the basal TST in mid-shelf areas, complicating recognition of the SB. However, this surface becomes increasingly recognizable into the outer shelf as the RST becomes muddier. The RST represents the depositional period during which the lowstand fan would normally be deposited in a passive margin setting where sea level had dropped below the shelf/slope break. The RST as designated here may be equivalent to ramp margin wedge of Van Wagoner et al. (1990). However, because we recognize the sequence boundary as overlying the strata in question, they are more closely aligned with the definition of falling stage systems tract of Plint and Nummedal (2000), also referred to as
regressive systems tract by Brett et al. (1990) and Naish and Kamp (1997). We use the later term because we think these strata represent deposition during regression, the counterpart to deposition during rising sea level represented by the transgressive systems tract.

3.2 Description of sequences and component cycles

Parasequences: comprise one of the smallest scales (5th-order) of cyclicity in the study interval, yet in many cases are traceable across broad portions of the study area. Parasequences may share motifs of larger depositional sequences containing condensed shell beds at their bases, analogous to the TST, followed by a shallowing upward succession analogous to the HST and RST. Parasequences average around 1 m in thickness, though may be much greater (see discussion of the C1 sequence), with component systems tracts making up varying proportions of that thickness. Parasequences are grouped into parasequence sets based on similar lithologies forming the systems tracts of small-scale sequences (Fig. 5).

Small-scale sequences: comprise an intermediate scale of cyclicity (4th-order), composed of multiple parasequences and parasequence sets (Fig. 5). Parasequence sets form the systems tracts of small-scale sequences. Typically four to five parasequences make up a parasequence set. Three parasequence sets representing transgressive, highstand, and regressive systems tracts, typically comprise a small-scale sequence. Small-scale sequences contain the most readily identifiable systems tracts of any scale of cyclicity. Small-scale sequences as defined herein are approximately equivalent in thickness to “parasequences” defined for this stratigraphic interval by Pope and Read (1997), but differ in having a slightly different motif.
Depositional sequences: are the largest scale of cyclicity (3rd-order) positively identified in this study and make up the fundamental unit for describing changes in basin geometry in the following sections (Fig. 5). Depositional sequences are commonly 20–25 m in thickness, but may be much greater. They approximate in thickness “parasequence sets” defined by Pope and Read (1997a) for the Lexington Limestone. Duration of these depositional sequences is speculative. However, our best estimates (Brett et al., this volume) suggest that they are likely 1.0–1.5 Ma in duration, similar to 3rd-order sequences defined for more recent and better-dated successions (Van Wagoner et al., 1990). The very regular distribution of component small-scale sequences and parasequences throughout not only the Upper Ordovician, as detailed in this study, but in the Silurian and Devonian of the Appalachian basin (Brett et al., 1990, Brett and Baird, 1996) in general, leads us to suggest that the components of depositional sequences likely represent the signature of high-frequency eustatic fluctuations within the Milankovitch band. At a larger scale, depositional sequences cluster into small groups (depositional sequence sets) that show a general shallowing trend (e.g. approximating Holland and Patzkowsky’s (1996) M5 and M6 sequences).

3.3 Description of Event beds and their sequence stratigraphic significance

K-Bentonites: horizons of altered volcanic ash, occur as yellow weathering unctuous clay layers in outcrop and occasionally as reworked concentrations of biotite or other heavy minerals (Kolata et al., 1996). These horizons are traceable for 10s–10,000 km² in the study area and are typically concentrated at maximum flooding surfaces (Fig. 5).

Seismites: widespread zones of soft-sediment deformation are relatively common in rhythmite and silty calcarenite facies where bioturbation is minimal (Fig. 7d). The lateral extent
of deformed zones ranges from 100s to 1000s km$^2$ (Pope et al., 1997; McLaughlin and Brett, submitted) and are typically concentrated in HST and RST intervals (Fig. 5).

**Hardgrounds:** surfaces showing evidence of encrustation and/or boring, often display iron and/or phosphatic mineralization (Fig. 7e). These surfaces are most common in transgressive systems tracts and at maximum flooding surfaces (Fig. 5). Individual hardgrounds in the study interval have been traced for 10–100s km$^2$ (McLaughlin and Brett, 2001).

**Epiboles:** widespread intervals containing an abundance of a normally rare or otherwise absent taxa, occur at several horizons in the study interval (Fig. 7f; Brett and Baird, 1997; see Brett et al., in press for detailed description of epiboles in the C1 sequence). Epiboles can be correlated 10–1000s km$^2$. Most epiboles in the study interval are associated with maximum flooding surfaces (Fig. 5).

### 4. Vertical and lateral facies gradients of depositional sequences M5A–C1: evidence for the relative influences of eustasy and tectonics

The 3rd-order depositional sequences identified in the study interval are briefly outlined below. We discuss the lateral aspects of these sequences, including their bounding unconformities/flooding surfaces and facies of their systems tracts. On the basis of absolute depth analogs from the Silurian (Brett et al., 1993), we have estimated depth variations within and between sequences. These data indicate the amount of sea level fluctuation that accompanied formation of systems tracts and bounding surfaces. Sequences and their components are compared along a standard set of reference points, commencing with the Danville region in the south, northward to the Frankfort, Swallowfield, Cincinnati, and finally Dayton regions (Fig. 1a). In this way we are able to show that while the bounding surfaces that bracket each systems tract
Fig. 8. Facies and thickness distributions for depositional sequences from the Danville-Dayton regions. Note occurrence of grainstone-rudstone facies as thin tongues far to the north where they are surrounded by laminated shale facies. Note the progressive onlap of laminated shale facies from the Danville region into the Cincinnati region (M5A-M6A sequences) and subsequent retreat beginning in the M6B sequence. Note the presence of fenestral micrite and calcarenite facies in the Danville region in the M5C and M6A sequences and subsequent retreat of those facies to be replaced by deeper grainstone-rudstone facies and distal shaly nodular wacke-packstone and laminated shale facies beginning with M6B sequence. Note thinning of sequences in the Frankfort region (resulting in v-shape) with increase of shallow water facies beginning in the M6A sequence. Sections scaled from figure 7 with additional data from the Dayton region.
are persistent, the extent of lateral facies changes within each systems tract varies slightly from one region to the next, reflecting local tectonic effects of subsidence and uplift (Figs. 6, 8, 9).

4.1 M5A Sequence

The M5A sequence boundary is an erosional unconformity, at the base of the Lexington Limestone, across much of the study area (Fig. 3; Cressman, 1973; Holland and Patzkowsky, 1996; Pope and Read, 1997; Hohman, 1998). This sequence boundary is a sharp, slightly wavy surface, which demonstrably cuts out several decimeters of the underlying strata (Cressman, 1973). The fenestral micrites of the upper M4 sequence (Tyrone Formation) contain a series of K-bentonites (Conkin and Conkin, 1983, Huff et al., 1992), which form useful markers that demonstrate the regional truncation at the base of the M5A sequence (Cressman, 1973; Pope and Read, 1997a; Hohman, 1998). Locally, in the Frankfort region, the uppermost of the K-bentonites (Millbrig), was removed, by erosion at the sequence boundary. However, in nearly all other sections the M5A sequence boundary lies close above the Millbrig K-bentonite. This indicates that the erosion surface is not highly irregular and that only a minor, regionally angular, discordance exists at this level.

The M5A TST (Curdsville Member of the Lexington Limestone) is composed of *Tetradium* coral-bearing calcarenite near its base in the Danville region, which grades vertically and laterally to the north into relatively massive, crinoidal grainstone-rudstone (Fig. 8). The M5A TST contains hardgrounds at several levels. Near the middle of the TST two K-bentonites are present within rhythmite facies and can be traced in several outcrop sections across Danville and Frankfort regions (Shaker Creek and Capitol metabentonites of Conkin and Desari, 1986). In many sections the Capitol metabentonite just underlies or is incorporated into a widely traceable
Fig. 9. Series of basin profiles along A-A' (south is left in each case), inferred from facies distributions within each depositional sequence (density of arrows representative of rate of basin shape change; depositional regions abbreviated, i.e. DV=Danville region). Relative position of sea surface line represents level at maximum sea level highstand. (A) Rocklandian M5A sequence, note that the basin is essentially flat with a shallowing in the south near Danville and the beginnings of deepening in the very northern portion of the study area. (B) and (C), Kirkfieldian M5B and early Shermanian M5C sequences respectively; note subsidence in the north (right) proceeding south, possible uplift in the Danville area. (D) Mid-Shermanian M6A sequence; note subsidence of area south of Frankfort, uplift of the Frankfort area and continued deepening in the north. (E) Mid-late Shermanian M6B sequence; subsidence of the Danville area, continued uplift of the Frankfort area, and subsidence in the north. (F) Late Shermanian M6C sequence; continued subsidence in the south and north and progradation of sediments on the flanks of the Winchester-Frankfort high. (G) Late Shermanian and Edenian C1 sequence; similar to M6C sequence, continued subsidence in the south and north and progradation of sediments on the flanks of the Winchester-Frankfort high.
seismite. Additionally, the M5A TST contains an unusual echinoderm fauna, which also forms a distinct stratigraphic marker. The M5A MFS is marked by multiple mineralized hardgrounds and a facies offset from skeletal grainstone-rudstone facies of the TST to rhythmite facies of the overlying HST. An early Chatfieldian (Rocklandian) age carbon isotope excursion identifiable in many disparate sections across eastern North America (Bergström, 2001) occurs near the M5A MFS (basal Logana Member).

The M5A HST (Logana Member) exhibits a very uniform distribution of rhythmite facies across much of the study area. However, in the Danville region it grades into shaly nodular wacke-packstone facies and in the northern Cincinnati region it grades into brown laminated shale facies.

The M5A RST (lower Grier Member) is predominantly composed of shaly nodular wacke-packstone facies across the study area. Unlike later RSTs, this interval is widely traceable, suggesting deposition was uniform. The FRS is sharp, marking an offset from rhythmite to shaly nodular facies. However, there is little to no evidence of erosion at this contact.

The vertical distribution of facies within the M5A sequence into discrete packages with sharp boundaries enables easy recognition of small-scale sequences across the entire study area. Unfortunately, parasequences are difficult to resolve within rhythmite facies, which are predominant in the HST across much of the study area. Recognition of parasequences within the HST is restricted to shaly nodular facies in the Danville region.

Lateral facies distribution within the M5A sequence suggests that the Danville region was relatively shallow, but deepened rapidly toward the Frankfort region. North of Frankfort the ramp dipped little as indicated by the nearly uniform facies of those regions (Figs. 3, 8). Only in the northernmost extent of the Cincinnati region are there signs of further deepening. The Sebree
Trough in north-central Ohio was therefore likely only a very narrow bathymetric depression during this time. The pattern of lateral facies change remains consistent throughout the M5A sequence, indicating nearly uniform subsidence across the study area.

4.2 M5B sequence

The M5B is one of the thinnest sequences in the study interval, attaining a maximum thickness of 13 m in the Danville region, gradually thinning to approximately 9 m in the Cincinnati region (Fig. 3). The sequence boundary is a correlative conformity and is recognized as a sharp facies dislocation across the study area.

At its fullest development in the Swallowfield region, the M5B TST (lower-middle Grier Member) is as much as 5 m thick. However it more commonly ranges between 2–3 m in thickness. It consists of calcarenite facies in the Danville region, which grade into skeletal grainstone-rudstone facies to the north. The M5B TST contains several mineralized hardgrounds near its top that become more prominent toward the Sebree Trough (Cincinnati region). Here the surface is marked by several centimeters of reworked pyrite and phosphate horizons that identify an important condensed interval.

The M5B HST (upper-middle Grier Member) is composed of shallow shaly nodular facies with a diverse fauna typified by the domal bryozoan Prasopora, which grade into rhythmite facies in the Cincinnati region. In the Dayton region the M5B HST is represented by an interval of laminated shale facies. An unusual feature of the M5B sequence is that the RST is not
recognizable. However, this may be accounted for by the paucity of outcrop exposures of this sequence.

The vertical differentiation of facies in the M5B sequence into small-scale sequences is difficult to recognize even in the more distal facies of the Cincinnati region. Surprisingly, thin parasequences are readily recognizable across nearly the entire study area.

The lateral distribution of facies within the M5B sequence suggests a significant depth range, perhaps as great as 50 to 100 m from the shallowest facies in the Danville region to the most distal facies in the Dayton region. This pattern contrasts with subtle lateral shifts in facies characteristic of the underlying M5A sequence, suggesting minor uplift in the Danville region and increased subsidence during this time in the northern Cincinnati and Dayton regions (Fig. 9).

4.3 M5C sequence

The M5C sequence (Fig. 3) is a maximum of 19 m thick in the Danville region, thinning to 10 m in the northern portion of the Cincinnati region. The M5C sequence boundary is a subaerial unconformity in the Danville region, which becomes conformable in the Frankfort region.

The TST of the M5C sequence (lower–upper Grier Member) is composed of 2–3 m of fenestral micrite and calcarenite facies in the Danville region. These facies grade into grainstone-rudstone facies as much as 5 m thick in the Swallowfield region. In the Dayton region the TST is only recognizable as a discrete interval of thin grainstones composed of deep-water dalmanellid brachiopods (Fig. 3). Mineralized hardgrounds near the top of the M5C TST mark the MFS, and an abrupt change into rhythmite facies of the lower HST. The M5C MFS is one of the most laterally persistent and diagnostic contacts of the study interval.
In the Danville region the lower HST rhythmite facies (Macedonia bed of the Grier Member) change upward rapidly into shallow shaly nodular and calcarenite facies (highest upper Grier Member). However, in the Swallowfield region the HST is dominated by rhythmite facies, which grade into laminated shale facies in the Cincinnati region.

The forced regression surface of the M5C sequence is difficult to pick in the few sections in the Frankfort region in which it is exposed. The surface is picked within an interval of shaly nodular facies, at a horizon across which the degree of fragmented skeletal grains greatly increases and the amount of shale greatly decreases. The RST (Faulconer bed) is composed of 5–10 m of shallow shaly nodular facies in the Danville region containing *Tetradium* corals and abundant ostracodes.

The vertical facies distribution within the M5C sequence is subtle across much of the study area. Small-scale sequences and component parasequences are most readily identifiable in the southern Cincinnati region. However, recognition of the division between the HST and RST is difficult across the entire study area. The lateral distribution of facies in the M5C sequence suggests continued minor uplift in the Danville region, with continued subsidence in the Cincinnati and Dayton regions (Fig. 9).

4.4 M6A sequence

The M6A sequence is as much as 16 m thick in the northern portion of the Danville and Swallowfield regions, but averages 12 m throughout the study area, thinning to 10 m in the Cincinnati region (Fig. 3). The sequence boundary at the base of the M6A sequence is a karstic unconformity developed on fenestral micrite near the base of the Salvisa bed of the Perryville Member of the Lexington Limestone across much of the Danville region. This horizon becomes
a planar submarine erosion surface in the Frankfort and Swallowfield regions, finally recognized as a sharp facies dislocation (correlative conformity) in the Cincinnati region.

In the Danville region the TST of the M6A sequence (Salvisa bed of the Perryville Member) is composed of fenestral micrites and minor desiccation-cracked shaly dolostones, which abruptly change upward across a heavily mineralized, pitted surface to wavy/nodular stromatoporoid-bearing grainstone-rudstone facies (Cornishville bed of the Perryville Member). This contact was interpreted as a karst surface, representing subaerial exposure and subsequent transgression (sequence 2 sequence boundary of Pope and Read, 1997). However, this contact is traceable between many closely spaced outcrops into deeper water facies to the north, where it retains the characteristics described above. Therefore, we propose an alternative interpretation. We suggest that the contact at the base of the Cornishville bed represents a marine flooding surface marked by dissolution and condensation. Indeed, this contact shares many characteristics in common with numerous hardgrounds throughout the study interval. The fenestral micrite facies of the lower TST (Salvisa bed) are laterally continuous southward to the Nashville Dome, Tennessee (Mackey, 1972; Pope, 1995). However, on the northern margin of the Danville region the micrite facies grades rapidly into calcarenite and tabular grainstone-rudstone facies accompanied by loss of nearly all stromatoporoids from the upper portion of the TST (Cornishville bed). The depth sensitive stromatoporoids are abundant again in the southern portion of the Frankfort region, but absent further to the north. The entire TST becomes medium to thin-bedded grainstone-rudstone facies with numerous iron mineralized hardgrounds in the northern Swallowfield region, which persist into the Cincinnati region. The M6A TST is therefore one of the more extensive in the study interval (Figs. 3, 8). Nonetheless, in the northern portion of the Cincinnati region the skeletal grainstone-rudstone facies grade into a cluster of
fine-grained grainstones, which are replaced in the Dayton region by thin pyrite-rich shell hash horizons surrounded by dark brownish-gray laminated shales.

The MFS of the M6A sequence is marked by two heavily mineralized corrosion surfaces below an abrupt shift into cherty, rhythmite facies (Brannon Member) across most of the study area. This contact is associated with a widespread K-bentonite (basal Brannon K-bentonite of Black et al., 1965; and Kulp 1995; Quisenbury Road metabentonite of Conkin and Conkin, 1992). In the Danville and Frankfort regions the HST (Brannon Member) coarsens upward rapidly into bryozoan-rich proximal shaly nodular facies. Otherwise this interval is predominantly composed of rhythmite facies, deepening into laminated shale facies in the Dayton region. Similar to the underlying TST, the HST contains some of its most distal facies in the northernmost Danville region (Fig. 8). Cressman (1973) was the first to recognize this anomalously abrupt deepening; later Kulp (1995) and Ettensohn et al. (2002) suggested that this area was part of a narrow fault-bounded basin.

The RST of the M6A sequence (lower Sulphur Well Member) is composed of calcarenite facies from the northern Danville to southern Swallowfield regions, which grades away from these areas into proximal shaly nodular facies. The RST is typically as much as 4 m thick; however this thickness is highly variable, as the FRS at its base often forms a highly irregular, channeled surface.

Several horizons of soft-sediment deformation are traceable across the study area within the HST and RST (Kulp, 1995; Pope et al., 1997b; Ettensohn and Stewart; 2002; Fig. 3). Rhythmite facies are typically highly deformed in the upper few meters of the HST (Fig. 6d; Cressman, 1973; Jewell, 2001). This zone of seismites is one of the most widely traceable stratigraphic markers in the study interval, conspicuous in outcrop and core.
The vertical facies distribution within the M6A sequence is easily differentiated into small-scale sequences across the study area. Parasequences are identifiable in all sections except the most distal HST facies in the Cincinnati and Dayton regions.

Sequence M6A shows the most complete lateral facies transition in the study interval (Fig. 8). The total bathymetric range from subaerial exposure to basinal facies represents depth variation of perhaps >100 m and abrupt lateral facies changes record one of the steepest gradients in the study interval. This extreme lateral facies change records complimentary local uplift along the Kentucky River fault zone in the Danville region and subsidence of the Sebree Trough in the Swallowfield and Cincinnati regions (Mackey, 1972; Cressman, 1973; Ettensohn et al., 2002). Significant lateral facies changes are observed in the Frankfort region for the first time during the M6A sequence, including subsidence in the southern part of that region and uplift in the north. Regardless of the tectonic instability, small-scale sequences can be correlated confidently through all outcrop sections in this area.

4.5 M6B sequence

The M6B sequence is as much as 35 m thick in the Danville region, thinning to 22 m toward Frankfort, thickening again to 27 m toward Swallowfield, before thinning to only 9 m in the Cincinnati and Dayton regions (Figs. 3, 6). The sequence boundary is an inferred subaerial exposure surface in the Frankfort region that underlies oncolitic, ostracode micritic wackestones, and flaser bedded shales, evidence of deposition in shallow, tidally influenced environments. The sequence boundary becomes a correlative conformity away from this region. The M6B sequence shows increased lateral facies complexity similar to the M6A sequence.
In the Frankfort region, the TST (Sulphur Well Member) deepens upward through micritic wackestone and calcarenite into grainstone-rudstone facies. A meter thick succession of wavy grainstone-rudstone facies containing heavily reworked stromatoporoids and stacked hardgrounds and firmgrounds cap the TST forming the MFS. Micritic wackestone and calcarenite facies of the lower TST grade to the north and south into grainstone-rudstone facies.

The HST of the M6B sequence (Stamping Ground, Strodes Creek, and Greendale members; Figs. 3, 6) is dominated by medium dark gray highly fossiliferous, shaly nodular facies containing abundant rhynchonellid brachiopods and bryozoans. In the Frankfort region these beds grade to shallow shaly nodular wacke-packstone and calcarenite facies containing abundant stromatoporoids (Fig. 4c). To the north and south the HST becomes deep shaly nodular and rhythmite facies. The HST contains a prominent K-bentonite (here designated the Swallowfield K-bentonite), as well as locally traceable deformed horizons in outcrops of the Frankfort and Swallowfield regions (Figs. 3, 6). Parasequences of the M6B sequence are traceable in new, closely spaced outcrops across a steep facies gradient from the Frankfort and Swallowfield regions (Fig. 6).

An FRS truncates the HST to varying degrees across the study area. In the Frankfort region strata assigned to the M6B RST (lower Devils Hollow Member) were studied multiple times (Cressman, 1973; Etter, 1975; Kasl, 2001). The base of the RST is marked by dessication-cracked, green mudstone facies. This facies grades to the north and south into calcarenite facies with mud drapes and herringbone cross bedding, indicative of tidal deposition. The calcarenite facies shows extensive deformation in the Swallowfield and Cincinnati regions. Similar to the M6A sequence, the FRS of the M6B sequence is channel-form (long axis north-south directed), as exposed in roadcuts of the southern Cincinnati region.
The vertical distribution of facies within the M6B sequence shows rapid fluctuations suggesting multiple scales of cyclicity (Fig. 6). Indeed, the M6B sequence contains some of the most readily identifiable parasequences within the study interval.

The lateral distribution of facies in the M6B sequence is distinct from the underlying sequences in that the locus of peritidal deposition shifted from the Danville region to local highs in the Frankfort region (Fig. 8). Surprisingly the Danville region contains some of the thickest and most distal deposits in the study interval during this time. As with the underlying succession, the facies gradient to the north of Frankfort is steep.

4.6 M6C sequence

The M6C sequence attains a maximum of 25 m thickness in the Danville region, thinning to 8 m in the Cincinnati region (Fig. 3). In the Frankfort region, the sequence boundary is an inferred subaerial unconformity developed below a gastropod grainstone-rudstone lag deposit, containing green mudstone rip up clasts.

The M6C TST (upper Devils Hollow Member) in the Frankfort region is composed of a few meters of alternating grainstone-rudstone and calcarenite facies. The TST grades into more distal facies to both the north and south of Frankfort. To the south this transition is rapid, whereas the transition is much more gradual to the north. The MFS is best exposed in outcrops of the Swallowfield region where four tightly stacked mineralized hardgrounds form a condensed section, approximately half a meter in thickness at the contact of skeletal grainstone-rudstone facies (upper Devils Hollow Member) and overlying rhythmite facies (Bromley shale; informal unit).

The M6C HST (Bromley shale) ranges from as little as 10 m thick in the Frankfort region, where it is dominated by calcarenite and proximal shaly nodular facies, to as much as 25 m thick
in the Danville and Swallowfield areas where it grades into distal shaly nodular facies. The HST is dominated by rhythmite and laminated shale facies in the Cincinnati and Dayton regions.

The M6C FRS forms north-south directed channels, similar to the FRS of the M6A and M6B sequences, well exposed in the southern Cincinnati region (Fig. 7c). The RST of the M6C sequence (Tanglewood Member) is composed predominantly of silty calcarenite facies. Deformed beds of fine-grained calcarenite of the RST appear to record rapid progradation of near-shore sediments followed by their seismic deformation (McLaughlin and Brett, 2002, submitted). The increased proportion of silt contained within the M6C RST over previous RSTs within the study interval is suggestive of overall progradation.

The vertical facies distribution of the M6C sequence is similar to the M6B sequence, however facies are slightly more distal. Small-scale sequences, and parasequences are well developed and readily identifiable in most regions.

The lateral facies distribution of the M6C sequence displays less variation than does the M6B sequence (Fig. 8). This likely indicates that the bathymetric difference between the Sebree Trough and the adjacent Lexington Platform was becoming more subdued.

4.7 Sequence C1

The stratigraphically highest and by far the thickest sequence discussed here is the C1. It ranges from approximately 70 m thickness in the Frankfort region (Weir et al., 1984) to nearly 90 m thick in the Cincinnati region (Fig. 8). The sequence architecture is similar to that of underlying sequences, although the proportion of siliciclastics (especially mud and silt) is greatly increased. This sequence roughly conforms to the description of the C1 (first Cincinnatian sequence) designated by Holland (1993) and Holland and Patzkowsky (1996). Holland (1993)
did not specify the precise placement of the C1 sequence boundary. However, we place this boundary at the sharp base of an interval of skeletal grainstone-rudstone, which make up the Point Pleasant Formation. The C1 sequence boundary is a submarine erosion surface in the Frankfort and southern Swallowfield regions that becomes a correlative conformity away from that area.

In the Frankfort region the TST (Point Pleasant Formation) is represented by calcarenite facies. However, away from that area the TST thickens and is dominated by tabular grainstone-rudstone facies. In the Swallowfield and Danville areas the C1 TST differs from those of underlying sequences in that parasequences contain a component of shaly nodular facies, which greatly increases their thickness and ease of identification. An epibole of cyclocrinitid green algae is widely traceable in outcrops of the Frankfort and Swallowfield regions in the basal parasequence of the TST. The TST also contains several encrusted, bored, and mineralized firmgrounds and hardgrounds (Figs. 6, 7e; see Brett et al., in press, for further discussion).

Higher, in the TST, epiboles of the trilobite *Triarthrus* and the rhombiferan echinoderm *Cheirocystis* can be traced along the margin of the Sebree Trough in the Cincinnati region (Sumrall and Schumacher, 2002; Brett et al., in press; Fig. 3) into the northern Swallowfield and Danville regions. Detrended correspondence analysis, following the techniques outlined in Holland et al. (2000), of fossils present within the grainstone-rudstone facies (Point Pleasant Formation) displays a clear deepening upward trend. A thin (1–3 cm), but widespread pyrite, phosphate, and conodont-rich packstone bed at the top of the TST marks the MFS.

The C1 HST (Kope Fm) is 55 m-thick in the Frankfort region, thickening rapidly northward to 80 m in the Cincinnati and Dayton regions. Across this transect the HST grades from shallow shale and limestone facies to laminated shale facies. A complimentary increase in
the amount of shale and decreasing amount of silt is also observed across this transect (Weir et al., 1984). The C1 HST is well exposed in a large number of thick exposures across the Cincinnati and northern Swallowfield regions. In this area, detailed correlation of small-scale sequences, parasequences, and even individual beds has been accomplished for nearly every exposure across the 40 km outcrop belt (Brett and Algeo, 2001). Multiple event beds within the C1 HST have aided in correlation. One of the most spectacular is a *Triarthrus* epibole, which occurs near the middle of the HST and has been traced from the Cincinnati region into the laminated shale facies of the Dayton region (Brett and Algeo, 2001; Kohrs et al. 2002; Kohrs, 2002). The use of faunal gradient analysis in the C1 HST in the Cincinnati region (Holland et al., 2001; Miller et al., 2001) confirms the presence of small-scale sequences and parasequences (Fig. 8).

The RST of the C1 sequence (Garrard Formation) is represented by 15 to 30 m of strata dominated by very silty calcarenite facies. The C1 RST is best exposed in the Frankfort and Danville regions, but becomes increasingly difficult to recognize to the north. This interval records rapid progradation of coarser sediments during a sea level drop (RST). Similar to the TST and HST, the RST of the C1 sequence is thicker than RSTs of the previous sequences. Alternatively, this interval was interpreted as a ramp margin wedge by Pope and Read (1997a).

The vertical distribution of facies within the C1 sequence is similar to those patterns observed in underlying sequences, though much thicker. The great increase in sedimentation rate allows for much easier recognition of small-scale sequences and parasequences, as amalgamation and erosion are lessened.

The lateral distribution of facies within the C1 sequence suggests that the basin gradient decreased along the north-facing ramp into the Sebree Trough. In the Frankfort region the TST
appears to be more distal than the TSTs of lower sequences M6A,B,C, and indeed there are no associated peritidal facies. However, this interval persists as a coarse, skeletal limestone further basinward than do any of the TSTs above the M5C sequence (Fig. 8). Yet, in the Dayton region the TST grades into thin phosphatic shell hash beds. The upper HST in particular records only slight basinward deepening, with perhaps a few tens of meters of water depth change from the Frankfort region to the Dayton region.

5. Lateral facies gradients of sequences M5A–C1: Evidence for far-field tectonics

Lateral tracing of depositional sequences and component small-scale sequences and parasequences has allowed for detailed temporal mapping of facies distributions across an ancient tectonically active ramp (Fig. 9). The Danville region remained a relatively shallow platform area throughout the M5A, B, C, and into the M6A sequences, possibly exhibiting minor uplift as fenestral micrite facies become increasingly dominant features of TSTs over this time. However, the area rapidly subsided, perhaps coincident with the large number of seismites in the later M6A sequence, as evidence by the much more distal facies of the M6B and overlying sequences. A narrow area at the Danville-Frankfort region contact continuously recorded deep-water facies throughout the study interval, forming a critical interface between these two structural regions.

The Frankfort region recorded deeper water facies north of the Danville region and was in continuity with the regions north to Cincinnati throughout the M5A, B,C, and initial M6A sequences. However, rapid shallowing of the Frankfort region occurs during the later M6A sequence concomitant with deepening to the south in the Danville region. The Frankfort region
continues to show marked shallowing into the M6C sequence. During this period of rise facies gradients steepened to the north.

The Swallowfield, Cincinnati, and Dayton regions subsided continuously throughout the study interval, though the subsidence rate was variable. Subsidence was greatest in the Dayton region initially, however throughout the M5B through C1 sequences subsidence increased southward toward Frankfort. However, increasing sedimentation rates, especially in the C1 sequence appear to have overcome subsidence resulting in reducing the gradient of the ramp to the north.

The topographic instability described above seems to be associated with known basement structures (Fig. 1b) activated by pulses of thrust loading on the continental margin (Ettensohn, 2002). Seismites and K-bentonites are nonrandomly distributed throughout the study interval, coincident with periods of change in basin geometry. The northern margin of the southern tidal flat belt (Danville region) and subsequent area of rapid subsidence are closely aligned with the Kentucky River Fault Zone and the eastern expression of the Rough Creek Graben (both served as the northern margin of the Cambrian–Middle Ordovician Rome Trough). The Frankfort region high is developed parallel to a known fault zone, but is more closely aligned directly between two minor gravity anomalies in northern Kentucky and southern Ohio (Fig. 1b). Similarly, Ettensohn et al. (2002) have suggested multiple basement structures suspected of controlling trends on the Sebree Trough.

6. Implications of sequence-based reconstruction:

The results of this study suggest that the individual signatures of eustasy and tectonics are recognizable within a high-resolution sequence stratigraphic framework for seven Upper
Ordovician depositional sequences. Tracing of these sequences from shallow shelf environments of the Lexington Platform into basinal environments of the Sebree Trough reveals a distinct facies gradient. Detailed correlation suggests that vertical patterns of abrupt facies change are primarily eustatic in origin whereas gradual to relatively rapid lateral facies gradients are primarily related to differential subsidence resulting from far-field tectonic effects. The complex vertical variation in facies represents nesting of eustatically driven cyclicity of at least three scales. Subtidal depositional sequences, small-scale sequences, and parasequences, identifiable in this epicontinental succession are extremely robust, traceable across adjacent areas of penecontemporaneous subsidence and uplift, reinforcing the conclusion that allocyclic processes, probably eustasy rather than tectonism provided the driving mechanism of cyclicity.

Sediment starvation and subsequent sediment influx in response to sea level rise and fall provided a major control on vertical facies distribution in this mixed carbonate-siliciclastic succession. The energy level of a given depositional environment across the basin gradient provided a secondary modification to that signature, followed by modification by burrowing organisms.

Sequence boundaries are traceable from subaerial into submarine erosion surfaces, becoming correlative conformities in deep subtidal facies. Sequence boundaries in subtidal settings show little evidence of substantial erosion when compared to the channeling observed at the forced regression surface in many of the sequences described above, though the sequence boundary is positioned above the shallowest facies. Forced regression surfaces are responsible for the greatest amount of erosion within depositional sequences, yet evidence for subaerial exposure in most cases is absent, suggesting that the erosion is predominantly submarine, likely the expression of sediment bypass and removal during sea level fall.
Recognition of similar biofacies between grainstone-rudstone beds and shaly nodular facies indicates that both faces could accumulate at similar water depths and energy levels. In turn this suggests that sea level rise of perhaps as little as a few meters is sufficient to initiate sediment starvation and formation of grainstone-rudstones in ramp settings positioned well away from the active orogenic belt.

Far-field tectonic effects, while having a profound event signature (i.e. seismites) appear to generate low magnitude topographic changes. Periods of tectonic uplift and subsidence appear to span several million years duration in contrast to depositional sequences, small-scale sequences, and parasequences, which form over much shorter time spans.

Acknowledgements

The authors are greatly indebted to the numerous students (notably Alex Bartholomew) from the department of geology at the University of Cincinnati who accompanied us on many outings to the field area. We would like to thank the Kentucky Geological Survey (particularly Ray Daniel, Mark Eversole, and Patrick Gooding) and the Ohio Geological Survey (especially Ron Rea) for assisting us with access to drill core. Special thanks to Steve Holland, Gordon Baird, and Frank Ettensohn for creative theoretical discussions. This work incorporates aspects of masters and doctoral work of McLaughlin, masters work of Taha McLaughlin, and doctoral work of Cornell at the University of Cincinnati. It was funded in part by the Department of Geology at the University of Cincinnati, and student research grants to McLaughlin from the Geological Society of America Student Research Grants and the American Association of Petroleum Geologists Grants in Aid program. Acknowledgement is also made to the donors of The American Chemical Society Petroleum Research Fund for partial support of this research.
Thoughtful reviews by Mark Kulp, J. Fred Read, and Mike Pope helped improve the quality and focus of this manuscript.
References


Mackey, R.T., 1972. Lithostratigraphy and depositional environment of the Perryville Member of the Lexington Limestone (Middle Ordovician, Kentucky). Masters thesis, University of Kentucky, Lexington, Kentucky.


Pope, M.C., Read, J. F., 1997a. High-Resolution surface and subsurface sequence stratigraphy of the Middle to Late Ordovician (late Mohawkian–Cincinnatian) foreland basin rocks, Kentucky and Virginia. AAPG Bulletin 81, 1866–1893.


W.C. et al. (Eds.), Sea level changes: an integrated approach: Society of Economic
Paleontologists and Mineralogists Special Publication 42, 39–45.

Weir, G.W., Greene, R.C., 1965. Clays Ferry Formation (Ordovician)- A new map unit in south-
central Kentucky. US Geological Survey Bulletin 1224-B.

Weir, G.W., Peterson, W.L., Swadley, W.C., 1984. Lithostratigraphy of Upper Ordovician strata
exposed in Kentucky. US Geological Survey Professional Paper 1151-E.

history of the Trenton Limestone (Ordovician) and adjacent strata in northwestern Ohio.

Witzke, B. J., 1990. Palaeoclimatic constraints for Paleozoic palaeolatitudes of Laurentia and
Euramerica. In: McKerrow, W.S., Scotese, C.R. (Eds.), Paleozoic palaeogeography and
CHAPTER 2

Comparative Sequence Stratigraphy of Two Classic Upper Ordovician Successions, Trenton Shelf (New York-Ontario) and Lexington Platform (Kentucky-Ohio):

Implications for Eustasy and Local Tectonism in Eastern Laurentia

CARLTON E. BRETT\textsuperscript{a}, PATRICK I. McLAUGHLIN\textsuperscript{a}, GORDON C. BAIRD\textsuperscript{b}, AND SEAN R. CORNELL\textsuperscript{a}

\textsuperscript{a} H. N. Fisk Laboratory for Sedimentary Geology, Department of Geology, University of Cincinnati, Cincinnati, Ohio 45221, U.S.A.

\textsuperscript{b} Department of Geology, State University of New York-Fredonia, Fredonia, New York 14063

(submitted to Palaeogeography, Palaeoclimatology, Palaeoecology; December, 2003; published in volume 210, June, 2004)
Abstract

Comparison of the classic Upper Ordovician (Mohawkian to lower Cincinnatian; Caradoc to lower Ashgill) Black River and Trenton Groups in New York State/southern Ontario with the Tyrone-Lexington Formations exposed in the Jessamine Dome (northern Cincinnati Arch) in north-central Kentucky and southern Ohio reveals striking similarities. Previous emphasis on complex local facies mosaic has obscured widespread regional patterns. Biostratigraphy and K-bentonites provide broad constraints on inter-regional correlations; however, an allostratigraphic approach permits higher resolution correlations and a partial test of eustatic vs. strictly local tectonic models to explain stratigraphic patterns. Upper Mohawkian to lower Cincinnatian (~455-445 Ma) depositional sequences, previously recognized for the Jessamine Dome and Nashville Dome areas are correlated between the two main study areas and further refined; Chatfieldian (Rocklandian to Shermanian or Cobourgian of traditional terminology) sequences M5 and M6 of previous workers are interpreted to be composite sequences and are each subdivided into three smaller-scale sequences, which also have counterparts in the New York-Ontario strata. In turn, these correlations indicate at least partial allocyclic control on sedimentary cycles. Complex lateral variations within depositional sequences indicate that tectonically controlled patterns of basinal subsidence and uplift of crustal blocks (perhaps reflecting forebulge migration) exerted a strong influence on the local facies and motif of depositional sequences. These tectonic features, however, did not obliterate the underlying allocyclic pattern. Indeed, high-resolution sequence stratigraphy enables detailed resolution of shifting patterns of minor uplift and subsidence.
1. Introduction

The well preserved Upper Ordovician strata of eastern North America provide an outstanding testing ground for assessing the interplay of eustatic oscillations and local tectonism in an active foreland basin and cratonic platform. The degree to which each of these agents controlled sedimentation can be assessed from regional patterns of change within sedimentary cycles, but only if these are correlated at high resolution. Changes in the degree of local modification of through-going cycles, if they can be recognized, may also provide a subtle signature of events in the orogenic hinterland and their far-field tectonic effects in the craton.

The Black River and Trenton, Groups of New York State (Caradoc; Mohawkian Series; recently reassigned from Middle to Upper Ordovician; Mitchell et al., 1997) comprise some 50 to 100 m of peritidal to shallow shelf carbonates overlain by fossiliferous limestones with increasing proportions of shale; they are overlain by black Utica Shale facies of Cincinnatian age. This interval, typically exposed in the Mohawk Valley of New York State (Figs. 1–4; Hall, 1847; Beecher and Hall, 1886; Prosser and Cummings, 1897; Kay, 1937, 1953, 1960; Chenoweth, 1952; Fisher, 1965, 1980; Cameron and Mangion, 1977), is widespread in the eastern US, and southern Ontario, Canada. The coeval Tyrone, Lexington and Kope (Clays Ferry) Formations of the Jessamine Dome area (Cincinnati Arch) of north central Kentucky and southern Ohio form a comparable, well studied succession of carbonate to shale facies (Figs. 1–4; Ulrich, 1888; Ulrich and Bassler, 1914; Black et al., 1965; Cressman, 1973). Despite over a century of study, many details in both areas remain obscure, and there has been little attempt at detailed correlation between these areas. In a genetic sense it is important to crosscut artificial differences and to establish regional tie lines. One approach to this problem is to attempt to define and correlate allostratigraphic depositional sequences instead of local lithostratigraphic
Figure 1. Study area for comparative sequence stratigraphy of Upper Ordovician successions. A) outline map of eastern North America; boxes show the two areas compared in this paper. B) Jessamine Dome (southern Cincinnati Arch) in north central Kentucky, southern Indiana, and southern Ohio. Note names of important localities. C) southern Ontario, Canada, and New York State with names of major localities referred to in this paper.
units. More importantly, rocks in both areas have been subdivided and interpreted using different approaches and paradigms: in the early history of study strata were treated from the standpoint of widely correlatable stratigraphic units, somewhat akin to depositional sequences (Ulrich, 1888; Bassler, 1906; Ulrich and Bassler, 1914; Cushing et. al., 1910; McFarlan and White, 1948). Conversely, many recent authors in both areas have characterized the strata as a complex mosaic of local facies (Cressman, 1973; Cressman and Noger, 1976; Ettensohn, 1992; Joy et al., 2000; also see Davis and Cuffey, 1998, and papers therein). In actuality, strata in both areas exhibit a mixture of these characteristics. Through-going surfaces, event beds, cycles, and consistent facies dislocations provide a regionally widespread framework for detailed correlations in each area, and perhaps between these regions, as we suggest herein. However, local effects of minor uplifts and depressions have produced subtle mosaic patterns of lateral facies change within the confines of through-going allostratigraphic units. Complex facies relationships in the Upper Ordovician Lexington Limestone of the Jessamine Dome area of northern Kentucky were well documented by Cressman (1973; see Ettensohn, 1992). These result from local uplift and subsidence of small crustal blocks; Ettensohn et al. (2002) have inferred that this resulted from far-field tectonics associated with the later (Vermontian) tectophase of the Taconic Orogeny. However, emphasis on complex local facies mosaics may have obscured some widespread regional patterns. Detailed correlation permits resolution of shifting patterns of minor uplift and subsidence (see Chapter 1).

Recently, researchers attempted to divide Upper Ordovician strata of both the New York-Ontario area and those of the Nashville Dome–Cincinnati Arch into allostratigraphic intervals using a sequence stratigraphic approach (Fig. 2). First, Holland and Patzkowsky (1996, 1998) subdivided Mohawkian to Cincinnatian strata into a series of twelve large-scale, unconformity-
Figure 2. Chronostratigraphic charts, showing sequence stratigraphy, adapted from Holland and Patzkowsky (1996, 1998) for the Jessamine Dome (left), compared with chronostratigraphy and Ontario/New York (right) as inferred by Joy et al. (2000). Compare with figure 14.
bounded depositional sequences, labeled M1 to M6 and C1 to C6. Pope and Read (1997a, b) further refined some of these sequences and correlated them from the Jessamine Dome into the Taconic Basin in Virginia. These authors suggested that the sequences were widely correlative and had isochronous boundaries (Fig. 2). Conversely, Joy et al. (2000) documented a series of four unconformity-bounded sequences in the New York portion of the Black River and Trenton Groups and the overlying Indian Castle (Utica) Shale (Fig. 2). In contrast to other workers, Joy et al. (2000) argued that these sequences, especially the upper two, were largely controlled by pulses of tectonism associated with the Taconic Orogeny, and that key surfaces and bracketed systems tracts were diachronous, at least in some areas. These authors also presented evidence for major lateral changes in subsidence and bathymetry; they further suggested that water depths exceeded 500 m in the Taconic basin.

In this paper we provide the first results of an on-going detailed comparison of two coeval platform-ramp areas, within a sequence stratigraphic framework (Figs. 2, 4). The stratigraphy of the type Black River and Trenton Groups in central New York State and Ontario has recently been revised substantially based on new bio-, K-bentonite, and sequence stratigraphic interpretations (Noor, 1989; Brookfield and Brett, 1988; Melchin et al., 1994; Delano et al., 1994; Goldman et al., 1994; Mitchell et al., 1994; Lehmann et al., 1994, 1995; Armstrong, 1997; Joy et al., 2000; Baird and Brett, 2002; Brett and Baird, 2002). This provides a new basis for comparison with other areas. Moreover, our measurement of new roadcuts north of Frankfort, Kentucky, combined with the data from Cressman (1973), and Pope and Read (1997a, b) provides evidence for consistent sequence stratigraphic interpretation and detailed regional correlations in the Jessamine Dome (see McLaughlin et al., this volume).
We begin by discussing the general biostratigraphic, K-bentonite, and sequence stratigraphic basis for correlations. We then make a detailed comparison of depositional sequences (sensu Van Wagoner et al., 1988; Vail et al., 1991) in the: a) upper Black River-Tyrone, b) Trenton-Lexington, and c) basal Indian Castle Shale–Kope Shale successions in the New York-Ontario vs. Jessamine Dome areas, respectively (Figs. 1, 2). These comparisons not only emphasize the similarities of approximately coeval sequences, but also point out key differences in each area between successive packages.

While independent criteria for temporal correlation are still incompletely demonstrated, we document a series of similarities of pattern in intervals bracketed by well-characterized K-bentonites and corroborated by biostratigraphic data. This method enables us tentatively to identify the regional trends and shelf-to-basin transitions, and will ultimately contribute to understanding the paleogeographic evolution of portions of eastern Laurentia during the Taconic Orogeny.

This work ties together the two most classic North American Upper Ordovician sections: the type regions of the Mohawkian and Cincinnatian series, respectively. These results, although preliminary, indicate that depositional sequences can be correlated over geographically widespread areas and are approximately coeval. Moreover, both meter- and decameter-scale sequences appear to be laterally persistent, albeit variable across shelf to basin transects. As such, we suggest, contrary to Joy et al. (2000), that while they were locally influenced by tectonics, these Upper Ordovician depositional sequences were more strongly influenced by allocyclic and probably eustatic fluctuations. This work corroborates earlier conclusions of Pope and Read (1997b) that Mohawkian cycles in Kentucky and the southern Appalachians were produced by a hierarchy of sea level fluctuations on the order of 10–50 meters. Significantly,
Figure 3. Paleogeographic map showing full development of Taconic foreland basin and Sebree Trough during the middle Shenianian (see Figure 2 for time scale). Dashed rectangles outline study areas in New York and the central Kentucky-southern Indiana and Ohio region shown in Figure 1.
however, different sequences show dramatically differing lateral patterns across regional
transects, ranging from minor, subtle facies variations and nearly “layer cake” stratigraphy in the
Turinian to early Chatfieldian, to pronounced lateral changes, from shelf carbonates to basinal
muds during late Chatfieldian to Edenian time. Our ability to correlate key surfaces and systems
tracts regionally permits precise discrimination of tectonically induced changes in basin
topography and has implications for the evolution of the Taconic orogen and foreland basin.

2. Paleogeographic and tectonic setting

The Black River and Trenton Groups of New York State and the time equivalent Tyrone
(upper High Bridge Group) and Lexington Formations of Kentucky and Ohio record the
transition of a widespread carbonate platform into a shelf-to-basin system. In the Late
Ordovician the southeastern (present eastern) margin of Laurentia underwent a shift from passive
to active margin with the onset of the Taconic Orogeny (Dewey and Kidd, 1974; Stanley and
During the first (Blountian) tectophase of the Taconic Orogeny (Chazyan–early Mohawkian), the
southeastern margin of the Appalachian Basin (present day eastern Tennessee, and Georgia)
became an active margin (Fig. 3). Collision of an island arc at the Virginia Promontory
(Ettensohn, 1991, 1992) caused thrust loading and the formation of the Sevier Basin in
However, the Black River shallow carbonate platform persisted over much of Laurentia through
Turinian to earliest Chatfieldian time. The later Vermontian (or Taconian) tectophase (middle
Mohawkian–Cincinnatian) affected areas farther north and appears to record overthrusting of an
accretionary wedge and Amonoosuc Volcanic Arc onto the northeastern margin of Laurentia,
with collision first at the New York Promontory (Fig. 3; Ettensohn, 1991, 1992). The Vermontian tectophase appears to have been much more widespread than the Blountian with not only reactivation of the Sevier Basin, but also development of a steep-sided peripheral foredeep, the Taconic (or Vermontian) Basin, oriented NE–SW through present day eastern Quebec, eastern New York State (Cisne et al. 1982, 1984; Bradley and Kidd, 1991; Diecchio, 1991; Lehmann et al., 1994, 1995) and into central Pennsylvania, where it broadened into a large basin, the Pennsylvania Basin (Keith, 1988). As noted by McLaughlin et al. (this volume), evidence for deepened facies also appears nearly synchronously in the Sebree Trough, a NE–SW trending intracratonic basin through western Kentucky, southern Indiana, NW Ohio, which passed into the western side of the Pennsylvania Basin (Fig. 3; see Diecchio, 1991, 1993; Mitchell and Bergström, 1991; Bergström and Mitchell, 1992; Wickstrom et al., 1992; Kolata et al., 2001; Ettensohn et al., 2002). The coincidence in timing suggests that basement faults were reactivated by far-field tectonics associated with Taconian thrust loading resulting in localized subsidence (Diecchio, 1993; Ettensohn et al., 2002). These subsiding areas were bordered by a series of smaller carbonate ramps and sub-basins (Fig. 3). In particular, the Ontario, Canada-to-central New York State outcrop belt closely approximates a dip parallel (northwest-to-southeast) transect from the Trenton carbonate shelf southeastward into the Taconic Foreland Basin (Fig. 3). Similarly, the central Kentucky-to-southwestern Ohio transect (south-to-north) approximates an oblique to dip, ramp-to-basin cross-section from the Lexington Platform into the Sebree Trough.

Both the Trenton Shelf and Lexington Platform regions lay approximately along the same paleolatitude, in the tropics at ~20–25° south latitude (Fig. 3; Scotese and McKerrow, 1990; Witzke, 1990). Despite this position previous researchers have suggested, based upon
sedimentological features, that eastern Laurentia underwent a change in climate from warm and perhaps slightly arid to cooler, more humid climates during the Turinian to Chatfieldian, with concomitant faunal changes (Brookfield, 1988; Patzkowsky and Holland, 1993, 1996, 1997; Pope and Read, 1997a, b, 1998). Railsback et al. (1990) discussed evidence for deep circulation by warm, saline bottom water, which would have given basinal areas a propensity to stagnation and oxygen depletion. The entire study area lay within the southerly trade wind belt; as such, normal winds would have been southeast to northwest (modern directions). Southwesterly-directed hurricanes, originating south of the inter-tropical convergence zone (near the equator), would have impinged on these two ramp-to-basin regions with different trajectories (Witzke, 1987, 1990; Jennette and Pryor, 1993). More specifically, with a rotated northwest-to-southeast ramp-to-basin transect (Ontario–New York State), storm-winds would have blown surface waters obliquely from shallow ramp settings toward deeper basin areas. Conversely, storm-winds would have encroached onto the Ohio-Kentucky basin-to-ramp area with surface waters directed from deep to shallow waters. As a result, these disparate paleo-wind trajectories could have had significant impact and control on the distribution and development of shallow shelf carbonates in these transects. Nonetheless, the general similarity of facies and faunas suggests that these differences were not a major control on deposition.

Moreover, given a southern hemispheric position, paleocurrents within the Taconic Foreland would have developed counterclockwise gyre. As such, intermediate-to-deep, bottom currents would have a preferred flow direction, if unimpeded by surface winds, from the south-southwest out of the Sebree Trough and across the Lexington Platform and into deeper waters toward the north-northeast. Conversely, within the northern Taconic Foreland, paleocurrents would have been reinforced by surface winds and favored, at the least, strike parallel (northeast-to-
southwest) trajectories. Given that the Lexington Platform presumably was developed with a northwest-facing ramp, the potential of nutrient-rich, oxygen-depleted deep waters to upwell against west-facing ramp settings was high. Undoubtedly, this process impacted deposition in this region, more so than in the Ontario–New York region. Yet, previous studies have possibly over emphasized its significance. Our work suggests that the impact was not significant enough to obscure similarities between the two areas.

3. Facies of the Upper Ordovician (Turinian–Edenian) strata in eastern North America

Turinian and Chatfieldian (= Rocklandian, Kirkfieldian, Shermanian) strata are assigned to the Black River and Trenton Groups in New York and the coeval Tyrone and Lexington Formations in Kentucky (Fig. 4). These units display close facies similarities between the study areas indicating similar ranges of water depths and depositional environments. Micritic facies with fenestral fabrics and shaly desiccation-cracked dolostones are predominant in the Turinian (Black River) interval and occur very locally within the Chatfieldian interval. More offshore facies typical of highstand systems tracts of both areas comprise a consistent onshore-offshore spectrum of carbonate to siliciclastic facies, ranging from carbonate wacke-, pack- and grainstones, through nodular calcareous mudstones and argillaceous limestones, to olive, dark gray, and black shales (Table 1). Intervals interpreted as transgressive systems tracts appear to be represented by skeletal grainstones and packstones across broader areas; these pinch out basinward into thin skeletal hash and phosphatic beds that appear to represent strong sediment starvation. For further detailed description and discussion of these facies in the Lexington Platform, see McLaughlin et al. (this volume).
Table 1. Description and interpretation of Upper Ordovician Trenton-Lexington lithofacies; BA = benthic assemblages, standardized depth-related biofacies groupings. *BA = benthic assemblage; a standardized onshore-offshore succession of faunas. ** Water depths are best estimates based upon combined data of photic, wave base, storm, and paleontological (e.g., algal) indicators.
4. Biostratigraphy and K-bentonite event stratigraphy of the Mohawkian

4.1 Biostratigraphy

The upper Black River Group, as recognized herein for Ontario and New York State, is, by definition, assigned to the Turinian provincial Stage (essentially equivalent to the term Blackriveran). Biostratigraphic correlations of the Black River are hampered by the absence of zonally diagnostic taxa from the shallow carbonate platform during the Turinian and early Chatfieldian or Rocklandian Stages (Fig. 4). Most coral and brachiopod species are clearly facies controlled and are of relatively little time significance. However, conodonts have been obtained from some samples of these carbonates from southern Ontario (Barnes, 1964, 1967) and Manitoulin Island (Barnes et al., 1978). These were used by Barnes et al. (1978) to assign the bulk of the Black River in southwestern Ontario to conodont assemblage zone 7 of Sweet and Bergström (1971), and Sweet (1984).

Biostratigraphic control in the Trenton–Lexington interval is also weak, as this interval only encompasses parts of two major conodont zones (Leslie and Bergström, 1995a,b). Work by Schopf (1968) and Sweet and Bergström (1971) established a conodont zonal boundary, between the *Amorphognathus tvaerensis* and *A. superbus* zones low in the Denley Formation (Poland Member) in sections near Trenton Falls (Fig. 4). Sweet and Bergström (1971) Mitchell and Bergström (1991) placed this boundary within the upper-middle Lexington Limestone, approximately between the Macedonia and Brannon intervals in the Cincinnati Arch area (Fig. 4).

Graphic correlation of sections in the Mohawk Valley using fingerprinted K-bentonites, permits calibration of the conodont zonation of Trenton shelf facies with graptolite biozones of the Taconic basin (Flat Creek Formation; Mitchell et al., 1994; Joy et al., 2000). The *A.
**Figure 4.** Terminology and chronostratigraphy for the New York/Ontario and Kentucky/Ohio (Jessamine Dome) study areas. Conodont zones = conodont assemblages and zones of Bergström (1971); Ma = approximate millions of years BP; Grapt. zones = graptolite zone/occurrence; graptolite zonation of Mitchell and Bergström (1991); C. spinif. = Climacograptus spiniferus, G. pyg. = Geniculograptus pygmaeus; extended range of Climacograptus spiniferus within the Sebree Trough relative to New York sections is shown to right of column. Note the positions of key marker horizons: Seismites (S), * shows position; K-bentonites (K-b), X shows position, abbreviations for specific K-bentonite beds are: C = Capitol; D = Deicke (Pencil Cave); HF = Hounsfield; HH = High Falls; HH = Happy Hollow; Kh = Kuyahoora; M = Millbrig (Mud Cave); MH = MH K-bentonite, MH; MR = MR K-bentonite; MX = MX (Barriefield) K-bentonite; QR = Quinsberry Road; SC = Shaker Creek; SF = Sherman Fall; Sw = Swallowfield; U = unnamed; WB = Westboro. Other abbreviations: B.R. = Black River Group, Formation = Formation, Mb = Member; T = Turonian. * indicates occurrence of a probable seismite at a particular level; x indicates occurrence of a K-bentonite at a particular level.
tvaerensis–A. superbus boundary, thus appears to fall in about the middle of the Corynoides americanus graptolite Zone (Fig. 4).

In more basinal facies of the Mohawk Valley the C. americanus–Orthograptus ruedemanni zonal boundary lies high in the Flat Creek Formation (formerly “Canajoharie Shale”; Mitchell et al. 1994; Joy et al., 2000) in a dark shale, informally termed “Valley Brook shale” by Brett and Baird (2002), and slightly below the thick interval of interbedded shale and tabular calcisiltites of the Dolgeville Formation (Joy et al. 2000). Using a sequence stratigraphic approach Brett and Baird (2002) tentatively correlated the Valley Brook black shale with the relatively shaly, calcisiltites and wackestones, the Walcott Quarry beds of the Rust Formation, in the Trenton Falls area (Brett et al., 1999; Brett and Baird, 2002). The Dolgeville Formation, an interval of thin tabular bedded calcisiltites and dark shales, belonging to the O. ruedemanni to lower Climacogratus spiniferus zones, is then tentatively correlated with the upper Rust through Steuben limestones, an interval of shallow water pack- to grainstone facies in the Trenton shelf (Figs. 4, 5; Joy et al., 2000; Brett and Baird, 2002). Identification of the O. ruedemanni–C. spiniferus zone boundary in the uppermost Dolgeville or the base of the overlying Indian Castle Black Shale (Utica Shale of most earlier reports) in the Mohawk Valley indicates the Chatfieldian (Shermanian)-Edenian Stage boundary near this position. The base of the overlying Geniculograptus pygmaeus Zone lies within the Indian Castle Shale above a distinctive cluster of fingerprinted K-bentonites.

Unfortunately, recent correlations show that the first appearance of C. spiniferus is not synchronous between the New York and Sebree Trough sections in Ohio (Fig. 4). Indeed, Orthograptus ruedemanni actually may not be present in the Sebree Trough (C. Mitchell, pers. com. 2001). Furthermore, Climacograptus spiniferus occurs lower, relative to the A. tvaerensis-
A. superbus conodont boundary in the Sebree Trough than in New York (Fig. 4). C. spiniferus first appears well below the conodont boundary in the Ohio subsurface and well above it in New York. Obviously, then, one or both of these biostratigraphic zonal boundaries is diachronous. Mitchell et al. (1994) suggested that C. spiniferus immigrated into the Taconic basin substantially later than its first appearance in the Sebree Basin and elsewhere, globally.

The C. spiniferus–G. pygmaeus boundary was identified within the Kope Formation near Fort Thomas, Kentucky (Figs. 4, 5, 13) by Mitchell and Bergström (1991). Assuming that this graptolite zonal boundary is synchronous across the Taconic Basin and mid-continent region, an assumption not as yet tested, this implies correlation between the Indian Castle and Kope Formations.

4.2 K-bentonite Event Stratigraphy.

In the New York/Ontario region a number of K-bentonites have been recognized and traced throughout the Mohawkian of the New York-southern Ontario outcrop area and, to a limited extent, in the subsurface (Fig. 4; Kay, 1935, 1953; Liberty, 1969; Adhya et al., 2000; Cornell, 2001). A greenish clay bed previously identified as the MX bentonite is now thought to be equivalent to the Deicke K-bentonite, a marker bed widespread throughout the Midwest (Huff, 1983; Haynes, 1994; Kolata et al., 1996). This bed, also recognized and named the Barriefield Hill Metabentonite by Conkin (1991) in Kingston, Ontario, occurs near the base of the Lowville Formation, the classic “birdseye limestone” of the Black River Group. A second, yellowish smectitic clay, the MH horizon (of Liberty, 1969), forms a prominent marker bed about 4–7 m above the MX in sections throughout Ontario and the Watertown area of New York. Although this horizon has not been geochemically fingerprinted using phenocrysts, this marker bed shows
Figure 5. Comparative sequence stratigraphy of Upper Ordovician strata in New York/Ontario (adapted from Brent and Baird, 2003) and the Jessamine Dome of Kentucky-Ohio (modified from McLaughlin et al., this volume). The left (NY/ON) column is based on a composite section in the western Mohawk Valley near Bovinaville NY; the right column is based on a composite section in the Huron Valley area, near Troy OH. Relative geologic time scales are aligned with the New York (left) column. Interbedded patterns and faunal episodes, see Figure 4 for K-bentonite terminology. Note similarity between the columns. See text for discussion.
a clay mineralogy that indicates its origin from a volcanic ash. The MH bed is consistently located about 60–100 cm below a sharp facies change (flooding surface; see below) that marks the top of the lower (“birdseye”) member and the base of the House Creek Member of the Lowville Formation (Cornell, 2001). The exact position of the well-known Millbrig K-bentonite remains somewhat uncertain; however, a K-bentonite with similar mineralogy, the Hounsfield bentonite (Kay, 1931), was identified as the Millbrig on the basis of apatite phenocryst geochemistry by Adhya et al., (2000). There is some confusion as to where (stratigraphically) this bentonite was collected, but based on the descriptions given by Kay (1931, 1935) the Hounsfield is located at the base of the Glenburnie shale, a shaly interval, at the top of the House Creek Member and slightly below the sharp basal contact of the Watertown Formation (Figs. 4–6). Confusion was further exacerbated by the presence of several closely spaced K-bentonites within the overlying Watertown to Selby interval.

Four K-bentonites (termed the T-2, T-3, “unnamed”, and T-4 K-bentonites, Wilson, 1946; Smith et al., 1971) have been identified in the Tyrone Formation (High Bridge Group) in central Kentucky and elsewhere in the southern Appalachian (Haynes, 1994; Huff and Kolata, 1990; Kolata et al., 1996). The Deicke (T-3), “unnamed” and, Millbrig (T-4) K-bentonites of Kentucky (Pencil Cave and Mud Cave bentonites of Cressman, 1973) occur within the middle to upper Tyrone Formation (Fig. 4). These K-bentonites are considered to be equivalent to the MX, MH and Hounsfield K-bentonites of New York and Ontario (Cornell, 2001). If these correlations are correct, they imply that the Deicke, MH, and Millbrig K-bentonites, sometimes referred to as the Hagan complex, are Turinian and can not be of Rocklandian age, as implied in several previous works (Kolata et al. 1996, 1998), because these beds lie well below the base of Rocklandian strata (Selby and Napanee Formations; Kay, 1965) in the type area of that stage in New York and
Ontario (Fig. 6). Moreover, if the Hounsfield K-bentonite is equivalent to Millbrig this further indicates that the overlying Watertown is early Chatfieldian in age and not Turinian, as previously suggested by Walker (1973). This chronology simplifies previous chronostratigraphic assessments for units in all areas concerned (Cornell and Brett, 2000). However, it also indicates the Watertown (and probably the Curdsville) occupy a pre-Rocklandian part of the Chatfieldian Stage, as the latter is defined as beginning at the Millbrig K-bentonite, while the Rocklandian was defined as beginning above the Watertown at the base of the Selby Limestone (Figs. 4–6).

K-bentonites, above the Millbrig, have not been well studied in Kentucky. However, Conkin and Desari (1986) and Conkin and Conkin (1992) have identified a number of putative K-bentonites in the Lexington Limestone of the Jessamine Dome. These have potential for correlation into New York-Ontario, but, as yet, they have not been geochemically fingerprinted using phenocrysts. The primary K-bentonites of the Lexington and their possible correlations with New York are given in Figures 4 and 6. a) The Capitol metabentonite (Conkin and Desari, 1986), and the Shaker Creek metabentonite (Conkin and Desari, 1986) occur in the Curdsville Member of the Lexington Limestone (Figs. 4-6). We suggest a possible linkage with two K-bentonites recently recognized in the uppermost Watertown and Selby Formations (Rocklandian) in the vicinity of Watertown, New York, and the MR K-bentonite of Ontario (Liberty, 1969). b) A series of closely spaced (~ 60 cm apart) thin K-bentonites occur in the Macedonia beds (Grier Member), likely equivalent to the Westboro K-bentonite beds of southwestern Ohio (Schumacher and Carlton, 1991). These could correlate with the upper and lower Sherman Falls K-bentonites in the Poland Member (Denley Formation) at (Sherman Falls), West Canada Creek Trenton Falls, New York (Figs. 4, 5). The latter beds have been correlated rather widely in the
Figure 6. Comparative schematic columns of the uppermost Turinian to lower Chattfieldian (Rocklandian to Kirkfieldian) Stages in northwestern New York State and in Jessamine Dome area of Kentucky. Left column based on section Roaring Brook near Lowville, Lewis County, New York; right column based on sections in Frankfort, Franklin County, Kentucky. * G = approximate position of Guttenburg carbon isotopic excursion. KB = K-bentonite. Sequence stratigraphic abbreviations include: HST = highstand systems tract; FRS = forced regression surface; FS = flooding surface; MFS = maximum flooding surface; RST = regressive (or late highstand) systems tract; SB = sequence boundary (for composite third order sequences); SSB = small scale (4th-order sequence) sequence boundary; TST = transgressive systems tract.
Taconic basin sections. c) In the lower part of the Brannon Member shaly calcsiltites, thin clay layers, termed the Quinsberry Road metabentonites by Conkin and Conkin (1992; also see Black et al., 1965; Ettensohn et al., 2002) possibly are equivalent to the Bear Creek K-bentonite beds of southwestern Ohio (Schumacher and Carlton, 1991). d) The Happy Hollow bed of Conkin and Conkin (1992) occurs low in the Sulphur Well Member. This bed could also correlate with the High Falls K-bentonite located at the base of a shaly interval presently assigned as the basal unit of the Rust Formation in the Mohawk Valley of New York (Brett and Baird, 2002). e) At least one locally developed K-bentonite in the upper Stamping Ground Member, herein termed Swallowfield, slightly below its contact with the Strodes Creek Member (terminology as modified by Taha McLaughlin, personal communication.; see Brett et al. 2002). This bed may tie with one of the patchy K-bentonites in the upper portion of the Mill Dam Member of the Rust Formation in gorge of West Canada Creek at Trenton Falls. We have not, as yet, identified K-bentonites in the Kope Formation, despite the abundance of ash beds in presumably coeval lower Indian Castle Shale in New York. Possibly any K-bentonites here have been mixed with muds by bioturbation, whereas an absence of burrowing in black shales of the Indian Castle may have aided in preservation of bentonites as discrete event beds.

Thus, in summary, the temporal correlation of the Lexington-Kope Formations with the New York Trenton-Indian Castle is presently imprecise, although both the base and top of the study interval are fairly securely tied to the New York sections.

5. General sequence architecture

Several levels of cyclicity are recognizable in the Upper Ordovician mixed carbonate-siliciclastic strata (see Figs. 5, 6 for examples). The smallest correlatable cycles are meter-scale
shale-limestone successions, interpreted as parasequences (*sensu* Van Wagoner et al., 1988). In subtidal shelf facies emphasized herein, these cycles commence with thin-bedded calcisiltites/lutites and shales and pass upward into bioturbated nodular to wavy-bedded wacke- and packstones and finally into amalgamated pack- and grainstone (Brett and Baird, 2002). It is not clear whether meter-scale cycles identified in portions of the sections can be correlated between regions and therefore these parasequences will not be further considered in this paper. However, detailed tracing within regions strongly suggests that these parasequences are at least regionally widespread (Brett and Baird, 2002; Brett et al., in press). Larger discontinuity bounded depositional sequences of at least two orders of magnitude are also recognizable. Decameter-scale sequences, show motifs comparable to larger sequences, discussed below, including thin analogs of transgressive and highstand systems tracts. They are typically 5 to 15 m thick, and are thought to record depositional cycles of a few hundred thousand years, comparable to fourth-order cycles of Vail et al. (1991). Larger scale sequences have thicknesses of 10s of meters and inferred durations of about 1–2 million years, falling within the envelope of third-order sequences (Vail et al., 1991). They are composite in that most show two or more smaller-scale sequences. Major sequences recognized herein are subdivisions of the third-order sequences M5 and M6 recognized by Holland and Patzkowsky (1996, 1998) and Pope and Read (1997a, b). They have similar in thickness and temporal magnitude to third order “parasequence sets” previously recognized by (Pope and Read, 1997b). We agree with Pope and Read that these probably represent cycle durations of approximately one million years and thus should be considered as 3rd-order cycles. However, in some instances, we have subdivided these packages differently. For example, Pope and Read considered many thick grainstones within the Lexington Limestone to be upper portions of shallowing-upward successions (parasequences). Conversely,
we have found widely traceable erosional disconformities at the bases of several such grainstones that we interpret as sequence boundaries and also recognize a subtle upward deepening condensation pattern within these limestone successions and therefore interpret them as the basal portions of transgressive systems tracts. In the following, we discuss these interpretations in more detail; also see McLaughlin et al., this volume).

Large, composite, and smaller scale sequence boundaries are marked in most study sections by a sharp contact at which significant erosion or at least a facies dislocation occurs (see, for example, Fig. 6). These surfaces may actually be erosion/transgression (E/T) surfaces, wherein the sequence boundary and transgressive ravinement surfaces are merged to form one surface. Such surfaces are laterally extensive and on a regional scale can be seen to overstep underlying truncated beds.

Lowstand deposits (LST) are generally lacking in shallow shelf Black River and Trenton facies, but may be present in basinal facies as bundles of allodapic carbonates intercalated with dark shales (e.g. Dolgeville Formation in the Mohawk Valley of New York; see Baird and Brett, 2002).

Thin, glauconitic lag beds followed by retrograding successions of clean, fenestral micrites to fossiliferous wackestones occur in the transgressive systems tracts (TSTs) of shallow water areas, especially in the Black River/Tyrone sequences. Intervals interpreted as TSTs in Trenton-Lexington shallow subtidal facies are marked by widespread intervals of pelmatozoan-brachiopod pack- or grainstone (and commonly rudstones). We suggest that the clean carbonate nature of the TSTs reflects sequestering of siliciclastics during rising sea level. In the past, many of these skeletal limestones were interpreted as the caps of shallowing upward parasequences (Pope and Read, 1997a, b, 1998). However, pack- and grainstones of the TST may be
distinguished from those of the underlying late highstand, or regressive systems tract by their more intact preservation of fossils, and poorer sorting. Highly abraded, cross-bedded calcarenites are more typical of RSTs. While the putative TSTs may appear merely massive and homogeneous, careful examination shows evidence for increased condensation and deepening upward within the carbonates. Nearly all of the skeletal limestones inferred to represent TSTs show stacking of mineralized hardgrounds and increasing content of reworked nodules, phosphatic, glauconitic and/or chamositic grains toward their upper contacts, indicating increased sediment starvation. Quantitative gradient analyses of fossil assemblages also reveal distinct vertical faunal changes, indicative of upward deepening (P. McLaughlin, unpublished data). Meter-scale parasequences within these limestones exhibit subtle retrogradational patterns, consistent with their interpretation as transgressive systems tracts. Moreover, TST grainstones are more consistent in thickness and more regionally persistent than inferred late HST beds and overlie surfaces of regionally angular discordance. Hence, we argue that the TST grainstones may be slightly diachronous and formed as transgressive blankets of skeletal debris during initial sea level rise.

Major flooding surfaces at the tops of the TST grainstones are marked by abrupt shifts to condensed, finer-grained (and presumably deeper, but not deepest, water) facies (see Figs. 5, 6, 8A). In most cases these contacts occur at hardgrounds/corrosion surfaces with pyritic or phosphatic coatings; see detailed discussion of similar corrosion surfaces within condensed intervals by Loutit et al. (1988) and Baum and Vail (1988). Some previous workers (e.g. Titus and Cameron, 1976) have apparently interpreted these surfaces as subaerial unconformities (sequence boundaries), but this seems unwarranted as they show evidence for deepening and a high degree of condensation. Thus, these sharp contacts represent flooding surfaces, commonly
with evidence of submarine erosion/corrosion. These contacts are interpreted, herein, as surfaces of maximum sediment starvation (SMS; sensu Baum and Vail, 1988) associated with maximal rates of sea level rise. Such sediment starvation may result from essentially no siliciclastic input, combined with deepening that is sufficiently rapid and/or intense to inhibit carbonate production (Baum and Vail, 1988). The corrosion surface on top of the limestone may represent a considerable amount of time with no sedimentation, as there is evidence for a strong facies shift from relatively shallow water carbonate facies to offshore, shaly sediments. This degree of water depth change would seemingly require many thousands of years, for which there is no sedimentary record. Typically, SMS contacts are overlain by a few decameters to a few meters of calcareous shale and fine-grained, argillaceous limestones commonly with hardgrounds and corrosion surfaces. These thin intervals are considered to mark a condensed interval, rather than an entire TST, as apparently inferred by some previous workers (e.g., Pope and Read, 1997b). Sharp contacts at the tops of these intervals may be marked by thin concentrations of comminuted fossil fragments and phosphatic or pyritic debris. These coincide approximately with peaks on gamma ray profiles noted by Pope and Read (1997a) and used by them to infer maximum flooding zones and we consider them to record the position of deepest water conditions and associated sediment reduction.

Early highstands in proximal sections are characterized by a few meters of thin, wavy bedded pack- to grainstones with shaly partings (Fig. 5). In more down-ramp sections they consist of wackestones, fine grained calcisiltites, calcilutites, and shales, or simply dark, organic-rich shales. Highstands may show general bed thickening- and coarsening-upwards trends; meter-scale cycles show aggradational to progradational patterns. The intervals we identify as typical highstands, show a high proportion of mud and silt, indicating a renewed influx of these
terrigenous sediments during times of stable to slightly falling sea level. The content of coarser siliciclastics appears to be slightly to markedly elevated in later highstands or regressive systems tracts (see below).

A notable feature of many Trenton–Lexington sequences is a sharp facies dislocation within the later highstand. This sharp and locally irregular and erosional contact within late highstand is considered to represent a forced regression surface (FRS), submarine erosion associated with a rapid drop in sea level (see Plint and Nummedal, 2000). This surface is sharply overlain by coarser skeletal wacke- to grainstone beds, in some cases with abundant siliciclastic silt that exhibit a shallowing-upward (progradational) pattern. In proximal sections of the Lexington Formation the grainstones-packstones may grade into fenestral micrites with minor caps of desiccation cracked, greenish shaly dolostones (Fig. 6). This interval, herein identified as a late highstand or regressive systems tract (RST), occurs between the FRS and the overlying sequence boundary. Because these beds show an abrupt base and shallowing upward internal motif, even where the interval is thin and not evidently progradational, we prefer to use the term regressive systems tract. This interval is, in turn, overlain sharply by coarse skeletal packstones and grainstones of the TST of the next sequence.

Hence, we recognize a total of three systems tracts in most Trenton-Lexington depositional sequences: transgressive (TST), including an upper, condensed section, (early) highstand (HST), and late highstand or regressive systems tracts (RST). Moreover, many of the third-order highstands can be considered as composite sequences, being comprised of two or more, smaller packages with sequence-like rather than parasequence motifs (see McLaughlin et al., this volume, for discussion of high frequency cyclicity in the Lexington Formation).
Figure 7. M4-M5A boundary sections in (A,B) New York/Ontario and (C, D) the Jessamine Dome, Kentucky. (A) Erosional channel in top of House Creek Limestone infilled with Watertown Limestone, note Phytopsis burrows, apparent M4-M5 sequence boundary; East Canada Creek at Inghams Mills, NY. (B) Section of upper House Creek (Black River Group) and Watertown Formations showing position of sequence boundary and of Hounsfield K-bentonite; roadcut on NY Rte. 54 near Brownville, NY. (C, D) Section of upper Tyrone Formation (High Bridge Group) and overlying Curdsville Member of Lexington Formation, showing position of sequence boundary and of Millbrig K-bentonite; (C) Detail of M4-M5 sequence boundary showing light gray Tyrone limestone overlain by Curdsville Limestone Member of Lexington Limestone; note Millbrig (Mud Cave) K-bentonite in crevice occurs just below the erosive sequence boundary; (D) Overview of Tyrone and Curdsville units; roadcut on KY Rte. 34 at Marcellus, KY.
6. Comparison of depositional sequences between the Trenton shelf and Lexington Platform

In the following sections we briefly discuss the major sequences as presently recognized in the upper Black River and Trenton Limestone to Indian Castle Shale, in New York State and Ontario, and in the coeval, upper Tyrone, Lexington Limestone, and Kope Formations of the northern Cincinnati Arch area (refer to Fig. 1 for locations). We note that certain of the sequence boundaries overlap with those previously recognized by Holland and Patzkowsky (1998) and Pope and Read (1997a, b) in their work in the Nashville Dome, Jessamine Dome, and southern Appalachians, respectively (Fig. 2). However, we also subdivide their larger sequences M5 and M6 into several depositional sequences of inferred 3rd-order status (Fig. 5).

Recently, we have made detailed studies of the upper Black River, Trenton and Indian Castle (upper Utica) Shale and their lateral correlatives in southern Ontario, northern New York, and the Mohawk Valley in central New York (Cornell, 2001; Brett and Baird, 2002; Baird and Brett, 2002) using a combination of intermediate to small scale cycles, condensed beds, and event beds, including K-bentonites to establish more detailed correlation within existing biostratigraphic framework. We have also established a series of depositional sequences that appear to be relatively isochronous within the constraints provided by the few fingerprinted K-bentonites and conodont–graptolite biostratigraphy. Sequences appear to be correlative from central New York into southern Ontario. Some portions of the section, particularly the basal and upper units, are better constrained than the middle portion of the Trenton-Lexington succession.

The New York-Ontario depositional sequences relate to sequences previously identified in the central-southern Appalachians and Cincinnati Arch, by Holland and Patzkowsky (1996): Mohawkian M5 to M6 and Cincinnatian C1 (Figs. 2, 5). Comparison of approximately coeval
intervals (based on conodont-graptolite biostratigraphy) in north central Kentucky and the classic Trenton sections of central New York reveal striking similarities in facies (summarized in Fig. 5 and Tables 2-4). In both of these areas we were able to subdivide the previously recognized sequences M5 and M6, each into a series of three smaller sequences. Nonetheless, we do recognize that Groups of these sequences may be clustered into the still larger composite sequences previously recognized; hence, we retain the M5, M6 numbering scheme of Holland and Patzkowsky, but divide each into three component third order sequences, viz. M5A-C and M6A-C.

At present, we have not fully tested all the sequence-based correlations suggested in the following section or Figure 5. However, the similarity of patterns of major and minor sequences between the upper and lower constrained endpoints strongly suggest that these are local manifestations of allocycles that can be correlated widely, at least in eastern Laurentia. We present the following as a preliminary comparison of pattern and suggest, based on these similarities, that the sequences discussed are both widespread and correlative between the study areas. This comparison also gives rise to a series of new correlational hypotheses, each of which is ultimately subject to testing on independent grounds, if additional K-bentonites can be extracted and processed from the New York and Kentucky sections. Based on present constraints, however, we are able to discern a comparable number of sequences in both areas and find that they do have substantial shared similarities.
Figure 8. M5A highstand facies; Napanee Formation of New York-Ontario and the Logana Member of Lexington Limestone of the Jessamine Dome. A) Thin-beded micrites and calcisiltites of Selby and Napanee Formations, with Selby resting sharply at surface of maximum starvation on Watertown Formation, Black River at Boonville, NY. (B) Logana Member, note rhythmic calcisiltites and shales very similar to Napanee and division of Logana into lower (L) and upper (U) rhythmic calcisiltite-shale intervals separated by middle (M) pack- to grainstone bed, Rte. 127 just north of Frankfort, KY. Roadcut along Rte. 127 north of Frankfort, Kentucky.
In the following sections we outline the late Turinian to Edenian depositional sequences of the Trenton Shelf and Taconic Foreland Basin in central New York State and southern Ontario and attempt to tie them to those recognized in the Lexington Limestone and Kope Formation of the Lexington Platform to Sebree trough in northern Kentucky and southern Ohio; for details, see Tables 2-4. A more detailed analysis of the similarities and differences among probable coeval sequences will be presented elsewhere.

6.1. Late Turinian: Sequence M4

The recognition of a series of correlated K-bentonites and other distinctive marker beds in the upper Turinian of New York, Ontario, and central Kentucky, as noted above, permits detailed correlation and revised comparison of sequence stratigraphic pattern within this interval among these areas (Conkin and Desari, 1986; Huff and Kolata, 1990; Figs. 4–6). Specifically, the Deicke (or Pencil Cave), unnamed K-bentonite, and Millbrig (or Mud Cave) K-bentonites in the Tyrone Limestone of the Jessamine Dome (Cressman, 1973; Huff and Kolata, 1990), are believed to correlate with the MX (Barriefield), MH, and Hounsfield beds, respectively, of the Lowville Formation in New York and Ontario (Figs. 4, 6).

In the late Turinian Stage, intervals bracketed by the Deicke and Millbrig K bentonites demonstrate that both regions were sites of extensive shallow lime mud-dominated platforms. Both the upper Black River Group and the coeval Tyrone Formation show comparable thicknesses of cyclic peritidal facies. In both regions a flooding surface slightly above an unnamed K-bentonite (designated MH in New York–Ontario) juxtaposes strongly burrowed, *Tetradium*-rich wacke- to packstones over dove gray fenestral micrite facies (Fig. 6). Together, these divisions appear to record the TST and HST, respectively, of the M4 depositional
SEQUENCE M5C
MID SHERMANIAN: C. americanus Zone
(5C LATE HIGHSTAND SYSTEMS TRACT)
upper Poland-Russia, wavy-
laic packstone and grainstone
MID SHERMANIAN: C. americanus Zone
(5C EARLY HIGHSTAND SYSTEMS TRACT)
basal Poland Mmb (Densley Formation); shale,
thin-bedded calcisiltites, wackestones, shales
MID SHERMANIAN: C. americanus Zone
(5C TRANSGRESSIVE/CONDENSED INTERVAL)
Rathbun grainstones, unnamed grainstones
large Prasopora epibole

SEQUENCE M5B
KIRKFIELDIAN-LOWER SHERMANIAN: L. C. americanus Zone
(M5B EXTENDED TRANSGRESSIVE SYSTEMS TRACT)
Sugar River Formation, shaly middle upper Grier Member; shaly
nodular limestones, rich in nodular limestones, rich in
Prasopora simulatrix Prasopora simulatrix
KIRKFIELDIAN (LOWER CHATFIELDIAN)
(M5B TRANSGRESSIVE SYSTEMS TRACT)
Kings Falls-Kirkfield Formation lower Grier Member
brachiopod-bryozoon brachiopod-bryozoon
pack-and grainstone of pack-and grainstone of
calcs of is, basement rock
calcs of is, basement rock

SEQUENCE M5A
ROCKLANDIAN/KIRKFIELDIAN (LOWER CHATFIELDIAN)
(M5A: LATE HIGHSTAND SYSTEMS TRACT)
upper Napanee Formation upper Logana Member
shallows upward to shallows upward to
thin pack- grainstones thin pack- grainstones
ROCKLANDIAN (LOWER CHATFIELDIAN)
(M5A: EARLY HIGHSTAND)
Napanee Formation Logana Member
rhythmic, thin bedded shale rhythmic, thin bedded shale
calcsilitite and packstone; middle calcisilitite and packs; middle.
dalmanellid grainstone contains Guttenburg 13C excursion
contains Guttenburg 13C excursion
ROCKLANDIAN (LOWER CHATFIELDIAN)
(M5A: MAXIMUM FLOODING ZONE)
abrupt upper contact of Selby Limestone abrupt upper contact of Selby Limestone Member
multiple hardgrounds stacked hardgrounds
in upper Selby in upper Curdsville
ROCKLANDIAN (LOWER CHATFIELDIAN)
(M5A: TRANSGRESSIVE SYSTEMS TRACT)
Watertown/Selby Formations Curdsville Limestone Member
massive, bioturbated thick bouded bioturated
cherty packstone rarely cherty packstone
with corals with corals, abundant
nautiloids
LOWER ROCKLANDIAN (BASAL CHATFIELDIAN)
(M5A: SEQUENCE BOUNDARY)
sharp sequence bounding sharp sequence bounding
erosion surface erosion surface
below Watertown Ls. below Curdsville Limestone Member
Watertown overlies Curdsville overlies
fenestral micrites above fenestral micrites above
Hounsfield k-bentonite Millbrig k-bentonite

Table 2. Comparison of Rocklandian to lower Shermanian sequences and systems tracts in the lower portion of the Trenton Group in central New York State vs. the lower part of the Lexington Formation of the Jessamine Dome, central Kentucky.
sequence. Moreover, the number of meter-scale cycles, and their relative thicknesses are roughly comparable in both areas though slightly thinner on average in the Jessamine Dome. In both the Lexington Platform-Sebree trough (KY, Ohio) and the Black River shelf to Kingston Embayment (New York–Ontario), lateral changes in the depositional sequences are relatively subtle and, at most, show transitions in the HSTs from fenestral micrites to bioturbated wackestones. This evidence from lateral facies variation indicates that depositional topography was very subdued during this time with total relief only amounting to a few meters based on paleobathymetric indicators. Strong allocyclic control on cycle generation is indicated.

6.2. Sequence M5A: Early Chatfieldian (Rocklandian)

The basal Lexington and Trenton carbonates show strong similarities in the two regions (Figs. 6-9; Table 2). The Watertown Limestone previously has been assigned to the upper Black River Group (Fisher, 1977; Walker, 1973; Cameron and Mangion, 1977), but the erosion surface at the base of this unit regionally truncates underlying beds, indicating a discontinuity between the units. This important contact is most dramatically developed in the central Mohawk Valley of New York. A channeled erosion surface at the base of the Watertown Formation, along east Canada Creek at Inghams Mills, truncates more than 3 meters of Lowville Formation. Northward into the Black River Valley the sequence boundary is more subtle, as less material is removed below it. However, it can be recognized through the superposition of the Watertown on the shaly-stromatolitic micrites (Weaver Road beds of Cornell, 2001) or grainstones and shales (Glenburnie) in the underlying Black River Group (Figs. 6, 7 a, b).

The unconformity between the Tyrone and Lexington Formations (Curdsville Member) in Kentucky has been identified consistently as a major erosion surface (Cressman, 1973; Pope and
Read, 1997a, b; Fig. 6), and termed the M5 sequence boundary by Holland and Patzkowsky (1996, 1998). We concur that it represents a significant sequence boundary and further infer that it is equivalent to the erosional unconformity at the base of the Watertown Formation in New York and Ontario (Cornell, 2001).

In both areas, massive, locally cherty skeletal wacke- to packstones (Watertown Formation-Selby Member in NY and Ontario Curdsville Member of Lexington Limestone in Kentucky–Ohio) sharply and unconformably overlie bioturbated wackestones and locally cut out the Millbrig K-bentonite and upper Tyrone or Lowville, respectively (Cressman, 1973; Pope and Read, 1997a; Cornell, 2001; Fig. 6). The Curdsville also contains several of the same fossils as the Watertown-Selby succession in New York, including the brachiopods Sowerbyella curdsvillensis and Hesperorthis tricernaria. Present evidence strongly favors interpretation of the Watertown and Curdsville as coeval and early Chatfieldian (post-Turinian) in contrast to their previous assignment to the Turinian and Kirkfieldian, respectively, by Holland and Patzkowsky (1996).

In Ontario and northwestern New York, the Selby rests sharply on the underlying Watertown at an inferred surface of sediment starvation. In the central Mohawk Valley the shaly calcisiltites of the Napanee, in turn, overlie the Selby–Watertown interval with unconformity (Figs. 6, 8a; Cameron and Mangion, 1977). This sharp corrosion surface was interpreted previously as a sequence boundary separating Black River and Trenton sequences. However, we infer that this surface records submarine corrosion, including dissolution, associated with a maximum flooding surface, rather than a sequence boundary (Figs. 6, 8). Likewise, the sharp upper contact of the Curdsville Member in Kentucky is inferred to represent a maximum flooding surface, with shales
and thin calcilutites/calcisiltites of the Logana Member, interpreted as the highstand systems tract of sequence M5A (Figs. 6, 8b; Table 2).

The highstand intervals of the M5A sequence, Napanee Formation of New York–Ontario and Logana Member of Kentucky, bear striking resemblances to one another. In both areas these intervals are dominated by rhythmically-interbedded calcilutite/calcisiltite and dark shale facies. Both show a thick, amalgamated, middle bed of dalmanellid-rich grainstone that may represent a minor condensed interval (Figs. 6, 8).

The Logana and Napanee both show evidence for a positive carbon isotopic excursion near their bases (Bergström et al., 2001). This is interpreted as the Guttenburg excursion, originally recognized in the upper Mississippi Valley; Ludvigson et al., 1999). If so, this suggests that the Logana and Napanee are coeval, and Rocklandian (Early Chatfieldian) not Kirkfieldian in age. Outcrop and subsurface studies show the persistence of the distinctive rhythmically bedded shales and fine grained limestones over much of Ohio and in Southern Tier New York subsurface (based on logging of a drill core from Chemung County, NY; unpublished data). Based on recent biostratigraphy of Melchin et al. (1994), the Napanee also appears to correlate with shaly, thin-bedded facies of the middle Bobcaygeon Formation in Ontario (Armstrong, 1997). Moreover, acritarch and chitinozoan assemblages from that unit indicate deeper shelf environments (Melchin et al., 1994). Thus, in contrast to most previous interpretations of the Napanee Formation as shallow lagoonal sediments (Titus and Cameron, 1976), we infer that both it and the coeval Logana record deep offshore shelf sediments, that show an aggradational to slightly progradational pattern. The sharp base of the Napanee-Logana represents a major, synchronous, probably eustatic deepening pulse that affected both areas. Broad similarities of the typical Napanee–Logana facies also indicate that depositional topography remained relatively
Figure 9. Sedimentary structures and fauna of the M5B sequence in New York. (A) Type section of Kings Falls Limestone along Deer River near Kings Falls, Copenhagen, New York; note the prominent ledge at the base of the Kings Falls Formation overhanging thin bedded limestones and shales of the Napanee Formation. (B) Bedding plane covered with large specimens of the bryozoan Prasopora simulatrix (lens cap for scale); upper Sugar River Limestone, City Brook (Wolf Hollow Creek) north of Middleville, New York.
subdued, although minor local uplifts, e.g., near Middleville, New York, and south of the Kentucky River Fault Zone produced local shallowing, as evidenced by abrupt transitions to shaly, nodular wacke- to packstone facies (see McLaughlin et al., this volume).

6.3. Kirkfieldian-Early Shermanian: Sequence M5B

In both the Lexington Platform and Trenton shelf, the Rocklandian rhythmite facies are succeeded rather abruptly by skeletal grainstone facies of probable Kirkfieldian Age (basal Grier Member in KY and Kings Falls-upper Bobcaygeon in NY and Ontario; Table 2; Fig. 9a). This sharp facies dislocation marks the M5B sequence boundary in the Jessamine Dome area. This surface is locally erosional in central New York, where the basal Kings Falls contains clasts of Lowville lithologies, as well as Grenville basement rocks. This indicates erosion of locally uplifted highs during a lowstand (Bradley and Kidd, 1991). The overlying Sugar River Formation (NY) and middle Grier are similar wavy bedded pack- to grainstone facies particularly noted for beds containing the domal bryozoan Prasopora (Titus and Cameron, 1976; Fig. 9b). Echinoderm skeletal pack- and grainstones are overlain by shaly nodular wacke- to packstones with abundant Prasopora in the south-central New York subsurface, central Kentucky, and central Pennsylvania (Cuffey, 1997). This again indicates relatively uniform topography over a substantial tract of eastern Laurentia.

Only rather minor lateral changes occur in the TST of this sequence going down ramp into the incipient Taconic Foreland Basin and Sebree Trough, respectively (see Fig. 14). However, the highstand facies of the Sugar River and correlative Glenns Falls Formations in eastern New York show an abrupt upward change to dark Flat Creek Shale (Mitchell et al., 1994; Joy et al., 2000). The sharp contact between these units shows phosphatic-pyritic staining and clasts at
upper Prospect Quarry: “Locust Creek” beds
member of Rust Formation: Millersburg Member, Lexington Ls.
platy to wavy packstones: lower Devils Hollow Member
packstones and shale: locally deformed lower Devils Hollow Member
laminated fine grainstones: heavily deformed locally deformed
ball-and-pillow deformed channel fills: locally deformed
packstones and shale: deformed Spillway beds locally deformed
UPPER SHERMANIAN: O. ruedemanni Zone?
(M6A: SEQUENCE BOUNDARY)
erosive contact at base: sharp contact at base of Sulphur Well Member
major deformed zone: local brecciation of upper Russia clasts: Brannon clasts
wavy bedded packstones: pyritic corrosion surface at top
rudstone: massively deformed locally deformed
UPPER SHERMANIAN: upper C. americanus Zone
(M6A: TRANSGRESSIVE SYSTEMS TRACT)
abrupt upper contact: abrupt upper contact of Cornishville Bed; pyritic packstones (stromatoporoid bed)
condensed bed: of Cornishville Bed; pyritic packstones (stromatoporoid bed)
platy to wavy packstones: locally deformed lower Devils Hollow Member
packstones and shale: locally deformed
laminated fine grainstones: heavily deformed locally deformed
UPPER SHERMANIAN: upper C. americanus Zone
(M6B LATE TRANSGRESSIVE-EARLY HIGHSTAND)
red algae-bryozoan-stromatoporoid: red algae-bryozoan-stromatoporoid"; crude packstones and nodular wackestone/shale: wavy bedded packstones
pack- and equivalent grainstone: pyritic corrosion surface at top
Shannon clasts: of Cornishville Bed; pyritic packstones (stromatoporoid bed)
local breccia of upper Russia clasts: Brannon clasts
wavy bedded packstones: pyritic corrosion surface at top
UPPER SHERMANIAN: upper C. americanus Zone
(M6B SEQUENCE BOUNDARY)
local breccia of upper Russia clasts: Brannon clasts
major deformed zone: local breccia of upper Russia clasts: Brannon clasts
wavy bedded packstones: pyritic corrosion surface at top
rudstone: massively deformed locally deformed
UPPER SHERMANIAN: upper C. americanus Zone
(M6B: MAXIMUM FLOODING ZONE)
abrupt upper contact: abrupt upper contact of Cornishville Bed; pyritic packstones (stromatoporoid bed)
condensed bed: of Cornishville Bed; pyritic packstones (stromatoporoid bed)
platy to wavy packstones: locally deformed lower Devils Hollow Member
packstones and shale: locally deformed
laminated fine grainstones: heavily deformed locally deformed
UPPER SHERMANIAN: upper C. americanus Zone
(M6B: MAXIMUM FLOODING ZONE)
abrupt upper contact: abrupt upper contact of Cornishville Bed; pyritic packstones (stromatoporoid bed)
condensed bed: of Cornishville Bed; pyritic packstones (stromatoporoid bed)
platy to wavy packstones: locally deformed lower Devils Hollow Member
packstones and shale: locally deformed
laminated fine grainstones: heavily deformed locally deformed
UPPER SHERMANIAN: upper C. americanus Zone
(M6B LATE TRANSGRESSIVE-EARLY HIGHSTAND)
abrupt upper contact: abrupt upper contact of Cornishville Bed; pyritic packstones (stromatoporoid bed)
condensed bed: of Cornishville Bed; pyritic packstones (stromatoporoid bed)
platy to wavy packstones: locally deformed lower Devils Hollow Member
packstones and shale: locally deformed
laminated fine grainstones: heavily deformed locally deformed
UPPER SHERMANIAN: upper C. americanus Zone
(M6B: SEQUENCE BOUNDARY)
local breccia of upper Russia clasts: Brannon clasts
major deformed zone: local breccia of upper Russia clasts: Brannon clasts
wavy bedded packstones: pyritic corrosion surface at top
rudstone: massively deformed locally deformed
UPPER SHERMANIAN: upper C. americanus Zone
(M6B: TRANSGRESSIVE SYSTEMS TRACT)
abrupt upper contact: abrupt upper contact of Cornishville Bed; pyritic packstones (stromatoporoid bed)
condensed bed: of Cornishville Bed; pyritic packstones (stromatoporoid bed)
platy to wavy packstones: locally deformed lower Devils Hollow Member
packstones and shale: locally deformed
laminated fine grainstones: heavily deformed locally deformed
UPPER SHERMANIAN: upper C. americanus Zone
(M6B: MAXIMUM FLOODING ZONE)
abrupt upper contact: abrupt upper contact of Cornishville Bed; pyritic packstones (stromatoporoid bed)
condensed bed: of Cornishville Bed; pyritic packstones (stromatoporoid bed)
platy to wavy packstones: locally deformed lower Devils Hollow Member
packstones and shale: locally deformed
laminated fine grainstones: heavily deformed locally deformed
UPPER SHERMANIAN: upper C. americanus Zone
(M6B LATE TRANSGRESSIVE-EARLY HIGHSTAND)
abrupt upper contact: abrupt upper contact of Cornishville Bed; pyritic packstones (stromatoporoid bed)
condensed bed: of Cornishville Bed; pyritic packstones (stromatoporoid bed)
platy to wavy packstones: locally deformed lower Devils Hollow Member
packstones and shale: locally deformed
laminated fine grainstones: heavily deformed locally deformed
UPPER SHERMANIAN: upper C. americanus Zone
(M6B: SEQUENCE BOUNDARY)
local breccia of upper Russia clasts: Brannon clasts
major deformed zone: local breccia of upper Russia clasts: Brannon clasts
wavy bedded packstones: pyritic corrosion surface at top
rudstone: massively deformed locally deformed
UPPER SHERMANIAN: upper C. americanus Zone
(M6B: TRANSGRESSIVE SYSTEMS TRACT)
abrupt upper contact: abrupt upper contact of Cornishville Bed; pyritic packstones (stromatoporoid bed)
condensed bed: of Cornishville Bed; pyritic packstones (stromatoporoid bed)
Canajoharie, New York, indicative of an interval of sediment starvation associated with tectonically enhanced deepening. Interestingly, a similar phosphatic corrosion surface is observed at approximately this position (middle Grier) within the Sebree trough (Mitchell and Bergström, 1991; Bergström and Mitchell, 1992; see also McLaughlin et al., this volume).

6.4. Early Shermanian: Sequence M5C

Abrupt lateral changes are seen in sequence M5C in the upper Grier and Macedonia beds interval of the Lexington Limestone in the Jessamine Dome area of Kentucky, and in its probable correlative in the Rathbun Member (upper Sugar River Formation)-Denley Formation succession of the Trenton shelf (Table 2; Figs. 4, 5). In both areas, crinoidal grainstones are locally developed and interpreted as TST successions that overlie relatively sharp sequence boundaries. These are again overlain by several small-scale cycles, or parasequences, that commence with thin-bedded calcilutites and dark gray shales that pass upward into pack- and grainstones. Similar cycles occur within the Macedonia Bed of sequence M5C in the Lexington area, although, at present it is not known whether these cycles can be correlated precisely.

A series of K-bentonites occurs within the upper Sugar River Limestone and the Denley Formation (Figs. 4, 5); these have permitted precise correlation into foreland basinal facies to the southeast of the type Trenton area (Goldman et al. 1994; Mitchell et al., 1994; Brett and Baird, 2002). On this basis it was demonstrated that the Rathbun-Denley carbonates pass abruptly, in the central Mohawk Valley area, into dark gray to black calcareous, graptolitic shales of the Flat Creek Formation, formerly termed Canajoharie Shale (Goldman et al., 1994; Mitchell et al., 1994).
NW NEW YORK

SEQUENCE C1

EDENIAN: *C. spiniferus* Zone
(C1: HIGHSTAND)

Hillier /Lindsay Fms, lower Kope Formation
gray shales and calcisiltite
and packstone
correlates with Utica
(Indian Castle) black shale
with *Triarthrus*-graptolites.
in Utica trough

*Triarthrus*-graptolites.
in Sebree Trough

EDENIAN: *C. spiniferus* Zone
(C1: MAXIMUM FLOODING SURFACE)

phosphatic-pyritic
hardground at top
Steuben-Dolgeville fms.

UPPER SHERMANIAN: *O. ruedemanni-C. spiniferus* Zone?
(C1: TRANSGRESSIVE SYSTEMS TRACT)

upper Rust Formation/ Steuben Formation
medium-massive bedded
locally trough cross-bedded
grainstone and packstone
correlates to Dolgeville Formation
bundle of thin-bedded
calcisiltites and shales
in Sebree trough

upper Point Pleasant Fm.
medium-thick bedded,
locally trough cross-bedded
grainstone, minor shale
correlates to unnamed
formation, thin-bedded
calcisiltites and shales
in Utica trough (foreland)

UPPER SHERMANIAN: *O. ruedemanni* Zone?
(C1: SEQUENCE BOUNDARY)

sharp truncation
at top of flat to
deformed beds of
Spillway Member

sharp truncation
at top of flat to
deformed beds of
Locust Creek beds

Table 4. Comparison of uppermost Shermanian–Edenian sequences of the uppermost Trenton Group in central New York State vs. the upper Lexington-Point Pleasant-Kope Formations of the Jessamine Dome, central Kentucky.
The upper part of the Macedonia Bed, likewise, contains several thin K-bentonites, termed Westboro K-bentonite zone (Fig. 4; Schumacher and Carlton, 1991). Sequence-based correlations indicate a similarly abrupt change of sequence M5C into dark gray, graptolitic shales in the Sebree Trough (see Fig. 14; also McLaughlin et al., this volume).

The appearance of K-bentonites in this sequence in both areas suggests intensified tectonism. Furthermore, in contrast to lower units, this sequence exhibits abrupt lateral variations, which suggest increased dissection of the platform into local highs and basins. In New York this can be attributable to subsidence of the foreland basin due to tectonic loading (Joy et al., 2000). It appears that the Sebree Trough underwent increased subsidence at this time, as well.

6.5. Late Shermanian: Sequence M6A

In the southern Jessamine Dome near Danville, Kentucky, the base of the next sequence is well developed with a sharp karstic contact separating nodular beds from peritidal (Salvisa Bed) fenestral micrite facies of the Perryville Member (Table 3; Figs. 4, 5). Thus, this contact was recognized as a major sequence boundary (base of M6) by Holland and Patzkowsky (1996, 1998). However, farther north, near Lexington and Frankfort, the basal sequence boundary is more subtle to cryptic; the same is true of the corresponding sequence in the Denley Formation (Russia Member) in the New York Trenton, suggesting that the central Kentucky area was shallower than central New York due to local tectonic uplift (see Fig. 14). However, in nearly all areas, the maximum flooding zone and highstand of this sequence are well defined as thin-bedded calcilutites and shales (upper High Falls submember of Russia Member in New York and similar Brannon Member in Kentucky), recording a widespread deepening (Fig. 10). This facies is widespread and analogous to that seen in the Logana- Napanee units, the early highstands of
Figure 10. (A) Upper Russia (Upper High Falls beds) at Upper High Falls on West Canada Creek, Trenton Falls, NY; note thin bedded calcilutites and shales. Height of outcrop is about 4m. (B) Upper Russia Formation; Taylor Mill phosphatic bed erosionally overlying deformed channel fill in Upper High Falls beds; Upper High Falls on West Canada Creek, Trenton Falls, NY. (C) Section of the Brannon Member of Lexington Limestone; note deformed channel-fill in the upper part of the section; Blue Grass Parkway, Lawrenceburg, KY; height of view is about 6 m. (D) Highly deformed strata in upper Brannon Member (Cane Run Bed); I-75 south of Georgetown, KY.
sequence M5A. The occurrence of K-bentonites and widespread soft sediment deformation (seismites) within these highstand facies in both New York and Kentucky signals intensified Taconic tectonism that seemingly affected both the Trenton Shelf and Lexington Platform during this time (Rast et al., 1999; Ettensohn et al., 2002). We do not mean to imply that these are record precisely the same events of faulting, but rather that this was a time of increased frequency of faulting both in the proximal foreland and in intracratonic areas.

It is notable that the sequence M6A highstand, both in New York and in Kentucky–Ohio, shows very abrupt lateral change from carbonates into dark shales in adjacent basins (see Fig. 14). This implies that substantially steepened ramps had developed by this time (near the A. tvaeensis–A. superbus conodont zonal boundary and the later part of the C. americanus graptolite zone; Fig. 4). Increasing siliciclastic content of the shelf carbonates may similarly record influx of sediments from uplifted orogenic areas.

6.6 Late Shermanian: Sequence M6B

In the Jessamine Dome the sequence boundary of sequence M6B is well demarcated at the base of the Sulphur Well bryozoan-rich grainstones (Table 3; Figs. 5, 11) and there is evidence for erosive truncation beneath this contact (Cressman, 1973; Ettensohn et al., 2002). The putative counterpart of this surface on the Trenton shelf is equally sharp at the base of the Mill Dam Member of the Rust Formation, although evidence for erosion is minimal. The upper surface of the Sulphur Well TST grainstone shows an abrupt shift to dark, shaly, nodular limestones (Stamping Ground Member) marked in most localities by pyrite and phosphate impregnated hardgrounds (Fig. 11). A comparable phosphatic corrosion surface is noted at the top of the lower Mill Dam division in several localities in the Mohawk Valley (Brett and Baird, 2002).
Figure 11. (A) M6B sequence boundary at Clays Ferry, Kentucky; Sulphur Well Member sharply overlies Brannon thin-bedded limestones at an erosion surface. (B) Section of Sulphur Well and overlying Stamping Ground in cut along I-75 near Georgetown, KY. Note sharp discontinuity (flooding surface) separating the units and occurrence of large stromatoporoids in the Stamping Ground Member.
The four-part sequence M6B (tri-partite Mill Dam Member and Walcott-Rust Quarry beds) can be compared to the Sulphur Well-Stamping Ground-Strodes Creek-Greendale member succession in the Lexington Platform (Fig. 5). In both areas, sequence M6B appears to be complex and comprised of two distinct smaller (fourth-order) sequences. In the Jessamine Dome this interval is particularly rich in stromatoporoids and solenoporid algae. Stromatoporoids are lacking in the Rust Formation of New York, which, however, displays common cyclocrinitid algae. Both suggest an interlude of altered conditions, possibly increased temperatures, during deposition of this sequence. The dark shaly packstones and calcisiltites of the Greendale Lentil (or member) in the Jessamine Dome may be matched by an interval of fossiliferous packstones (Walcott Quarry beds), in the Trenton Gorge area (see Brett et al., 1999).

Sequence M6B in Kentucky-Ohio and its counterpart in New York both undergo rapid thinning and gradation into dark gray, calcareous shales with thin bryozoan-rich limestones marking the TSTs (see Fig. 14). Strongly deformed strata (seismites) occur in the upper part of this sequence (Spillway Member of Rust formation) in New York and minor deformation occurs also in the highstand facies (Greendale lentil) of the Jessamine Dome. The abrupt facies shifts and seismites indicate episodes of tectonism, probably associated with the major thrust-loading in the Taconic Orogen during mid-Shermanian time.

6.7 Late Shermanian: Sequence M6C

The M6C succession in both areas begins with grainstones or packstones (Devil’s Hollow Member in Kentucky and upper Rust, Prospect Quarry member in New York) that are overlain by thin bedded, shaly carbonates, that pass upward abruptly into regressive, fine-grained grainstones that are extensively deformed (Table 3; Figs. 5, 14). Similarities between the two
Figure 12. Epibole taxa of the Kope Formation in Kentucky and Ohio. (A) Articulated specimens of the rhombiferan Cheirocystis fultonensis, basal Kope, Fulton beds; cut along KY Rte. 1159 near Brookville, KY; x 2; photo courtesy of C. Sumrall. (B) the crinoid Ectenocrinus, Kope Formation (Pioneer Valley submember), note intact crowns; Alexandria, Kentucky; x 1. (C) Triarthrus becki; cluster of exuviae; middle Kope Formation (Alexandria submember); Sycamore Creek; Indian Hill, Ohio; x 1.5. (D) "Logjam" of columns of Merocrinus, basal Kope, Fulton beds; quarry along Rte. 8 at Bradford, KY; x 1.
areas are not as great at this level as in the preceding intervals. However, in both the Jessamine Dome and in New York, deformed intervals appear to be most numerous and widespread within this sequence; particularly notable is the Locust Creek deformed interval, which has been traced over approximately 6,000 km$^2$ in northern Kentucky and southern Ohio (McLaughlin, 2002; McLaughlin and Brett, 2002). Deformation is also notable in the upper Rust Formation and equivalent portions of the Dolgeville ribbon limestone-shale strata in the Mohawk Valley. These patterns may suggest far-field tectonic instability and reactivation of faults.

The platform-to-ramp transitions in both New York and Kentucky-Ohio remained abrupt during deposition of sequences both M6B and M6C. Zones of deformed strata in the upper portions of both successions suggest another series of episodes of seismicity that affected both the Trenton Shelf and Lexington Platform during this interval. The abrupt appearance of *Triarthrus becki* and several other taxa (e. g., the crinoid *Merocrinus*; Fig. 12) in the Bromley Shale of the Sebree Trough and immediately adjoining Lexington Platform (first noted by Ulrich, 1888) indicates that deeper, dysoxic biotas in the Taconic foreland and Sebree Trough were interchanged by this time. This interbasinal connection, presumably through the Pennsylvania embayment, may have allowed the (delayed) entrance of *Climacograptus spiniferus* into the Taconic Basin from the southwest. Moreover, widespread dysoxic conditions, recorded in the basal Bromley Shale, extended even onto the proximal Lexington Platform.

6.8. Cincinnatian (Edenian): Sequence C1

Dramatic changes in sedimentation and basin configuration appear to have occurred in the early Edenian Stage, during deposition of the C1sequence of the Lexington Platform-Sebree trough area and its probable counterpart in New York (Table 4; Figs. 5, 12-14). In both areas a
Figure 13. Highstand deposits of sequence C 1. (A) Thick black shales of the Indian Castle Shale near *C. spiniferus-G. pygmaeus* boundary in East Canada Creek at Dolgeville, New York; light gray bands are calcilitites. Thickness of exposed shale is about 20 m. (B) Thick gray shales of the Kope Formation exposed along KY Rte. 445 at Brent, Kentucky; approximate *C. spiniferus-G. pygmaeus* boundary is shown at arrow. Note meter- and decameter scale cycles. Height of outcrop is approximately 35 m.
widespread interval of crinoidal grainstone, the Steuben Limestone of New York and Ontario, and the Point Pleasant Limestone (also termed the uppermost tongue of Tanglewood Member of the Lexington Limestone) in Ohio and Kentucky, signals relatively shallow, but transgressive conditions. Both units show evidence of local development of cross-stratified crinoidal sand shoals. Both pass basinward into a succession of fine-grained turbiditic calcarenites-calcisiltites and interbedded black shales (Dolgeville Formation in New York). The upper contact, where conformable, shows an abrupt transition to a back-stepping succession of shales and argillaceous packstones, and then into a major shale-rich succession, which carries graptolites of the *C. spiniferus* and lower *G. pygmaeus* zones (Hillier-Frankfort, Indian Castle Shale of New York: Baird and Brett, 2002; Kope Formation in Ohio-Kentucky: Mitchell and Bergström, 1991; Bergström and Mitchell, 1992). There is evidence of major local tectonism during deposition of this sequence in the Taconic foreland, as indicated by very extensive deformation in the upper Dolgeville, rapid subsidence into deep, dysoxic/anoxic conditions, abundant, closely spaced K-bentonites, and evidence of synsedimentary growth faulting (Mitchell et al., 1994; Baird and Brett, 2002). Perhaps most importantly, there is evidence of substantial influx of siliciclastics into the foreland basin. In central New York this is seen as a final transition from the Trenton-Dolgeville carbonates into black shale of the Indian Castle Formation (Fig. 13). Dark, organic rich muds (“Utica shale facies”) also prograded into the Sebree Trough and eventually out onto the Lexington Platform to form the “Utica” and Kope Formations (Bergström and Mitchell, 1992; Fig. 13). Isopachs indicate that this mud originated from the (present) northeast and thins out to the southwest (Ettensohn et al., 2002). This mud presumably filled the Sebree Trough and began to level out its topography: both the Point Pleasant and the Kope Formations show less lateral change into the Sebree trough than underlying Lexington units (Fig. 14). Overlying
Maysvillian and Richmondian strata show evidence of only minor deepening into the largely filled Sebree trough.

Significantly, the appearance of dysoxic muddy facies in the early Edenian (Point Pleasant and Kope Formations) is associated with a second brief, but widespread incursion of shelf to deeper basin taxa (e.g., the trilobite *Triarthrus*, the crinoids *Merocrinus, Ectenocrinus* and the rhombiferan *Cheirocystis*; Fig. 12) onto the Lexington Platform. This may also imply altered water mass circulation during this time.

The Kope Formation is broadly interpreted as a 3rd-order HST (Holland, 1993). However, it is clearly subdivisible into a series of meter- and decameter-scale cycles that also show sequence-like motifs (see Brett and Algeo. 2001; Brett et al., in press for detailed discussion). Moreover, in proximal areas, upper portions of the Kope-equivalent Clays Ferry Formation exhibit a transition into a thick body of heavily-deformed siltstones, referred to as the Garrard Member. Package appears to Pope and Read (1997a, b) interpreted the Garrard siltstones as a lowstand deposit. We suggest that these beds record a progradation of siliciclastics from a southerly source during a rapid lowering of base level and form a late highstand or regressive systems tract (Fig. 14); the abrupt input of siliciclastic silt also suggests a pulse of tectonic uplift. The heavy deformation is inferred to represent a series of seismites, which might be related to this tectonism (McLaughlin and Brett, 2002). It is notable that comparable package of prograded silts, the Hasenclever Siltstone, occurs in the an approximately coeval upper Utica or Frankfort Formation in New York State (Fig. 14; Lehmann et al., 1994).
Figure 14. Revised schematic chronostratigraphic chart for Upper Ordovician (Turnian to Edenian) sequences recognized herein for central New York State and the Jessamine Dome of Kentucky. Light to dark shading in HSTs shows relative changes from oxic carbonates and gray shales to dysoxic dark shales and calcisiltite facies. Compare with Figure 2.
7. Discussion: Implications of sequence stratigraphic correlations

The Trenton Shelf-Taconic Foreland Basin and Lexington Platform to Sebree Trough were contemporary platform to basin ramps, but were separated by more than 1000 km and differed with respect to orientation, paleowind and current patterns, and proximity to active orogenic areas. Moreover, the tectonic settings of the two shelf areas were distinctly different from one another. The Lexington Platform lay on the southeast side of the intracratonic Sebree Trough, while the Trenton Shelf lay to the northwest of the active Taconic Foreland Basin (Fig. 3). Given these considerable differences in location and paleogeographic setting, these two areas might be expected to show substantially different facies and stratigraphic patterns. However, our comparative studies actually show remarkably similar vertical and lateral patterns of facies in coeval, Mohawkian to earliest Cincinnatian, sequences.

Widespread distribution of particular facies at certain levels may be partly attributable to widespread and unique climatic factors. Thus, as emphasized by Holland and Patzkowsky (1996), the abundant fenestral micrites in sequence M4 not only indicate widespread shallow water conditions, but also a high production of algally- or microbially-produced micrites, probably a result of relatively warm water. The rhythmic, thinly bedded calcilutites and shales in the highstands of sequences M5A and M6A in both areas may be attributable to similar climatic fluctuations and conditions poised between carbonate production and siliciclastic input. These climatic/environmental conditions apparently were pervasive over the most of study area, i.e. the eastern quarter of Laurentia.

Moreover, the number of sequences and their component subsequences matches well between the two areas, and the constraints of K-bentonites and biostratigraphy are at least permissive of the correlations suggested herein. In certain instances, the distinctive and detailed
patterns of individual sequences appear to match very closely at meter- to decameter-scales. These cycles are most readily identified in medial sections; they tend to become amalgamated at erosion surfaces in proximal areas. Distal basinal settings tend to be dominated by dark shales and cycles are very subtle. Nonetheless, careful attention to detail permits recognition and correlation of these cycles across these major facies transitions, both in the Lexington Platform-Sebree Trough (see McLaughlin et al., this volume) and in the Trenton Platform-Taconic Foreland (see Brett and Baird, 2002, Baird and Brett, 2002). This evidence indicates a strong allocyclic and, in part, eustatic control on the development of these third- and fourth-order sequences and possibly on higher order cycles.

Thus, we do not agree entirely with Joy et al. (2000) that the major cycles documented in the Taconic Foreland are primarily tectonic in origin and asynchronous. That said, however, the comparison of resolved time-slices, in the form of sequences and their components, across regional gradients suggests an increasingly significant overprint of tectonics in both areas through the Mohawkian-early Cincinnatian epochs (Fig. 14).

As noted, the earlier Mohawkian sequences (M4, M5A, and M5B) show the lowest degree of lateral facies change over ~ 150 km transects in Ontario–New York and Kentucky–Ohio (Fig. 14). These sequences display an approximate “layer-cake” pattern with similar thicknesses and regionally extensive facies. In sharp contrast, sequences M5C through M6C show much more abrupt lateral facies changes in both areas (Fig. 14). For example, the middle Trenton sequences in New York show an abrupt transition, across less than 10 km, from pack- and grainstone facies near Trenton Falls to dark graptolitic shales (Brett and Baird, 2002). Sequences M6A and M6B have been traced from peritidal fenestral micrite facies in southern parts of the Lexington Platform to black, graptolitic shales in the Sebree Trough, with much of the transition occurring
again across only about 15–20 km (McLaughlin et al., this volume; Fig. 14). Finally, sequence C1 shows more widespread similarity of shale-rich facies, possibly because of widespread influx of siliciclastic sediments during this time.

A key counterintuitive observation of this study is that the magnitude of shallowing of cycles in platform areas of both regions is not a good predictor of which carbonate units will persist farthest into the basin. For example, the shallowest portions of lower sequences M5A-M5B are represented by shallow shelf skeletal grainstone and packstone facies in the southern Lexington Platform, and are recorded as shaly nodular, deeper shelf facies in the Sebree Trough. Conversely the Perryville Member, in sequence M6A, shows the shallowest, peritidal facies of the Lexington Formation in the south. However, these facies pass laterally into dark shaly calcisiltites in the Sebree Trough. Thus, the total range of facies is much greater at this level than in the lower sequences (Fig. 14). This observation suggests increased partitioning of subsiding basins and local highs.

Furthermore, it is notable that those sequences which show the greatest extent of lateral facies change are also the ones which show the highest frequency of regionally deformed beds, interpreted as regional seismites (McLaughlin and Brett, 2002, submitted). This association suggests that both are responses to increased tectonism. Moreover, these tectonically-influenced sequences occur synchronously in both study areas. This similarity of pattern obviously cannot be attributed to eustasy. Rather, we suggest that tectonic loading in the Taconic Orogen influenced far-field tectonism in both areas (see Rast et al., 1999; Ettensohn et al., 2002). Increased subsidence occurred in the Taconic Foreland and Sebree Trough at approximately the same time. Lithospheric loading may have indirectly reactivated older deep-seated faults causing local uplift and/or subsidence of crustal blocks. It is clear that regular flexural migration did not
occur in the Sebree Trough, but patterns of progressive uplift and subsidence are being documented in the foreland basin (Cornell and Brett, 2003).

These patterns indicate that tectonically controlled patterns of basinal subsidence and uplift of blocks (perhaps reflecting forebulge migration) exerted a strong influence on the local facies and motif of depositional sequences both in the foreland basin and on the Lexington Platform. However, these local tectonic features did not obliterate the underlying allocyclic pattern.

8. Conclusions

Detailed comparison of depositional sequences in the classic Upper Ordovician Black River and Trenton Groups in New York State and Ontario with those recognized by Holland (1993) and Holland and Patzkowsky (1996, 1998) and refined by Pope and Read (1997a, b) and Brett et al. (2002) in the coeval Tyrone–Lexington–Kope succession reveals striking similarities. These comparisons indicate at least partial allocyclic control on sedimentary cycles. Past emphasis on local facies variation, especially in the Jessamine Dome, has inhibited recognition of these striking regional similarities in patterns. The salient conclusions of this paper can be summarized as follows.

a) The Upper Ordovician (upper Mohawkian–lower Cincinnatian) upper Black River, Trenton and Utica Groups in the type area of New York State/Ontario are divisible into some eight depositional sequences. These sequences and systems tracts can be correlated tentatively into the shale-dominated Taconic foreland basin succession to the east although sequences are less clearly defined in this area.

b) The coeval Tyrone, Lexington, Point Pleasant and Kope Formations exposed in the Jessamine Dome can, likewise, be divided into eight depositional sequences. These sequences and their
component systems tracts appear to correlate with those of the New York section on the basis of preliminary conodont-graptolite and K-bentonite data, although precise correlations remain to be established.

c) Boundaries of composite sequences in both areas are thought to coincide approximately with M5, M6, and C1 recognized by Holland and Patzkowsky (1996) and Pope and Read (1997a, b). However, we also recognize at least three internal subdivisions of the two upper Mohawkian sequences: M5A-M5C, M6A-M6C; these are comparable in scale to the “parasequence sets” of of Pope and Read (1997b), but different in detail of boundaries.

d) Sequences and component systems tracts are most regionally similar and widespread during the Turinian to early Chatfieldian (Rocklandian–Kirkfieldian). The facies and faunal patterns are highly comparable and indicate that these sequences reflect allocyclic, probably eustatic fluctuations during a time of reduced topographic contrasts in eastern Laurentia.

e) Sequences are still recognizable in the late Chatfieldian (Shermanian) and show some intriguing similarities in both areas (e.g. abundant “seismites” in M6A and M6C). However, sequences M5C, M6A-C, and C1 also show more local variability than do the older sequences. Relatively abrupt lateral changes within the upper sequences imply tectonic control on local development related to ongoing Taconic tectonism, both in the Taconic Foreland Basin and in the intracratonic Sebree Trough. Similarities of lateral pattern suggest episodes of nearly synchronous subsidence in both basins.

f) Thus, regional study of precisely defined allostratigraphic sequences across individual shelf to basin transects shows that successive cycles do not show consistent lateral patterns. In some cases, changes are very minor and subtle, whereas other sequences undergo very abrupt lateral changes encompassing major facies shifts. These may reflect pulses of basinal subsidence and
gentle local uplift. Despite such local influences, the widespread nature of major sequences and their components also suggests a strong allocyclic control on their formation.

**Acknowledgments**

We have benefited greatly from discussions with many colleagues, including, especially John Delano, Frank Ettensohn, Steve Holland, Bob Jacobi, Dave Lehmann, Chuck Mitchell, Mark Patzkowsky, Mike Pope, Fred Read, and Bob Titus. Their views, while differing from ours in some details, helped considerably in the honing of our own ideas and arguments. Brett also acknowledges with deep appreciation, the support given by the Department of Geology at University of Cincinnati during initial studies of the Ordovician in the Tri-states area and help with graphics from Evelyn Pence. Finally, we wish to thank the organizers of this symposium, Mike Pope and Mark Harris for encouraging us to submit this work. Acknowledgment is made to the Donors of the American Chemical Society Petroleum Research Fund for partial support of this research.
References


(Eds.) Sequences, Cycle and Event Stratigraphy of Upper Ordovician and Silurian Strata of the Cincinnati Arch Region. Kentucky Geological Survey Guidebook 1, pp. 47-64.


Variations related to complete and incomplete suturing: Geology 2, 543-546.

Ordovician geology. In C. R. Barnes and S. H. Williams (Eds.) Geological Survey of Canada
Paper 90-9, 225-234.

Diecchio, R. J., 1993. Stratigraphic interpretation of the Ordovician of the Appalachian basin and
implications for Taconian flexural modeling. Tectonics 12(6), 1410-1419.

Emmons, E., 1842, Geology of New York, Part II, comprising the survey of the Second

and stratigraphic sequences, central and southern Appalachians, U.S.A. In: Advances in
Ordovician geology. In Barnes, C.R. and Williams, S.H., eds.. Geological Survey of Canada
Paper 90-9, 225-234.


possible far-field responses to the Taconian Orogeny: Middle-Late Ordovician Lexington

Schenectady area meeting, Field Trip Guidebook, State Geological Association.


Pope, M.C., Read, J.F., 1997a. High-resolution surface and subsurface sequence stratigraphy of the Late Middle to Late Ordovician (Late Mohawkian–Cincinnatian) foreland basin rocks, Kentucky and Virginia. American Association of Petroleum Geologists Bulletin 81, 1866-1893.

Pope, M.C., Read, J.F., 1997b. High resolution stratigraphy of the Lexington Limestone (Late Middle Ordovician), Kentucky, U.S.A.: A cool-water carbonate-clastic ramp in a tectonically


CHAPTER 3

Signatures of Sea Level Rise in Mixed Carbonate-Siliciclastic Foreland Basin Successions

PATRICK I. MCLAUGHLIN and CARLTON E. BRETT

H.N. Fisk Laboratory for Sedimentary Geology,
Department of Geology
University of Cincinnati, Cincinnati, OH 45221-0013

(submitted to Palaios; April, 2005)
ABSTRACT

Many mixed carbonate-siliciclastic successions contain widespread skeletal limestones that are commonly interpreted as representing discontinuous winnowed lag deposits of skeletal sand shoals and are thus assigned to regressive parasequence caps. Analysis of the Upper Ordovician (Chatfieldian-Edenian) Point Pleasant-Fulton interval in Kentucky and southwestern Ohio and resemblance to other middle Paleozoic successions in eastern North America challenges these paradigms. The limestone-rich Point Pleasant-Fulton interval is extremely widespread and contains sedimentological, taphonomic, and faunal evidence suggestive of upward deepening. Hardgrounds and condensed beds are widespread and numerous throughout the study interval, concentrated relative to surrounding strata. Taphonomic and faunal gradient analysis reveals both lateral within-bed gradients and parallel vertical gradients within the Point Pleasant; the Fulton, surprisingly, shows little vertical change. The interval also contains a number of bioevent beds (epiboles) that are interpreted as a response to rapidly shifting environmental parameters. Comparison of the Point Pleasant-Fulton interval with other widespread middle Paleozoic skeletal limestone successions in eastern North America reveals similarity of vertical and lateral trends. These patterns suggest that widespread limestones formed primarily in response to sea level rise and offshore siliciclastic sediment starvation, rather than simply through regression and increased winnowing. Further, this study suggests that such transgressive limestone successions may be relatively common in the sedimentary rock record and preserve a biotic response to a portion of the sea level cycle that has been largely overlooked in mixed carbonate-siliciclastic foreland basin successions.
INTRODUCTION

Setup of the Problem

Mixed carbonate-siliciclastic, middle Paleozoic foreland basin successions of eastern North America contain widespread skeletal limestones whose genesis has been interpreted sequence stratigraphically either as resulting from regressive shoaling or transgressive siliciclastic sediment starvation (Fig. 1-5). These opposing views not only suggest different trends in sea level fluctuation, but also in the scale of cyclicity and expression of transgressive systems tracts in mixed carbonate-siliciclastic systems in general (Fig. 3). The regressive shoaling model advocates that the sediment-water interface comes within the range of normal wave base through relative sea level fall (regression), resulting in concentration of carbonate grains by intensive hydraulic sorting; the distribution of such skeletal-lag deposits should therefore be limited to the areal extent of the upper shoreface zone. Alternatively, the transgressive limestone model suggests siliciclastic sediment starvation, the by-product of relative sea level rise (Loutit et al., 1988; Posamentier et al., 1988; Van Wagoner et al., 1990; Catuneanu, 2002), results in stimulation of the carbonate factory (Sarg, 1988; Handford and Louks, 1993) and formation of a skeletal sand deposit with little to no primary siliciclastic component; the distribution of transgressive limestones consequently should be widespread. Support in the literature for the transgressive nature of widespread skeletal limestones in mixed carbonate-siliciclastic successions is still in the early stages and hinges primarily on: (A) distribution of the limestone unit over tens of thousands of square kilometers, (B) position upon a regional unconformity, (C) a geometry indicative of terrigenous sediment starvation, (D) inclusion of multiple discontinuity surfaces such as hardgrounds, (E) taphonomic signature suggestive of condensation, and (F) faunal gradients indicative of deepening (i.e. McCave, 1973; Baum and Vail, 1988; Brett and
FIGURE 1 - Outcrop pictures of two typical mixed carbonate-siliciclastic successions. Examples are drawn from the Lower Silurian of the Niagara, New York area (A, C) and the Upper Ordovician near Maysville, Kentucky. Each outcrop picture is labeled for both shoal (A, B) and sediment starvation (C, D) models. The shoal model interpretation regards the entire package of strata pictured as simply shallowing upward, from a maximum flooding level near the base (base of Rochester Shale and Greendale Member), to a regressive ravinement surface (DeCew/Rochester and Greendale/Devils Hollow contacts), with continued shallowing upward (through the Gasport Dolostone and Devils Hollow Member). The siliciclastic sediment starvation model regards the shale-rich intervals (Rochester Shale and Greendale Member) as shallowing upward highstand systems tract, the sharp contact with the overlying carbonate as forced regression surface (Rochester/Decew and Greendale/lower Devils Hollow contacts), the lower, heavily reworked part of the carbonate succession as a falling stage systems tract (DeCew Dolostone and lower Devils Hollow Member), overlain by a sequence boundary/correlative conformity (DeCew/Gasport and lower/upper Devils Hollow contacts), and a switch to deepening upward through the remaining carbonate succession (Gasport Dolostone and upper Devils Hollow Member).
FIGURE 2- Close-ups of interpreted sequence boundaries within the successions highlighted in Figure 1. The two successions contain a striking similarity of lithology and bedding from parallel to slightly flaser laminated, highly amalgamated, fine- to medium-grained, slightly argillaceous calcarenite, upward across a sharp, slightly irregular contact to relatively clay free, tabular- to herringbone cross-bedded, skeletal grainstone.
Baird, 1985, 1996; Brett et al., 1990; Banerjee and Kidwell, 1991; Fürsich et al., 1991, 1992; Abbot, 1998; Fürsich and Pandey, 2003; Heckel, P.H., 1994; McLaughlin et al., 2004). Further, there is even less support in the literature that Paleozoic widespread limestones were generated in a fashion similar to Mesozoic and Cenozoic mollusc-dominated shell beds.

The geometry of sedimentary deposits is a critical piece of data in sequence stratigraphic interpretation (Van Wagoner et al., 1988; Fig. 4). Currently accepted facies models for mixed carbonate-siliciclastic deposits (James and Kendall, 1992) suggest that the geometry of a deposit generated by regressive shoaling would be thinnest in inner ramp areas where constant winnowing within fair-weather wave base and bypass results in a condensed lag deposit (Fig. 4A, 5C, D). This deposit would become less condensed in a down-ramp direction attaining its thickest (compacted) level in the middle ramp where storm transported carbonate sediments and flocculated clays are interbedded with fair-weather hemipelagic clays (Kreisa and Bambach, 1982; Aigner, 1985; Pope and Read, 1998). On the outer ramp and basinal areas this deposit would commonly thin slightly and be composed primarily of hemipelagic clays with only a minor component of distal storm beds. The geometry of a deposit generated by transgressive siliciclastic sediment starvation shows a number of similarities, but also key differences (Fig. 4B, 5E; Sarg, 1988). Transgressive limestones are similar to regressive deposits in the inner ramp where winnowing is a dominant process, though it is primarily fine-grained carbonate grains that are winnowed rather than a combination of siliciclastic and carbonate grains. The middle, outer, and basinal parts of the ramp are predicted to be much thinner than those of the regressive model as terrigenous input is much more limited (Baum and Vail, 1988). The resulting condensed deposits are primarily composed of autochthonous carbonate sediments (Brett et al., 1990; Fürsich et al., 1991, 1992; Kidwell, 1991; Brett and Baird, 1996). A transgressive origin for
FIGURE 3- Schematic diagram showing variable interpretations of mixed carbonate-siliciclastic successions. (A) Pope and Read (1997) interpreted "meter-scale" cycles of the Upper Ordovician Lexington Limestone as simply shallowing upward successions. (B) McLaughlin et al. (2004) proposed that the "grainstone cap" actually represented both initial shallowing and subsequent deepening separated by a sequence boundary.
FIGURE 4-Hypothetical models of thickness distribution across a mixed carbonate-siliciclastic foreland ramp profile and correspondence to systems tracts and depositional systems. (A) Winnowing model. Stratigraphic package is thinnest in the shallowest areas where it is dominated by a lag of skeletal sand. Near the middle of the ramp the package thickens to a maximum where it is balanced between autochthonous skeletal material and allochthonous skeletal material transported from up-ramp shoals by storms, and hemipelagic clay deposits. The interval thins down-ramp into pure muds deposited as hemipelagic rain and distal storm beds. (B) Sediment starvation model. Similar to the winnowing model the section is condensed by the effects of winnowing in up-ramp shoal areas. The primary differences are seen in middle and outer ramp where siliciclastic sediments are much more scarce. Thin, organic-rich, clay shale in the outer ramp typically contains an abundance of pyrite and pelagic skeletal elements.
widespread limestones does not restrict the regressive shoal model from use in mixed carbonate-siliciclastic successions; in fact it is a key part in understanding the genesis of highstand and falling stage systems tracts.

One of the impediments to studying widespread skeletal limestones is the difficulty of dissecting and sampling these locally massive limestone units, due to their amalgamated and heavily cemented nature. However, certain widespread limestones have a slightly higher proportion of shale partings than others, which allow for separation of individual beds, permitting high-resolution analysis at the centimeter-scale. The following study documents the internal complexity and lateral variability of one such widespread skeletal limestone unit (Point Pleasant-Fulton interval) from the Upper Ordovician of eastern North America (Fig. 6). It has been interpreted, variably, as the cap of a meter-scale, shallowing upward cycle (parasequence of Pope and Read, 1998) and alternatively as the transgressive systems tract of a third-order depositional sequence (McLaughlin et al., 2004). New data on the regional facies distribution, occurrence and characteristics of hardgrounds and other diastems, taphonomic and faunal gradient analysis, and bioevents of this interval are presented. Additionally, comparison is made of this case study to other Phanerozoic examples and the implications for transgressive systems tracts in mixed carbonate-siliciclastic successions are considered generally.

Study Area

For this study we chose to focus upon a distinctive and well-exposed interval in the Upper Ordovician (Chatfieldian-Edenian; Fig. 6) of the Cincinnati Arch region (Jessamine Dome) of northern Kentucky and southwestern Ohio (Fig. 7). This interval is significant for several reasons. (1) The middle of the study interval (Point Pleasant/Fulton contact) marks the Mohawkian/Cincinnatian series boundary (the northern part of the study area occupying its type
A. EPEIRIC CARBONATE RAMP

IRWIN (1965)

B. EPEIRIC SILICICLASTIC RAMP

C. EPEIRIC MIXED CARBONATE-SILICICLASTIC RAMP


D. EPEIRIC MIXED CARBONATE-SILICICLASTIC RAMP

MCLAUGHLIN ET AL. (2004) HIGHSTAND AND FALLING STAGE SYSTEMS TRACTS

E. EPEIRIC MIXED CARBONATE-SILICICLASTIC RAMP

MCLAUGHLIN ET AL. (2004) TRANSGRESSIVE SYSTEMS TRACT

section), which is well documented in the local literature. (2) This contact is also marked by an easily identifiable lithologic transition between a lower limestone-dominated succession (Lexington Limestone) and an overlying shale-dominated interval (Kope Formation) across most of the study area. (3) The base of the interval is underlain by a series of widely traceable deformed beds (Locust Creek member seismites of McLaughlin and Brett, 2004). (4) Analysis of faunal gradients, event beds, and small-scale cyclicity within the Kope Formation are the subject of several recent high-resolution (centimeter-scale) studies (Holland et al., 1997; Holland et al., 2001; Brett and Algeo, 2001; Meyer et al., 2002; Webber, 2003; Brett et al., 2005).

Methods of Study

The study commenced with bed-by-bed measurement of the 6- to 13-meter thick Point Pleasant-Fulton interval and adjacent strata (Fig. 6, 8,9). A total of approximately 50 outcrop sections in northern Kentucky and southwestern Ohio (Fig. 7) were examined in this study. Particularly useful were a series of newly created roadcuts along US highway 127 between Owenton and Frankfort, Kentucky (Fig. 7, sections A1-A5), which permitted examination of the study interval in large (0.5 to 2.0 kilometer long), closely spaced sections along a proximal-to-distal transect near the margin of the Lexington Platform. Each section was photographed and measured at centimeter-scale. In addition, a series of 10 drillcores were photographed and measured at similar resolution for patterns in color, bed thickness, and larger skeletal elements. All beds were noted, measured, and assigned to Dunham (1962) categories; dominant skeletal elements were identified; and features of bed bases and tops (e.g., sharp to gradational boundaries, irregularities, hardground features) were carefully noted.
FIGURE 6—Upper Ordovician (Turonian-Maysvillian) stratigraphy for Kentucky. Composite stratigraphic column displaying depositional sequences, facies distribution, event beds (after McLoughlin et al., 2004), lithostratigraphic units (modified from Cressman, 1973), and biostratigraphy (zonation of Mitchell and Bergström, 1991). Note the regular alternation of limestone-dominant and shale-dominant facies. The Point Pleasant member of the Lexington Limestone and Fulton submember of the Kope Formation are shown at the top of the column and form the transgressive systems tract of the C1 sequence. Symbols: S, system; s, stage; G, graptolite zone; DS, depositional sequence designation; ST, systems tract; Facies, vertical facies distribution; Event, vertical distribution of event beds; TST, transgressive systems tract; HST, highstand systems tract; FSST, falling stage systems tract.
Correlation involved, first, the identification of major marker beds (i.e. thick, laterally continuous limestone beds, heavily mineralized limestones, limestones having unusual color, limestones containing reworked concretions or abundant mudstone rip-up clasts), significant contacts (i.e. major litho-, tapho-, or biofacies change), small-scale limestone-shale cycles, faunal epiboles, hardgrounds, and deformed beds (cf. Brett et al., 2003). Secondly we employed an iterative procedure of pattern matching between localities, which involved: (1) comparing high-resolution outcrop photographs and drafted sections between adjacent localities to make preliminary correlations, and (2) revisiting sections (in some cases on multiple occasions) to re-examine specific intervals for the presence of marker beds, epiboles, etc. identified in adjacent exposures. Finally, all sections were physically re-examined in close succession to confirm (or deny) correlation of markers in relation to one another. Only in this way and by matching of nearest neighbor sections is it possible to establish firm correlations in the absence of biostratigraphic zonal boundaries and K-bentonites. Further, taphofacies and faunal gradient analysis were employed in a number of outcrop and core sections for identification of subtle gradients within the study interval.

GENERAL GEOLOGICAL SETTING

Paleogeography and Paleoclimates

Upper Ordovician (Mohawkian) strata of eastern North America record the collapse of a relatively flat, widespread, shallow marine tropical carbonate platform (Black River) during the onset of Taconic orogenesis. Stresses produced by thrust loading in the Taconic hinterland reactivated basement faults across eastern Laurentia, which resulted in the formation of the Taconic peripheral foreland basin on the margin of the craton, the broad Lexington Platform, and the intracratonic Sebree Trough (Stanley and Ratcliff, 1985; Kolata et al., 2001; Ettensohn et al.,
FIGURE 7-Location/paleogeographic map for the tristate area of Kentucky-Ohio-Indiana. Paleogeographic reconstruction of the study area for the latest Mohawkian/early Cincinnatian based on facies analysis of marked outcrops and core from the present study and regional analysis of the Kope Formation by Jennette and Pryor (1993), Holland et al. (2001), and Brett and Algeo (2001). Inset map shows 200 and 250 south paleolatitude lines. Sections designated A1-A5, B1-B11, and C1-C5 are presented in figures 13A-C. Outcrop and core sections studied but not included in the cross sections are assigned lowercase letters and numbers, respectively. Faint gray lines represent inferred distribution of equivalent facies. Symbols: C, Cincinnati; F, Frankfort; LX, Lexington; M, Maysville; O, Oxford; and W, Winchester.
The distal effects of Taconic orogenesis are recorded in the sediments of the Lexington Platform as: (1) an abrupt influx of clay, (2) numerous altered volcanic ash horizons (K-bentonites; Kolata et al., 1996), and (3) several widespread intervals of soft-sediment deformation (seismites; Pope et al., 1997; Rast et al., 1999; McLaughlin and Brett, 2004). Additionally, rapid lateral facies change noted within some portions of the Lexington Limestone have been attributed to local fault reactivation as a result of far-field tectonics (Cressman, 1973; Ettensohn et al., 2002, 2004; McLaughlin et al., 2004).

In the study area (paleolatitude between 20 and 25 degrees south of the equator; Fig. 7), the widespread peritidal carbonates of the upper Black River Group, locally the Turinian age Tyrone Formation, are abruptly overlain by approximately 100 meters of Chatfieldian age Lexington Limestone (Fig. 6). In contrast to the Black River Group, the Lexington Limestone is composed of argillaceous skeletal limestones and shales, which alternate with discrete intervals of widespread skeletal limestone containing little to no argillaceous content. The characteristics and interpretation of the latter form the focus of this paper.

Sediments within the study interval were deposited during the onset of glaciation in Gondwana (Frakes et al., 1992). Based on outcrop studies and modeling, Read (1998) considers the Upper Ordovician a transitional time between the greenhouse period of the Late Cambrian to Early Ordovician and the end Ordovician glaciation. Meter-scale cycles of Pope and Read (1998) and depositional sequences of McLaughlin et al. (2004), comprising largely the same packages of strata within the Lexington Limestone (Fig. 3), are interpreted to represent sea level fluctuations of greater than 20 meters. Both Holland and Patzkowsky (1996) and Pope and Read (1997b) interpreted the switch from carbonate mud-dominated sediments of the Black River Group to deposition of the skeletal-dominated sediments of the Lexington Limestone as a change
FIGURE 8-Schematic cross-section compiled from data of McLaughlin et al. (2004) and Algeo and Brett (2001) showing facies distribution through the upper Lexington Limestone, Clays Ferry/Kope Formation, and Garrard Siltstone from central Kentucky (Frankfort-Winchester area) into southwestern Ohio (Oxford area; Fig. 7). Facies gradients from calcarenite to organic-rich brown shale depicted by the gray scale. Condensed limestones extend as tongues well down ramp. Note similarity in stratigraphic architecture of the cyclic shallowing upward, shaly Bromley through Gratz members of the Lexington Limestone and Kope/Clays Ferry formation. However, while the proportion of skeletal packstones and grainstones is relatively the same, the Clays Ferry/Kope Formation contains nearly ten times more clay and quartz silt than the Bromley. Note that the Point Pleasant-Fulton interval extends well out into the basin as a relatively clean limestone tongue, even though shoreface deposits are confined to the upper ramp.
from tropical- to temperate-style carbonate production during an arid to humid climactic shift. Although an alternate interpretation invoking major sea level rise has also been suggested (Taylor and Allison 1998; McLaughlin et al., 2004). The Upper Ordovician has also been classified as a time of calcite seas, when lowered seawater Mg/Ca ratios favored the early precipitation of low-Mg calcite (Palmer and Wilson, 2004).

Upper Ordovician Facies and Environments

McLaughlin et al. (2004) define a facies series for the Upper Ordovician (Chatfieldian) strata of the Jessamine Dome from central Kentucky to southwestern Ohio, which builds upon previous work of Cressman (1973), Pope and Read (1997a), and Holland and Patzkowsky (1998), and facies and sequence stratigraphic concepts outlined in Brett et al. (1990) and Brett and Baird (1996). One of the key findings of the McLaughlin et al. (2004) study is that the Upper Ordovician (Chatfieldian-Edenian) shallow marine facies of the Jessamine Dome can be divided into two broad groups: (1) a grainstone-rudstone facies gradient, and (2) a shaly-nodular wacke-packstone facies gradient (Fig. 5D, E). The shaly-nodular wacke-packstone facies gradient progresses from argillaceous calcarenite, to shaly-nodular wacke-packstone, to shale-calcisiltite couplets, finally into carbonate laminated olive shales, in a down-ramp direction (Fig. 5D). The grainstone-rudstone facies gradient progresses from clean calcarenite, to grainstone-rudstone, to fine-grained grainstone-shale couplets, finally into skeletal stringers in organic-rich shales, in a down-ramp direction (Fig. 5E). The Point Pleasant-Fulton interval generally displays the grainstone-rudstone facies gradient, described in detail below.
FIGURE 9-Stratigraphic units discussed in the text and characteristic facies. (A) Locust Creek member near Peaks Mill, Kentucky (section A2); note the presence of soft-sediment deformation in the upper of these thin, platy-bedded calcarenites. (B) Point Pleasant to lower Fulton interval in a road cut on I-64 northeast of Winchester, Kentucky (section aa); massive calcarenite facies of the Point Pleasant is overlain by tightly stacked grainstone-rudstone beds of the Fulton (contact at arrow). (C) Fulton submember near Monterey, Kentucky (section A5); contact with the underlying Point Pleasant marked at A-arrow. Note the dark fissile shales and concretions of the lower Fulton (rhythmite-laminated shale facies) abruptly overlain by rhythmically bedded pack-grainstone beds and shales of the upper Fulton. B-arrow marks the Cheirocystis epibole bed in the basal Fulton. C-arrow marks a prominent hardground in the upper Fulton shown in figure 15F. (D) Kope Formation (middle Brent up to Alexandria submembers) from Brent, Kentucky (section B10). Note that this portion of the Kope Formation is dominated by shale (approximately 80%) in contrast to the basal Kope Fulton submember shown in image C (approximately 40%). (E) Kope Formation equivalent exposed near Winchester, Kentucky (locally referred to as portion of the Clays Ferry Formation) overlain by the Garrard Siltstone (contact marked at arrow). The contact is sharp and represents a change from interbedded thin silty packstones (approximately 20%), siltstone (approximately 30%), and shale (approximately 50%) to siltstone-dominated (greater than 80%).
SEQUENCE STRATIGRAPHY AND REGIONAL VARIATION OF THE POINT
PLEASANT-FULTON INTERVAL

Large-scale Sequence Stratigraphic Patterns

A number of recent studies have designated large-scale depositional sequences within the Upper Ordovician strata of the Jessamine Dome. Pope and Read (1998) divided the Ordovician of eastern North America into three 2\textsuperscript{nd}-order supersequences (10-30 m.y. duration each). The Point Pleasant-Fulton contact marks the position of the inferred maximum flooding surface approximately halfway through the uppermost (Taconic) supersequence. They additionally divided this Taconic supersequence into four 3\textsuperscript{rd}-order sequences (1-10 m.y. duration each).

Holland and Patzkowsky (1996) divided this same stratigraphic interval into a series of eight 3\textsuperscript{rd}-order sequences (M5, M6, and C1-C6; 1-3 m.y. duration each). Sequence 1 of Pope and Read (1998) overlaps almost completely with Holland and Patzkowsky’s (1996) M5 (3\textsuperscript{rd}-order) sequence, however, Sequence 2 includes both Holland and Patzkowsky’s M6 and C1 sequences. It appears to be the case that neither set of authors recognized a repeating stratigraphic pattern common to all 3\textsuperscript{rd}-order sequences. McLaughlin and Brett (2004), however, did propose such a 3\textsuperscript{rd}-order sequence stratigraphic pattern for strata included within Holland and Patzkowsky’s (1996) M5 and M6 sequences, subsequently dividing them into M5A-M5C and M6A-M6C (∼1 m.y. duration each) and only slightly modifying the boundaries of the C1 sequence (including the Point Pleasant-Fulton interval; Fig. 6). These sequence designations are very similar to Pope and Read’s (1998) “meter-scale” cycles (as noted in the introduction). McLaughlin and Brett (2004) mapped the distribution of the M5A-C1 sequences across a significant portion of central and northern Kentucky and southwestern Ohio and Brett et al. (2004) made a case for the great
FIGURE 10-C1 sequence boundary. (A) Point Pleasant/Locust Creek contact near Peaks Mill, Kentucky (section A3). The irregular contact (dotted line) of the sparry, fine-to-medium grained, calcarenites of the Point Pleasant member erosionally overlie the slightly deformed, laminated, argillaceous, fine-grained calcarenites of the Locust Creek member at an iron-mineralized surface, which may represent diagenesis during subaerial exposure. (B) Point Pleasant/Locust Creek contact in a core from approximately 100 kilometer to the north of the section in panel A (section 3). Note the sharp contact between the skeletal grainstone bed at the base of the Point Pleasant member with deformed carbonate laminae of the underlying Locust Creek member.
similarity between these seven sequences and seven deposition sequences of the same age and sharing the same sequence stratigraphic architecture in New York and Ontario.

The Point Pleasant-Fulton interval forms a fundamental part of the C1 sequence (Fig. 6, 8-11). A more argillaceous interval underling the Point Pleasant member includes the (informal) Bromley, Peaks Mill, and Gratz members of the Lexington Limestone, which cumulatively form the M6C highstand systems tract (HST). The overlying Locust Creek member forms the M6C falling stage systems tract (FSST). The contact between the Locust Creek and the Point Pleasant members is sharp in outcrop sections across the study area; near Frankfort, Kentucky the contact is irregular (>10 centimeters relief), erosive, and possibly karstic on underlying Locust Creek deformed calcarenites (Fig. 10). This contact is interpreted as a sequence boundary, which becomes a correlative conformity away from the Frankfort area (Fig. 11A). Stratigraphically higher, the Point Pleasant-Fulton contact is marked by a conspicuous condensed bed (Fig. 11B), designated the C1 maximum starvation surface (McLaughlin and Brett, 2004) and a lithologic change in outcrops and core across the majority of the study area as an abrupt shift from approximately 20 percent gray shale to approximately 50+ percent organic-rich brownish gray shale (Fig. 8, 9C). This contact is also marked by a well-documented faunal change (Ulrich, 1888; McFarlan and Freeman, 1935; Lattman, 1954), designated the Mohawkian-Cincinnatian series boundary. The top of the Fulton marks the C1 maximum flooding surface (Fig. 5, 7). The siliciclastic-dominated strata of the remaining Kope Formation and laterally equivalent Clays Ferry Formation and Garrard Siltstone compose the highstand and falling stage systems tracts of the C1 sequence (McLaughlin et al., 2004; Fig. 6, 8). Hence, the Point Pleasant is interpreted to occupy the position of an early transgressive systems tract and the Fulton is interpreted to occupy the late transgressive systems tract within the 3rd-order C1 sequence (Fig. 6, 8). The general
FIGURE 11-Maps of the study area showing the distribution of (A) irregular and mineralized contact of the Point Pleasant on the Locust Creek member suspected as a karst surface, and (B) iron and phosphate mineralized polymictic conglomerate at the Point Pleasant-Fulton contact.
characteristics of small-scale cyclicity in the Point Pleasant-Fulton interval are common to other widespread limestones in mixed carbonate-siliciclastic successions of eastern North America (see concluding sections for specific examples), thus detailed discussion of these cycles is provided below.

Sedimentology and Small-scale Cyclicity of the Point Pleasant-Fulton Interval

Analysis of the Point Pleasant-Fulton interval from central Kentucky northward to southwestern Ohio (Figs. 7) reveals that this mixed carbonate-siliciclastic succession is highly ordered both laterally along the basin profile and stratigraphically (Fig. 12-14). Within this interval 13 major lithologic alternations between packages of skeletal grainstone-rudstones and packages containing shales with argillaceous limestones are designated as small-scale cycles (see left-hand column of figure 12 for idealized motif). The majority of these cycles are widespread, being identifiable in nearly all sections (Fig. 13). Several characteristics were used to establish the identity of individual cycles within particular sections, including: a) position relative to the upper and lower contacts of the Point Pleasant and Fulton, b) cycle thickness, c) proportion of the cycle occupied by the limestone hemicycle, d) presence of hardgrounds, and e) presence of faunal epiboles.

Many of the thicker limestone beds, which make-up the limestone hemicycle, can be traced across the study area, internally recording a gradient of lithologies (Fig. 13, 14). For example, along the A1-A5 (Fig. 13A) cross section the limestone hemicycles grade laterally within the Point Pleasant from calcarenite (sections A1-A2), to skeletal grainstone-rudstone (sections A3-A4), and finally into fine-grained grainstones (sections A5). This gradient can be extended northward in cores containing the Point Pleasant from section C3 (having very similar facies to section A5) into very thin beds and laminae of fine-grained pelletal grainstone of section C5.
FIGURE 12-Key for stratigraphic columns (example from section B4). Note the representation of lithologies, bedding, and component taxa in the stratigraphic column on the left and corresponding horizons in outcrop picture. The motif of small-scale cycles is given at the lower left and cycles are labeled in the stratigraphic column. Rank abundance of common and abundant taxa is represented by ordering of faunal symbols, moving from higher to lower abundance away from the column.
(Fig. 13C). The same gradient is recognizable in the limestone hemicycles of the Fulton with a few exceptions including: 1) limestone hemicycles rarely reach calcarenite grade in the scattered exposures of even the most up-ramp areas between Frankfort and Winchester, and 2) calcareous packstones and wackestones and fine-grained pelletal packstone to calcisiltite in down-ramp areas are associated with abundant carbonate concretions that appear to be lacking in much of the outcropping Point Pleasant (though this may be an artifact of a general lack of down-ramp outcrop exposures of the Point Pleasant). Bedding and sedimentary structures also vary along this lithologic gradient (Fig. 15) from: 1) herringbone cross-bedding and closely spaced and small-scale wavy amalgamation surfaces in calcarenites, 2) more irregular amalgamation surfaces, trough cross-bedding, mudstone rip-up clasts, and irregular bases and megarippled tops in the grainstone-rudstones, 3) little amalgamation, planar to irregular bases, starved ripples, and general lack of cross-bedding in the calcareous packstones and wackestones, and 4) thin discontinuous pods of fine-grained pelletal grainstone to calcisiltite in many cases underlain by concretions.

Similarly, shale-rich hemicycles of the Point Pleasant along the A1-A5 and C3-C5 transects grade laterally from argillaceous calcarenite (sections A1-A2), to argillaceous packstones and wackestones interbedded with shale (sections A3-A4), to interbedded calcisiltites and shale (sections A5 and C3), ending eventually in shale with only thin carbonate laminae (section C5). A distinctive lateral trend in the geometry, thickness, and number of individual limestone beds occurs in the shale hemicycle: 1) the thickest, most numerous, and most highly amalgamated limestone beds occur in the argillaceous calcarenites, typical of central Kentucky, 2) to the north nodular bedded argillaceous packstones and wackestones interbedded with shale are typically thinner and fewer in number, 3) semi-tabular to tabular bedded calcisiltites interbedded with
FIGURE 13-High-resolution cross sections of the Point Pleasant-Fulton interval. Datum is the Point Pleasant-Fulton contact. (A) South-to-north cross section parallel to inferred depositional dip from calcarenite facies at Frankfort (A1) to deep subtidal facies at Monterey, Kentucky (A5), approximately 25 kilometer to the north. This transect records significant facies change over a short distance, yet note the continuity of major bedding packages/cycles and event beds (i.e. hardgrounds, deformed horizons, and epiboles). Faunal distribution appears to track lithofacies change, note that the lateral lithofacies and faunal variations closely parallel the vertical variations. (B) Southeast-to-northwest cross section from Maysville (B1), Kentucky to Cincinnati (B12), Ohio (approximately 80 kilometers). The majority of this cross section documents only subtle lateral facies change within the Point Pleasant-Fulton interval, however the degree of lateral facies change increases between Waterworks Creek (B11) and Cincinnati (B12). Much of this transect is interpreted to occur along a deeper portion of the ramp than the A1-A5 cross section (Fig. 7). Note that many of the thinner wake-packstone and fine-grained grainstone beds are the first to pinch out into Triarthrus-dominated shale with concretions in a down-ramp direction. Grainstone-rudstone beds that form the bases of small-scale cycles are the most laterally persistent, some extending to the margin of the Seabree Trough; others pinch out, but their position is marked laterally by concretion horizons. (C) C1-C5 cross section (southeast-to-northwest) across southwestern Ohio toward the Indiana state line. Data in this cross section are taken completely from drill core. The combination of more widely spaced sections and much reduced surface area for study resulted in lower accuracy correlations across this transect. However, the observed distribution of strata and epibole taxa, such as Merocrinus and Triarthrus, are very similar to observed patterns toward the northwestern end of B1-B12 cross section.
shale are thin and still fewer in number, and 4) in the Cincinnati area and in cores to the north pinch and swell bedding is characteristic of relatively rare thin limestones interbedded with pure shale. The discontinuous nature of many of the beds within the shale hemicycles makes tracing them laterally much more difficult than the largely continuous beds in the limestone hemicycles.

Vertical trends in the shale hemicycle composition through the Point Pleasant parallel the limestone hemicycle trends and range from argillaceous calcarenite at the base to nodular packstones and wackestones toward the top or, in more northern sections, from nodular packstones and wackestones at the base to interbedded calcisiltites and shale at the top. Unlike the Point Pleasant, the Fulton shows little overall vertical trend in lithology within the limestone hemicycles (primarily skeletal grainstone-rudstone across much of the study area), though shale hemicycles do show an increase in quartz silt content upward (Fig.13, 14).

Changes in cycle thickness occur both laterally and vertically within the Point Pleasant-Fulton interval. Lateral tracing of cycle thickness within the Point Pleasant reveal marked thinning to the north, accounted for by thinning of both the limestone and shale hemicycles. Surprisingly, the cycles of the Fulton show slight increase in shale hemicycle thickness, whereas the limestone hemicycles thin similar to those of the Point Pleasant (Fig. 13, 14). Cycles of the Point Pleasant also thin vertically, again primarily do to reduced thickness of the shale-hemicycle. The Fulton shows little to no systematic vertical change in cycle thickness.

Interpretation of regional patterns within the Point Pleasant-Fulton interval

Lateral variation in small-scale cycles, similar to variations in the geometry of systems tracts, can yield insight into the genesis of stratigraphic packages, even at the scale of bed bundles. The observed lateral trends of the Point Pleasant of decreasing cycle thickness (especially the hemicycles) from central Kentucky into southwestern Ohio and lateral change in
sedimentary structures confirm interpretations of deepening from inner ramp to outer ramp and basinal deposits and a reduction in both siliciclastic and carbonate sediments suggestive of TST. The slight lateral thickening of shale hemicycles within the Fulton suggests early HST.

Beyond stratal geometry many sequence stratigraphic analyses, especially in pure carbonates, rely on secular changes in cycle thickness to designate systems tracts (Sarg, 1988). Result of the analysis presented here show a systematic change in Point Pleasant small-scale cycle thickness, primarily manifest reduction of the thickness of shale hemicycles upward consistent with the designation of TST. The reduced thickness of the shale hemicycles is interpreted here as a reduction in the influx of siliciclastics to the Lexington Platform through the deposition of the Point Pleasant. Interestingly, the Fulton does not show a similar trend, cycle thicknesses show little secular change, though some of the thickest cycles are near the top. This primarily aggradational geometry could fit the definition of a TST or early part of the HST.

Competing models of the genesis of similar small-scale cycles in the middle and upper Kope Formation agree on a Milankovitch-scale climatic control on the influx of fine siliciclastic sediments onto the Lexington Platform, however they differ on whether the climate cycles are controlling fluctuations in storm intensity (Holland, in review) or eustasy (Brett et al., 2003). Resolving this debate is beyond the scope of this cycle analysis, the important point to be made at this point in the study is that each of the small-scale cycles likely formed over a similar period, therefore differences in thickness between cycles represents shifts in siliciclastic and/or carbonate sedimentation rates.
FIGURE 14-Schematic cross section of the Point Pleasant member of the Lexington Limestone and Fulton and Brent submembers of the Kope Formation from central Kentucky to west-central Ohio (approximately 100 kilometers). Stratigraphic data derived from ten sections representing shoal to basinal facies (see Figure 7 for locality information). Vertical bars mark stratigraphic distribution of each section. Note the variations in geometry between the Point Pleasant, Fulton and Brent submembers as well as increasing siltstone content in the upper Fulton into the Brent. Limestone hemicycles colored in white-blue-gray gradient (equals calcarenite-grainstone/rudstone-fine grained grainstone), limestone within shale hemicycles colored as gray, siltstones colored yellow.
DISCONTINUITIES IN THE POINT PLEASANT-FULTON INTERVAL: HARDGROUNDS, REWORKED CONCRETIONS, AND DIACLAST BEDS

Description of discontinuities

Discontinuity surfaces are unusually abundant and diverse within the Point Pleasant-Fulton interval by comparison with surrounding strata (Fig. 13A,B, 16). No recognizable discontinuity surfaces have been identified within the shaly Bromley or Gratz members of the Lexington Limestone underlying the Point Pleasant-Fulton interval. The more limestone-rich Peaks Mill and Locust Creek members do contain hardgrounds, but they are few and only well expressed locally. Overlying the Point Pleasant-Fulton interval, the remaining 60+ meters of the Kope Formation contains multiple discontinuities (primarily marked concretion horizons), though these are lower in number and much lower in concentration than those in the Point Pleasant-Fulton interval. Within the Point Pleasant-Fulton interval there can be found a high density and high diversity of discontinuity surfaces marked by in situ and reworked concretions, simple and complex hardgrounds, mineralized crusts, conglomeratic limestones, and diaclast beds.

Concretion Beds: Beds of simple ellipsoidal concretions of cemented mudstone are especially common in shale hemicycles of the Fulton near Cincinnati. Up to 14 concretionary horizons occur within this 5 to 7-meter thick interval, many directly underlying thin beds and laminae of skeletal packstone and fine-grained grainstone (concretionary underbeds of Brett et al., 2003). The concretionary horizons diminish in more proximal sections. Reworked concretions and concretionary internal molds of mollusks (Fig. 16G) are common within many of the limestone hemicycles of the Fulton. These concretionary sediments are commonly encrusted, occasionally on all sides, by sheet-like bryozoans and crinoids; borings, however, are rare.
FIGURE 15-Bedforms and sedimentary structures of the Point Pleasant-Fulton interval. (A, B) Herringbone cross-bedding (A) and a firmground (B) in calcarenite near the top of the Point Pleasant member near Frankfort, Kentucky (sections A2, A3) characteristic of shoal facies. (C) Megaripples in the upper part of the Point Pleasant member on Bullskin Creek, near Point Pleasant, Ohio (cycles P7-P8). Note that the long axis of ripple crests of consecutive beds show markedly different orientations, characteristic of storm-dominated mid-ramp facies. (D) Thin, rippled grainstones with concretionary underbeds in the lower Fulton submember near Cincinnati, Ohio, characteristic of outer ramp facies.
Hardgrounds: Simple and complex hardgrounds are scattered throughout the Point Pleasant-Fulton interval. Simple hardgrounds are typified by well-preserved encrusting organisms, which occur in low diversity assemblages. The most common type of simple hardground is developed on shell pavements and exhibits thin crusts of bryozoans and/or *Trypanites* borings. Simple hardgrounds have limit aerial extent and are difficult to detect as they most commonly occur at amalgamation surfaces within limestone hemicycles. Complex hardgrounds commonly cap limestone hemicycles within the Point Pleasant and more rarely the Fulton. The primary characteristics of complex hardgrounds are multiple states of encruster preservation (Fig. 16E). Two complex hardgrounds within the Point Pleasant (cycles P2 and P5) are especially widespread (Fig. 13A, B). In places, hardgrounds occur along edges or bases of the cemented layers. Such complete undermining of sections of cemented limestone led to production of detached slabs and cobbles.

Mineralized crusts: Iron and phosphatic crusts coat many limestone surfaces, especially within and capping limestone hemicycles. These crusts are typically only a few millimeters thick, though locally may become greater than a centimeter thick. The thicker crusts tend to be heavily bored, which obliterates any primary features such as lamination. At some localities the crusts are overgrown by bryozoans and pelmatozoans (cycle P2, section A5).

Conglomeratic limestones: Conglomeratic limestone beds are also common at several levels within the Point Pleasant-Fulton interval (Fig. 16A-C, F) and represent reworking of hardgrounds and/or exhumation of cemented limestone layers, including concretions and their subsequent boring and encrustation on the sea floor, before incorporation into the final sedimentary layer. Conglomeratic limestones can be subdivided into monomictic and polymictic. Monomictic limestone conglomerates can further be subdivided into large angular blocks,
FIGURE 16-Condensed beds from the Point Pleasant-Fulton interval. (A) "Sugar Creek Bed"; a polymictic limestone conglomerate from the top of the Point Pleasant member, collected near Glencoe, Kentucky. Note the variable sizes and textures of the more than 4 different limestone clast types included within this skeletal grainstone bed. (B) Monomictic limestone conglomerate from near the middle of the Point Pleasant member (cycle P4; section B4). Many of these poorly sorted fine-grained skeletal grainstone clasts are encrusted by very small (approximately 2 millimeter diameter)ptycheldictid bryozoan holdfasts (out of resolution of image). Hammer for scale. (C) Monomictic limestone conglomerate/breccia. Large angular limestone slabs welded to the base of thick calcarenite beds from the base of the Point Pleasant (cycle P1, section m). Hammer for scale. (D) Polished bedding parallel section through the Duck Creek Bed from the top of the Fulton (cycle F8, section B9). Large crinoid columnals belong to Merocrinus, light colored hash is pyrite-replaced burrows, skeletal debris, and steinkerns. (E) Anomalocrinus encrusted hardground from the Fulton (cycle F3; section A5). This hardground shows holdfasts in a variety of preservation states encrusted an irregular surface composed of relatively well-preserved whole and fragmented brachiopod valves and columnals of pelmatazoans. (F) Monomictic limestone conglomerate. Clasts are reworked concretions encrusted on all sides by small sheet-like bryozoans and Anomalocrinus holdfasts (cycle F3, section A5). Note imbrication of concretions. Knife is 9 centimeter long. (G) Reworked, 3-dimensional, spar filled, internal mold of a small orthocone cephalopod (cycle F2, section A5). Note the presence of encrusting sheet-like bryozoan directly on spar and partially encrusting mold of the siphuncle.
smaller rounded clasts, and reworked concretions. The lower portion of the Point Pleasant locally contains large slabs (up to 50 centimeters across) of cemented skeletal pack- to grainstone (Fig. 16B, C). Such monomictic conglomeratic beds are not restricted to shallow water deposits, they also occur in down-ramp, mud-rich sections where storms processes exhume concretions (or concretionized skeletal elements; Fig. 16G). For example, a coarse-grained skeletal grainstone bed in the middle Fulton (Fig. 16F) contains abundant reworked concretions, which are encrusted on all sides. Polymictic conglomeratic beds contain mixtures of angular to rounded cobbles and/or concretions, some of which are encrusted. For example, the uppermost bed of the Point Pleasant contains at least four different limestone cobble lithologies (Fig. 16A).

Pyrite/diaclast beds: Thin beds and laminae of reworked and concentrated small phosphatic and pyritic diaclasts including small molluscan steinkerns and granules occur at the top of the Fulton in the Cincinnati area and are more common throughout the Point Pleasant and Fulton in the C4 and C5 sections (Fig. 16E,D). These diaclast beds/laminae also commonly contain abundant conodonts, orbiculoid brachiopods, and ostracods.

Interpretation of discontinuity surfaces

The various discontinuity surfaces described above, with their increasing levels of complexity, are here interpreted to represent a general hierarchy of increasing degrees of sedimentary condensation. Simple hardgrounds only required exposure of a hard substrate to form; this may have entailed as little as cementation of sediments at the sea floor or may have involved exhumation of a bed cemented within the zone of sulphate reduction tens of centimeters down into the sediment column. Carbonate concretions within organic-rich shales of the Fulton might have formed over a similar duration within the zone of sulphate reduction. Complex hardgrounds were certainly of a longer duration as they contain evidence of multiple encrustation
events. Monomictic limestone conglomerates and reworked concretions beds share many similarities with complex hardgrounds and probably formed over a similar duration, perhaps as much a few thousand years to incorporate lithification, erosion, undercutting, rounding, and resedimentation. Mineralized crusts share many physical attributes in common with those described from the Jurassic (Palmer and Wilson, 1990) and Cretaceous (Soudry and Lewy, 1990). In these Mesozoic examples, the crusts are interpreted as the byproduct of microbial (possibly bacterial, algal, or fungal) activity on the sea floor during periods of sediment starvation. As for the time it would take to form such crusts, no estimates have been given, however, the two Mesozoic examples are components of major discontinuities. It is likely that the Upper Ordovician examples formed by similar processes and were generated during periods of sediment starvation. Polymictic conglomeratic beds are the most complex discontinuities in the study interval and are commonly associated with mineralized crusts. Surrounding matrix frequently contains a mixture of well-preserved and reworked skeletal fragments and phosphatic granules and steinkerns, indicating time averaging and prolonged exposure at the sea floor. Thin beds pyrite/diaclast beds appear to represent intervals of condensation in more offshore, deep-water setting where carbonate production was minimal. For example, in section C5 limestone hemicycles of the Point Pleasant are represented only by similar thin pyrite/diaclast beds surrounded by mostly barren organic-rich shale.

The high abundance of discontinuity surfaces in the Point Pleasant-Fulton interval in contrast to surrounding strata suggests that it is relatively condensed. The large angulate cemented blocks associated with the basal limestones of the Point Pleasant (Fig. 15C; cycle P1) suggest erosional reworking of thick, cemented layers by relatively high-energy events. The high frequency of complex hardgrounds, monomictic limestone conglomerates and mineralized crusts in the Point
Pleasant are closely associated with the limestone hemicycles and suggest episodic sediment starvation. The polymictic limestone conglomerate bed at the top of the Point Pleasant member (Fig. 8A; cycle P8) appears to represent the most highly condensed horizon within the study interval and indicates a propensity toward more prolonged condensation toward the middle of the Point Pleasant-Fulton interval. This interpretation is supported by the distribution of concretions within the Fulton, which are commonly most concentrated near its base.

TAPHONOMIC GRADIENTS OF THE POINT PLEASANT-FULTON INTERVAL

Taphonomic Data Collection and Analysis Methods

The established cycle framework for the Point Pleasant-Fulton interval facilitates semi-quantitative assessment of trends in sorting, skeletal grain preservation (fragmentation and abrasion), and the degree of mixing of preservation-states over a broad area. One major difficulty encountered in taphonomic analysis is that time-averaging can generate beds which display mixed populations of well- to poorly-preserved taxa, making assessment of the overall preservation state of the skeletal grains for a given bed inaccurate. To deal with this problem fragmentation and abrasion were judged on the most prevalent skeletal grain size (i.e. coarse sand) with the added appraisal of the degree of mixing of different preservation-states to ensure a representative measurement of the taphonomic variability within the bed. These five categories (sorting, fragmentation, fragmentation mixing, abrasion, and abrasion mixing) were evaluated at the bed level in five sections (A3, A4, A5, and B10) across the study area. Each sample was scored as high, moderate, or low for each of the five categories. Using this methodology, the taphonomic characteristics of a total of 233 bedding planes (predominantly limestone), each ranging from 10 to 40 squares centimeters, were recorded. The data were
FIGURE 17-Examples of biotic and taphonomic composition of limestone hemicycles (A-C) shown across from adjacent shale hemicycles (D-F), arranged stratigraphically for the lower (D, F), middle (B, E), and upper (A, C) Point Pleasant. (A) Skeletal grainstone from the uppermost bed of the Point Pleasant at Swallowfield, Kentucky (section A4). Note the moderate to high degree of mixing of various preservation states of this Ectenocrinus, Merocrinus, thin bifoliate, and thin ramose bryozoan-dominated assemblage and association with one of four different lithologies of reworked limestone clast (lower part of the image) that have been documented in this bed at this locality; contrasted against, (D) an argillaceous packstone containing abundant gastropod (g) and lesser bivalve (b) and cephalopod (c) clay-filled internal molds that display little to no abrasion or fragmentation and little evidence of mixing of preservation types. (B) Well-sorted Glyptocrinus grainstone displaying a moderate to heavy degree of fragmentation and abrasion, with a moderate degree of mixing of preservation types; contrasted against, (E) adjacent argillaceous packstones containing moderately fragmented and little abraded thick ramose and thick bifoliate bryozoans, Rafinesquina, and a minor component of Glyptocrinus columnals, which show a low degree of mixing of preservation types. (C) Well-sorted, highly fragmented and abraded skeletal grains from a lower Point Pleasant calccarenite; contrasted against (F) slightly argillaceous sparry packstone containing moderately to heavily fragmented and abraded Cyclonema and thick ramose and thick bifoliate bryozoans in a skeletal hash with a moderate degree of mixing of preservation types.
FIGURE 18-Examples of biotic and taphonomic composition of limestone hemicycles (A-C) shown across from adjacent shale hemicycles (D-F), arranged stratigraphically for the lower (D, F), middle (B, E), and upper (A, C) Fulton. (A) Sowerbyella-dominated grainstone exhibiting low levels of sorting and abrasion and moderate levels of fragmentation, with low to moderate levels of preservation-state mixing; contrasted against (D), a thin calcareous siltstone containing a low to moderate degree of sorting and orientation of skeletal grains, a low degree of fragmentation and abrasion with a low degree of mixing of preservation states. (B) Typical skeletal grainstone of the middle and lower Fulton displaying a low degree of sorting and abrasion, low to moderate levels of fragmentation, partially mineralized skeletal grains, and a moderate to occasionally high degree of mixing of preservation states; contrasted against (E) argillaceous packstone displaying a low degree of sorting, however with some orientation of skeletal grains, a low amount of abrasion, moderate to low fragmentation, and a low to moderate degree of mixing of preservation types. (C) Merocrinus-Cheirocrystis "log jam" capping a 15 centimeter-thick crinoid grainstone. The articulated and oriented nature of these echinoderm elements suggests rapid burial under high-energy conditions, however, it forms only a thin veneer on a grainstone containing skeletal grains that are poorly sorted, show low abrasion and low to moderate fragmentation, and a moderate to high degree of mixing of preservation types. This type of juxtaposition of time-poor on time-rich taphofacies is typical of the tops of many limestone hemicycles within the upper Point Pleasant and Fulton interval and suggests a final catastrophic event served to bury the bed permanently. Contrasted against (F) an argillaceous packstone displaying a low degree of sorting, low abrasion and fragmentation, with little mixing of preservation states.
FIGURE 19-Characteristic biota and preservation of taxa from the Point Pleasant-Fulton interval in cores from southwestern Ohio. (A) Bedding plane view of dense grouping of fragmented valves of small orbiculoid brachiopods, ostracods, and conodont elements in the Point Pleasant equivalent brown shales of the Sebree Trough (section C5). (B) Cross-sectional view of bed shown in A. (C) Dense accumulation of cephalons from the blind trilobite Cryptolithus preserved in dark brown shale of the upper Fulton interval (section B11). (D) Typical preservation of Triarthrus molts in dark brown shale of the lower Fulton (Reilly Township core).
analyzed in PC-ORD version 4.0 using nonmetric multidimensional scaling (NMS) ordination; the results are displayed in figure 20.

Results of Taphonomic Analysis

Lateral tracking of taphonomic variables within the Point Pleasant-Fulton interval across the study area reveals a gradient from high values in central Kentucky to successively lower values in the surrounding areas; this pattern is especially true of the Point Pleasant (Fig. 17-20). In sections from Frankfort eastward to Winchester, nearly the entire Point Pleasant is composed of highly sorted and heavily abraded and fragmented skeletal grains with little to no mixing of preservation-states (Fig. 7, 17E-F, 20). In the same sections, the overlying Fulton shows a much more heterogeneous taphonomic composition and never reaches the same extreme degrees of sorting and skeletal degradation shown in the Point Pleasant (Fig. 18E-F, 20, 21). Phosphate, occurring as fine-silt to sand size light bluish grey amorphous masses and steinkerns of the diminutive gastropod Cyclora, small bivalves, and bryozoan zooecia, is disseminated throughout the Point Pleasant in this area. Away from the Frankfort-Winchester area the phosphate content decreases concomitant with a change to increasingly more heterogeneous preservation of skeletal grains, more similar to the Fulton. By contrast, at Cincinnati and in core sections to the north, skeletal grains generally exhibit low abrasion and only moderate degrees of fragmentation, however thin mineral coatings are more common (see below); in these sections small skeletal grains and steinkerns are occasionally altered to authigenic pyrite, as in one of the thin capping beds of the Fulton at section B9 (Fig. 16D-E).

Vertical heterogeneity in taphonomic composition is especially prominent in sections intermediate between Frankfort and Cincinnati where small-scale lithologic cyclicity is well expressed both in the Point Pleasant and Fulton, though each unit shows slightly different
FIGURE 20-Q-mode NMS scatter plot of taphonomic data from sections A3, A4, A5, and B11. Samples are coded by stratigraphic unit, locality, and limestone versus shale hemicycle. Clustering of sample points is an artifact of the low number of variables and preservation categories (i.e. low, medium, high) in the analysis. Note, rather, the overall spread of sample points along axes 1 and 2. The majority of Point Pleasant samples from near Frankfort plot near the upper left-hand corner, whereas, the remaining Point Pleasant and Fulton limestone hemicycle samples are arrayed along Axis 2. Samples from shale hemicycles of the Point Pleasant and Fulton are scattered along Axis 1. Size and position of light gray arrows indicate relative contribution of different variables (derived from NMS correlation coefficients) to axes 1 and 2.
overall taphonomic trends (Fig. 21). In these areas limestone hemicycles generally contain high to moderately sorted skeletal grains displaying a broad range of preservation, whereas moderate to poorly sorted and relatively well-preserved skeletal grains dominate the shale hemicycles (Fig. 17, 18, 20, 21). Additionally, skeletal grains within the limestone hemicycles, especially brachiopods, bryozoans and trilobites display a color range from blackened (phosphatized?) to reddened (iron-rich cements?), that are in sharp contrast to coloring in adjacent well-preserved specimens, more representative of the shale hemicycles. The taphonomic composition of the hemicycles differs further as internal molds found within the shale hemicycles are largely composed of mudstone (or micrite?), whereas internal molds found within the limestone hemicycles are much more commonly spar-filled (Fig. 16G, 17B). Spar-filled internal molds of orthoconic nautiloids are especially common in limestone hemicycles of the Fulton, which show a relatively consistent overall taphonomic signature vertically (Fig. 18, 20, 21). Conversely, limestone hemicycles of the Point Pleasant show a marked upward change from heavily processed skeletal grains more representative of the interval at Frankfort, including concentration of phosphate steinkerns, to more heterogeneous preservation of skeletal grains, similar to the Fulton limestone hemicycles (Fig. 17, 20, 21).

Interpretation of Taphonomic Gradients

Patterns revealed by taphonomic analysis suggest multiple competing local, basinal, and extrabasinal controls on the distribution of skeletal grain preservation types. High levels of mixing of preservation states suggest that the majority of the beds making up the limestone hemicycles are time-averaged (sensu Kidwell and Bosence, 1991; Speyer and Brett, 1991). This interpretation is supported by color variability of the skeletal grains, interpreted as authigenic mineralization in response to variable residence times on the sea floor. However, a patchy
distribution of preservation-types is also sometimes observed within limestones at the bed level. This patchiness, found within both limestone and to lesser extent shale hemicycles, is interpreted as the localized, random effects of storm generated currents. The prominence of megaripples and sole marks in addition to the presence of gutter casts and aligned skeletal grains is well documented in this interval outside of the present study (Weiss et al., 1965) and genesis of these features by storm-generated currents is a long-standing interpretation in overlying strata (Jennette and Pryor, 1993). The broader, lateral taphonomic gradients documented across the study area are interpreted as a response to regional-scale control on sea floor topography generated by far-field tectonic stresses imposed by the Taconic Orogeny (Ettensohn et al., 2002; McLaughlin et al., 2004). Alternatively, the vertical taphonomic gradient is interpreted as a response to two different scales of eustatic fluctuations. Small-scale cyclicity, corresponding to limestone and shale hemicycles, corresponds to the most common vertical variation in preservation-state. The larger scale of cyclicity becomes apparent only when the small-scale variability is removed by vertically comparing only the limestone hemicycles or only the shale hemicycles to one another. This larger scale of cyclicity shows a marked increase in the general preservation-state of skeletal grains upward within the Point Pleasant, interpreted as decreasing levels of hydraulic processing. Alternatively, the small degree of upward change in the Fulton submember suggests little net change in relative sea level, interpreted as a near balance between continued, but slowing, eustatic rise and increasing siliciclastic sedimentation.

FAUNAL GRADIENT ANALYSIS

Methods of Faunal Data Collection and Analysis

Faunal gradient analysis, like taphonomic gradient analysis, yields a wealth of information about strata that show little obvious net lithologic change. It has been pointed out that the
distribution of taxa within a given succession may be a more sensitive bathymetric indicator than lithology alone (Holland et al., 2001). The two techniques are easily combined, as faunal gradient analysis first requires a faunal census at which time the conditions of the skeletal grains can be recorded. Faunal census data were collected, generally following the technique of Miller et al. (2001), from the same 233 samples as those in the taphonomic analysis. The relative abundance of all taxa (bryozoans were lumped into categories based on colony morphology) was recorded as rare (1-2 specimens), common (3-12 specimens), or abundant (>12 specimens).

Several permutations of the data were statistically analyzed in PC-ORD version 4.0 before the data set was finalized. Statistical analysis utilized detrended correspondence analysis (DCA) ordination, which is commonly used to identify trends in ecological data sets (McCune and Mefford, 1999). Analysis of various culled versions of the data set revealed little variation in pattern, therefore the results presented here represents the initial 233-sample data set.

Results and interpretations of the faunal gradient analysis

Faunal gradient analysis the Point Pleasant-Fulton interval yields patterns of faunal change which support an interpretation of deepening upward, though with slightly different trends for the Point Pleasant and Fulton individually. The results of the DCA analysis on the final faunal data set are shown in two scatter plots (Fig. 22). Axis 1 scores for the taxa (Fig. 22B), representing approximately 70% of the variation in the data (eigen value of 0.67), are interpreted to represent distribution along a water depth gradient; proceeding from shallow (low scores) on the left to deeper (high scores) on the right. In general, the platyceratid gastropod Cyclonema, small modiomorphid bivalves, massive bryozoans, and the robust brachiopods Hebertella and Platystrophia yield low scores interpreted to represent relatively shallow water conditions, whereas, the more delicate thin-shelled strophemenid brachiopod Sowerbyella, small lingulid
FIGURE 21- NMS Axis 2 taphonomic scores (light gray lines) and DCA Axis 1 faunal scores (black lines) plotted against stratigraphic position for four sections (A3-A5 and B10) across the study representing a gradient of lithofacies. Note that the overall trends in both the taphonomic and faunal data are very similar suggesting an upward trend toward lower environmental energy (interpreted from NMS Axis 2 taphonomic scores, see figure 20) and greater water depths (interpreted from DCA Axis 1 faunal scores, see figure 21). Variability in small-scale oscillations between the two curves reflect a closer correspondence of the taphonomic data to the lithologic small-scale cyclicity designated in figures 12 and 13.
brachiopods, and the trilobite Triarthrus becki show the highest scores. The dominance of phosphatic internal molds of Cyclora on the shallow end of the gradient is a taphonomic artifact of high concentrations of phosphate in shoal deposits of the Lexington Limestone as noted above in the taphonomy discussion.

The distribution of variables along DCA Axis 1 is similar, though not identical, to results from faunal gradient analyses of the surrounding strata by Holland et al. (2001) and Holland and Patzkowsky (2004). Comparison of the distribution of taxa along DCA Axis 1 common to the present study and the Holland and Patzkowsky (2004) faunal gradient analysis of the entire Lexington Limestone in central Kentucky (including a few samples that overlap with this study) yields a Spearman’s Rank correlation coefficient of +0.76 (N=18; +1.0 being a perfect positive correlation and −1.0 being a perfect negative correlation). Comparison of the distribution of taxa along DCA Axis 1 common to this study and the Holland et al. (2001) study of the entire Kope Formation (though samples of the Fulton submember were relatively rare and do not coincide with any samples from this study) and lower Fairview Formation in the Cincinnati, Ohio area yield a much lower correlation coefficient of +0.48 (N=23). However, comparison of only the most abundant taxa common to both studies (N=13) yields a much higher correlation coefficient of +0.88. It is important to note that previous studies, primarily focusing on lithofacies analysis, suggested that the shale-rich Kope Formation was deposited in a substantially deeper, quieter water depositional setting than that of the limestone-rich Point Pleasant (Cressman, 1973; Weir et al., 1984; Pope and Read, 1997). However, comparison of faunal gradient analysis results from this study with those of Holland et al. (1991) indicates similar paleobathymetry for portions of these two, lithologically dissimilar, successions.
Plotting of DCA Axis 1 scores from the faunal analysis against stratigraphic position reveals both small-scale variations and larger-scale unidirectional trends (Fig. 21). Small-scale alternations in DCA Axis 1 scores only partially correspond to lithologic cyclicity and likely record an overprint of original sea floor patchiness preserved in the fossil record (sensu Webber, 2003). Larger-scale trends suggest overall deepening upward in both the Point Pleasant and the Fulton, though typically with slightly different trajectories. These large-scale trends vary systematically across the sampled transect, with vertical trajectories representing threshold effects both on the shallow (i.e. the Point Pleasant at section A3) and deep (Fulton at section B10) ends of the gradient.

BIOEVENTS (EPIBOLES)

Epiboles are defined as the abundant occurrence of normally rare or absent taxa across a broad geographic area over a narrow stratigraphic interval (Brett and Baird, 1997). They are typically concentrated within specific parts of depositional sequences, though are not restricted to any specific systems tract (Brett, 1995). Examples of proliferation and incursion epiboles occur within the Point Pleasant-Fulton interval. A proliferation epible occurs when an organism that is present in low numbers suddenly becomes highly abundant. Incursion epiboles, by contrast, occur when a locally absent organism suddenly appears, commonly in large numbers, is only present through a thin stratigraphic interval, over broad areas, and then is absent once again.

Characteristics of epiboles within the Point Pleasant-Fulton interval

Merocrinus epible: The first and most prolific epible taxa to appear within the study interval is that of the cladid crinoid Merocrinus. It occurs within the uppermost beds of the Point Pleasant and in the overlying Fulton (Fig. 13), an occurrence that was well documented by McFarlan and Freeman (1935). Merocrinus has a long stem, up to 1 meter, capped by a relatively
small cup and pinulate branching arms, which form a relatively low-density filter. This morphology is common amongst Upper Ordovician cladid crinoids, which commonly score on the deep end of faunal gradient analysis (Fig. 22; Holland et al., 2001; Meyer et al., 2002). *Merocrinus* is largely absent from the underlying and overlying strata, with only a few scattered columnals recognized near the underlying Devils Hollow/Bromley contact (Davies, 1958). Additionally, Meyer et al. (2002), used columnals to document the occurrence of crinoid genera throughout the Kope Formation; these authors did not identify *Merocrinus* in any samples above the Fulton. Within the study interval Merocrinus co-occurs with a wide variety of taxa. The distribution of *Merocrinus* as a common to abundant skeletal element of limestones within the Point Pleasant-Fulton interval closely parallels the margin of the Sebree Trough and can be traced over an area of greater than 5500 square kilometers (Fig. 23A).

Cheirocystis *epibole*: The small rhombiferan cystoid *Cheirocystis fultonensis* occurs within a narrow stratigraphic interval, typically a single, thin, argillaceous packstone bed (typically 1- to 2-centimeters thick) and surrounding shales. The bed occurs near the base of the F1 cycle, is widespread along strike (Fig. 23B), paralleling the Sebree Trough, similar to the *Merocrinus* epibole. In many localities in the Ohio River Valley the bed is commonly composed of densely packed thecal plates (Sumrall and Schumacher, 2002; Brett et al., 2003). The cystoid bed also includes other typical Fulton taxa (although in much lower concentrations) including *Aspidopora, Hallopora, Onniella, Merocrinus, Ectenocrinus, and Flexicalymene*. The *Cheirocystis* epibole can be traced for over 3000 square kilometers across the study area (Fig. 23B).

Triarthrus *epibole*: *Triarthrus*, a small olenid trilobite, occurs in dark brownish gray, organic-rich shales of the Fulton (cycles F1-F4; Fig. 13B-C, 19D). When they occur in large numbers
FIGURE 23—Location maps marking the distribution of Point Pleasant-Fulton epiboles. (A) Abundant *Mericrinus* distribution. (B) *Cheirocystis* distribution. (C) Common and rare *Triarthrus* distribution. Solid lines demarcate areas with close outcrop or core control; dashed lines are extrapolations of trends based on facies distribution. Note how the distributions of epiboles closely parallel the Sebree Trough.
(typically the case) they are often found in monospecific assemblages, predominately as molted cephalons, but scattered individuals co-occur with lingulid and orbiculoid inarticulate brachiopods, *Dalmenella, Merocrinus, Isotelus, Cryptolithus* and nuculid bivalves (Fig. 19A-C). Examination of exposures near Cincinnati and southwestern Ohio core sections suggest that the underlying organic-rich shales of the Bromley and Gratz members of the Lexington Limestone largely lack *Triarthus*. Similarly, all but a thin zone in the Alexandria submember of the overlying Kope Formation, approximately 40 meters up section, lack *Triarthus*. The Fulton *Triarthus* epibole occurs in outcrop over an area of approximately 25 square kilometers and can be traced an additional 75 kilometers to the north in drill core (Fig. 23C).

**Interpretation of Point Pleasant-Fulton epiboles**

The occurrence of multiple epiboles within the Pleasant-Fulton interval strongly suggests an organismal response to shifting environmental conditions. Further, the close vertical clustering of these epiboles is interpreted as a response to sedimentary condensation. It seems significant that these epibole taxa first appear in the study area near the Point Pleasant-Fulton contact. As stated above, the sedimentary features surrounding this contact suggest that it is the most condensed part of the study interval and represents the most rapid facies change.

The epibole taxa appear to be largely endemic to deeper portions of the Taconic foreland basin and their immigration onto the Lexington Platform during deposition of the later Point Pleasant and Fulton strata is interpreted as a response to rising sea level. The spread of *Merocrinus* and other dysoxic tolerant faunas as far as the shallow Frankfort area supports the interpretation that dysoxic waters of the Sebree Trough expanded into adjacent ramp areas during deposition of the upper Point Pleasant-Fulton interval (cycles P7-F8). Indeed the main prolific cystoid bed grades down ramp into an interval showing the first incursion of *Triarthus*. 
The immediately overlying dark brown shales are mostly barren and we suspect that this marks a major onlap of dysoxic water, which may have extinguished the cystoids from the Lexington Platform. *Triarthrus* belongs to the olenids, a holdover group typical of the Cambrian evolutionary fauna (Fortey, 2004) that was largely restricted to low oxygen conditions (Henningsmoen, 1957), some members of which may have been symbiotic with sulfur bacteria (Fortey, 2000). Hence, unusual appearances of these trilobites in dark shales of the Fulton strongly support the interpretation that a dysoxic and possibly cool, nutrient-rich water mass shifted up ramp in response to rising sea level.

The connection of epiboles to the transgressive systems tract is in part predicted by modeling of first and last occurrences of taxa (Holland and Patzkowsky, 2002). Following the conclusions of Holland and Patzkowsky (2002) one could argue that the occurrence of these epiboles simply represents extreme range offsets generated by the additive effects of sedimentary condensation on rare and facies specific taxa. This explanation would be adequate if the organisms were truly rare both stratigraphically and spatially. However, many epibole taxa, including those discussed here, are common to abundant skeletal elements of the horizons in which they occur. Additionally, the epibole taxa are no more facies specific than the stratigraphically broad-ranging taxa with which they co-occur. For example, Holland et al. (2001) included samples of the Fulton within their analysis of the Kope Formation and found that *Merocrinus* shows a very similar peak abundance and depth tolerance to associated taxa. An interesting twist is that the epibole taxa discussed here are known to be relatively long-ranging taxa in the laterally equivalent and overall deeper water facies of the Trenton-Utica strata of central New York State (www.mcz.Trenton.com). Thus, it appears that the TST, especially the late TST, is marked by
brief incursions of exotic faunas, possibly in response to shifting water masses in the epicontinental sea.

**DISCUSSION AND COMPARISON: TOWARD A GENERAL MODEL OF TRANSGRESSIVE LIMESTONES**

The case study presented above provides several lines of evidence that the Point Pleasant-Fulton interval is not simply a shoal deposit capping a parasequence, but rather contains several characteristics suggestive of transgression. Widespread distribution, regional geometry, vertical cycle stacking pattern, relatively high abundance of discontinuity surfaces, and close agreement between faunal and taphonomic gradient analysis (Fig. 21), all give strength to the interpretation of deepening upward through this interval, though with the caveat of division into an early and late phase of the TST. Lateral tracing of small-scale cycles within the Point Pleasant-Fulton interval reveals that it is nowhere equivalent to a thicker (down ramp) shale-dominated interval representing an expanded equivalent resulting from redeposition of muds winnowed from up-ramp areas. Therefore, we propose that while winnowing plays an important role in the formation of skeletal grainstone-rudstone successions, especially in shallow portions of the ramp, the widespread nature of the Point Pleasant-Fulton interval requires siliciclastic sediment starvation on the foreland ramp as a primary control in its genesis. Thereby, during sea level rise, skeletal sediment blankets may cover large portions of the ramp/shelf, and are not just restricted to high-energy “shoals”, although such shoals form an important component of TSTs in up-ramp settings.

Having made the transgressive case for the Point Pleasant-Fulton interval the question then becomes, are the characteristics described above shared with other widespread limestone-rich intervals that occur in mixed carbonate-siliciclastic foreland basin deposits? Table 1 shows a
<table>
<thead>
<tr>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>Series</td>
<td>Caradoc</td>
<td>Caradoc</td>
<td>Caradoc</td>
<td>Ashgill</td>
<td>Wenlock</td>
<td>Wenlock</td>
<td>Eifelian</td>
<td>Givetian</td>
<td>Givetian</td>
</tr>
<tr>
<td>Name</td>
<td>Pt. Pleasant-Fulton</td>
<td>Curdsville</td>
<td>Strodes Creek</td>
<td>Fairview</td>
<td>Irondequiot</td>
<td>Gasport</td>
<td>Onondaga</td>
<td>Givetian</td>
<td>Hungry Hollow</td>
</tr>
<tr>
<td>Locality</td>
<td>KY, OH, PA, NY, ON</td>
<td>KY, OH, PA, NY, ON</td>
<td>Kentucky</td>
<td>KY, OH</td>
<td>KY, OH, IN, PA</td>
<td>KY, OH, IN, PA</td>
<td>ON, MD, WV, VA</td>
<td>ON, NY, PA</td>
<td>ON, NY, PA</td>
</tr>
<tr>
<td>Orogeny</td>
<td>Taconic</td>
<td>Taconic</td>
<td>Taconic</td>
<td>Taconic</td>
<td>Salinic</td>
<td>Salinic</td>
<td>Acadian</td>
<td>Acadian</td>
<td>Acadian</td>
</tr>
<tr>
<td>FACIES FEATURES</td>
<td>Thickness</td>
<td>Shoal Facies</td>
<td>Inner-ramp Facies</td>
<td>Mid-ramp Facies</td>
<td>Heterolithic, skeletal grainstone &amp; shale</td>
<td>Down-ramp Facies</td>
<td>Concretionary, thin, fine-grained grainstone</td>
<td>Surrounding Strata (mid-ramp)</td>
<td>Overlying facies</td>
</tr>
<tr>
<td></td>
<td>7-14 m</td>
<td>yes</td>
<td>yes</td>
<td>yes</td>
<td>yes</td>
<td>yes</td>
<td>yes</td>
<td>yes</td>
<td>shoal facies</td>
</tr>
<tr>
<td></td>
<td>4-6 m</td>
<td>no</td>
<td>thin bedded</td>
<td>yes</td>
<td>yes</td>
<td>yes</td>
<td>yes</td>
<td>yes</td>
<td>trough t-bedded skeletal shoal</td>
</tr>
<tr>
<td></td>
<td>1-2.5 m</td>
<td>yes</td>
<td>yes</td>
<td>yes</td>
<td>yes</td>
<td>yes</td>
<td>yes</td>
<td>yes</td>
<td>inner-ramp facies</td>
</tr>
<tr>
<td></td>
<td>6-10 m</td>
<td>yes</td>
<td>yes</td>
<td>yes</td>
<td>yes</td>
<td>yes</td>
<td>yes</td>
<td>yes</td>
<td>calcilutite frazilite</td>
</tr>
<tr>
<td></td>
<td>3-5 m</td>
<td>yes</td>
<td>yes</td>
<td>yes</td>
<td>yes</td>
<td>yes</td>
<td>yes</td>
<td>yes</td>
<td>mid-ramp facies</td>
</tr>
<tr>
<td></td>
<td>2-6 m</td>
<td>yes</td>
<td>yes</td>
<td>yes</td>
<td>yes</td>
<td>yes</td>
<td>yes</td>
<td>yes</td>
<td>mid-ramp facies</td>
</tr>
<tr>
<td></td>
<td>1-7 m</td>
<td>yes</td>
<td>yes</td>
<td>yes</td>
<td>yes</td>
<td>yes</td>
<td>yes</td>
<td>yes</td>
<td>down-ramp facies</td>
</tr>
<tr>
<td></td>
<td>0.5-2 m</td>
<td>yes</td>
<td>yes</td>
<td>yes</td>
<td>yes</td>
<td>yes</td>
<td>yes</td>
<td>yes</td>
<td>surrounding strata (mid-ramp)</td>
</tr>
<tr>
<td></td>
<td>0.3-1 m</td>
<td>yes</td>
<td>yes</td>
<td>yes</td>
<td>yes</td>
<td>yes</td>
<td>yes</td>
<td>yes</td>
<td>overlying facies</td>
</tr>
</tbody>
</table>

TABLE 1—Comparison of widespread skeletal limestones throughout the middle Paleozoic of eastern North America. Point Pleasant, Curdsville, and Strodes Creek are members of the Lexington Limestone (see discussions in Cressman, 1973; Pope and Read, 1997a; McLaughlin et al., 2004; Brett et al., 2004). Fairview refers to the basal, limestone-dominated 3 to 5 meters of the Fairview Formation (see discussions in Jennette and Pyor, 1993; Holland, 1993; Holland et al., 2001; Algeo and Brett, 2001). Irondequiot refers to the Irondequiot Formation of the Clinton Group (see discussions in Brett et al., 1990; LoDuca and Brett, 1994). Gasport refers to the Gasport Formation of the Lockport Group (see discussions in Brett et al., 1990; Goodman and Brett, 1994). Onondaga refers to the Edgecliff Member of the Onondaga Formation and Union Springs Member of the Marcelus Shale (see discussions in Baird et al., 2000; Ver Straeten and Brett, 2000). Hungry Hollow is a member of the Widder Formation (see discussion in Bartholomew et al., 2006). Tichner is a member of the Moscow Formation (see discussion in Grabeau, 1899; Baird, 1979). p: pelmatazoan. by: bryozoan, b: brachiopod. t: trilobite. c: coral. prods: large burrow fillings projecting downward from base of bed. g-rstn: grainstone-rudstone. sl. relief: slight relief. L: Lower. M: Middle. U: Upper. Ord: Ordovician. Sil: Silurian. Dev: Devonian. NA: information not available. NR: information not reported in literature, though further fieldwork may reveal more data.
comparison between characteristics noted in the Point Pleasant-Fulton interval and other major limestone units in eastern North America, which suggests that indeed these intervals share much in common in terms of their lateral facies distribution, sequence stratigraphic features, taphonomic, and biotic feature. However, some examples contain interesting departures as well.

As noted in the introduction, few widespread limestone-rich intervals are as argillaceous as the Point Pleasant-Fulton interval. Even greater departures can be seen in both the Upper Ordovician Strodes Creek Member of the Lexington Limestone and the Middle Devonian Hungry Hollow and Tichner limestones. Similar to the other examples they are widespread, however, they are also relatively thin lacking small-scale cyclicity. Both show evidence of time-averaging (e.g. see Baird, 1979), but analysis of vertical taphonomic or faunal gradients is very difficult to impossible as they commonly occur as massive beds. Especially notable in the Tichner is the extremely high diversity of this coral-rich bed compared to the surrounding calcareous mudstones and thin limestones (Grabau, 1899; Baird, 1979). The Strodes Creek is interpreted as a 4\textsuperscript{th}-order TST (McLaughlin et al., 2004), whereas, the Tichner and Hungry Hollow are interpreted as 3\textsuperscript{rd}-order TSTs. As the lack of small-scale cyclicity is likely an artifact of condensation in these examples its absence is not as easily explained in Fairview example. The lower limestone-rich portion of the Fairview Formation forms the C2 TST, the first major limestone unit following the Point Pleasant-Fulton interval, and other than a lack of small-scale cyclicity is quite similar. Rather than differentiation into limestone and shale hemicycles the lower Fairview is largely composed of a largely varying succession of medium bedded (approximately 10 centimeter) grainstones split by thin shale beds and partings. The lack of small-scale cyclicity displayed by the lower Fairview is common to many (Upper Ordovician) Maysvillian to Richmondian age limestone-rich intervals. In total the comparative analysis
provided in Table 1 leads us to the conclusion that not all transgressive limestones are created equally and their final stratigraphic signature will be in part dependent on basin dynamics (i.e. siliciclastic sedimentation rates and subsidence), and the magnitude (and thus rate) and frequency of sea level fluctuations.

CONCLUSION

The discussion above suggests a new model for the genesis of widespread limestones/limestone-rich intervals in mixed carbonate-siliciclastic foreland basin successions is required. Such a model, requiring extensive testing for generalities of pattern, could help to explain a number of interrelated physical, geochemical, and biotic patterns. It appears that the combined effects of relative sea level rise and resultant low siliciclastic sedimentation rates have several synergistic effects on benthic faunas and their taphonomy. First, the direct effect of sea level rise results in a shift of bathymetrically-related taxa in an up-ramp direction. Second, relatively clear water resulting from siliciclastic sequestration permits proliferation of oligotrophic, filter feeders. Third, low siliciclastic sedimentation rates, coupled with skeletal accumulation and early cementation results in substrate modification and taphonomic feedback (sensu Kidwell and Jablonski, 1983). To summarize, a general model of transgressive limestones within mixed carbonate-siliciclastic foreland basin systems would include:

(1) An abundance of hardgrounds or other internal discontinuities relative to the surrounding strata, with increasingly complex beds at more significant discontinuity surfaces, e.g. the maximum starvation surface.

(2) Increased abundance of authigenic mineralization.

(3) An increase in abundance and/or diversity among organisms that are inferred to be specialized for hard substrates and low turbidity environments.
(4) Clustering of epiboles, especially close to the MSS.

(5) A two-part division of transgressive limestones into early and late phases separated by the maximum starvation surface.

(6) An early transgressive systems tract which contains:

   (a) A general overall geometry, which is condensed by winnowing in shoal areas, expands slightly into the mid-ramp, and pinches markedly down ramp into a thin condensed interval.

   (b) Taphonomic gradients of upward improving preservation and mixing of preservation-states of skeletal grains, manifested poorly in the massive carbonates of shallow water areas, but more prominently in mid-ramp heterolithic facies.

   (c) Biotic gradients of upward change from generally robust forms to more thin shelled or gracile forms.

(7) A late TST comprising a shift to more offshore, siliciclastic mud-rich facies; these intervals may be organic-rich and contain:

   (a) A general overall geometry of slight thinning by winnowing in up-ramp areas, expanding into the mid-ramp, and expanding still further toward the basin center accommodated by the shale hemicycles while the limestone hemicycles pinch out.

   (b) An upward increase in cycle thickness and/or siliciclastic grain size.

   (c) Near absence of net vertical change in taphonomic gradient.

   (d) Near absence of net vertical change in taxonomic gradient.

Finally, the specificity of this study to foreland basin strata is intentional; it is not clear at this point that mixed carbonate-siliciclastic successions form on passive margins and in foreland basins in the same manner. A few important aspects to consider when comparing these two
systems include: (1) much greater subsidence rates in passive margin systems, (2) sediment transport direction across depositional dip (unidirectional on passive margins versus bi-directional in foreland basins), and (3) variability in carbonate production rates and types of carbonate sediments produced on a shelf facing the open ocean (including possible upwelling) versus within a foreland basin (possibly stratified and/or stagnant at times).

ACKNOWLEDGMENTS

The authors would like to thank the Kentucky and Ohio geological surveys for assisting us with access to drill core. Special thanks to: Gordon Baird and Alex Bartholomew for thoughtful discussion and keen eyes in the field, Steve Felton for many stimulating and insightful conversations of Upper Ordovician stratigraphy and paleontology, Dr. Mark Wilson for his enthusiasm and support, and Stefano Dominici for his encouragement and patience. The paper benefited greatly from the detailed review of Steven Holland. This work incorporates aspects of masters and doctoral work of PIM at the University of Cincinnati and was funded in part by the Department of Geology at the University of Cincinnati. Acknowledgement is also made to the donors of The American Chemical Society Petroleum Research Fund for partial support of this research.
REFERENCES


Bartholomew, A., Brett, C.E., DeSantis, M., Baird, G., Tsujita, C., 2006, Sequence stratigraphy of the Middle Devonian at the border of the Michigan Basin: correlation with New York and


Dunham, R.J., 1962, Classification of carbonate rocks according to depositional texture:


McLaughlin, P.I., Brett, C.E., Taha McLaughlin, S.L., and Cornell, S.R., 2004, High-resolution sequence stratigraphy of a mixed carbonate-siliciclastic, cratonic ramp (Upper Ordovician; Kentucky-Ohio, USA): insights into the relative influence of eustasy and tectonics through


Pope, M.C., and Read, J. F., 1997a, High-Resolution surface and subsurface sequence stratigraphy of the Middle to Late Ordovician (late Mohawkian–Cincinnatian) foreland basin rocks, Kentucky and Virginia: AAPG Bulletin, v. 81, p. 1866–1893.


APPENDIX 1
Locality Descriptions

A1. **Frankfort Northwest composite**—This composite section contains stratigraphic data taken from two closely spaced large outcrops. (1) The Route 421 section, located on the northwest side of Frankfort, Kentucky, approximately 300 meters north of the intersection of Kentucky Route 127 and 421, is one of the largest outcrops in the Bluegrass Region (38° 12' 52" N, 84° 53’ 24" W). Stratigraphically it covers more than 80 meters, from the top of the Curdsville Member of the Lexington Limestone at the base upward into the lower Fulton submember of the Kope (Clays Ferry) Formation at the top. (2) The Frankfort West road cut lies opposite the Route 421 road cut, approximately 500 meters to the south, across the Benson Creek valley (38° 12’ 10” N, 84° 53’ 37” W). The Frankfort West road cut begins in the Grier Member of the Lexington Limestone and ends in the middle part of the Point Pleasant member.

A2. **Peaks Mill South**—Large road cut located approximately 5 kilometers south of Elkhorn Creek and approximately 0.5 km south of Shadrick Ferry Road on Kentucky Route 127, southwest of Peaks Mill, Kentucky (38° 16’ 6” N, 84° 50’ 55” W). The upper beds of the Peaks Mill member of the Lexington Limestone form the base of the exposure and the lower meter of the Fulton submember of the Kope Formation forms the top. A significant portion of the low Fulton and its contact with the Point Pleasant is well exposed in a small road cut a few hundred meters north of the main Peaks Mill road cut.

A3. **Old Owenton Road**—Cluster of four large road cuts along Kentucky Route 127 on the southern side of the Elkhorn Creek valley at the intersection of Old Owenton Road, approximately 3 kilometers northwest of Peaks Mill, Kentucky (38° 18’ 22” N, 84° 50’ 40” W). The lowest strata exposed in the road cuts (northernmost cut) belong to the Sulphur Well Member of the Lexington Limestone; the highest strata (southernmost cut) belong to the Brent submember of the Kope Formation. Partially exposed small natural outcrops below the road level may extend down as low as the Perryville Member of the Lexington Limestone.

A4. **Swallowfield North**—Large road cut along Kentucky Route 127, 2.0 kilometers north of Swallowfield, Kentucky (38° 20’ 58” N, 84° 51’ 22” W). This very long road cut (approximately 2 kilometers long) contains strata of the Brannon Member of the Lexington Limestone at its base up to partially covered exposures of the Brent submember of the Kope Formation at its top. The Point Pleasant-Fulton contact is easily accessible in a low road cut, which forms the northernmost exposure of this large road cut on the western side of Route 127.

A5. **Monterey East**—Large road cut and creek section along Kentucky Route 127 at Sawdridge Creek, 1.5 kilometers east of Monterey, Kentucky (38° 25’ 15” N, 84° 51’ 18” W). The base of the creek section begins in the Perryville Member of the Lexington Limestone and continues up a vertical cliff face to the Strodes Creek Member where it intersects the road cut. Partially exposed strata in the northernmost portion of the road cut continue well up into the Brent submember of the Kope Formation.

B1. **KGS C-215**—4.75-centimeter diameter diamond drill core from Mason County Kentucky (38° 42’ 47” N, 83° 53’ 22” W) housed at the Kentucky Geological Survey.
(KGS) core repository in Lexington, Kentucky. Base of Point Pleasant placed at 191 feet, Point Pleasant-Fulton contact at 171 feet, top of the Fulton placed at 146 feet.

B2. **Augusta West**—Road cut on southern side of Kentucky Route 8 approximately 3.5 kilometers west of Augusta, Kentucky (38° 45’ 53” N, 84° 2’ 52” W). Exposes upper Locust Creek member of the Lexington Limestone up through lower Brent submember of the Kope Formation.

B3. **Woolcott composite**—Stratigraphic data from measurements of road cut on Kentucky Route 1159 just south of intersection with Kentucky Route 9 (AA highway) approximately 0.5 kilometers south of the Woolcott covered bridge (38° 43’ 59” N, 84° 6’ 6” W) and a steep creek section perpendicular to Salem Ridge Road, approximately 50 meters south of Kentucky Route 9; additional faunal data taken from bedding plane exposures of the Point Pleasant member on adjacent Poe Creek (38° 43’ 22” N, 84° 6’ 43” W). Combined exposure from middle Bromley member of the Lexington Limestone upward to lower Brent submember of the Kope Formation. Many adjacent exposures along Kentucky Route 9 extend from Fulton submember upward to the Snag Creek submember of the Kope Formation (see Brett and Algeo, 2001 for further locality information).

B4. **Bradford Quarry**—Abandoned quarry on Kentucky Route 8, 0.2 kilometers north of intersection with Kentucky Route 1109 (38° 47’ 4” N, 84° 8’ 45” W). Road cut 0.5 kilometers to the south displays similar stratigraphic interval. Stamping Ground Member of the Lexington Limestone up to Brent submember of the Kope Formation exposed in quarry.

B5. **Holst Creek Road cut**—Road cut on the east side of Kentucky Route 9 (AA Highway), a few hundred meters north of its intersection with Holst Creek Road (38° 46’ 54” N, 84° 12’ 39” W). The base of the cut is near the middle of the Point Pleasant member of the Lexington Limestone and the top extends well up into the Brent submember of the Kope Formation. This outcrop marks the southernmost known occurrence of *Triarthrus* (Fulton submember) in the study area.

B6. **Point Pleasant North composite**—Composite section derived from measurement of two partially covered steep creeks/small abandoned quarries on Possum Hollow (38° 53’ 53” N, 84° 14’ 11” W) and an adjacent creek 2 kilometers to the south; approximately 4 kilometers north of Point Pleasant, Ohio, exposed on east side of Ohio Route 52. Type section of the Point Pleasant member of the Lexington Limestone.

B7. **Boat Run**—Bedding plane and low cut bank exposures on north branch of Boat Run (creek), approximately 250 meters northeast of its intersection with Ohio Route 52 (approximately 400 meters northwest of Clermontville-Laurel Road) in Clermontville, Ohio (38° 55’ 54” N, 84° 15’ 14” W). The exposed creek section is accessible from Clermontville Spur. Deformed beds of the Devils Hollow Member of the Lexington Limestone form some of the lowest exposures; the lower Fulton submember of the Kope Formation forms the highest exposures.

B8. **Twelve Mile Creek**—Twelve Mile Creek crosses Ohio Route 52 280 meters south of its intersection with Ohio Route 749 in New Palestine, Ohio (38° 58’ 7” N, 84° 15’ 35” W). This small creek exposure displays the middle to upper Point Pleasant member of the Lexington Limestone and the lower beds of the Fulton submember of the Kope Formation. The northern bank of Twelve Mile Creek, approximately 700 meters northeast of Route 52 is the location of diamond drill core OGS-2536, described below.
B9. **New Palestine**—Gullies and partially covered small road cuts along the east side of Ohio Route 52, just north of New Palestine, Ohio (39° 0’ 58” N, 84° 18’ 20” W). Sections expose upper portion of the Point Pleasant member of the Lexington Limestone and the entire Fulton submember of the Kope Formation.

B10. **Duck Creek**—Duck Creek is located on the south side of Kentucky Route 1998, 3 km east of its intersection with Kentucky Route 27 in Alexandria, Kentucky (39° 0’ 58” N, 84° 18’ 20” W). The lowest exposures in the creek begin with the base of the Point Pleasant member of the Lexington Limestone. The upper section of the creek exposes the Fulton and Brent submembers of the Kope Formation. Duck Creek forms the lower part of the Kope Formation composite type-section of Brett and Algeo (2001).

B11. **“Water works” creek**—Small nameless creek section on the west side of Kentucky Route 8, approximately 4 kilometers north of the intersection with Kentucky Route 445 and the Interstate 275 overpass (39° 5’ 20” N, 84° 26’ 20” W). Base of the section has continuous exposure of the upper Point Pleasant member of the Lexington Limestone (in culvert under Route 8) through the Fulton and into the basal portion of the Brent submember of the Kope Formation. This section represents the northernmost outcrop exposure of the Point Pleasant-Fulton interval in the study area.

B12. **OGS-3020**—4.75-centimeter diameter diamond drill core from Cincinnati Township (Ohio; 39° 12’ N, 84° 27’ W) taken by the Ohio Department of Natural Resources Division of Geological Survey (ODNR-DGS), housed at the ODNR-DGS core repository in Columbus, Ohio. Base of Point Pleasant placed at 157 feet, Point Pleasant-Fulton contact at 153 feet, top of the Fulton placed at 147 feet.

C1. **OGS-2626**—4.75-centimeter diameter diamond drill core from Concord Township (Ohio; 39° 4’ N, 84° 40’ W) taken by the ODNR-DGS, housed at the ODNR-DGS core repository in Columbus, Ohio. Base of Point Pleasant placed at 937 feet, Point Pleasant-Fulton contact at 917 feet, top of the Fulton placed at 889 feet.

C2. **OGS-2620**—4.75-centimeter diameter diamond drill core from Marion Township (Ohio; 39° 17’ N, 84° 0’ W) taken by the ODNR-DGS, housed at the ODNR-DGS core repository in Columbus, Ohio. Base of Point Pleasant placed at 697 feet, Point Pleasant-Fulton contact at 681 feet, top of the Fulton placed at 651 feet.

C3. **OGS-860**—5.7-centimeter diameter diamond drill core from Lemon Township (Ohio; 39° 24’ N, 84° 20’ W) taken for the Texas Eastern Company, housed at the ODNR-DGS core repository in Columbus, Ohio. Base of Point Pleasant placed at 395 feet, Point Pleasant-Fulton contact at 382 feet, top of the Fulton placed at 352 feet.

C4. **OGS-2537**—4.75-centimeter diameter diamond drill core from Wayne Township (Ohio; 39° 34’ N, 84° 28’ W) taken by the ODNR-DGS, housed at the ODNR-DGS core repository in Columbus, Ohio. Base of Point Pleasant placed at 709 feet, Point Pleasant-Fulton contact at 702 feet, top of the Fulton placed at 669 feet.

C5. **OGS-3023**—4.75-centimeter diameter diamond drill core from Jackson Township (Ohio; 39° 50’ N, 84° 44’ W) taken by the ODNR-DGS, housed at the ODNR-DGS core repository in Columbus, Ohio. Base of Point Pleasant placed at 920 feet, Point Pleasant-Fulton contact at 914 feet, top of the Fulton placed at 878 feet.

a. **Banklick Creek**—Exposure along the southern side of Banklick Creek, parallel to Grand Avenue at the intersection with Readlin Rd., in Winston Park, Kentucky (39° 2’ 1” N, 84° 30’ 2” W). At this locality a small monocline reveals the upper 15 meters of the Lexington Limestone. This locality represents one of the northernmost and distal facies
of the Lexington Limestone in Kentucky. Point Pleasant member of the Lexington Limestone exposed in stream bank just below the level of Grand Avenue; dark organic-rich shale, lateral equivalent of the Stamping Ground Member, form lowest exposure.
b. **DeCoursey Creek**—Cutbank exposure on DeCoursey Creek approximately 100 meters from its confluence with the Licking River (39° 0’ 32” N, 84° 28’ 53” W). The partially covered section contains the Fulton and Brent submembers of the Kope Formation.
c. **Ivor North**—Road cut on Kentucky Route 8 nearly 1.0 kilometer north of its intersection with Ivor Rd. (38° 52’ 7” N, 84° 14’ 4” W). Partially covered road cut exposing the Locust Creek through Point Pleasant members of the Lexington Limestone.
d. **Carntown Composite Section**—Composite section derived from measurement of the Carntown North section and Carntown Quarry section. The Carntown North section is a small road cut and overlying partially exposed hill slope located on Kentucky Route 8, approximately 1 kilometer north of the intersection with Kentucky Route 154 (38° 50’ 49” N, 84° 14’ 34” W). It ranges stratigraphically from the Bromley member of the Lexington Limestone upward into the Fulton submember of the Kope Formation. The Carntown Quarry section is an abandoned quarry located on the west side of Kentucky Route 8, approximately 150 meters south of the intersection with Kentucky Route 154 (38° 50’ 9” N, 84° 14’ 32” W). The section is partially covered and extends from the Devils Hollow Member of the Lexington Limestone upward into the basal part of the Brent submember of the Kope Formation. The lower portion of the quarry is now the site of a power conversion grid.
e. **Sterling Materials Quarry**—Exposures on the road entering the underground limestone mine at Sterling Materials Quarry located on the north side of Kentucky Route 42, approximately 5.5 kilometers northeast of its intersection with Kentucky Route 127 (38° 49’ 6”, 84° 46’ 43” W). The lowest exposures at the mine entrance are in the Stamping Ground Member of the Lexington Limestone and the top of the cut is in the Fulton submember of the Kope Formation.
f. **Bear Creek Quarry**—Bear Creek Quarry is an abandoned quarry located on Bear Creek Road approximately 200 meters north of its intersection with Ohio Route 52, 1.5 km west of Chilo, Ohio (38° 48’ 7” N, 84° 9’ 31” W). The quarry has been partially backfilled, however, lower sections containing the Devils Hollow Member of the Lexington Limestone are exposed on the south end of the quarry; the northern part of the quarry contains exposures up into the Fulton submember of the Kope Formation.
g. **Bullskin Creek**—Bedding plane and cutbank exposures on Bullskin Creek adjacent to Felicity Cedron Road approximately 1 kilometer north of its intersection with Smith Landing Road near the village of Cedron, Ohio (38° 48’ 50” N, 84° 3’ 36” W). The creek section exposes nearly the entire Point Pleasant member of the Lexington Limestone and the lower few meters of the Fulton submember of the Kope Formation.
h. **Butler East**—Partially covered road cut on Kentucky Route 27, just over 1.0 kilometer east of Butler, Kentucky (38° 47’ 18” N, 84° 21’ 3” W). The slightly discontinuous section contains the Locust Creek member of the Lexington Limestone upward through the Brent submember of the Kope Formation.
i. **Kentucky Route 42 Road cut**—Road cut along Kentucky Route 42 extending from near the intersection with Kentucky Route 127 westward for nearly 1.0 kilometer (38° 49’ 6” N, 84° 46’ 43” W). Lowest exposures are in the Gratz member of the Lexington...
Limestone and extend upward into the lower meter of the Fulton submember of the Kope Formation.

j. **Sterman Creek**—Creek section adjacent to Kentucky Route 42, located 3.6 kilometers west of the intersection of Kentucky Routes 42 and 127 (38° 46’ 59” N, 84° 51’ 7” W). This partially covered section exposes the upper meter of the Point Pleasant member of the Lexington Limestone upwards into the Brent submember of the Kope Formation. The top of the Fulton at this locality is marked by a heavily pitted and scalloped, and phosphate encrusted fine-grained grainstone.

k. **Boston Methodist Church**—Small outcrop behind the Methodist church along Kentucky Route 27, just north of the village of Boston, Kentucky (38° 46’ 2” N, 84° 21’ 11” W). Partially covered section contains the Locust Creek member of the Lexington Limestone upwards through the Fulton submember of the Kope Formation.

l. **Big Sugar Creek**—Low bank and bedding plane exposures of the Point Pleasant member of the Lexington Limestone on Big Sugar Creek, near the intersection of Kentucky Route 127 and Tapering Point Road, approximately 3.0 kilometers north of the intersection of Kentucky Route 127 and Interstate 71 (38° 45’ 39” N, 84° 49’ 13” W).

m. **Menzie South**—Outcrop along the CSX Railroad approximately 0.5 kilometers south of Menzie, Kentucky (38° 44’ 2” N, 84° 20’ 30” W). Section contains the Locust Creek member of the Lexington Limestone at the base and extends upward into the Point Pleasant member.

n. **Falmouth Composite**—Road cuts along Kentucky Route 27 at both the north (38° 40’ 46” N, 84° 20’ 45” W) and south (38° 39’ 52” N, 84° 19’ 55” W) margins of Falmouth, Kentucky. Both sections are partially covered. The northern section is the more extensive and begins in the Devils Hollow member of the Lexington Limestone at the base and continuing upward into the Fulton submember of the Kope Formation. (See Lorenz, (198X) for additional data).

o. **Sunrise Road cut**—Small road cut and adjacent creek section on Kentucky Route 1284 (Sunrise-Claysville Road; 38° 28’ 32” N, 84° 57’ 31” W). The section contains at its base the Locust Creek member of the Lexington Limestone and extends upward into the Point Pleasant member.

p. **Gratz Northwest**—Small road cut on Kentucky Route 389 on the western wall of the Kentucky River valley, approximately 4.0 kilometers northwest of Gratz, Kentucky (38° 29’ 27” N, 84° 59’ 48” W). Section contains the Locust Creek member of the Lexington Limestone at its base and extends upward into the lower part of the Fulton submember of the Kope Formation.

q. **Gratz North**—Partially covered road cut on Kentucky Route 355 on the eastern wall of the Kentucky River valley, on the northern margin of Gratz, Kentucky (38° 28’ 32” N, 84° 57’ 31” W). The exposure extends from the Bromley member of the Lexington Limestone at the base upwards into the Fulton submember of the Kope Formation.

r. **Blue Licks Composite**—Series of four closely spaced road cuts on Kentucky Route 68 extending from the entrance to Blue Licks State Park southwestward for approximately 4.0 kilometers (38° 25’ 47” N, 83° 59’ 31” W). Primary section is a large road cut on the west side of Kentucky Route 68 at the entrance to Blue Licks State Park. This section exposes the upper two meters of the Point Pleasant member of the Lexington Limestone and the entire Fulton and the lower few meters of the Brent submembers of the Kope Formation. Sections to the southwest lack the Fulton and are primarily in the Bromley
through Point Pleasant members of the Lexington Limestone. An addition section exposing the Point Pleasant down into the Devils Hollow Member of the Lexington Limestone is located on the park road just inside the entrance to Blue Licks State Park.

s. **Cynthiana North**—Road cut on Kentucky Route 27 a few hundred meters north of its intersection with North Church Street on the northern margin of Cynthiana, Kentucky (38° 23’ 59” N, 84° 17’ 36” W). Exposure of the lower several meters of the Point Pleasant and entire Locust Creek members of the Lexington Limestone are located at the hillcrest. Discontinuous exposure leading downhill to the northeast contains the Gratz and Bromley members of the Lexington Limestone.

t. **Sadieville Northwest**—Large road cut along Interstate 75 approximately 2.5 kilometers north of Exit 136 and 4.0 kilometers northwest of the village of Sadieville, Kentucky (38° 24’ 2” N, 84° 34’ 36” W). This cut exposes strata extending from the Peaks Mill member of the Lexington Limestone up to the lower few meters of the Fulton submember of the Kope Formation.

u. **Sadieville West**—Road cut on Kentucky Route 32 just west of the village of Sadieville in the Elk Lick Creek valley (38° 23’ 19” N, 84° 32’ 32” W). The road cut extends upward from the Bromley member of the Lexington Limestone into the upper part of the Point Pleasant member.

v. **Depleplain West**—Large road cut on the east side of Interstate 75 at the northbound ramp of Exit 129 (38° 16’ 20” N, 84° 33’ 21” W). This road cut display the Locust Creek member of the Lexington Limestone at the base and extends upward to near the top of the Point Pleasant member. Several spectacular hardgrounds are present within the Point Pleasant member at this locality.

w. **Bridgeport West**—Road cut on Kentucky Route 60 and exposures in bottom and cutbanks of Benson Creek (38° 9’ 43” N, 84° 57’ 21” W). Section contains, at its base, the Brannon Member of the Lexington Limestone and discontinuously continues upward into the Fulton submember of the Kope Formation.

x. **I-64 Kentucky River Valley Composite Section**—Series of large road cuts along the eastbound and westbound lanes of Interstate 64, extending for approximately 6.0 kilometers westward from its intersection with Kentucky State Route 127, in the Kentucky River valley. The section contains nearly the entire Lexington Limestone, from the lower Grier Member at the base to the Point Pleasant member at the top. The Point Pleasant member forms the highest exposure in the westbound lane on the eastern side of the Kentucky River valley (38° 10’ 20” N, 84° 50’ 0” W). Initially the section included the Fulton submember of the Kope Formation, however it is now completely covered.

y. **Lawrenceburg Northwest**—Road cuts on the east and west side of Kentucky Route 127 bypass around Lawrenceburg, Kentucky, approximately 1.0 kilometer south of the junction with North Main Street (38° 3’ 39” N, 84° 55’ 17” W). Exposures contain the Bromley member of the Lexington Limestone upward into the Point Pleasant member. Deformation within the Locust Creek member forms a prominent aspect of these road cuts.

z. **Lawrenceburg West**—Low road cuts on Kentucky Route 44 at the intersection with Kentucky Route 127 and adjacent road cuts approximately 200 meters to the south on Kentucky Route 127 on the western margin of Lawrenceburg, Kentucky (38° 1’ 57” N, 84° 54’ 41” W). Exposures on Route 44 contain strata of the Locust Creek and Point
Pleasant members of the Lexington Limestone. Exposures on Route 127 extend from the Point Pleasant member upward into the Fulton submember of the Kope Formation.

aa. Mount Zion North—Road cut on both the north and south sides of Interstate 64 near the intersection with Kentucky Route 60 (38° 1’ N, 84° 5” W). Section contains most of the Point Pleasant member of the Lexington Limestone upward through the Fulton and Brent submembers of the Kope Formation.

bb. Ruckerville South—Road cut on east side of Kentucky Route 89 approximately 200 meters south of its junction with White Conkwright Road, near Ruckerville, Kentucky (37° 55’ 40” N, 84° 5’ 7” W). Section contains Fulton and Brent submembers of the Kope Formation.

c. Sinai Post Office—Road cut at the Sinai, Kentucky post office on Kentucky Route 53, approximately 200 meters south of its intersection with Kentucky Route 62 (37° 57’ 29” N, 85° 1” 28” W). Section contains the Fulton submember of the Kope Formation.

dd. Boonesborough South—Road cut on the eastern side of Kentucky Route 627, approximately 1.5 kilometers south of the Kentucky River (37° 53’ 40” N, 84° 16’ 22” W). This section contains a partially exposed and massive Point Pleasant member of the Lexington Limestone overlain by several meters of the Fulton submember of the Kope Formation. The middle and upper submembers of the Kope Formation rest directly on the Fulton at a high angle normal fault with approximately 30 meters of offset, which occurs in the middle of the exposure.

e. Clays Ferry Composite—The Clays Ferry section is a composite of closely spaced exposures. The lowest is a road cut on Kentucky Route 2328 on the southern wall of the steep Kentucky River valley, just below the Interstate 75 bridge (37° 52’ 55” N, 84° 20’ 20” W). This lowest exposure extends from above a prominent zone of fault gouge overlain by the Faulconer member of the Lexington Limestone upward into the Fulton submember of the Kope Formation, which is partially exposed at a sharp switchback in the road. This section partially overlaps with a small gully that runs up under the I-75 bridge partially exposing the Locust Creek member of the Lexington Limestone through the Fulton submember of the Kope Formation. A thin covered section in the lower Fulton is suspected as a small normal Fault. Above a covered section of approximately 3.0 meters the Fulton is exposed at the base of a large road cut on the east and west sides of I-75 that continues up through the Kope (Clays Ferry) Formation into the Garrard Siltstone. This large exposure is cut by high angle normal fault with approximately 50 meters of displacement, nearly doubling the thickness of the Kope.

ff. Pollard South—Road cut on River Road approximately 1.5 kilometers southeast of Pollard, Kentucky (37° 48’ 0” N, 84° 29’ 58” W). Section exposes the Locust Creek and Point Pleasant members of the Lexington Limestone.

gg. Caldwell Stone Quarry—Large limestone quarry located at the south side of the intersection of Stanford Road and Goose Pike in Danville, Kentucky (37° 37’ 54” N, 84° 45’ 5” W). Southern end of the quarry exposes strata from the base of the Lexington Limestone upward into the lower Kope Formation.

1. OGS-2982—4.75-centimeter diameter diamond drill core from Oxford Township (Ohio; 39° 33’ N, 84° 42’ W) taken by the ODNR-DGS, housed at the ODNR-DGS core repository in Columbus, Ohio. Base of Point Pleasant is tentatively placed at 940’ feet, Point Pleasant-Fulton contact is tentatively placed at 950’ feet, top of the Fulton placed is tentatively placed at 970 feet.
2. **OGS-2682**—3.5-centimeter diameter diamond drill core from Fairfield Township (Ohio; 39° 29' N, 84° 37' W) taken for Cominco American Inc., housed at the ODNR-DGS core repository in Columbus, Ohio. Base of Point Pleasant placed at 1018 feet, Point Pleasant-Fulton contact at 992 feet, top of the Fulton placed at 955 feet.

3. **OGS-2981**—4.75-centimeter diameter diamond drill core from Reily Township (Ohio; 39° 28' N, 84° 43' W) taken by the ODNR-DGS, housed at the ODNR-DGS core repository in Columbus, Ohio. Base of Point Pleasant is tentatively placed at 751 feet, Point Pleasant-Fulton contact is tentatively placed at 736 feet, and top of the Fulton is tentatively placed at 705 feet.

4. **OGS-2680**—3.5-centimeter diameter diamond drill core from Monroe Township (Ohio, 38° 42’ 24” N, 83° 30’ 39” W) taken for Cominco American Inc., housed at the ODNR-DGS core repository in Columbus, Ohio. Base of Point Pleasant placed at 512 feet, Point Pleasant-Fulton contact at 502 feet, top of the Fulton placed at 482 feet.

5. **KGS C-173**—3.5-centimeter diameter diamond drill core from Cynthiana quadrangle (Kentucky; 38° 22’ N, 84° 21’ W) taken by the Kentucky Geological Survey and the United States Geological Survey (KGS-USGS) as part of the geologic quadrangle-mapping program, housed at the Kentucky Geological Survey (KGS) core repository in Lexington, Kentucky. Base of Point Pleasant placed tentatively placed at 65 feet, Point Pleasant-Fulton contact tentatively placed at 55 feet, top of the Fulton tentatively placed at 31 feet.

6. **KGS C-199**—3.5-centimeter diameter diamond drill core from the Lexington East quadrangle (Kentucky; 38° 0’ N, 84° 27’ W) taken by the Kentucky Geological Survey and the United States Geological Survey (KGS-USGS) as part of the geologic quadrangle-mapping program, housed at the Kentucky Geological Survey (KGS) core repository in Lexington, Kentucky. Base of Point Pleasant is tentatively placed at 174 feet, Point Pleasant-Fulton contact at 158 feet, top of the Fulton placed at 133 feet.

7. **KGS C-209**—4.75-centimeter diameter diamond drill core from Montgomery county (Kentucky; 37° 54’ N, 83° 58’ W) taken for Cominco American Inc., housed at the Kentucky Geological Survey (KGS) core repository in Lexington, Kentucky. Base of Point Pleasant is tentatively placed at 515 feet, Point Pleasant-Fulton contact tentatively placed at 497 feet, and the top of the Fulton tentatively placed at 460 feet.

8. **KGS-115**—3.5-centimeter diameter diamond drill core from Lincoln county (Kentucky; 37° 36’ N, 84° 40’ W) taken for Humble Oil and Refining Co., housed at the Kentucky Geological Survey (KGS) core repository in Lexington, Kentucky. Base of Point Pleasant is tentatively placed at 223 feet, Point Pleasant-Fulton contact at 220 feet, top of the Fulton placed at 204 feet.
CHAPTER 4
Forced Regression in Foreland Basin Settings: Synchronized Responses from the Carbonate- and Siliciclastic-Dominated Margins

PATRICK I. MCLAUGHLIN AND CARLTON E. BRETT

University of Cincinnati, Cincinnati, Ohio 45221, USA

(submitted to the Journal of Sedimentary Petrology; May, 2006)
**ABSTRACT:** Foreland basin successions are an ideal setting within which to test concepts of sequence stratigraphy as carbonate- and siliciclastic-dominated facies are well represented on either side of the foreland basin. Falling stage systems tracts (FSSTs) in particular, are a rather new concept in sequence stratigraphy that require further development. In terms of foreland basin sequence stratigraphy, FSSTs have primarily been reported from the Cretaceous Sevier Basin in western North America. The following study documents similar siliciclastic-dominated FSSTs from the basal Cincinnatian Series (Upper Ordovician; C12 and C1 sequences) Taconic foreland basin in central Kentucky and southwestern Ohio. The fossiliferous nature of these siliciclastic-dominated deposits yields greater insight into the relative sea level history of FSSTs. Underlying sequences of the uppermost Mohawkian Series (M6C, M6B, and M6A sequences) show similar stratigraphic architecture but lack siliciclastics of silt size or greater. These represent some of the first carbonate-margin FSSTs documented in the literature. They are dominated by argillaceous calcarenite and have sharp erosive bases (forced regression surface) that remove significant portions of the underlying highstand systems tracts (HSTs). These Upper Ordovician carbonate-margin FSST deposits bare striking resemblance not only to regressive strata from other periods transitional between icehouse and greenhouse periods (i.e. Lower Silurian, Middle Devonian, Upper Cretaceous), but to upper Carboniferous Midcontinent cyclothem deposited during full icehouse conditions. The juxtaposition and similarity of Upper Ordovician FSST’s characteristic of the siliciclastic-margin on those characteristic of the carbonate-margin provide a unique view of processes effecting both sides of the foreland basin during relative sea level fall, in one vertical succession.
INTRODUCTION

Forced regressive strata, deposited during relative sea level fall, are common components of foreland basin successions, though their recognition is still largely restricted to the siliciclastic margin. New data presented here from the Upper Ordovician of Kentucky together with previously documented regressive deposits both from the siliciclastic- and carbonate-dominated margins of middle Paleozoic foreland basins in eastern North America provides insight into sedimentation patterns across foreland basins resulting from relative sea level fall. The siliciclastic strata of these middle Paleozoic foreland basins not only exhibit patterns comparable to those previously presented from the Cretaceous Western Interior Seaway, but provide additional observations of shallowing/deepening trends from biofacies analysis, stacking of small-scale cycles, lateral facies distribution, and characteristics of bounding surfaces. Furthermore, we provide detailed examples from the carbonate margins of these foreland basins that surprisingly display many analogous features, yet also contain important differences.

Foreland basins typically display distinct carbonate- and siliciclastic-dominated margins, which form in response to the very different terrigenous sedimentation rates on either side of the basin. The siliciclastic margin of foreland basins accommodates large amounts of terrigenous input from erosion of high relief terrains and high rates of subsidence generated by loading at the continental margin. Alternatively, carbonate margins of foreland basins form in response to marine flooding, low subsidence rates, and very low levels of terrigenous input from adjacent low relief cratonic areas. Further, during periods of continental flooding these low relief carbonate ramps/platforms may extend for thousands of kilometers cratonward before reaching the shoreline; the total area of the craton draining into epicontinental seas in these cases is also greatly reduced. The basin center acts as a trap to terrigenous sediments derived from the
siliciclastic dominated margin, prohibiting coarse sediments from reaching the carbonate margin. Yet, the carbonate margin is typically composed of a mixture of shales and argillaceous limestones in areas proximal to the foreland basin, becoming more purely carbonate cratonward. This suggests that clays are not simply being delivered to the basin via gravity flows, but rather are also transported long distances as hyperpycnal flows (*sensu* O’Brien et al., 1998). Thus, carbonate margins of foreland basins are very sensitive to changes in clay output from the siliciclastic margin.

**Previous Studies**

The recognition of a distinctive phase of deposition during sea level fall is not a new concept; a regressive phase was incorporated into the earliest models of sequence stratigraphy (e.g. Posementier et al., 1988; Fig. 1A). Central to all sequence stratigraphic models are the concepts of relative sea level (eustasy plus subsidence) fluctuation and its effect on creation and removal of accommodation. In particular, Posementier et al. (1988) point to the rising and falling inflection points in the relative sea level curve as marking important periods in the formation of depositional sequences. The rising inflection point marks the most rapid rate of relative sea level rise during a sea level cycle (Fig. 1A). Likewise, the falling inflection point (key to the following discussion) marks the most rapid rate of relative sea level fall and the period of most rapid seaward migration of the shoreline. Posementier and Vail (1988) suggested that little to no sediments were deposited on the shelf during sea level fall. Rather this period was marked by bypass and incision with deposition occurring primarily below the shelf-slope break as deep sea fans (lowstand fans). These early sequence stratigraphic models were based largely on seismic sections of siliciclastic-dominated passive margin strata. This low resolution approach left the
FIGURE 1. Models of system tract development in relation to relative sea level fluctuation. Curves A and B are developed for siliciclastic-dominated passive margin (Posementier et al., 1988) and foreland basin (Catuneanu, 2002) settings respectively. Curves C and D are developed for mixed carbonate-siliciclastic passive margin (Baum and Vail, 1988) and foreland basin (this study) settings respectively.
door open to subsequent outcrop studies, which provided greater detail and subsequent refining of the sequence stratigraphic model.

High-resolution sequence stratigraphy was largely developed from analysis of siliciclastic-dominated foreland basin deposits (e.g. Van Wagoner et al., 1990). In particular, the studies of the Cretaceous Western Interior Seaway by Plint (1988), Hadley and Elliot (1993), Van Wagoner et al. (1995), Fitzsimmons and Johnson (2000) and Edwards et al., 2005) have strongly influenced the general sequence stratigraphic model in documenting the characteristics of strata deposited during sea level fall. However, these studies have also sparked some disagreement as to the position of the sequence boundary, summarized by Posementier and Morris (2000). Plint and Numedal (2000) point out that timing of the most rapid basinward shift of facies (the falling inflection point) is not coincident with greatest basinward shift of subaerial exposure (marking the sequence boundary). Thus, most workers currently place the sequence boundary stratigraphically at the top of the most strongly regressive strata within a sequence, coincident with the most basinward position of the subaerial exposure surface in proximal areas (Fig. 1B; Catuneanu, 2002). These highly regressive deposits have been termed the “falling stage systems tract” (FSST). Plint and Nurmmedal (2000) define the falling stage systems tract as displaying: (1) offlap and a basinward shift in facies, (2) a position between the highstand and lowstand (i.e. the basal contact is the regressive surface of forced regression; the upper surface is the sequence boundary-correlative conformity), (3) a shallowing upward pattern, and (4) evidence for deposition during relative sea level fall. They state that in offshore areas of their Cretaceous examples “it is impossible to place the lower boundary of the FSST at a single surface, but, for practical purposes, it can be placed at the base of the first gutter-casted sandstone capping much thinner-bedded shelf facies”. Posamentier and Morris (2000) expand on the Plint and Nummedal
(2000) definition by adding (5) the presence of long-distance regression, and (6) “foreshortened” stratigraphic successions. They suggest that the stratal architecture of forced regressive deposits is controlled by: (1) the gradient of the sea floor, (2) the ratio of sediment flux to the rate of relative sea-level fall, (3) the “smoothness” of relative sea-level fall, (4) the variability of sediment flux, and (5) changes in sedimentary processes resulting from progressively more exposure of the shelf.

Hunt and Tucker (1993), working in deposits formed on carbonate shelves in open ocean settings, also recognized inconsistencies in early sequence stratigraphic models, especially in regard to deposition during falling sea level. From their studies of carbonate shelves in open ocean settings they saw the need to designate a systems tract specifically deposited during relative sea level fall. For these deposits they proposed the term “forced regressive wedge systems tract” (equivalent to FSST), which was bounded below by the “basal surface of forced regression” and above by the sequence boundary. They defined the sequence boundary as the lowest point of sea level fall. They also proposed the term “lowstand prograding wedge” for deposits overlying the sequence boundary formed during initial sea level rise. Subsequent studies of Cenozoic and modern passive margin settings (e.g. Trincardi and Correggiari, 2000, Hernández-Molina et al., 2000) that show down stepping shoreline deposits supporting this model.

Few studies have proposed sequence stratigraphic models expressly for application in cratonic carbonate ramp deposits, especially those focusing on deposition during sea level fall. Brett et al. (1990) developed a general sequence stratigraphic model derived from analysis of Lower Silurian foreland basin strata of the northern Appalachian basin that considered, in part, deposits of the carbonate margin. This model was refined by McLaughlin et al. (2004) following
analysis of Upper Ordovician mixed carbonate-siliciclastic strata deposited on the carbonate-
margin of the Taconic foreland basin in present day central Kentucky. Tucker et al. (1993)
proposed a sequence stratigraphic model for homoclinal carbonate ramps, such as those
bounding the cratonward side of foreland basins, and demonstrated its application in the mid-
Triassic Muschelkalk intracratonic carbonate platforms of eastern Spain. Surprisingly, Tucker et
al. (1993) suggest that in these settings sedimentation takes place primarily during the TST and
HST, with the “LST” (approximately equivalent to the forced regressive wedge of Hunt and
Tucker (1993) and FSST of this study) contributing little. Further, Tucker et al. (1993)
suggested that no specific environments are representative of the “LST”; rather it is simply
marked by a down-ramp shift in facies. However, they suggest that a significant carbonate sand
body may form in some cases (similar to the late highstand deposits of Brett et al., 1990 and
FSST of McLaughlin et al., 2004). They note that if siliciclastics are available they may overlie
more carbonate-rich facies of the HST.

In the following section we provide additional insights into sedimentation patterns indicative
of both the carbonate- and siliciclastic-margins of foreland basins during relative sea level fall
from Paleozoic strata of eastern North America. Siliciclastic-dominated Upper Ordovician strata
in central Kentucky and southwestern Ohio are discussed initially for recognition of similarities
to previously described siliciclastic-dominated FSST deposits and to add paleoecological data
largely missing from Cretaceous examples. To further add to the model of falling stage systems
tracts case studies from slightly older Upper Ordovician strata deposited on the carbonate margin
of a foreland basin are provided. Comparisons are made to studies of younger strata and a
general model is proposed for genesis of regressive deposits from the siliciclastic margin to the
basin center and across to the carbonate margin of foreland basins.
CASE STUDIES FROM THE UPPER ORDOVICIAN CINCINNATI ARCH

A series of abbreviated case studies are presented below based upon strata deposited on the Cincinnati Arch during a brief period of the Late Ordovician (~5 Ma) Taconic orogeny (Fig. 1). The abundant exposures of this interval in the tri-state area of Ohio, Kentucky, and Indiana are some of the most fossiliferous middle Paleozoic strata in the world. In central Kentucky the uppermost Mohawkian age strata are assigned to the carbonate-dominated Lexington Limestone, a mixed carbonate-siliciclastic unit composed of relatively tabular grainstone-rudstone intervals that alternate with more argillaceous intervals dominated by wackestone-packstones and interbedded shales (Fig. 2). The overlying strata of the basal Cincinnatian Series (Kope and Fairview formations and lateral equivalents) mark a sharp change to siliciclastic-dominated sedimentation. The influx of these allochthonous siliciclastics signifies a breach of the Taconic foreland basin, which previously acted as a trap to quartz silts and sands (Weir et al. 1984). Thus, the Cincinnati Arch underwent a rapid transformation from carbonate-margin of a foreland basin to more closely resembling the siliciclastic-margin of the Taconic foreland basin. For example, siliciclastic sedimentation rates in the C1 sequence are nearly an order of magnitude higher than they are in the underlying upper Mohawkian sequences (McLaughlin et al. 2004). Thus, the unique paleogeographic evolution of this area during the Upper Ordovician resulted in a succession that provides insights into the sedimentary dynamics of both the carbonate and siliciclastic margins of foreland basins. Additionally, siliciclastic sedimentation during the early Cincinnatian Series was not so great as to preclude intermittent colonization of the sea floor by an abundant and relatively diverse assemblage of marine invertebrates, the fossil record of which is a valuable tool in reconstructing relative sea level history (e.g. see Cisne and Gildner, 1988). The following section proceeds from these youngest siliciclastic-dominated strata of the C2 and
FIGURE 2. Generalized stratigraphic column for the middle and upper Lexington Limestone and overlying Kope and Fairview Formations. Note the repeating lithologic motif of depositional sequences, even in the face of change from carbonate- to siliciclastic-dominance with the C1 maximum flooding surface.
C1 sequences, followed by carbonate-dominated examples from the M6C, M6B, and M6A sequences.

Siliciclastic-Dominated Falling Stage Deposits from the Basal Cincinnatian Series

The C2 Sequence.—The lower Maysvillian age C2 sequence is composed almost entirely of the Fairview formation in northern Kentucky and southwestern Ohio (second composite depositional sequence of the Cincinnatian Series; terminology of Holland and Patzkowsky, 1996). Near Maysville, Kentucky the Fairview contains three readily recognizable divisions: (1) a lower skeletal grainstone-dominated interval, (2) a middle interval containing slightly more shale than siltstone with rare packstones and (3) an upper portion dominated by siltstone with some beds reaching nearly a meter in thickness (Fig. 3). These division represent the transgressive, highstand, and falling stage systems tracts (TST, HST, FSST) respectively. The contact between the lower and middle Fairview is sharp and marks an abrupt decrease in limestone content, replaced by shales, thin siltstones and silty packstones (Fig. 4A). The abundance and thickness of the middle Fairview siltstones changes only slightly upward. The contact with the upper division of the Fairview is sharp, marked by a thick skeletal grainstone bed. Just above this grainstone large siltstone gutter casts up to 30 cm deep and 20 centimeters wide mark an increase in silt influx and erosion levels (Fig. 4C). Bioturbation in the upper Fairview is locally heavy (Fig. 5A) and body fossils are dominated by robust brachiopods (Fig. 5B). Interference ripples cap many of the thick siltstones. The thick siltstones also display a distinctive banding alternating between fine muddy laminations and more massive calcareous bands (Fig. 5D). Rarely, well developed hardgrounds, dominated by encrusting cryptostome and ptylodictyid bryozoans, are developed on the tops of these calcite cemented siltstones (Fig. 5E).
FIGURE 3. Stratigraphic column and location map for the Fairview Formation at Maysville, Kentucky.
FIGURE 4. Images showing the major bedding features of the middle and upper Fairview Formation at Maysville, Kentucky. A) Contact (white dotted line) between the thinner bedded and shalier middle Fairview and the upper, thick siltstone-dominated Fairview. B) Close-up of thin tabular and lenticular beds in shale typical of the middle Fairview. C) Thicker and more frequently amalgamated siltstones of the upper Fairview. Note the presence of large gutter casts forming continuous corrugated erosion surfaces (arrows). Also note that the thick siltstone overlying the upper corrugated surface is an amalgam of several decimeter-thick beds. D) Close-up of the side of a single isolated gutter cast showing well preserved longitudinal scratch marks. E) Tightly stacked siltstones of the upper Fairview. Note again the corrugated base of the thick siltstone. F) Close-up of thick siltstone from previous panel. Note that the thick siltstone overlying the upper corrugated surface is an amalgam of several decimeter-thick beds.
FIGURE 5. Sedimentary features of the upper Fairview. A) Heavily bioturbated olive colored shale and siltstone present locally on the upper surfaces of siltstones. B) Low density accumulation of typical siltstone biofacies dominated by strophomenid brachiopods Rafinesquina and Strophomena as well as robust orthids Hebertella and Plectorthis. C) Sharp crested interference ripples on top of thick siltstone. D) Alternating fine laminations and thicker bands in calcareous siltstone common to the thick siltstones. E) Hardground developed on upper surface and margin of calcareous siltstone pillow Trypanites borings, sheet-like encrusting bryozoans, and numerous conical holdfasts of ptylodictid bryozoans (note the near total obstruction of laminae by encrusters). F) Irregular casts on the base of deformed siltstone. Note rectangular outline.
Additionally, these thick siltstones of the upper Fairview often display soft-sediment deformation, typically associated with rectangular crack fillings (Fig. 5F) developed on the underside of pillows (see McLaughlin and Brett, 2004 for further discussion). A distinctive feature of the thick upper Fairview siltstones is that they fill broad shallow channels more than a meter deep and up to 10 meters wide (Fig. 4B) that are oriented with their long axes parallel to depositional dip. The thick siltstones of the upper Fairview appear to be separated into at least four cycles, each marked by a channeled horizon (Fig. 4B; Schumacher 2001). No thick siltstones occur in the Upper Ordovician of the Cincinnati Arch above the top of the Fairview and indeed siltstones become rare to absent in general (Weir et al. 1984). The C3 sequence boundary (correlative conformity) is marked by and abrupt decrease of silt of the upper Fairview and a return to mixed shales and limestones of the Miamitown Shale (LST), followed by the Bellevue Formation which at its base displays a limestone-dominated facies similar to that of the lower Fairview (North Bend informal member).

The C1 Sequence. —The stratigraphically lowest quartz siltstones to occur in the Upper Ordovician of the Cincinnati Arch are found in the C1 sequence. This is the first 3rd-order composite depositional sequence of the Upper Ordovician Cincinnatian Series (Fig. 2; Holland and Patzkowsky, 1996). The lowstand (LST) and transgressive systems tracts (TST) are composed of the skeletal grainstone-rich Point Pleasant (informal) member of the Lexington Limestone and the overlying Fulton sub-member of the Kope Formation (Fig. 6; Holland and Patzkowsky, 1996; McLaughlin et al., 2004; McLaughlin and Brett, in review). The stratigraphically lowest siltstones to occur in this area are found near the top of the Fulton (Fig. 6). The overlying shale-dominated Brent, Pioneer Valley, Alexandria and Grand View sub-members of the Kope Formation make up the highstand systems tract (HST) within which
FIGURE 6. Stratigraphic column for the Kope Formation composite reference section near Augusta, Kentucky (modified from Brett and Algeo, 2001). Curves are faunal gradient analysis score, with higher scores (to the right) interpreted as shallower water (modified from Holland et al., 2001).
FIGURE 7. Outcrop images of the near-shore Garrard to offshore Taylor Mill gradient forming the FSST of the C1 sequence. A) Proximal Garrard Siltstone at Agawam Station, Kentucky. Note the sharp planar contact between the shaly Clays Ferry facies of the Kope Formation (KP) and the overlying massive beds of the Garrard Siltstone (GA). B) Close-up of closely stacked deformed siltstone in the lower half of the Garrard. C) Large outcrop section at Sherburne, Kentucky showing closely stacked thin siltstones of the Taylor Mill sub-member sharply overlain by the brachiopod coquinas of the lower Fairview Formation. D) Mays Lick locality showing increasing shale content in the Taylor Mill below its contact with the Fairview. E) Kope/Fairview contact at Maysville, Kentucky showing thin siltstones and interbedded shales of the Taylor Mill sub-member sharply overlain by massive to medium bedded skeletal grainstones. F) Exposure on Rapid Run Creek at Cincinnati, Ohio showing the Kope/Fairview contact with the shale-dominated Taylor Mill sub-member containing only very thin and discontinuous siltstones.
FIGURE 8. Bedding features and faunas of the Garrard Siltstone and Taylor Mill sub-member of the Kope Formation. A) Meter-thick siltstone from the Garrard at Clays Ferry, Kentucky. Note the bands of more massive calcareous siltstone (standing in slight relief) in comparison to the finely laminated intervals (weathering dark brown). B) Slightly shaler facies of the Garrard at Clays Ferry showing alternating thin silty shales and thicker siltstones. C) Close-up of finely laminated layers from meter-thick siltstone in panel A. D) Diplocraterion “dumbbells” and small prod marks on the underside of a thin siltstone from the Taylor Mill Member at Mays Lick, Kentucky. E) Typical siltstone biofacies from the Grand Avenue Member at Rapid Run Creek showing low density accumulation of fragmented strophomenid brachiopods and branching bryozoans with lesser amounts of crinoid columnals, trilobite fragments, and orthid brachiopods. Note the even color distribution between skeletal material of the same genera and contrast with multicolored (reddened and blackened) highly concentrated skeletal material from a stratigraphically adjacent condensed grainstone (F). Also note the general absence of strophomenids in the grainstone.
siltstones occur with variable frequency, but typically make up only a small percentage. Siltstones within the HST are typically no more than a few centimeters thick and are often manifest as distal storm beds. Features indicative of storm deposition include interference ripples, small-scale hummocky cross stratification, gutter casts, and small flute marks. Near Cincinnati, the gutter casts typically do not exceed 10 centimeters in width or depth, but are especially noteworthy as many are sinuous, display well defined tool marks on their sides, and occasionally are shaped in a manner that demonstrates that the muds that they cut into were so stiff that they actually could form overhangs (Brett et al., 2003). Measurements taken on the long axes of gutter casts from the Kope Formation show a preferential north-south orientation parallel to depositional dip, indicating genesis by gravity controlled gradient currents (Jennette and Pryor, 1993). Other more enigmatic features are also preserved such as kinea and millimeter ripples. These siltstone storm beds show classic proximality trends from deeper ramp facies near Cincinnati, Ohio becoming more proximal southward toward Lexington, Kentucky (locally referred to as the upper Clays Ferry Formation). Along this transect the siltstones increase in number, maximum thickness (10-20 cm average thickness), and gutter casts give way to more planar erosive bases. Individual siltstones appear to be semi-discontinuous compared to thicker pack-grainstone beds, which have been traced between closely spaced exposures over more than 75 kilometers obliquely across depositional dip (Brett and Algeo, 2001). Stratigraphic trends in storm bed proximality features and siltstone abundance have also been reported from the Kope Formation (Jennette and Pryor, 1993; Holland et al., 1997; Brett and Algeo, 2001). For example, gutter casts within the Grand Avenue and siltstone-rich Taylor Mill sub-members in the Cincinnati area tend to be much larger than those within the underlying submembers, with dimensions similar to those observed in the upper Fairview, described above. The Grand
Avenue and Taylor Mill sub-members form the falling stage systems tract (FSST) of the C1 sequence in northern Kentucky and southwestern Ohio (Fig. 6-8). The fauna of the limestone-rich Grand Avenue sub-members indicates an abrupt shallowing (Fig. 6; Holland et al., 2001) that surprisingly marks both a period of siliciclastic sediment starvation as indicated by time-averaged faunas of the abundant grainstone beds (Fig. 8A), but also the beginning of increased influx of quartz silt (Fig. 8B). The overlying Taylor Mill sub-member contains abundant siltstones (more than twice as many as the Grand Avenue sub-member). Many siltstones in this interval display the distinctive spreiten structures of *Diplocraterion* (Fig. 7, 8C). Utilizing new outcrop exposures the Grand Avenue and Taylor Mill sub-members are correlated here from southwestern Ohio into central Kentucky, confirming their age equivalence to the Garrard Siltstone (Fig. 9), as proposed by Foerste (1905) based on bryozoan biostratigraphy. The Garrard ranges from 12-16 meters-thick, thickening to the southeast (Weir et al. 1984; Jacobs, 1986) and is composed of approximately 60-80 percent laminated calcareous siltstone that occurs in thin to massive beds, similar to those of the upper Fairview (Fig. 8A, B). The siltstones of the Garrard contain trace fossils of the skolithos ichnofacies (Fig. 8D). Intervening silty packstones are largely dominated by the robust brachiopod *Rafinesquina* (Jacobs, 1986). Soft-sediment deformation is also common in the Garrard Siltstone (Fig. 8B) primarily as ball-and-pillow structures, but also containing rare occurrences of the dewatering structure *aristophycus*. Ball-and-pillow structures are distributed between both thin and thick bedded siltstones with deformation restricted to certain stratigraphic horizons (at least within a given outcrop), while similar siltstone beds remained totally undeformed regardless of bed thickness. Further, many of the deformed beds contain primary sole marks on their bases, suggesting that they were deposited upon cohesive muddy sediments. These features combined with the observation that
FIGURE 9. Detailed stratigraphic cross section for the Kope-Garrard interval from central to northern Kentucky (~125 kilometers). Note the rapid decrease in siltstone concentration from south to north and the continuity of the pack-grainstone horizons.
many deformed horizons actually include several beds, supports the interpretation that
defformation was generated by seismic shaking (Pope et al., 1997). Weir et al. (1984) reported
that the Garrard Siltstone conformably overlies the upper Clays Ferry Formation (lower and
middle Kope-equivalent). However, this contact is sharp and mapping of small-scale cyclicity
within the Kope Formation across central and northern Kentucky indicates that it does indeed
become an unconformable contact in central Kentucky. In this area the thick amalgamated
siltstones of the Garrard give way upward into mixed shales and limestones with sparse thin
siltstones before reaching the base of the Fairview formation (Fig. 9). This rapid decrease in silt
influx marks the beginning of lowstand and thus the C2 sequence boundary is marked at the top
of the last prominent siltstone (Z-bed of Brett and Algeo, 2001). This sequence boundary
becomes highly erosive to the south near Nashville, Tennessee, removing the Garrard Siltstone
and nearly the entire underlying Kope Formation (Holland and Patzkowsky, 1998).

**Similarities to Other Siliciclastic-Dominated Foreland Basin Successions.**—The details
of the study above share much in common with siliciclastic falling stage systems tracts described
from the Cretaceous western interior seaway. For example, Fitzsimmons and Johnson (2000)
reported a series of features associated with regressive erosion surfaces in the siliciclastic-
dominated Upper Cretaceous strata of the Bighorn Basin, including: (1) marked basinward shift
in facies (dislocation), (2) abrupt increase in sand-shale ratio, (3) missing facies, (4) change in
parasequence stacking patterns, (5) *Glossifungites* ichnofacies, (6) precursor gutter casts, (7)
widespread soft sediment deformation and growth faulting, (8) change in paleocurrent directions,
(9) regional truncation, (10) regionally traceable surface. All but features 5 and 8 are present
near the base of the FSST in both the C1 and C2 sequences. Following Hadley and Elliot (1993),
the forced regression surface is interpreted to form during the maximum rate of sea level fall.

The main point in which these Cretaceous studies differ from those described from the Upper
Ordovician examples is the presence of a condensed bed marking the forced regression surface.
Several authors describe gutter casts occurring as “precursors” and firmground burrows at the
base of the to the falling stage systems tract, but no limestones or any other type of condensed
bed is discussed. However, Fitzsimmons and Johnson (2000) do report that in distal areas the
forced regression surface is overlain by an unusually thick sandstone with well-developed flutes
and scours on the base. A similar bed is reported from regressive deposits in the Lower Silurian
strata of central Pennsylvania. In this case, inner ramp sandstone facies of the lower Keefer
(FSST) abruptly overly outer ramp facies of the upper shaly division of the Rose Hill Formation
(HST; equivalent to the Williamson Shale of western New York State; Brett et al., 1990). The
Salmon Creek ironstone is a condensed bed that marks the contact between these two units.
Locally this ironstone grades into an unusually thick quartz arenite, similar to that described by
Fitzsimmons and Johnson (2000). Another well documented condensed bed associated with
regressive siliciclastic deposits comes from sections near the basin axis of the Acadian foreland
basin in the Middle Devonian of New York. In this succession a shell-rich bed containing
phosphate pebbles, designated the “precursor bed” (Brett and Baird, 1996) marks the forced
regression surface, separating black shales of the Butternut Shale at a sharp contact with the
overlying muddy siltstones of the lower Centerfield Formation. Brett and Baird (1996)
suggested that the juxtaposition of shoreface on mid- to outer-ramp facies suggests a gap in the
rock record marking at least a period of nondeposition and probable erosion. Thus, these
“precursor” condensed beds at the forced regression surface are interpreted to form as a result of
rapid sea level fall resulting in a period of disequilibrium (non-deposition and bypass) in the mid-ramp to basin before coarser sediments are transported into these areas.

**Carbonate-Dominated Falling Stage Deposits of the Uppermost Mohawkian Series**

The carbonate margins of foreland basins also display a distinctive, although commonly overlooked, sedimentary response to forced regression. Examples are given here of carbonate-dominated falling stage deposits from three consecutive depositional sequences in the Upper Ordovician Lexington Limestone in central Kentucky, immediately preceding the siliciclastic-dominated sequences described above. Their geographic and stratigraphic distributions have significant implications for the recognition of falling stage systems tracts on the carbonate margins of foreland basins.

**The M6C Sequence.** —The M6C composite depositional sequence is composed of the upper Devils Hollow, Bromley, Peaks Mill, Gratz, and Locust Creek members of the Lexington Limestone (Fig. 10). The upper Devils Hollow Member is composed of argillaceous limestones near its base that mark the LST. The remaining limestones of the Devils Hollow contain a much lower argillaceous content and form the TST. The argillaceous limestones and thin shales of the Bromley and Gratz members, separated by the prominent skeletal grainstones of the Peaks Mill member together form a composite HST. The argillaceous calcarenites of the Locust Creek member (informal) form the FSST (Fig. 10, 11). The contact of the Locust Creek calcarenites on the rhythmically bedded shales and marls of the Gratz (Fig. 11A, B) is highly erosive in the Frankfort area, removing the unit altogether in many sections (Fig.11E). In this area a sparry skeletal grainstone bed rests on this erosion surface and is clearly distinctive from the overlying argillaceous calcarenites of the Locust Creek and the underlying tabular shale-marl succession of
FIGURE 10. Stratigraphic column and location map for the carbonate-dominated M6C sequence at Monterey, Kentucky.
FIGURE 11. Outcrop images of the Locust Creek (M6C FSST) and surrounding (informal) members of the Lexington Limestone in central Kentucky. A) Stratigraphic relations of the Bromley, Peaks Mill, Locust Creek, and Point Pleasant members at Clays Ferry, Kentucky (sections 66). B) Sharp, planar contact of tightly stacked, slightly deformed argillaceous calcarenites of the Locust Creek member on the distal storm beds and concretions of the shaly Gratz member (section 28). C) Corrugated and mineralized Point Pleasant/Locust Creek contact at Peaks Mill, Kentucky (section 40). D) Light gray weathering, broad, shallow channel filled with deformed, thick-bedded, laminated argillaceous calcarenites and shale within the uppermost Locust Creek cutting into medium- and irregularly-bedded, argillaceous calcarenites and shale of the underlying Locust Creek member near Blue Licks State Park (section 27). E) Low-angle erosive contact of the Locust Creek on the Peaks Mill at Swallowfield, Kentucky (section 35). Note that the entire Gratz is removed at this locality. F) Vertical gradation from medium- and irregularly-bedded argillaceous calcarenites and shale in the lower Locust Creek upward into thick, laminated argillaceous calcarenites at the top of the Locust Creek (section 29).
FIGURE 12. Bedding structures of the Locust Creek member. A) Alternating reddish gray ferruginous and phosphatic planar and rippled laminae alternating with more massive tabular bands of heavily abraded and fragmented skeletal grains composing the more proximal argillaceous calcarenite facies (section 40). B) Less laminated and more heavily cross bedded argillaceous calcarenites at a slightly more distal locality (section 35). C) Corrugated erosion surface at the Point Pleasant/Locust Creek contact with coarse grained grainstones filling gutters cut into the underlying platy argillaceous calcarenites (section 34). D) Large grainstone filled gutter cutting into deformed shales and argillaceous calcarenites within the Locust Creek member (section 35). E) Small-scale cycles within the Locust Creek member alternating from clean skeletal grainstones, to shale, to argillaceous calcarenite. Note the similarity of these cycles to the motif of the 4th-order depositional sequences. F) Close-up of the iron mineralized hardground formed on deformed argillaceous calcarenite (left) and a mudstone diapir (right, under hand) just below the contact with the Point Pleasant (section 37).
the Gratz (Fig. 11E). Argillaceous calcarenites of the Locust Creek display well developed ripple lamination (reddish laminae composed of ferruginous-phosphatic lags) that alternates with more massive calcareous bands in proximal areas (Fig. 12A). In the mid-ramp the laminations become slightly more cross bedded (Fig. 12B). In the outer ramp amalgamation tends to be largely lacking, rather this heterolithic facies of the Locust Creek member is dominated by irregularly based distal storm beds that may partially truncate one another locally and thin interbedded shales (Fig. 11F). Small-scale cycles are present in several exposures of the Locust Creek. The most fully developed cycles begin with an erosive-based coarse grained grainstone, which is overlain by a thin shale and capped by argillaceous calcarenite (Fig. 12E). Commonly the middle shale is missing or thinned through erosion or nondeposition. In some cases the shale is partially or completely displaced laterally, forming diapirs in deformed horizons. The resulting effect is that the argillaceous calcarenites are commonly found resting directly upon the skeletal grainstone (Fig. 12D). The erosion surfaces at the bases of these small-scale cycles occasionally expand from simple gutter casts (Fig. 12D) to form broad shallow channels (Fig. 11D). The Locust Creek is thickest at Peaks Mill, Kentucky (inner ramp position of underlying highstand), composed of seven small-scale cycles (5th-order; Fig. 12E), and thins rapidly up-ramp and more gradually down-ramp in concert with a decrease in the number of cycles (Fig. 13). The contact with the overlying Point Pleasant member of the Lexington Limestone is sharp and typically slightly irregular. However, in sections near Peaks Mill (sections 37 and 40) the contact is highly irregular (Fig. 12C) and in all sections along Kentucky Route 127 and Interstate 75 (sections 28, 29, 35, 37, 39, 40) it is additionally marked by a hardground with a well developed ferric crust and crinoid holdfasts (Fig. 11C, 12F).
FIGURE 13. Schematic cross section of the M6C sequence from Frankfort to Monterey, Kentucky. Note the rapid truncation of the Gratz member (HST) under the Locust Creek member (FSST) between Monterey and Swallowfield.
The M6B Sequence. — The M6B sequence contains regressive deposits similar to those described above for the M6C sequence; however, some important differences do exist, such as evidence for subaerial exposure around Frankfort, Kentucky. In this 3rd-order composite depositional sequence the argillaceous limestones of the lower Sulphur Well Member form the LST, the clean skeletal grainstone-dominated upper Sulphur Well forms the TST (Fig. 14). The HST is represented by the argillaceous limestones and shales of the Stamping Ground and Greendale members split by the clean skeletal limestones of the Strodes Creek Member. The FSST is composed of the lower Devils Hollow Member (Fig. 14).

In order to evaluate the nature of the forced regression surface the underlying HST must be well understood. The Greendale is easily traceable across central Kentucky by the presence of multiple distinguishing characteristics: 1) heavily mineralized hardground at its basal contact with the underlying Strodes Creek Member, 2) an aggradational, highly fossiliferous, basal unit typically 0.5 meters thick (LTST), and 3) a middle, soft-sediment deformed, calcisiltite-dominated interval, containing (4) an epibole of Hindia sponges (Fig. 17). The contact between the Greendale and the lower Devils Hollow is sharp and typically flat, however at one locality in Frankfort (section 44) it has approximately 0.5 meters relief. Just below this contact at Monterey and Gratz East (sections 26 and 29) the upper Greendale contains “precursor” gutter cast horizons. The Greendale is increasingly truncated beneath this contact from the Old Owenton Rd. outcrop (section 37) toward Frankfort (section 44), though never becomes completely removed (Fig. 17). In this way the M6B sequence differs from the surrounding sequences. The lesser extent of erosion at the forced regression surface may have resulted from the greater proportion of skeletal limestone within the Greendale relative to HSTs of the M6C and M6A
Gratz East

seq-strat

<table>
<thead>
<tr>
<th>LST</th>
<th>FSST</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
</tr>
</tbody>
</table>

litho-strat

upper Devils Hollow

lower Devils Hollow

Greendale

Strodes Creek

Stamping Ground

Sulphur Well

Donerail

FIGURE 14. Stratigraphic section and location map for the M6B sequence.
FIGURE 15. Outcrop images of the Devils Hollow Member of the Lexington Limestone along a proximal to distal (A-F) transect from central to northern Kentucky. A) Lower Devils Hollow Constellaria beds and argillaceous calcarenites (i) overlain by recessive ostracode-rich green-mudstone facies (ii; section 47). Note the sharp contact with the sequence bounding fenestral micrites (iii) and deep lagoonal stromatoporoid-bearing micritic wackestones (iv) of the upper Devils Hollow. B) Very thin interval of laminated argillaceous calcarenites of the lower Devils Hollow (FSST) sharply overlain by a thick amalgamated and cross bedded sparry calcarenite at the base of the upper Devils Hollow, which contains abundant rip-up clasts of the former near its base (section 41). C) Thicker package of laminated argillaceous calcarenites and thin shales of the lower Devils Hollow (FSST) sharply overlain by the massive basal sparry calcarenite of the upper Devils Hollow, which is distinctive in this locality for its well developed herringbone cross bedding and presence of abundant green mudstone rip-up clasts at its base (section 36). D) Close-up of the lower Devils Hollow showing irregularly bedded argillaceous calcarenites and shale becoming heavily amalgamated upward. Note sharp erosive contact of coarse skeletal grainstones of the upper Devils Hollow on the deformed argillaceous calcarenites of the lower Devils Hollow (section 33). E) Large founded blocks of laminated argillaceous calcarenite into a mass of highly disrupted thin- to medium-bedded and irregularly bedded argillaceous calcarenites within the lower Devils Hollow. Note the founded block over-thrusting itself in the lower right of the image, suggesting that the sediments were already partially lithified at the time of deformation (section 15). F) Same stratigraphic interval as in E, but here the lower Devils Hollow is almost completely undeformed (section 6).
FIGURE 16. Bedding features of the lower Devils Hollow Member. A) Ostracode pavement from the green mudstone facies of the lower Devils Hollow (section 47). B) Close-up of tightly packed bryozoan branches of the Constellaria beds (section 49). C) Green mudstones (i) of the lower Devils Hollow sharply overlain by flaser-bedded micrites with cryptagal laminae (ii), fenestral micrite (iii) and stromatoporoid wackestones (iv) of the upper Devils Hollow. D) Trough cross bedded calcarenites of the Strodes Creek Member (i) overlain by amalgamated argillaceous calcarenites of the lower Greendale Member (ii) and Constellaria beds of the lower Devils Hollow (section 49). Note that the upper portion of the Greendale is missing at this locality. E) Large bifurcating gutters within calcarenites within the basal unit of the upper Devils Hollow (section 15). F) Characteristic alternating fine lamination and more massive calcareous bands of the argillaceous calcarenite facies within the lower Devils Hollow (section 15).
sequences that may have acted to armor the sea floor against erosion (i.e. taphonomic feedback
sensu Kidwell and Jablonski, 1983).

The lithologic characteristics of the M6B FSST vary slightly from those of the surrounding
sequences and yield a more complete understanding of thickness changes of the FSST into near
shore settings. Across much of the study area the lower Devils Hollow is dominated by
argillaceous calcarenite, made distinctive by an abundance of spar filled gastropods. The basal
contact of the lower Devils Hollow in northern Kentucky is slightly irregular to channeled and
the lower Devils Hollow may contain channels internally. The lower Devils Hollow shows soft
sediment deformation in a series of closely spaced sections in northeastern Kentucky (Fig. 14E;
localities 3, 6, 9, 15, and 33), otherwise it is lacking. Small-scale cyclicity is poorly developed in
the lower Devils Hollow, but a shallowing upward pattern is evident lithologically in many of the
northern sections (Fig. 14F). For, example in the section at Carntown (locality 6) the base of the
lower Devils Hollow is composed of distal storm beds of fine grained grainstone, diagnostic by
their irregular and sharp bases, hummocky cross stratification, and graded bedding. These beds
are overlain by laminated calcarenites more typical of the Lower Devils Hollow (Fig. 14F). The
underlying HST is differentiated from the FSST in these more distal portions of the ramp by a
near lack of limestone beds and presence of concretions; or shale/marl alternations, which lack
the irregular bedding, erosive bases, and other storm bed indicators. The lower Devils Hollow is
thickest in these northern (down ramp) outcrops and thins gradually toward Frankfort. It is
thinnest at the Old Owenton Rd. section (section 37) where it is as little as 60 cm-thick. Here it
overlain by clean calcarenite shoal facies containing closely spaced hardgrounds of the upper
Devils Hollow (Fig. 15B; 17). To the south within Frankfort and to the southeast near Versailles,
the lower Devils Hollow surprisingly thickens where it underlies low energy lagoonal micrites of
FIGURE 17. Schematic cross section of the M6B sequence from Frankfort to Monterey, Kentucky. Note that the lower Devils Hollow is present as supratidal green mudstones at Frankfort where it underlies low energy lagoonal facies, but is largely missing in more northerly sections where it underlies high-energy shoal facies of the basal upper Devils Hollow. The progressive thickening of the lower Devils Hollow to the north represents an increasing degree of preservation below the M6C sequence boundary. Cb=Constellaria bed
the upper Devils Hollow (Fig. 17). In this area the argillaceous calcarenites begin to alternate with gray mudstones packed with the fragmented colonies of the bryozoan *Constellaria* (*Constellaria* bed of Cressman, 1973), so much so that reentrants 10 to 20 centimeters thick may be composed completely of thousands of large, tightly-packed, and broken branches surrounded by as little as 5% mudstone (Fig. 16B). The *Constellaria* beds give way upward to green mudstones and micritic packstones (Fig. 15A) containing a restricted ostracod-dominated fauna (Fig. 16A) and dessication cracks, interpreted to represent shallow lagoonal to supratidal deposits (Etter, 1975; McLaughlin et al., 2004). Across Frankfort and in areas just to the southeast, the green mudstone interval is sharply overlain by a thick (~10-30 cm) micritic gastropod packstone to grainstone marking the sequence boundary and the base of the upper Devils Hollow.

Description the lower portion of the upper Devils Hollow is instructive in contrasting the deposits of the LST against the FSST. As described above, the sequence bounding bed of the basal upper Devils Hollow is a thick, white, micritic gastropod packstone that locally grades into grainstone and in some localities fenestral micrite. In a few locations on the southeast margin of Frankfort, this bed expands to form isolated dunes (~50 m-wide) of gastropod coquina up to a meter thick (Coquinite Unit of Etter, 1975). The lenticular nature of this unit and distinctive low angle cross bedding led previous authors to interpret it as a barrier beach deposit (Cressman, 1973; Etter, 1975). The basal contact is sharp, erosive and occasionally undulating, locally removing up to 60 cm of the underlying green mudstone interval. The irregular base is locally ornamented by casts of mud cracks, small gutter casts, and large hypichnial burrows. The top of the bed is occasionally capped by a ferric hardground. This basal unit of the LST is abruptly overlain by pink to light brown stromatoporoid- and coral-bearing micritic wackestones interbedded with thin organic rich shales and partings containing. Identical facies in the Strodes
Creek Member of the Lexington Limestone contain an abundance of carbonaceous remains of dasycladacean green algae, which represent a relatively deep lagoonal environment (McLaughlin et al., 2004). The juxtaposition of the deep lagoon facies on beach and shallow lagoon fenestral micrite across condensed surface suggests sea level rise that caused shoals to aggrade rather than retrograde at this time. The deep lagoonal deposits are sharply overlain by a sparry and occasionally glauconitic skeletal grainstone interval (highly micritic locally) containing low angle cross-beds. This unit bears some lithologic similarity to the dune-form coquinite within the lower Devils Hollow; however, it is far more laterally continuous, has an irregular, highly erosive base and a flat upper contact marked by a prominent ferric hardground. The erosional base (transgressive ravinement surface) of this unit displays over a meter of relief locally (sections 44) where it forms broad channels that cut out the underlying lagoonal beds to come into contact with sequence boundary (Fig. 17). Down-ramp to the north and east, these three basal units of the upper Devils Hollow change facies. The gastropod coquina at the base of the Upper Devils Hollow transitions into calcarenite, which becomes approximately 1 m-thick at Swallowfield (section 35). At Millersburg (section 36) this basal upper Devils Hollow unit shows distinctive herringbone cross bedding and contains an abundance of green mudstone rip-up clasts near its base, very similar to the green mudstone facies preserved in the Frankfort area. In sections in northern Kentucky the basal contact of the upper Devils Hollow also displays large gutter casts (Fig. 16C). The overlying deep lagoonal wackestones grade laterally into a thick (~2m) interval of argillaceous calcarenites (marking the position of the aggrading shoal facies; locality 37) and eventually interbedded packstones and shale at Monterey (section 29). The upper glauconitic grainstone unit transitions into a prominent yellowish white grainstone-rudstone unit capped by the prominent ferric hardground that is traceable throughout the study area.
The M6A Composite Sequence. — The final carbonate falling stage deposit discussed here is from the M6A sequence, which is notable for the very large facies dislocation across the forced regression surface. The M6A sequence contains the lower Salvisa, middle Salvisa, upper Salvisa, Cornishville, Brannon and Donerail members of the Lexington Limestone. The lower Salvisa forms the TST of the M6A 3rd-order composite sequence (Fig. 18). The argillaceous middle Salvisa and Brannon are split by the clean grainstone of the upper Salvisa forming the HST. The Donerail member forms the FSST.

As in the sections above, exploring the characteristics of the upper portion of the HST is key to understanding the nature of the forced regression surface. As in the M6C sequence the forced regression surface is highly erosional. The Brannon is easily traced across the study area through recognition of a series of characteristics, including: (1) it directly overlies stromatoporoid-bearing pack-grainstones of the Cornishville member, (2) its basal contact with the Cornishville is marked by a prominent ferric hardground, (3) it is dominated lithologically by tabular, rhythmically bedded shale/marl alternations, typically with a sparse fauna of small, thin-shelled brachiopods and bivalves, and a Chondrites ichnofacies (Fig. 19A-C, 20E), (4) it contains several horizons of soft-sediment deformation (Fig. 19A-E, 20A), and (5) it has a sharp upper contact with argillaceous calcarenites of the Donerail member (Fig. 20A, D). This unit was deposited below storm wave base across much of central Kentucky, though in some areas shows slight shallowing upward to include distal storm beds. The Brannon varies greatly in thickness across the study area resulting from a combination of eustatic drop and movement on a number of basement faults with adjacent areas of uplift and subsidence (Fig. 21; Cressman, 1973; McLaughlin et al., 2004). The contact with the overlying Donerail is sharp and ranges from channeled to flat (Fig. 17).
FIGURE 18. Stratigraphic column and location map for the M6A sequence at Frankfort, Kentucky (section 49).
FIGURE 19. Outcrop images of the Donerail member and surrounding units along a south to north transect. A) An extremely truncated and slightly deformed Brannon (BR) overlain by Donerail (DR) at Harrodsburg, Kentucky (marks on Jacobs staff are 10 cm increments; section 70). In this area the highly truncated Brannon is easily traceable as it rests sharply on a heavily mineralized hardground at the top of the Cornishville member (CV). The contact between the slightly deformed calcisiltites and shales of the Brannon and the heavily amalgamated argillaceous calcarenites of the Donerail is sharp and typically planar. B) To the north the Brannon thickens rapidly and is spectacularly incised by channels filled with argillaceous calcarenite of the Donerail at a large outcrop along the Bluegrass Parkway (section 59). Note the distinctive limestone marker bed within the upper Brannon and the iron mineralized contact of the Donerail and Sulphur Well (adjacent to “SW”). C) On the southern side of Frankfort the Brannon is heavily deformed and the Donerail becomes increasingly thin, removed under the M6B sequence boundary. D) Away from Frankfort the Donerail thickens rapidly, dominated by thin amalgamated calcarenite, that sharply overlies and increasingly thick, but highly deformed Brannon. E) At Swallowfield (section 35) the Brannon is approximately 2 meters thick, highly deformed, and sharply overlain by the nearly 3 meter-thick Donerail maintaining a lithology of argillaceous calcarenite, whereas the Perryville at the base of the sequence has transitioned into very distal facies. F) At its most northerly outcrop the Donerail is still dominated by laminated argillaceous calcarenites (section 26).
FIGURE 21. Schematic cross sections of the M6A sequence from Danville, Kentucky to the Bluegrass Parkway just east of its junction with Kentucky Route 127 (~40 X kilometers). The irregular thickness distributions are tied to both differential subsidence across faults and eustatic fluctuations (see McLaughlin et al., 2004 for further discussion of local tectonic effects).
The Donerail is composed of argillaceous calcarenites similar to those described above for the Locust Creek and lower Devils Hollow (Fig. 20F). In contrast to the other FSSTs, soft-sediment deformation is uncommon in the Donerail (Fig. 20B). The underlying, distal ramp facies of the Brannon, by contrast, are very commonly deformed, as noted above. Across the southern half of the study area (sections 59, 70, 72, 74) south of Frankfort, the Sulphur Well/Donerail contact displays a locally developed ferric crust and reworked iron mineralized Donerail clasts within the basal Sulphur Well (Fig. 19B, 20C) interpreted as the M6B sequence boundary. At the abandoned quarry at Stamping Ground, Kentucky the Donerail is overain by a lag of fenestral micrite clasts at the base of the Sulphur Well, suggesting that this contact represents both a sequence boundary and transgressive ravinement surface in this area. In more distal parts of the paleoramp the Sulphur Well is differentiated from the underlying Donerail as being more thickly bedded with a lower argillaceous content and contains a more diverse, deepening-upward fauna dominated by bryozoans and crinoids (Fig. 19F).

**Similarity to other studies.** Few studies have documented falling stage systems tracts from the carbonate-dominated side of foreland basins. However, in the northern Taconic foreland basin Brett and Baird (2002) and Brett et al. (2004) described similar vertical facies relationships and channels in age equivalent Upper Ordovician strata in New York and Ontario. Wilson (1948) also reported channels filled with argillaceous calcarenite from Richmondian age Upper Ordovician strata of the Nashville Dome. Zenger (1965), Brett et al. (1990) and Goodman and Brett (1994) described very similar facies successions in late Llandovery and Wenlock age Lower Silurian strata of the Salinic foreland basin of New York and Ontario. Similar forced regressive deposits are common features of mid-continent Carboniferous cyclothems (i.e. Heckel 1994, 2002). In the case of cyclothems, a sharp contact between outer-ramp fossiliferous
mudstones and inner ramp argillaceous calcarenites and/or oolites are often observed. The inner ramp facies are occasionally capped by exposure surfaces and caliche nodules, rhizoliths, laminar ferric crusts and other pedogenic features (Goldstein, 1988).

**Channels Associated with FSSTs**

Sharp erosion surfaces with channel-form geometries occur both at the base and within FSSTs in both carbonate- and siliciclastic-dominated strata (Fig. 22). However, stacking of channels internally within the FSST is less commonly known. Stacking of channels may be restricted to more distal heterolithic parts of the ramp such as those represented by the upper Fairview. Some Cretaceous workers have referred to similar erosion features as gutter casts. However, it is clear from the Fairview examples (Fig. 4B) that these erosional features are of a much larger scale than typical storm induced gutter casts and, indeed, that the erosional bases of these features are formed from the coalescing of many adjacent small gutter casts along single horizon. These channels have a north-south orientation parallel to small gutter casts within both the Kope and Fairview. The Lower Silurian Thorold Sandstone in southern Ontario displays a similar broad shallow channel at its base, which marks the forced regression surface (McLaughlin and Brett, in press B). Such channel-form erosion surfaces are also reported from the bases of Cretaceous FSSTs in distal ramp settings of Wyoming (Fitzsimmons and Johnson, 2000).

Channels also occur at the bases of and within carbonate-dominated FSSTs (Fig. 22). These channels are similar in size and shape to those observed in the siliciclastics and are also oriented parallel to depositional dip. Channels are most commonly observed at the base of the FSST. However, several examples within the M6A, B, and C composite sequences in central and
FIGURE 22. Channeled falling stage erosion surfaces. A-B) Channels at the Locust Creek-Gratz contact in northern (A; section 13) and central Kentucky (B; section 53). C) Low angle channel within the lower Devils Hollow in northern Kentucky (section 15). D) Highly irregular channel at the base of the Donerail in central Kentucky (section 59).
northern Kentucky occur where stacked channels filled with argillaceous calcarenite cut out underlying more distal FSST deposits (Fig. 11D, 22C). Rankey (2002) described a similar occurrence of a channel formed in the carbonate-dominated Keokuk Formation (Middle Mississippian), an intracratonic ramp succession in southeastern Missouri. The channel cut into cherty wackestones interpreted to be deposited below storm wavebase and was filled with a lower quartz silt-rich calcarenite unit with foreset beds up to 4 meters-thick. In this Mississippian example, the channel and its fill lack any evidence of subaerial exposure or tidal features. Rankey (2002) refers to this feature as a valley as it contains multiple sub-channels, however, it is typically not thicker than 5 meters. Like the Upper Ordovician examples described above, the long axes of the Mississippian channels are oriented parallel to depositional dip (east-west toward the deeper-water Illinois Basin). Similar to the M6C sequence described above, the base of the FSST is condensed. Rankey (2002) reports that portions of this channeled erosion surface shows borings and a basal pebbly lag bed, indicating a period of sediment starvation, likely resulting from erosion and bypass, occurred for some period prior to filling. Similar to the channels described above, Rankey (2002) interprets the Keokuk channels to have formed in response to sea level fall.

LATERAL RELATIONSHIPS BETWEEN CARBONATE- AND SILICICLASTIC-MARGIN FALLING STAGE SYSTEMS TRACTS

Although the Upper Ordovician examples described above yield a very detailed picture of falling stage processes in both a siliciclastic-dominated and carbonate-dominated facies, they do not provide a sense of how and where these two systems might interact across the foreland basin profile. At this time such a transect is not available for the Upper Ordovician; however, multiple
FIGURE 23. Images of the early Wenlock falling stage systems tract across the Salinic foreland basin in New York and Pennsylvania. A) Highly amalgamated argillaceous dolo-calcarenites of the Rockway Dolostone deposited on the carbonate margin of the Salinic foreland basin (Lewiston, New York). Williamson shale almost completely eroded away at this section. Sharp contact with the skeletal grainstone-rudstones of the Irondequoit at very top of image. B) Close-up of the Rockway-Irondequoit contact near the basin center (Rochester, New York). Note how shaly the Rockway has become. C) Thick amalgamated beds of the Rockway-equivalent Dawes Sandstone overlain by the Irondequoit Limestone equivalent. Note ball and pillow deformation of two horizons in the Dawes sandstones (arrows; Allenport, Pennsylvania).
examples are published from the Lower Silurian of eastern North America. For the sake of discussion only the Lower Silurian Rockway-Dawes-lower Keefer (Brett et al., 1990) lateral succession from the Salinic foreland basin the of Ontario and New York is summarized in the following. The argillaceous dolomitic calcarenite-dominated Rockway Dolostone (~ 4-5 meters thick) sharply overlies the organic-rich Williamson Shale and in turn is sharply overlain by clean and occasionally glauconitic skeletal grainstones of the Irondequoit Limestone in the Niagara region of western New York State (Fig. 23A). Toward Rochester, New York the Rockway becomes less amalgamated, grading into rhythmically interbedded calcarenites and shales where it is as little as 2 meters thick (Fig. 23B). Those shales become siltier to the east and the argillaceous dolomitic calcarenites become enriched in quartz silt and sand, so that in exposures east of Syracuse the Rockway has graded laterally into a distal facies of the Dawes Sandstone (~ 3-4 meters thick). The Dawes thickens rapidly to the east and south, although the outcrop belt ends in New York before it gains great thickness. However, exposures of the Dawes equivalent lower Keefer Sandstone in central Pennsylvania near Juniata is as much as 20 meters thick (Fig. 23C).

TEMPORAL DISTRIBUTION AND VARIABILITY OF FALLING STAGE SYSTEMS TRACTS

As described above falling stage systems tracts are readily identifiable and have diagnostic characteristics in the Upper Ordovician M6A-C2 sequences. However, surrounding sequences do not contain as distinctive features of FSST deposits. That is not to say that these surrounding sequences do not contain falling stage systems tracts, but they are not recognizable by the same criteria given above. In these sequences it is not only the falling stage systems tracts that are
more difficult to recognize, but also transgressive systems tracts. Additionally, small-scale
cycliclity is less well developed in these sequences.

In contrast to the underlying HST and overlying TST, the FSST appears to mark a period of
environmental stress in the foreland basin recognized by the prevalence of low diversity faunal
assemblages, as noted in the Upper Ordovician examples above, though this may be partly a
taphonomic artifact. Rapid decrease in accommodation leads to greatly increased sedimentation
rates on the siliciclastic margin and overall increased turbidity across foreland basins, which
could certainly result in disruption of fossil populations. Further, the erosive nature of stepped
forced regressive shoal deposits within the FSST may also be largely removing the record of
more diverse mid-ramp facies. In a few cases from the Lower Silurian and Middle Devonian
these more diverse assemblages are preserved, as described below.

The “Willowvale facies” of the Lower Silurian Rockway Dolostone (FSST) is highly
fossiliferous in contrast to its up- and down-ramp equivalents and many other FSSTs deposited
on the carbonate-margin of foreland basins. The Willowvale facies represents deposition in the
outer mid-ramp of the carbonate margin of the Salinic basin in central New York. This facies
contains high diversity assemblages of benthic invertebrates dominated by brachiopods,
bryozoans, crinoids, corals, sponges, and trilobites. The most diverse assemblages are associated
with small locally developed bioherms.

Another highly fossiliferous FSST deposit occurs in the lower half of the Middle Devonian
Centerfield Member of the Ludlowville Formation in western New York. Here the strata
representing the shallow carbonate margin of the Acadian foreland basin for this sequence have
been largely removed by subsequent erosion associated with forebulge migration (Brett and
Baird, 1996). However, the outer-(carbonate) ramp facies to the siliciclastic margin are
preserved and well exposed. The carbonate-margin outer-ramp facies of the lower Centerfield are highly fossiliferous and contain diverse assemblages of benthic invertebrates similar to those described above for the Lower Silurian Willowvale facies. In this case the highly diverse FSST stands in sharp contrast to the barren dark gray to black shales of the underlying Butternut Shale. Another unusual aspect is that the vertical shallowing observed in the faunas of the lower Centerfield FSST reverse in the upper Centerfield TST and display a deepening upward pattern (symmetrical cycles of Brett and Baird, 1986). This pattern changes abruptly into the more siliciclastic-dominated basin center to the east where the lower Centerfield becomes sparsely fossiliferous and heavily Zoophycos bioturbated silty mudstone, but still sharply overlain by the fossiliferous upper Centerfield (asymmetrical cycles of Brett and Baird, 1986).

**DISCUSSION**

The general model for genesis of falling stage systems tracts proposed by Plint and Nummedal (2000) suggests that these deposits begin to form during the initial stage of relative sea level fall (see Fig. 1B). However, in trying to clarify inconsistencies and problems in the definition of lowstand systems tract, they generate or at least highlight a problem with the definition of maximum flooding surface. Maximum flooding surface is defined as the point of greatest relative sea level rise and maximum transgression of the shoreline. As such, it separates transgressive from regressive strata. Interpretation of the basal surface of forced regression as forming at the initiation of fall in relative sea level causes it to overlap with the maximum flooding surface and thereby eliminate the highstand systems tract. In their model they place the maximum flooding surface on the rising arm of the relative sea level curve and suggest that highstand deposition occurs wholly during relative sea level rise. High resolution integrated
sequence stratigraphic studies which utilize paleoecological data from the mixed carbonate-
siliciclastic margins of the Upper Ordovician Taconic, Lower Silurian Salinic, and Middle
Devonian Acadian foreland basins in eastern North America provide a different interpretation of
highstand systems tracts and an alternative position for the basal surface of forced regression
(Fig. 1D).

Falling stage systems tracts are commonly most difficult to identify in the basin center in
foreland basins with relatively high subsidence rates. Because differential subsidence is
commonly not balanced by differential sedimentation in the basin center the rate of subsidence
may equal or even exceed the rate of eustatic fall. However, it is the rate of relative sea level
change at the bayline (sensu Posementier et al., 1988) of the siliciclastic margin (creating or
destroying near shore sediment traps) that dictates the amount and grain size of allochthonous
sediments delivered to the basin center. Thus, the falling inflection point on the eustatic curve
may be marked by an abrupt increase in siliciclastic sedimentation rate and grain size in the basin
center, yet may contain a benthic fauna that suggests deepening upward. Its upper contact will
be capped by a condensed bed when eustatic rise resumes.

CONCLUSION

In conclusion, foreland basins record falling stage deposits on both their siliciclastic- and
carbonate-dominated margins. The shared characteristics between FSSTs of the carbonate- and
siliciclastic-margins of foreland basins include: (1) sharp forced regression surface at their base
(planar in up-ramp areas and highly irregular to channeled in the lower ramp), (2) a sharp facies
dislocation across the forced regression surface, (3) progradational stacking of small-scale cycles
(if present), (4) long distance regression, (5) a deformation prone sedimentary architecture, and (6) the upper surface is the sequence boundary or correlative conformity.

Notable differences in the FSSTs of carbonate- versus siliciclastic-margins are also noteworthy and include: A) a much greater absolute thickness on the siliciclastic margin, B) a greater proportion of the thickness of the depositional sequence is dominated by the FSST on the siliciclastic margin, C) small-scale cycles of the FSST tend to be more highly amalgamated on the carbonate-margin, D) channels formed at the forced regression surface tend to be more well developed on the carbonate-margin.

Similarity in motif of FSSTs on opposing sides of foreland basins and recognition of that motif in both transitional (Upper Ordovician-Middle Mississippian; Cretaceous) and ice house (Upper Mississippian-Permian) periods of the Paleozoic and Mesozoic reinforces the robustness of current sequence stratigraphic models. That both margins of foreland basins should develop such similar stratigraphic architectures in the face of such markedly different rates of subsidence and sedimentation indicates supports the original hypothesis put forth by Vail et al. (1977) that extrabasinal processes, namely eustasy, are the primary mechanism driving the formation of depositional sequences.

ACKNOWLEDGEMENTS

PIM would like to recognize the important contributions of many colleagues who unselfishly shared observations and insights; in particular, Steve Holland, Susie Taha McLaughlin, Sean Cornell, Arnie Miller, Tom Algeo, Mike Pope, Brian Witzke, Norlene Emerson, Frank Ettensohn, Andrew Webber, and Kate Bultinski for insightful discussions of Upper Ordovician stratigraphy and paleoecology. Thanks to Marcus Johnson, Axel Munnecke, and Mikael Calner

for discussion of Silurian stratigraphic patterns. Thanks to Alex Bartholomew, Gordon Baird, Mike DeSantis, Chuck Ver Straeten, Jed Day and Jeff Over for discussions of Devonian stratigraphy, sedimentology, and paleoecology. We are also indebted to the kindness and generosity of the Kentucky Geological Survey Core Repository personnel. Acknowledgement is also made to the donors of the American Chemical Society Petroleum Research Fund for partial support of this research.
REFERENCES


HECKEL, P.H., 1994, Evaluation of evidence for glacio-eustatic control over marine Pennsylvanian cyclothems in North America and consideration of possible tectonic effects, in


MCLAUGHLIN, P.I., C.E. BRETT, S.L. TAHA Mclaughlin AND S.R. CORNELL, 2004, High-resolution sequence stratigraphy of a mixed carbonate-siliciclastic, cratonic ramp (Upper Ordovician; Kentucky-Ohio, USA): insights into the relative influence of eustasy and


Pope, M.C. and Read, J. F., 1997, High-Resolution surface and subsurface sequence stratigraphy of the Middle to Late Ordovician (late Mohawkian-Cincinnatian) foreland basin rocks, Kentucky and Virginia: AAPG Bulletin, v. 81, p. 1866-1893.


Weir, G.W., Peterson, W.L., and Swadley, W.C., 1984, Lithostratigraphy of Upper
Ordovician strata exposed in Kentucky: US Geological Survey Professional Paper 1151-E,
121 p.

Wilson, C.W., 1948, Channels and channel-filling sediments of Richmond age in south-central

APPENDIX 1
Locality Descriptions

1. **K445**—Exposure along Kentucky Route 445 (River Rd.) just west of the intersection with Kentucky Route 8 (Mary Ingles Highway; 39° 3’ 20” N, 84° 26’ 4” W). Middle portion of Kope Formation composite reference section of Brett and Algeo (2001).

2. **Banklick Creek**—Exposure along the southern side of Banklick Creek, parallel to Grand Avenue at the intersection with Readlin Rd., in Winston Park, Kentucky (39° 2’ 1” N, 84° 30’ 2” W). At this locality a small monocline reveals the upper 15 meters of the Lexington Limestone. This locality represents one of the northernmost and distal facies of the Lexington Limestone in Kentucky. Dark organic rich shales of the Bromley member form the lowest part of the section, the Locust Creek member is represented by a thin interval of deformed argillaceous calcarenites, and all but the uppermost beds of the Point Pleasant member are exposed in the stream bank just below the level of Grand Avenue.

3. **Boat Run**—Bedding plane and low cut bank exposures on north branch of Boat Run (creek), approximately 250 meters northeast of its intersection with Ohio Route 52 (approximately 400 meters northwest of Clermontville-Laurel Road) in Clermontville, Ohio (38° 55’ 54” N, 84° 15’ 14” W). The exposed creek section is accessible from Clermontville Spur. Deformed beds of the Devils Hollow Member of the Lexington Limestone form some of the lowest exposures; the lower Fulton submember of the Kope Formation forms the highest exposures.

4. **Point Pleasant North composite**—Composite section derived from measurement of two partially covered steep creeks/small abandoned quarries on Possum Hollow (38° 53’ 53” N, 84° 14’ 11” W) and an adjacent creek 2 kilometers to the south; approximately 4 kilometers north of Point Pleasant, Ohio, exposed on east side of Ohio Route 52. Type section of the Point Pleasant member of the Lexington Limestone.

5. **Ivor North**—Road cut on Kentucky Route 8 nearly 1.0 kilometer north of its intersection with Ivor Rd. (38° 52’ 7” N, 84° 14’ 4” W). Partially covered road cut exposing the Locust Creek through Point Pleasant members of the Lexington Limestone.

6. **Carntown Composite Section**—Composite section derived from measurement of the Carntown North section and Carntown Quarry section. The Carntown North section is a small road cut and overlying partially exposed hill slope located on Kentucky Route 8, approximately 1 kilometer north of the intersection with Kentucky Route 154 (38° 50’ 49” N, 84° 14’ 34” W). It ranges stratigraphically from the Bromley member of the Lexington Limestone upward into the Fulton submember of the Kope Formation. The Carntown Quarry section is an abandoned quarry located on the west side of Kentucky Route 8, approximately 150 meters south of the intersection with Kentucky Route 154 (38° 50’ 9” N, 84° 14’ 32” W). The section is partially covered and extends from the Devils Hollow Member of the Lexington Limestone upward into the basal part of the Brent submember of the Kope Formation. The lower portion of the quarry is now the site of a power conversion grid.

7. **Bear Creek Quarry**—Bear Creek Quarry is an abandoned quarry located on Bear Creek Road approximately 200 meters north of its intersection with Ohio Route 52, 1.5 km west of Chilo, Ohio (38° 48’ 7” N, 84° 9’ 31” W). The quarry has been partially backfilled, however, lower sections containing the Devils Hollow Member of the Lexington
Limestone are exposed on the south end of the quarry; the northern part of the quarry contains exposures up into the Fulton submember of the Kope Formation.

8. **Sterling Materials Quarry**—Exposures on the road entering the underground limestone mine at Sterling Materials Quarry located on the north side of Kentucky Route 42, approximately 5.5 kilometers northeast of its intersection with Kentucky Route 127 (38° 49’ 6”, 84° 46’ 43” W). The lowest exposures at the mine entrance are in the Stamping Ground Member of the Lexington Limestone and the top of the cut is in the Fulton submember of the Kope Formation.

9. **Foster East**—Road cut along northbound lane of Kentucky Route 8, approximately 0.5 km east of the intersection with Route 1019 and the village of Foster, Kentucky (38° 47’ 59” N, 84° 12’ 34” W). Greendale Member mostly covered at base of cut overlain by nearly the total thickness of the lower and upper portions of the Devils Hollow Member.

10. **Holst Creek/AA Highway**—Road cut along the northbound lane of Kentucky Route 9 (AA Highway) just south of where it crosses Holst Creek (38° 46’ 50” N, 84° 12’ 39” W). This series of closely spaced cuts discontinuously exposes the upper Point Pleasant member of the Lexington Limestone upward through the lower Fairview Formation.

11. **Glencoe North**—Road cut along Kentucky Route 42 extending from near the intersection with Kentucky Route 127 westward for nearly 1.0 kilometer (38° 49’ 6” N, 84° 46’ 43” W). Lowest exposures are in the Gratz member of the Lexington Limestone and extend upward into the lower meter of the Fulton submember of the Kope Formation.

12. **Butler East**—Partially covered road cut on Kentucky Route 27, just over 1.0 kilometer east of Butler, Kentucky (38° 47’ 18” N, 84° 21’ 3” W). The slightly discontinuous section contains the Locust Creek member of the Lexington Limestone upward through the Brent submember of the Kope Formation.

13. **Bradford Quarry**—Abandoned quarry on Kentucky Route 8, 0.2 kilometers north of intersection with Kentucky Route 1109 (38° 47’ 4” N, 84° 8’ 45” W). Road cut 0.5 kilometers to the south displays similar stratigraphic interval. Upper portion of the Devils Hollow Member of the Lexington Limestone up to Brent submember of the Kope Formation exposed in quarry.

14. **Boston Methodist Church**—Small outcrop behind the Methodist church along Kentucky Route 27, just north of the village of Boston, Kentucky (38° 46’ 2” N, 84° 21’ 11” W). Partially covered section contains the Locust Creek member of the Lexington Limestone upwards through the Fulton submember of the Kope Formation.

15. **Woolcott NW (Route 1951)**—Road cut on Kentucky Route 1951 (Woolcott-Johnsville Rd.) just northwest of Woolcott, Kentucky (38° 45’ 6” N, 84° 7’ 6” W). Graptolitic brown shales of the uppermost Greendale Member of the Lexington Limestone partially covered at base and overlain by the total thickness of the lower and upper parts of the Devils Hollow Member.

16. **Menzie North**—Outcrop along the CSX Railroad approximately 1.0 kilometers north northwest of Menzie (38° 44’ 49” N, 84° 20’ 50” W). Section exposes the Devils Hollow member of the Lexington Limestone at the base and extends upward into the basal Point Pleasant member.

17. **Menzie South**—Outcrop along the CSX Railroad approximately 0.5 kilometers south of Menzie, Kentucky (38° 44’ 2” N, 84° 20’ 30” W). Section contains the Locust Creek member of the Lexington Limestone at the base and extends upward into the Point Pleasant member.
18. **Woolcott composite**—Stratigraphic data from measurements of road cut on Kentucky Route 1159 just south of intersection with Kentucky Route 9 (AA highway) approximately 0.5 kilometers south of the Woolcott covered bridge (38° 43’ 59” N, 84° 6’ 6” W) and a steep creek section perpendicular to Salem Ridge Road, approximately 50 meters south of Kentucky Route 9; additional faunal data taken from bedding plane exposures of the Point Pleasant member on adjacent Poe Creek (38° 43’ 22” N, 84° 6’ 43” W). Combined exposure from middle Bromley member of the Lexington Limestone upward to lower Brent submember of the Kope Formation. Many adjacent exposures along Kentucky Route 9 extend from Fulton submember upward to the Snag Creek submember of the Kope Formation (see Brett and Algeo, 2001 for further locality information).

19. **Kincaid Lake Section**—Road cut on the western entrance road to Kincaid Lake State Park from Kentucky Route 159 (38° 43’ 15” N, 84° 17’ 59” W). Exposes the Locust Creek and Point Pleasant members of the Lexington Limestone.

20. **Falmouth Composite**—Road cuts along Kentucky Route 27 at both the north (38° 40’ 46” N, 84° 20’ 45” W) and south (38° 39’ 52” N, 84° 19’ 55” W) margins of Falmouth, Kentucky. Both sections are partially covered. The northern section is the more extensive and begins in the Devils Hollow member of the Lexington Limestone at the base and continuing upward into the Fulton submember of the Kope Formation. (See Lorenz, (198X) for additional data).

21. **Maysville North (Rte. 3071)**—Enormous roadcuts on the north and south sides of Kentucky Route 3071 just north of Maysville at Moranburg, Kentucky (38° 40’ 34” N, 83° 48’ 43” W). Base of cut exposes upper portion of the Brent sub-member of the Kope Formation, overlain by Fairview, Miamitown, Bellevue, and Corryville formations.

22. **Maysville Central (Rte. 11)**—Road cut on the south side of Kentucky Route 11 in Maysville, Kentucky (38° 37’ 35” N, 84° 45’ 3” W). Exposes lower Fairview upward through Corryville formations.

23. **Sunrise Road cut**—Small road cut and adjacent creek section on Kentucky Route 1284 (Sunrise-Claysville Road; 38° 28’ 32” N, 84° 57’ 31” W). The section contains at its base the Locust Creek member of the Lexington Limestone and extends upward into the Point Pleasant member.

24. **Mays Lick**—Road cut on the north side of Kentucky Route 68 just southwest of its intersection with Route 324, just east of Mays Lick, Kentucky (38° 31’ 2” N, 83° 50’ 17” W). Partially exposed Pioneer Valley sub-member of the Kope Formation at base, cut continues upward through basal Fairview Formation.

25. **Gratz Northwest**—Small road cut on Kentucky Route 389 on the western wall of the Kentucky River valley, approximately 4.0 kilometers northwest of Gratz, Kentucky (38° 29’ 27” N, 84° 59’ 48” W). Section contains the Locust Creek member of the Lexington Limestone at its base and extends upward into the lower part of the Fulton submember of the Kope Formation.

26. **Gratz East**—Large road cut on Kentucky Route 355 adjacent to the Kentucky River, approximately 2.5 km east of Gratz, Kentucky (38° 27’ 56” N, 84° 55’ 32” W). Section exposes at its base the Doneraile member of the Lexington Limestone and continues upward through the uppermost part of the Devils Hollow Member.

27. **Blue Licks Composite**—Series of four closely spaced road cuts on Kentucky Route 68 extending from the entrance to Blue Licks State Park southwestward for approximately 288
4.0 kilometers (38° 25’ 47” N, 83° 59’ 31” W). Primary section is a large road cut on the west side of Kentucky Route 68 at the entrance to Blue Licks State Park. This section exposes the upper two meters of the Point Pleasant member of the Lexington Limestone and the entire Fulton and the lower few meters of the Brent submembers of the Kope Formation. Sections to the southwest lack the Fulton and are primarily in the Bromley through Point Pleasant members of the Lexington Limestone. An addition section exposing the Point Pleasant down into the Devils Hollow Member of the Lexington Limestone is located on the park road just inside the entrance to Blue Licks State Park.

28. **Sadieville Northwest**—Large road cut along Interstate 75 approximately 2.5 kilometers north of Exit 136 and 4.0 kilometers northwest of the village of Sadieville, Kentucky (38° 24’ 2” N, 84° 34’ 36” W). This cut exposes strata extending from the Peaks Mill member of the Lexington Limestone up to the lower few meters of the Fulton submember of the Kope Formation.

29. **Monterey East**—Large road cut and creek section along Kentucky Route 127 at Sawdridge Creek, 1.5 kilometers east of Monterey, Kentucky (38° 25’ 15” N, 84° 51’ 18” W). The base of the creek section begins in the Perryville Member of the Lexington Limestone and continues up a vertical cliff face to the Strodes Creek Member where it intersects the road cut. Partially exposed strata in the northernmost portion of the road cut continue well up into the Brent submember of the Kope Formation.

30. **Cynthiana North**—Road cut on Kentucky Route 27 a few hundred meters north of its intersection with North Church Street on the northern margin of Cynthiana, Kentucky (38° 23’ 59” N, 84° 17’ 36” W). Exposure of the lower several meters of the Point Pleasant and entire Locust Creek members of the Lexington Limestone are located at the hillcrest. Discontinuous exposure leading downhill to the northeast contains the Gratz and Bromley members of the Lexington Limestone.

31. **Frank Clark Rd.**—Road cut on the west side of Kentucky Route 127 (Bluegrass Corridor), just north of the intersection with Frank Clark Rd. (Route 607). Section exposes at its base the upper portion of the Sulphur Well Member of the Lexington Limestone and continues upward to the upper part of the Devils Hollow Member.

32. **Sadieville West**—Road cut on Kentucky Route 32 just west of the village of Sadieville in the Elk Lick Creek valley (38° 23’ 19” N, 84° 32’ 32” W). The road cut extends upward from the Bromley member of the Lexington Limestone into the upper part of the Point Pleasant member.

33. **Pleasant Valley**—Large road cut along Kentucky Route 32, just south of where it crosses Fleming Creek, approximately 0.5 km east of the village of Pleasant Valley (38° 22’ 40” N, 83° 55’ 55” W). Base of the section exposes the top of the Strodes Creek Member of the Lexington Limestone and continues upward through the basal Bromley member.

34. **Cynthiana Southwest**—Road cut on Kentucky Route 62 approximately 3.0 kilometers southwest of its intersection with Kentucky Route 27 on the southwestern outskirts of Cynthiana, Kentucky (38° 23’ 59” N, 84° 17’ 36” W). This small, but well exposed road cut contains most of the Locust Creek member and the lower portion of the Point Pleasant member of the Lexington Limestone.

35. **Swallowfield North**—Large road cut along Kentucky Route 127, 2.0 kilometers north of Swallowfield, Kentucky (38° 20’ 58” N, 84° 51’ 22” W). This very long road cut (approximately 2 kilometers long) contains strata of the Brannon Member of the Lexington Limestone at its base up to partially covered exposures of the Brent
submember of the Kope Formation at its top. The Point Pleasant-Fulton contact is easily accessible in a low road cut, which forms the northernmost exposure of this large road cut on the western side of Route 127.

36. **Millersburg NE**—Small road cut on Kentucky Route 68 at its intersection with Kentucky Route 36, approximately 5 km northeast of Millersburg, Kentucky (38° 19’ 30” N, 84° 6’ 19” W). Section exposes the uppermost Greendale Member of the Lexington Limestone at its base and continues upward into the upper portion of the Devils Hollow Member.

37. **Old Owenton Road**—Cluster of four large road cuts along Kentucky Route 127 on the southern side of the Elkhorn Creek valley at the intersection of Old Owenton Road, approximately 3 kilometers northwest of Peaks Mill, Kentucky (38° 18’ 22” N, 84° 50’ 40” W). The lowest strata exposed in the road cuts (northernmost cut) belong to the Sulphur Well Member of the Lexington Limestone; the highest strata (southernmost cut) belong to the Brent submember of the Kope Formation. Partially exposed small natural outcrops below the road level may extend down as low as the Perryville Member of the Lexington Limestone.

38. **Sherburne**—Large road cuts on Kentucky Route 11 on either side of the Licking River Valley, just south of the village of Sherburne (38° 16’ 44” N, 83° 48’ 17” W). Section exposes Grand View sub-member of the Kope Formation at its base and is exposed discontinuously upward through the Bellvue Formation.

39. **Deleplain West**—Large road cut on the east side of Interstate 75 at the northbound ramp of Exit 129 (38° 16’ 20” N, 84° 33’ 21” W). This road cut display the Locust Creek member of the Lexington Limestone at the base and extends upward to near the top of the Point Pleasant member. Several spectacular hardgrounds are present within the Point Pleasant member at this locality.

40. **Peaks Mill South**—Large road cut located approximately 5 kilometers south of Elkhorn Creek and approximately 0.5 km south of Shadrick Ferry Road on Kentucky Route 127, southwest of Peaks Mill, Kentucky (38° 16’ 6” N, 84° 50’ 55” W). The upper beds of the Peaks Mill member of the Lexington Limestone form the base of the exposure and the lower meter of the Fulton submember of the Kope Formation forms the top. A significant portion of the low Fulton and its contact with the Point Pleasant is well exposed in a small road cut a few hundred meters north of the main Peaks Mill road cut.

41. **Deleplain SW**—Road cut on the northbound lane of I-75 at mile marker 128, approximately 1.5 km southwest of the hamlet Deleplain, Kentucky (38° 13’ 31” N, 84° 33’ 7” W). Section begins in the uppermost Greendale Member of the Lexington Limestone and continues upward into the basal Bromley member.

42. **Oxford SW**—Series of small road cuts on either side of Kentucky Route 62, approximately 4 km southwest of Oxford, Kentucky (38° 14’ 20” N, 84° 31’ 46” W). Lowest exposed strata belong to the Strodes Creek Member of the Lexington Limestone and continue upward into the lower quarter of the Bromley member.

43. **Frankfort North**—Immensely large road cut along Kentucky Route 127, just north of its intersection with Wilkinson Boulevard on the northern margin of Frankfort, Kentucky (38° 13’ 11” N, 84° 51’ 8” W). Section begins in the Tyrone Formation and continues upward into the upper Devils Hollow member of the Lexington Limestone.

44. **Frankfort Northwest (421)**—The section is located on Route 421 on the northwest side of Frankfort, Kentucky, approximately 300 meters north of the intersection of Kentucky
Route 127 and 421. It is one of the largest outcrops in the Bluegrass Region (38° 12’ 52” N, 84° 53’ 24” W). Stratigraphically it covers more than 80 meters, from the top of the Curdsville Member of the Lexington Limestone at the base upward into the lower Fulton submember of the Kope (Clays Ferry) Formation at the top.

45. Frankfort West—Immensely large exposure on Kentucky Route 127, approximately 0.5 km south of its intersection with Devils Hollow Road on the west side of Frankfort, Kentucky (38° 12’ 2” N, 84° 53’ 38” W). This succession begins in the Grier Member of the Lexington Limestone and continues upward into the lower part of the Point Pleasant member.

46. Donerail North—Large road cut on I-75 approximately 1.5 km south of where it crosses Newtown Pike to the east of Georgetown, Kentucky (38° 12’ 6” N, 84° 31’ 43” W). The section begins in the lower Salvisa member of the Lexington Limestone and continues upward to near the base of the Strodes Creek Member.

47. Frankfort Days Inn—Small road cut on the east side of Kentucky Route 127, just south of its intersection with Kentucky Route 60 (Louisville Road), below the Days Inn in Frankfort, Kentucky (38° 10’ 49” N, 84° 53’ 40” W). The Stamping Ground Member of the Lexington Limestone is partially covered at the base of this exposure and continues upward to near the top of the Devils Hollow Member.

48. Versailles Exit I-64—Small outcrops on both north and southbound Exit 58 ramps of I-64 where it crosses Kentucky Route 60. Lowest exposures (in the northbound lane of KY Rte. 60 at eastbound onramp to I-64) are in the upper Stamping Ground Member of the Lexington Limestone and continue discontinuously upward into the lower part of the upper Devils Hollow Member.

49. Frankfort South (I-64 Kentucky River Valley Composite Section)—Series of large road cuts along the eastbound and westbound lanes of Interstate 64, extending for approximately 6.0 kilometers westward from its intersection with Kentucky State Route 127, in the Kentucky River valley. The section contains nearly the entire Lexington Limestone, from the lower Grier Member at the base to the Point Pleasant member at the top. The Point Pleasant member forms the highest exposure in the westbound lane on the eastern side of the Kentucky River valley (38° 10’ 20” N, 84° 50’ 0” W). Initially the section included the Fulton submember of the Kope Formation, however it is now completely covered.

50. Bridgeport West—Road cut on Kentucky Route 60 and exposures in bottom and cutbanks of Benson Creek (38° 9’ 43” N, 84° 57’ 21” W). Section contains, at its base, the Brannon Member of the Lexington Limestone and discontinuously continues upward into the Fulton submember of the Kope Formation.

51. Duckers SE—Small railroad cut where the Louisville and Nashville Railroad tracks cross Frankfort Pike, approximately 1.5 km southeast of the village of Duckers, Kentucky (38° 9’ 17” N, 84° 46’ 17” W). Section exposes the lower and upper portions of the Devils Hollow Member of the Lexington Limestone.

52. Grassy Springs Rd.—Small outcrop where Kentucky Route 60 intersects Grassy Springs Rd., approximately 4.5 km southeast of where Route 60 crosses I-64 (38° 8’ 10” N, 84° 47’ 14” W). Section exposes the Devils Hollow Member of the Lexington Limestone.

53. Lawrenceburg Northwest—Road cuts on the east and west side of Kentucky Route 127 bypass around Lawrenceburg, Kentucky, approximately 1.0 kilometer south of the junction with North Main Street (38° 3’ 39” N, 84° 55’ 17” W). Exposures contain the
Bromley member of the Lexington Limestone upward into the Point Pleasant member. Deformation within the Locust Creek member forms a prominent aspect of these road cuts.

54. Mt Zion—Small road cuts on both lanes of I-64 approximately 3 km east of where it crosses Kentucky Route 60 east of Mt. Zion, Kentucky (38° 3’ 10” N, 84° 3’ 13” W). Section exposes Garrard Siltstone upward into the basal Fairview Formation.

55. Wild Turkey Distillery West—Large partially covered road cut on Kentucky Route 62, just west of the Kentucky River and the Wild Turkey Distillery, just east of Lawrenceburg, Kentucky (38° 2’ 24” N, 84° 50’ 58” W). The section exposes the Grier Member of the Lexington Limestone at its base and continues upward into the Stamping Ground Member.

56. Lawrenceburg West—Low road cuts on Kentucky Route 44 at the intersection with Kentucky Route 127 and adjacent road cuts approximately 200 meters to the south on Kentucky Route 127 on the western margin of Lawrenceburg, Kentucky (38° 1’ 57” N, 84° 54’ 41” W). Exposures on Route 44 contain strata of the Locust Creek and Point Pleasant members of the Lexington Limestone. Exposures on Route 127 extend from the Point Pleasant member upward into the Fulton submember of the Kope Formation.

57. Winchester North—Road cut on the exit ramp (#96) and onramp to westbound I-64, just north of Winchester, Kentucky (38° 0’ 47” N, 84° 10’ 30” W). Section exposes Strodes Creek Member of the Lexington Limestone upward into the Devils Hollow Member.

58. Mountain Parkway/I-64 Intersection—Road cut on the onramp from I64 eastbound to the Mountain Parkway, east of Winchester, Kentucky (38° 1’ 11” N, 84° 85’ 5” W). Section exposes Strodes Creek Member of the Lexington Limestone upward into the Devils Hollow Member.

59. Bluegrass Parkway West—Immense roadcuts on either side of the Bluegrass Parkway on the western side of the Kentucky River Valley (37° 59’ 9” N, 84° 49’ 45” W). This section begins in the Logana Member of the Lexington Limestone and continues upward into the Sulphur Well Member.

60. Winchester South—Large road cut on the southeast side of Kentucky Route 627, approximately 4 km south of Winchester, Kentucky (37° 57’ 421” N, 84° 12’ 31” W). This section begins in the upper part of the Stamping Ground Member of the Lexington Limestone and continues upward into the Devils Hollow Member.

61. Mountain Parkway @ Stoner Cr. Valley—Small, partially covered exposures of the Garrard Siltstone on the northbound lane of the Mountain Parkway, approximately 7.5 km southeast of its intersection with I-64. This small cut only exposes a portion of the Garrard Silstone.

62. Boonesborough NW—Series of moderate size road cuts on Kentucky Route 627, approximately 3 km northeast of the Kentucky River and Boonesborough, Kentucky (37° 55’ 39” N, 84° 14’ 48” W). The series of cuts begins with the Grier Member of the Lexington Limestone in the southwest and continues upward into the Stamping Ground Member.

63. Athens SW (I-75 MM 102) —Road cuts on I-75 (mile marker 102) in both the northbound and southbound lanes, just southwest of Athens, Kentucky (37° 55’ 49” N, 84° 22’ 27” W). This small cut exposes the upper Salvisa through Brannon Members of the Lexington Limestone. The Donerail Member is exposed in a faulted outcrop just 0.5 km to the north in the southbound lane.
64. **Agawam Station**—Cuts on the CSX Railroad where it crosses under White Conkright Road, just northwest of Agawam, Kentucky (37° 55’ 14” N, 84° 5’ 50” W). This section exposes the uppermost Kope Formation (Clays Ferry facies) and the Garrard Siltstone. The upper Garrard Siltstone and its contact with the Fairview Formation are mostly covered, but accessible in small cuts on the railroad to the southeast.

65. **Boonesborough South**—Series of discontinuous exposures on Kentucky Route 627 beginning just south of the Kentucky River and Boonesborough, Kentucky and continuing for approximately 1 km (37° 53’ 26” N, 84° 16’ 48” W). The Point Pleasant Member of the Lexington Limestone and overlying Fulton and basal Brent sub-members of the Kope Formation meet the upper Kope Formation (Alexandria sub-member?) at a faulted contact in one of the larger cuts approximately 1.5 km southwest of the intersection of Route 627 and Boonesborough Road.

66. **Clays Ferry Composite**—The Clays Ferry section is a composite of closely spaced exposures. The lowest is a road cut on Kentucky Route 2328 on the southern wall of the steep Kentucky River valley, just below the Interstate 75 bridge (37° 52’ 55” N, 84° 20’ 20” W). This lowest exposure extends from above a prominent zone of fault gouge overlain by the Faulconer member of the Lexington Limestone upward into the Fulton submember of the Kope Formation, which is partially exposed at a sharp switchback in the road. This section partially overlaps with a small gully that runs up under the I-75 bridge partially exposing the Locust Creek member of the Lexington Limestone through the Fulton submember of the Kope Formation. A thin covered section in the lower Fulton is suspected as a small normal Fault. Above a covered section of approximately 3.0 meters the Fulton is exposed at the base of a large road cut on the east and west sides of I-75 that continues up through the Kope (Clays Ferry) Formation into the Garrard Siltstone. This large exposure is cut by high angle normal fault with approximately 50 meters of displacement, nearly doubling the thickness of the Kope.

67. **Salvisa**—Road cut on the northbound lane of Kentucky Route 127 just west Salvisa, Kentucky (3° 55’ 4” N, 84° 51’ 47” W). This section contains the lower Salvisa through Sulphur Well members of the Lexington Limestone.

68. **McAfee**—Road cut on north- and southbound lanes of Kentucky Route 127 approximately 0.75 km south of the intersection with McAfee Lane at McAfee, Kentucky (37° 52’ 45” N, 84° 51’ 15” W). This section exposes the Faulconer though the Donerail members of the Lexington Limestone.

69. **Pollard South**—Road cut on River Road approximately 1.5 kilometers southeast of Pollard, Kentucky (37° 48’ 0” N, 84° 29’ 58” W). Section exposes the Locust Creek and Point Pleasant members of the Lexington Limestone.

70. **Harrodsburg Bypass**—Series of low road cuts on Kentucky Route 127 Bypass just to the north and east of Harrodsburg, Kentucky (37° 45’ 46” N, 84° 50’ 18” W). Lowest exposures are in the lower Salvisa Member of the Lexington Limestone and continue upward through the Sulphur Well Member.

71. **Taylor Fork**—Spillway and adjacent creek section at the western end of Taylor Fork Lake, approximately 5.5 km southwest of Richmond, Kentucky (37° 42’ 21” N, 84° 21’ 30” W). The section is nearly continuous through the middle Kope Formation upward to the basal Fairview Formation.
72. **Perryville East**—Road cuts just east of Perryville, Kentucky on State Route 150 (37° 39’ 4” N, 84° 56’ 42” W). The section extends from the upper Salvisa Member of the Lexington Limestone into through the Sulphur Well Member.

73. **Perryville Quarry**—Small abandoned quarry within the village of Perryville, Kentucky (37° 38’ 58” N, 84° 56’ 55” W). Section extends from the Faulconer Member of the Lexington Limestone upward into the Brannon Member.

74. **Caldwell Stone Quarry (Danville)**—Large active quarry operated by Caldwell Stone located on the eastern margin of Danville, Kentucky (37° 37’ 53” N, 84° 45’ 5” W). The section extends from the Tyrone Formation at the base upward into the Kope Formation.

A. **KGS C-215**—4.75-centimeter diameter diamond drill core from Mason County, Kentucky (38° 42’ 47” N, 83° 53’ 22” W) housed at the Kentucky Geological Survey (KGS) core repository in Lexington, Kentucky.

B. **KGS C-199**—3.5-centimeter diameter diamond drill core from the Lexington East quadrangle (Kentucky; 38° 0’ N, 84° 27’ W) taken by the Kentucky Geological Survey and the United States Geological Survey (KGS-USGS) as part of the geologic quadrangle-mapping program, housed at the Kentucky Geological Survey (KGS) core repository in Lexington, Kentucky.

C. **KGS C-209**—4.75-centimeter diameter diamond drill core from Montgomery county (Kentucky; 37° 54’ N, 83° 58’ W) taken for Cominco American Inc., housed at the Kentucky Geological Survey (KGS) core repository in Lexington, Kentucky.

D. **KGS-115**—3.5-centimeter diameter diamond drill core from Lincoln county (Kentucky; 37° 36’ N, 84° 40’ W) taken for Humble Oil and Refining Co., housed at the Kentucky Geological Survey (KGS) core repository in Lexington, Kentucky.
CHAPTER 5

Hierarchy of Sedimentary Discontinuity Surfaces and Condensed Beds:

Implications for Sequence Stratigraphy on the Craton

Patrick I. McLaughlin\textsuperscript{a}, Carlton E. Brett\textsuperscript{a}, and Mark A. Wilson\textsuperscript{b}

\textsuperscript{a} H. N. Fisk Laboratory for Sedimentary Geology, Department of Geology, University of Cincinnati, Cincinnati, Ohio 45221

\textsuperscript{b} Department of Geology, The College of Wooster, Wooster, OH 44691

(submitted to Dynamics of Epeiric Seas: Sedimentological, Paleontological, and Geochemical Perspectives; Geological Association of Canada Special Volume; February, 2006)
Abstract

Sequence stratigraphic and sedimentological analyses of middle Paleozoic cratonic and foreland basin successions suggests a general hierarchy of discontinuity surfaces and condensed beds of increasing complexity, reflecting duration of seafloor exposure and physical/chemical conditions of formation. Simple firmgrounds and hardgrounds, relatively common in cratonic carbonate successions during periods of calcite seas, form the base of the hierarchy. Taphonomic analysis of composite hardgrounds, modified firmgrounds, reworked concretions, mineralized horizons, and monomictic intraformational conglomerates indicates more complex histories. Polymictic intraformational conglomerates, ironstones, and phosphorites form the most complex discontinuity surfaces/condensed beds. Complexity of discontinuities is closely linked to duration of sediment starvation and degree of reworking, which in turn is tied to stratigraphic position within depositional sequences. These surfaces form in response to carbonate and siliciclastic sediment starvation during periods of rapid environmental change associated with transgressive systems tracts, maximum starvation surfaces, and forced regression surfaces. The characteristics of discontinuity surfaces, such as lithology, thickness, and areal extent, are further tied to tectonic regime (clastic sedimentation rate), position within the basin (i.e. water chemistry), and rate of eustatic fluctuation.
Introduction

Sequence stratigraphy is aptly termed a stratigraphy of surfaces. The development of this paradigm has fostered a re-evaluation of numerous sedimentary contacts as discontinuities of varying scales. Of considerable importance is the recognition and correct interpretation of such surfaces. In particular, it is critical to evaluate the relative time significance of discontinuities and diastems (Loutit et al., 1988; Brett, 1995). The resolution of most geochronologic techniques is generally too low to permit a viable approach, especially for more ancient rocks. Among the few approaches to assessment of relative “time-richness” is the identification of scales of time-averaging based on taphonomy and geochemistry (Fürsich, 1978; Kidwell and Bosence, 1991).

Hardgrounds, concretions, ironstones, and phosphorites are among the most readily identified indicators of condensation and/or erosion on the sea floor. In the present contribution we examine a variety of taphonomic and digenetic indicators of the scale of condensation on discontinuity surfaces and propose a hierarchy of progressively more modified cemented beds, hardgrounds, and mineralized surfaces related to periods of sediment starvation associated with submarine cementation and erosion. The hierarchy is then related to a sequence stratigraphic model. Examples are primarily taken from analysis of middle Paleozoic strata in eastern North America with comparison to younger strata. In this area the Upper Ordovician, Lower Silurian, and Middle Devonian strata can be traced from the craton margin where they are composed mostly of clastic-dominated foreland basin sediments to the cratonic interior where they are almost entirely carbonate. This framework allows for analysis of individual discontinuity surfaces across a range of sedimentary regimes. The high frequency and diversity of Upper Ordovician discontinuity surfaces gives a unique view into the shorter end of the discontinuity
hierarchy that is much more difficult to detect in other Paleozoic successions, but shares similarity with Jurassic epicontinental successions described in the literature.

**Hierarchy of discontinuities/early cemented beds**

Comparative analysis of a variety of discontinuity surfaces and condensed beds forms the basis for the following hierarchy. Physical characteristics of these features are reviewed and illustrated by many previously unpublished examples from the Upper Ordovician of Kentucky. Further examples from the literature are compared for identification of trends in physical characteristics, depositional setting, and age of the deposits.

*Firmgrounds:* Firmgrounds are defined as firm but not fully indurated layers identifiable as irregular surfaces produced by sharply defined burrows, lacking borings and encrusters (Fig. 1; Bromley, 1990). Simple firmgrounds include upper surfaces of fine-grained limestone beds with incised burrows (<2-3 centimeters deep) such as the *Trichophycus* (“turkey track”) horizons common in mid-ramp storm bed facies throughout the Upper Ordovician of eastern North America (Osgood, 1970). More complex firmgrounds are typified by *Thalassinoides* burrow networks that may penetrate up to a meter into the underlying sediments. Complex firmgrounds are often closely spaced near major facies offsets and tend to be more laterally continuous than simple firmgrounds, likely representing more extensive periods of slowed sedimentation. Complex firmgrounds often occur as predecessors to composite discontinuity surfaces, discussed below. Both simple and complex firmgrounds are described from literature spanning the Phanerozoic and their implications for sequence stratigraphy are noted (Ghibaudo et al., 1996; Bertling, 1999).

*Simple Hardgrounds:* Hardgrounds are surfaces of synsedimentarily cemented carbonate layers that have undergone periods of exposure on the seafloor (Wilson and Palmer, 1992).
FIGURE 1. Simple firmgrounds from the Upper Ordovician Lexington Limestone of Kentucky. A) Oblique bedding plane view of heavily bioturbated calcarenite from the Point Pleasant Member. B) Cross-sectional view of sharp-sided burrows from the Perryville Member. Marks on staff at right are spaced at one centimeter increments.
Simple hardgrounds are typified by: 1) minimal modification of surface topography (e.g. extensive boring or thick encrustation), and in at least some instances; 2) excellent preservation of encrusters suggesting abrupt burial of terminal communities comprising one or two cohorts of encrusting populations; 3) little to no evidence for multiple generations of encrusters; and 5) a generally absence of *Trypanites*. *Sphenothallus* encrusted surfaces (Neal and Hannibal, 2000) are one of the most common types of simple hardgrounds found in the Upper Ordovician (Maysvillian-Richmondian), though they are often overlooked because of their subtle character. Cemented shell pavements form a more spectacular type of simple hardground, which exhibit thin crusts of bryozoans, crinoid holdfasts, edrioasteroids, sphenothallids, cornulitids, etc. (Fig. 2). For example, a *Rafinesquina* pavement described by Meyer (1990) from the Corryville Formation of northern Kentucky (Fig. 1) shows a simple hardground with exceptional preservation of the edrioasteroids and small trepostome bryozoans. The encrusted *Rafinesquina* vary from whole articulated specimens to broken and slightly abraded valves. A few of the edrioasteroids were present on the matrix between shells indicating that it too was stabilized and partially cemented. Meyer (1990) noted a bimodal distribution of edrioasteroids and suggested that the shell pavement remained exposed for only the duration of two spat falls (probably tens to a hundred years). This is one of the more widespread simple hardgrounds, traceable for more than 30 kilometers, in the Maysvillian (Upper Ordovician) of the Cincinnati region. Typically, simple hardgrounds are relatively restricted in their areal extent. Simple hardgrounds are most commonly found at limestone/shale contacts, but may also occur within amalgamated limestone beds at truncation surfaces that may crosscut bedding features such as trough cross bedding.
FIGURE 2. Simple hardground from the Upper Ordovician Corryville Formation of northern Kentucky and southwestern Ohio. A) Well-preserved edrioasteroid attached to cemented sediment. Remnant mudstone partially covering the surface represents the record of catastrophic burial of this layer. B) Two edrioasteroids and small bryozoan colony encrusting a brachiopod shell. Note that the encrusters come close to abutting, but do not overlap; also note the sparseness of borings.
*Concretion Beds:* Beds of simple ellipsoidal concretions of carbonate cemented mudstone or siltstone are common in some mixed carbonate-siliciclastic successions. Concretions typically form in gray to brown shales containing *Chondrites* trace fossil assemblages and thin, scattered laminae of skeletal hash (distal storm deposits) suggestive of outer ramp to deep subtidal deposits (Fig. 3). Concretionary horizons die out up dip into more proximal sections either due to changes in geochemistry of the muds (i.e. lower concentrations of organic matter, higher oxygen levels, etc.), greater degrees of resuspension and mixing of muds with other sediments (i.e. winnowing and increased bioturbation), higher (carbonate) sedimentation rates, or some combination of all three. Carbonate concretions form during periods of reduced sedimentation and are typically considered to accrete on the order of a few centimeters per thousand years (Raiswell, 1978; Canfield and Raiswell, 1991; Raiswell and Fisher, 2004).

In the Upper Ordovician of the Cincinnati region concretion horizons are most abundant at the base of the shaly Kope Formation (Fig. 3; Brett et al., 2003), but also occur higher in the formation, where they typically just underlie prominent skeletal grainstone beds. An interesting observation is that not all mixed carbonate-siliciclastic successions grade down-ramp into shales containing concretions, many times concretions are absent even when the “required” bio- and lithofacies are present. This suggests a strong lithologic component to their formation; concretions only occur in very shaly facies. In successions where shales are greater than 10 centimeters thick, concretions are much more common. For example, concretions are almost totally absent within the Lexington Limestone, which typically only contains thin shales (<10 centimeters thick) even though they occur within highly argillaceous offshore intervals (>60 percent terrigenous sediments). Concretions are much more common in the offshore facies of the overlying Kope Formation where limestones are commonly separated by tens of centimeters.
FIGURE 3. Concretions and reworked concretions from the Upper Ordovician Fulton sub-member of the Kope Formation. A) Concretion horizons in organic-rich shales; hammer for scale. Note that the concretions are rather large and nearly form a continuous band. B) Small reworked concretions in skeletal grainstone; 10 centimeter long knife for scale. Note the imbrication of some of the lower concretions. Many of these concretions are encrusted by bryozoans and crinoids on both their upper and lower surfaces.
of shale. This is a bit of a paradox as it suggests that successions with relatively high sedimentation rates (~10 cm/1 k.y.) also contain some of the most easily recognizable evidence of sediment starvation in the form of reworked concretions and hardgrounds.

Concretion layers are present throughout the Phanerozoic. Classic studies of the stratigraphic distribution of concretions are from the Middle Devonian (Baird, 1976), the Jurassic (Fürsich et al., 1992; where some concretions are similar in morphology to *Thalassinoides* burrows) and the Upper Cretaceous (Kauffman, 1977). In the Miocene of Egypt concretion beds are paired with hardgrounds which formed within a few tens of centimeters above the concretion levels (Malpas et al., 2004) as seen in the Kope Formation described above (Brett et al., 2003). It has been suggested that concretions and hardgrounds form through similar processes (Fürsich et al., 1992).

*Complex Hardgrounds*: Complex hardgrounds show evidence of multiple encrustation events (Fig. 4), but typically contain only a single biofacies. Bioerosion, corrosion, and abrasion on the sea floor all work to remove the record of encrusting organisms; the longer the exposure time on the sea floor the greater the degree of degradation and number of preservation states of encrusting organisms. Therefore, hardgrounds that show a variety of preservational states amongst adjacent encrusters suggest multiple periods/episodes of encrustation, in some instances even indicating intermittent burial of the hardground surface.

Many well developed complex hardgrounds occur within the Upper Ordovician Point Pleasant member of the Lexington Limestone from central Kentucky into southwestern Ohio. These surfaces commonly display multiple overlapping layers of *Anomalocrinus* holdfasts and encrusting bryozoans developed upon a highly abraded and occasionally heavily bored and iron mineralized fine- to medium-grained skeletal limestone (Fig. 4). These surfaces are irregular,
FIGURE 4. Composite hardground from the Upper Ordovician Point Pleasant Member of the Lexington Limestone in central Kentucky. A) Oblique view of mounded bryozoan and crinoid encrusted hardground; hammer for scale. B) Cut and polished section through the encrusting layers of the hardground shown in A. This small mound record eight separate encrustation episodes (letter A-H) some of which are separated by mud burial events (number 1-4). C) This crinoid encrusted hardground represents the same horizon located 55 kilometers to the northeast. Note the overlapping and variable preservation (denoted as grades 1-4). The variable grades suggest four separate encrustation events likely corresponding to those recorded in the mounded lateral equivalent.
having reliefs of up to 3 centimeters, when developed upon firmgrounds. Encrusters show multiple grades of preservation, from highly abraded to pristine in the uppermost layer (Fig. 4c). Color alteration is also a common feature of encrusters on these surfaces; for example, the most pristine crinoid holdfasts are pink whereas more heavily degraded encrusters are blackened. Consecutive layers of encrusting bryozoans and crinoids interspersed with thin mud drapes form mounds with relief up to 8 centimeters. In some extraordinary cases, the crests of the mounds were colonized by large edrioasteroids prior to rapid mud burial that ended hardground formation. In places, cemented layers were undercut by erosion leading to the formation of complex hardgrounds along edges or bases of the cemented layer. Crinoid and bryozoan holdfasts, borings and even edrioasteroids have been observed on such undercut, cavernous hardgrounds. Encrustation of such cavities, on an otherwise flat sea floor, is also described from the Jurassic (Palmer and Fürsich, 1974; Fürsich et al., 1992). It is not uncommon to find complex hardgrounds in which only elevated portions of the surface are heavily encrusted suggesting a barrier to colonization of lower area, likely sediment cover (sensu Palmer and Palmer, 1977; Brett and Liddell, 1978). Raised areas on Upper Ordovician hardgrounds in the upper Mississippi valley have been interpreted as formed by buckling in response to cement growth (Palmer, 1978), but alternatively may record deformation by earthquakes (see discussion in section below on monomictic conglomerates).

As mentioned above, composite hardgrounds also occur as modified firmgrounds. Excellent examples come from the Curdsville and Perryville members of the Lexington Limestone (Fig. 5). In both cases, closely spaced *Thalassinoides* firmground networks occur in fine-grained skeletal limestone. These burrows are very distinct as they are lined by thin, discontinuous ferric and phosphatic crusts (sensu Fürsich et al., 1992). Encrusting bryozoans are present, though
FIGURE 5. Composite firmgrounds from the Upper Ordovician Lexington Limestone in Kentucky. A-C form the caps of the Point Pleasant, Perryville, and Curdsville members respectively; hammer for scale. A, C) Bedding plane views of composite firmgrounds. Both of these examples are associated with intraformational conglomerates. B) Cross-sectional view through composite firmground showing phosphatic and dolomitic burrow filling. All of these examples display ferric oxides encrusting the burrow walls.
typically not in high numbers. Slightly ferruginous, dolomitic and phosphatic carbonate sand in fills the burrow networks providing further evidence that they remained open for an extended period (Pope and Read, 1997). Similar modified firmground occurrences are also described from age equivalent Upper Ordovician strata in Iowa (Palmer, 1978), the Upper Jurassic of western India (Fürsich et al., 1992), and Cretaceous of southwest England (Garrison et al., 1987) and Texas (Fürsich et al., 1981). These cases studies vary in the degree to which physical modification of the firmground occurred, but all were formed in carbonates at the interface between shallow water facies (within which the firmgrounds were developed) and overlying deep water facies above.

Complex hardgrounds typically occur at sharp facies offsets and are traceable over broad areas. Within Upper Ordovician mixed carbonate-siliciclastic successions of the Cincinnati Arch complex hardgrounds typically cap packstone-grainstone beds forming sharp contacts with overlying gray shales and argillaceous limestones. In these successions complex hardgrounds form major marker horizons and are traceable over hundreds to thousands of square kilometers (McLaughlin and Brett, in press). Levorson and Gerk (1972) reported a similar distribution of Upper Ordovician hardgrounds in Iowa. Over a broad area the character of a complex hardground does change. For example, Malpas et al. (2004) were able to use continuous exposures of Miocene strata in Egypt to trace hardground surfaces down ramp recording little change in encruster composition but a large change in density of encrustation.

**Reworked and Hiatus concretions:** Both reworked and hiatus concretions occur in downramp, mud-rich sections where storm processes exhume concretions (or concretionized skeletal elements; Fig. 3b), which are later reworked. They are the counterpart of complex hardgrounds for siliciclastic-dominated settings and the latter show borings, encrustation and more rarely
mineralized coating; both indicate removal of a layer of surficial sediment and thus a change from sediment starvation to submarine erosion.

Reworked concretions occur at various levels within the Upper Ordovician Kope Formation from central Kentucky to southwestern Ohio (Wilson, 1985; Algeo and Brett, 2001). As noted above, this shale dominated (~60-80%) interval contains multiple concretion horizons, but most are not reworked. Spectacular reworked concretions and concretionized skeletal grains occur at multiple levels within the basal unit (Fulton sub-member) of the Kope. Here concretions are often found reworked into skeletal grainstone beds (Fig. 3b). Similar to the complex hardgrounds described above, these concretions are encrusted both by sheet-like bryozoans and Anomalocrinus holdfasts that show varying states of preservation. In many cases, both sides of the concretion are encrusted, suggesting currents strong enough to flip these flat discs.

Similar occurrences of reworked concretions are described from the Devonian of New York (Brett, 1974) and the Jurassic of India (Fürsich, 1992). At a few localities reworked concretions are only surrounded by shale and thus mark subtle shale-on-shale discontinuities. Such contacts are also described from the Devonian of New York (Baird, 1981; Mayer et al., 1994), the L. Jurassic of England (Hallam, 1969; Hesselbo and Palmer, 1992) and the Cretaceous of Israel (Soudry and Lewy, 1990).

Hiatus concretions form as reworked concretions are buried and addition layers of concretionary cement are added (Kennedy and Klinger, 1974). Hiatus concretions have not been reported from the Ordovician in North America. They are known, however, from the Devonian of New York (Baird, 1978) the Jurassic of India (Fürsich et al., 1992) and England (Hesselbo and Palmer, 1992) and the Cretaceous of South Africa (Kennedy and Klinger, 1974).
FIGURE 6. Ferric crust hardground from the Upper Ordovician Point Pleasant Member of the Lexington Limestone. A) Ferric oxide crust developed on a skeletal grainstone with a beveled upper surface displaying large partially exhumed burrows (large pits), hammer for scale. B) Close-up of the ferric oxide crust showing detail of internal layer and upper surface. Internal layer contains densely grouped Trypanites borings. Abraded crinoid holdfast encrusts the upper surface of the ferric crust. Dotted line marks where upper layer of crust has been broken away.
**Ferric and phosphatic crusts:** Thin (1-5 millimeters) mineral crusts rich in iron (typically oxidized to limonite) and phosphate are common in some carbonate successions. Typically these crusts are widespread and occur at facies contacts. Thin ferric and phosphatic crusts are abundant in the Upper Ordovician (Chatfieldian) of eastern North America (Sardeson, 1898; Palmer, 1978; Pope and Read, 1997; McLaughlin et al., 2004). The presence of crinoid and bryozoan holdfasts observed directly cemented to iron and phosphatic crusts (Fig. 6) in the Upper Ordovician Point Pleasant member of the Lexington Limestone in Kentucky and southwestern Ohio strongly suggests that they formed early and were exposed as seafloor. *Trypanites* borings are occasionally found penetrating the crusts, suggesting that they were indeed initially hard substrates. Similar crusts are described from the Devonian of New York (Baird, 1979), the Jurassic of England (Palmer and Wilson, 1990) and India (Fürsich et al., 1992) and the Cretaceous of Israel (Soudry and Lewy, 1990) and are interpreted to represent structures formed by bacterial, fungal, or algal mats. Similar to the Upper Ordovician examples above, Fürsich et al. (1992) describe a Jurassic hardground developed on megarippled sandy pack-grainstone that displays oysters that both encrust and are covered by a thin ferruginous crust. They report another hardground lying 30 centimeters above that displays high relief (>10 centimeters) and is covered by a stromatolitic ferruginous crust several centimeters thick. This stromatolitic crust was also observed to extend down to encrust *Thalassinoides* burrow walls. A third hardground lying 60 centimeters higher showed similar characteristics. These authors suggest relatively low energy conditions during the formation of the ferric crusts because of the limited disruption of the laminae.

Not all mineralization represents colonization by microbial mats. Palmer (1978) suggested that iron mineralization in some Upper Ordovician hardgrounds of Iowa was, in part, the early
diagenetic replacement of calcite by pyrite; although he stressed that the mineralization did not simply represent a surface crust developed through modern weathering, but rather, was part of the fabric of the limestone. This type of formation of pyrite in carbonates is generally very slow owing to the scarcity of free iron, thus even primary pyrite replacement of calcite could indicate condensation.

Ferric and phosphatic crusts have a distribution similar to complex hardgrounds. Ferric and phosphatic crusts are common within some carbonate successions and commonly become more closely spaced near facies offsets. Many can be traced for hundreds to thousands of square kilometers. Though ferric and phosphatic crusts may be associated with complex hardgrounds, more often than not, evidence of encrustation is lacking. However, in mixed carbonate-siliciclastic successions, thicker and more continuous crusts cap grainstone intervals at their contacts with overlying shalier strata and transition down-ramp into intraformational conglomerates and condensed phosphorite beds.

*Monomictic intraformational conglomerates*: Monomictic intraformational conglomerates commonly occur as medium to coarse grained grainstone-rudstone (skeletal and/or ooid) containing incorporated limestone clasts and blocks, representing some of the most complex discontinuities found in carbonate successions. These beds contain clasts of a single limestone lithology, differentiated from concretions by their skeletal/ooid composition.

Monomictic intraformational conglomerates occur at multiple levels within the Upper Ordovician strata of Kentucky and Ohio, notably within the Lexington Limestone (Fig. 7). The clasts are commonly thin (~1 to 2 cm) and relatively small, ranging in size from 1 to as much as 30 centimeters in maximum diameter. They are typically oval in plan view, but may be round, especially the smaller clasts, and display rounded edges. Some clasts are morphologically very
FIGURE 7. Monomictic intraformational conglomerates and synsedimentary faulting from the Point Pleasant Member of the Lexington Limestone in Kentucky and southwestern Ohio. A) View of the broken away underside of a thick (~40 centimeter) amalgamated skeletal limestone from near the middle of the Point Pleasant Member. The middle of this complex bed is composed of densely packed and imbricated reworked, encrusted and bored limestone clasts. B) Large angular slabs of fine-grained grainstone within the base of a thick skeletal grainstone-rudstone; hammer for scale. C) Broken skeletal grainstone slab showing partial over-thrusting as a possible analog for disruption of partially cemented layers initiating formation of intraformational conglomerate; surrounding beds are laterally continuous. Note bifurcation of laminae toward fault; hammer for scale.
similar to *Thalassinoides* burrows and may represent infilling, lithification, exhumation, and reworking of the burrow fillings. Clasts are commonly encrusted and bored, though the density of borings and degree of mineralization may vary widely from clast to clast. In many instances both the top and the bottom of the clasts are encrusted and/or bored.

Similar examples are known from a variety of ages. Other Upper Ordovician examples come from crinoid encrusted clasts of the Galena Group in the Upper Mississippi Valley (Sardeson 1908) and coral encrusted clasts in the Liberty Formation in the Cincinnati region (Foerste, 1917). Monomictic intraformational conglomerates are also common in the age equivalent Rust Formation in New York (Brett and Baird, 2002) and the slightly younger Hillier Formation in Ontario (Brett and Brookfield, 1984). Lower Silurian examples are known from the Brassfield Formation of southern Ohio (McLaughlin, unpublished data), Maplewood-Neahga Shale in New York (LoDuca and Brett, 1994) and the McKenzie Formation in Pennsylvania (Sumrall et al., in press). Mesozoic examples are described from the Jurassic of India (Fürsich et al., 1992) and the Cretaceous of Southwestern England (Garrison et al., 1987).

Somewhat more enigmatic are relatively large angular blocks that occasionally form monomictic intraformational conglomerates. Unlike the smaller, rounded clasts described above, these blocks are typically thick (2-10 centimeters), broad (10-40 centimeters), rectangular to diamond shape, show little rounding, and are coarse-grained. Further, their distribution tends to be more localized. An example from the Upper Ordovician Point Pleasant member of the Lexington Limestone in Kentucky typifies this type of occurrence (Fig. 7b). Foerste (1917) also described similar blocks from the Liberty Formation in southern Ohio as did Fürsich et al. (1992) from the Jurassic Dhossa Oolite of India. In the latter case, the blocks were interpreted to have spalled off of nearby submarine cliffs. This does not agree with the Point Pleasant and Liberty
examples which developed upon a gently dipping cratonic ramp lacking evidence of submarine cliffs. However, in the case of the Point Pleasant, widespread horizons of soft sediment deformation (>6000 square kilometers; McLaughlin and Brett, 2004) are present just a few meters below, providing evidence that seismic activity was affecting the area during that time period. The lack of similar ball and pillow-type deformation features in the Point Pleasant was interpreted to be controlled by lithofacies, rather than the abrupt cessation of seismic activity (McLaughlin and Brett, 2004). These angular blocks may be another manifestation of earthquake activity affecting the sea floor as suggested by Fürsich et al. (1992), but not resulting from collapse of submarine cliffs, but rather by fracturing of the early cemented sea floor. Similar processes were invoked by Pratt (2002) for generation of flat pebble conglomerates in the Cambrian of Montana.

Taphonomic evidence suggests that monomictic intraformational conglomerate beds form over a duration similar to complex hardgrounds. Extended periods were required to complete the processes of lithification, erosion, undercutting, rounding and resedimentation. The thickness and ellipsoidal to discoidal shapes of the clasts resemble those of lenticular, somewhat concretionary fine-grained limestones that commonly occur as interbeds between thicker limestones. The rounded outlines of these clasts may have been pre-determined in part by the pattern of concretionary cementation, possibly tracing burrow networks. The aggregation of many clasts of similar lithology and limited rounding suggests early cementation a short distance below the sediment/water interface or at the sea floor followed by exhumation and reworking, but with little subsequent transport.

Polymictic Intraformational Conglomerates: Polymictic intraformational conglomerate beds resemble monomictic conglomerates, but contain mixtures of rounded limestone clasts of
FIGURE 8. Polymictic intraformational conglomerate from the tops of the Upper Ordovician Point Pleasant and Perryville member of the Lexington Limestone in Kentucky. A) Close-up of upper surface of bed capping the Point Pleasant Member at Swallowfield, Kentucky showing at least four different types of limestone clast of variable sizes weathering out (note many are still enclosed in matrix). B) Same capping bed of the Point Pleasant as shown in A, located 50 kilometers to the north. Note again the presence of at least four different lithologies and sizes of clast in a variety of orientations and localized presence of ferric and phosphatic mineralization. (horizon is also continuous with composite firmground shown in figure 5A). C) Heavily mineralized horizon capping the Perryville Member at Monterey, Kentucky which contains at least three different types of crinoid encrusted limestone clast (this horizon is the down-ramp equivalent of the modified firmground shown in figure 5 B); hammer for scale. D) Close-up of bored clasts (near head of hammer) and bored and blackened Prasopora bryozoan (near base of hammer handle) from mineralized discontinuity surface shown in C.
variable lithologies and sizes; large limestone blocks are almost always absent from polymictic beds. Clasts range in size from pebbles to cobbles, texture from fine- to coarse-grained, lithology from limestone to mudstone, and degree of biologic modification from pristine to heavily encrusted, mineralized, and bored. Polymictic intraformational conglomerates are rare by comparison with monomictic varieties.

Polymictic intraformational conglomerates within the Lower Ordovician Kanosh Formation of Utah contain at least three generations of clasts. In this unusual case, the clasts formed during the final phase of reworking enclose two earlier generations of clasts (Palmer and Wilson, 2004) having a similarly complex history to hiatus concretions. Polymictic intraformational conglomerates within the Upper Ordovician Lexington Limestone are typically associated with modified firmgrounds and thick ferric and phosphatic crusts (Fig. 8). The surfaces they rest upon are occasionally otherwise modified by abrasion and corrosion becoming very irregular. Their distribution is largely restricted to major facies offsets where they cap relatively shallow skeletal grainstone successions. The occurrence of polymictic conglomerates from Upper Ordovician strata above the top of the Lexington Limestone in eastern North America is rare. However, a few are found in the Maquoketa group of the Upper Mississippi Valley where they occasionally co-occur with condensed phosphorite beds (Raatz and Ludvigson, 1996).

The more recent record of polymictic conglomerates shows an irregular distribution. Polymictic intraformational conglomerates occur at a few levels within the Lower Silurian of the Appalachian basin. For example, a polymictic conglomerate occurs on the top of the Merriton Limestone in Ontario, which can be traced basinward to the east where the entire package transitions into a 2 to 3 cm-thick phosphorite bed (Second Creek Phosphorite). This is noteworthy as this interval shows vertical facies changes reminiscent of the Lexington.
Limestone, where such horizons are common. Polymictic conglomerates from the Upper Silurian of the Welsh Borderland also show very similar characteristics to the Lexington Limestone examples (Cherns, 1980). Analysis of the Middle Devonian strata of the Appalachian Basin also reveals a general absence of polymictic conglomerates. Survey of the literature otherwise indicates few polymictic intraformational conglomerates are described; however, many studies only note the presence of clasts and do not specify if the lithology is variable.

*Condensed phosphorites:* Also referred to as bone beds, condensed phosphorites (sensu Föllmi, 1996) generally form thin (<10 centimeters; more commonly 1-2 centimeters) discontinuous beds or nodule layers within organic-rich shale successions, but when formed on limestones they also may be associated with phosphatic crusts. Condensed phosphorites are generally composed of phosphatized skeletal grains, phosphate steinkerns, and teeth and bone material (Fig. 9). The stable fluoroapatite composition of phosphatized bones and teeth/conodont elements ensures that these materials remain little changed on the sea floor for up to millions of years (Donovan 1991; Lucas and Prevot, 1991). These phosphatic materials may act as nucleation points for further phosphate precipitation (Föllmi, 1996). Formation of coated phosphate grains occurs just below the sediment-water interface in the suboxic zone and requires low sedimentation rates (Pufahl and Grimm, 2003). Concentrations of pyrite as burrow tubes, steinkerns, and internal molds are often a dominant component within these beds, in addition to other refractory materials such as chert and silicified skeletal material (Baird and Brett, 1986; 1991; Brett et al., 2003). The presence of concentrated authigenic pyrite suggests highly dysoxic to anoxic depositional environments and moderate to intense winnowing in order to liberate the pyrite from the enclosing muds. Taphonomic studies of condensed phosphorites reveal a high degree of time-averaging (Baird and Brett, 1986 and references therein; Brett et al., 2003).
FIGURE 9. Condensed phosphorites from the middle Paleozoic of the Cincinnati Arch and northern Appalachian Basin. A) Bedding parallel polished section through a thin (~5 centimeter) lenticular phosphorite bed containing phosphate nodules and steinkerns, placoderm bone, pyritized burrow tubes and skeletal grains, quartz pebbles and granules, coalified and calcified wood and abundant conodont elements from the Middle/Upper Devonian boundary interval in central Kentucky. This bed represents one of the more time-rich condensed phosphorites within eastern North America spanning at least three conodont zones likely representing over 1 million years of sediment starvation and episodic deposition and reworking (Brett et al., 2003). B) Bedding plane view of the Lower Silurian Second Creek phosphate bed from central New York. This 3 centimeter thick lag bed includes a variety of phosphatic, glauconitic, and hematitic grains, quartz granules and pebbles, limestones, and skeletal grains such as rugose corals and crinoid columnals. The duration of this bed is estimated on the order of a few hundred thousand years (Brett et al., 1990). C) Polished core section from west-central Ohio containing the contact of the Upper Ordovician Curdsville and Logana (overlying shale) members of the Lexington Limestone. The contact is marked by a 3 to 5 centimeter thick bed composed of phosphate granules, phosphatized and pyritized limestone grains with irregular margins, bored limestone clasts, chert pebbles, and pyrite steinkerns. Based on lateral stratal relations this bed is estimated to span approximately a few to ten thousand years (McLaughlin et al., 2004).
Many well-studied condensed phosphorites occur at “drowning surfaces” capping carbonate successions; multiple examples are known from the Upper Ordovician of eastern North America. A notable condensed phosphorite caps the Curdsville Member of the Lexington Limestone in the subsurface of Indiana and Ohio (Fig. 9; McLaughlin et al., 2004). This layer is typically 5 to 20 millimeters in thickness (Keith and Wickstrom, 1993) and contains coated phosphate grains, small phosphatized limestone clasts with sharp corroded margins, iron mineralized clasts, and small chert nodules developed on a modified firmground. It is preceded by as many as four phosphatized hardgrounds within the underlying two meters of primarily skeletal grainstone and it is sharply overlain by organic-rich, barren shales. These characteristics are very similar to the “depauperate zone”, a condensed phosphorite that caps the Upper Ordovician Galena Group at its contact with the organic-rich shales of the Maquoketa Group in the Upper Mississippi Valley; however, it is biostratigraphically much younger than the Curdsville. Biostratigraphy further suggests that tens of meters of section are missing (eroded/never deposited) beneath the depauperate zone unconformity increases from northern Iowa toward the Ozark Dome in eastern Missouri (Brown, 1966; Kolata and Graese, 1983; Black, 1985).

Condensed phosphorites are also described from multiple levels within younger Paleozoic strata of North America. Lower Silurian and Middle Devonian condensed phosphorites in the Appalachian basin are associated with thick successions of shale representing deposition near the centers of foreland basins (Baird and Brett, 1986; Brett et al., 1998). Similarly, Pennsylvanian and Permian age condensed black shales, bordering on phosphorites are common components of mid-continent cyclothems in North America representing extended sediment starvation (Heckel, 2002; Algeo et al., 2004). Alternatively, condensed phosphorites within Upper Devonian strata of the Cincinnati Arch appear to have been deposited at relatively shallow depths within a highly
restricted basin (Conkin, 1986; Schieber and Riciputi, 2004; Brett et al., 2003). Phosphorites were common on the stable Russian Platform during the Late-Jurassic to Late Cretaceous in the Volga-Ural and Trans-European Seaways (Ilyin, 1998) as well as the Tethyan margin (Funk et al., 1993). Many authors contend that phosphorites form below productive surface waters as an explanation for the source of phosphate. Macquaker et al. (1996) suggest that apatite authigenesis predates sulphate reduction, implying precipitation of phosphate just above the oxic-suboxic/sulphidic interface.

In many foreland basin settings during transitional times between greenhouse and ice-house periods, condensed phosphorites deposited near the basin center grade laterally toward the siliciclastic source area into ironstones. A classic example is found in Lower Silurian strata of the northern Appalachian Basin where the Second Creek Phosphorite (1-5 cm-thick) grades laterally into the Westmoreland Ironstone (~40 cm-thick; Brett et al., 1998). Typically, phosphorites are relatively thin compared to laterally equivalent ironstones. The observed lateral equivalence of these two deposits runs counter to the sequence stratigraphic model developed by Maquaker et al. (1996), which states that phosphorites and ironstones form under similar conditions differing only in the availability of free iron.
Ironstones: Ironstones are beds of hematite, goethite, and/or siderite cemented sediments that may contain abundant hematite and phosphate coated grains and clasts as well as skeletal grains and sand to pebble size quartz grains (Fig. 10). Primary mineralization within ironstones is early diagenetic, forming within a few centimeters of the sediment/water interface (Taylor et al., 2002). Petrographic data suggest that chamosite and bertherine iron-rich clays are precursors of hematite or goethite that commonly make up the present composition of many ironstones (Cotter and Link, 1993). Bertherine and siderite pre-date pyrite in ironstones and thus must have formed
FIGURE 10. Ironstones from the Lower Silurian of New York. A) Bedding plane image of the Keefer fossil ironstone. Skeletal grains within this bed are coated and cemented with hematite. B) Bedding plane image of Westmoreland oolitic ironstone. Westmoreland ironstone is lateral equivalent of the second creek phosphate bed shown in figure 9B.
in the suboxic zone before sulphate reduction, in this way they are similar to condensed phosphorites (Macquaker et al., 1996). Suboxic diagenesis requires low sediment accumulation rates and extensive physical reworking (Taylor et al., 2002) and ironstones commonly grade laterally into other authigenic mineral deposits such as glauconitic sands/ carbonates and phosphorites whose genesis requires similar environmental conditions. However, ironstones are largely restricted to the clastic side of foreland basins and passive margins where there is run-off from iron-rich basement rocks. Macquaker et al. (1996) suggests that ironstone formation is governed primarily by low sedimentation rates and availability of reduced iron. High levels of bioturbation are indicative of formation in relatively well-oxygenated waters (Brett et al., 1998; Taylor et al., 2002). Further, ironstones commonly exhibit cross-stratification, ripples, sharp bases, stringers of coarse quartz sand and pebbles, and sparry cements, suggestive of deposition within relatively high-energy environments (near normal wave base). They are typically widespread and range in thickness from only a few centimeters to over a meter in thickness, but are much more commonly in the ten-centimeter range and vary little in thickness across strike. Ironstones are most abundant during regional tectonic quiescence between tectonic pulses in greenhouse periods (Ordovician-Devonian and Jurassic-Paleogene; Van Houten, 1985, 1989, 1990; Young, 1992; Brett et al., 1998).

Ironstones are prominent components of Lower Silurian successions in the Appalachian basin. Widespread ironstones cap shallowing-upward sandstone successions at the contact with overlying shales, while more localized lenticular ironstones may be scattered throughout sandstone intervals. Lenticular ironstones also occur near contacts that mark the lateral interface of inner shoal margins and lagoons where skeletal limestones laterally grade into siliciclastic mudstones (LoDuca, 1988). Widespread ironstones in the Appalachian basin are highly
fossiliferous; component skeletal grains commonly show a high degree of within habitat time-averaging. Some of the thickest and most widespread ironstones show a mixing of shallow and deep-water faunas and are thus considered “ecologically time-averaged” (sensu Kidwell and Bosence, 1991). For example, the Westmoreland ironstone at the base of the Silurian Williamson Shale in New York State shows a time-averaged assemblage (Brett et al., 1998). This ironstone is extremely widespread along strike, traceable for almost 2000 kilometers along the length of the Appalachian basin.

**Sequence stratigraphic implications of discontinuity surfaces and condensed beds**

The increasing levels of complexity, lateral extent, and stratigraphic distribution of the discontinuity surfaces and condensed beds described above support a modified sequence stratigraphic model for epeiric seas (McLaughlin and Brett, 2005; Fig. 11). This model goes beyond other sequence stratigraphic models for foreland basins (Van Wagoner and Bertram, 1995; Plint and Nummedal, 2000) in describing sedimentation patterns on both the siliciclastic-dominated and the carbonate-dominated margins, including: a) Detection of early and late (lower and upper in a stratigraphic sense) portions of the transgressive systems tract on the carbonate margin, separated by a distinct discontinuity surface, termed the “maximum starvation surface” (MSS; sensu Baum and Vail, 1988)). This surface is marked by a sharp facies change which occurs across an erosion-corrosion surface. It lies well below the true maximum flooding surface or zone (MFS). The MSS is considered to reflect the maximum rate of relative sea level rise. b) Recognition of distinct surfaces and, in some locations, condensed beds (“precursor beds” of Brett, 1995), associated with abrupt sea level fall and traceable from the carbonate margin, to the basin center, and up-ramp into the distal portion of the siliciclastic margin. c) The
FIGURE 11. Schematic diagram representing the variation in motif of depositional sequences and component discontinuity surfaces and condensed beds across a foreland basin profile.
characterization of third-order depositional sequences as composites of fourth order cycles (400 ka durations), which contain systems tracts made-up of fifth-order cycles, that, on the carbonate margin, display sequence-like motifs, as opposed to parasequences. Examples from the Upper Ordovician Lexington Limestone of eastern North America, given below, illustrate the correspondence between such nested layers of cyclicity and discontinuity surfaces on the carbonate margin of the Taconic foreland basin.

Simple firmgrounds and hardgrounds within the Lexington Limestone occur sporadically and do not regularly correspond to any specific portion of sedimentary cycles. Cemented layers that never underwent seafloor exposure are also probably common, but these “incipient hardgrounds” are typically difficult to recognize. Such early-cemented carbonate layers are analogous to calcite-cemented concretions, but, whereas incipient hardgrounds form in carbonate-rich sediments, including skeletal pack- and grainstones, concretions form by interstitial cementation within siliciclastic muds in more distal portions of the carbonate margin. In both cases, early lithification probably takes place within the sediment, most likely at a small distance below the sediment-water interface (0.5-1 m), primarily within the zone of sulfate reduction, wherein the development of elevated alkalinity favors calcite/aragonite precipitation (Wilson and Palmer, 1992). Ironically, short-duration condensation may be more evident in muddy highstand sediments in these down-ramp areas than in the up-ramp carbonate-dominated sections, as concretions are readily identified in siliciclastic mudstones, whereas incipient early-cemented layers may be difficult to differentiate from later diageneric cementation in pure carbonates.

Complex hardgrounds, ferric and phosphatic crusts, and monomictic conglomerates mark maximum starvation surfaces within small-scale cycles (Fig. 12). However, rarely do all three of these features co-occur at such contacts. Small-scale cycles (partly analogous to parasequences)
FIGURE 12. Generalized diagram showing the hierarchy of cyclicity observed many mixed carbonate-siliciclastic (carbonate margin) middle Paleozoic strata in eastern North America. A) Compound depositional sequences (3rd-order; sensu Van Wagoner et al., 1990) are composed of two smaller 4th-order depositional sequences (sub-sequences of Brett et al., 1990). B) Depositional sequences contain a well defined TST and HST. FSST are typically more well developed in the upper of the two depositional sequences. C) Small-scale cycles (5th-order) form components of systems tracts (similar to parasequences on the clastic margin) and thus their motif is influenced by the systems tract in which they occur. Note the position and degree of development of discontinuity surfaces within this hierarchy.
are nested within transgressive, highstand, and falling stage systems tracts of 4th-order depositional sequences (Fig. 12). Membership of a small-scale cycle to one of the various systems tracts manifests itself not only in the lithologic composition of the cycle, but also in the development of discontinuities. Small-scale cycles within third- and fourth-order transgressive systems tracts (TSTs) display the most well developed and regularly occurring hardgrounds. Hardgrounds are more sporadically developed within the highstand and falling stage systems tracts, but when present, show a similar complexity to those in the TST. The hardgrounds of the HST and FSST also tend to show the patchiest lateral distribution as compared to those of the TST which are typically traceable for a few thousand square kilometers. Fürsich et al. (1992) suggested that hardgrounds cap regressive deposits and represent the ensuing transgressive phase that was largely condensed into a single horizon. Although this may be true in some cases, we argue that many of the best developed hardgrounds occur within transgressive deposits thus forming at large- and small-scale maximum starvation surfaces.

Polymictic intraformational conglomerates commonly co-occur with ferric and phosphatic crusts, and complex hardgrounds at maximum starvation surfaces of 4th-order depositional sequences (Fig. 12, 13). Complex hardgrounds and firmgrounds and ferric and phosphatic crusts typically cluster near these major facies offsets. For example, Malpas et al. (2004) recognized a clustering of Thalassinoides firmgrounds below flooding surfaces within a Miocene synrift succession. Clustering of firmgrounds and hardgrounds suggests that the enclosing carbonates are becoming similarly condensed. Tracing of polymictic conglomerates down-ramp reveals that they merge with underlying hardgrounds as they grade into condensed phosphorites (Fig. 11). Thus, condensed phosphorites generally represent a combination of the maximum starvation surface and much, if not all, of the TST toward the foreland basin center. Under the right
FIGURE 13. Outcrop images from the middle Paleozoic of eastern North America showing maximum flooding surfaces of composite depositional sequences. A) Middle Devonian Onondaga Limestone/Marcellus Shale contact in central New York. This sharp change from micritic limestones into organic-rich shale, similar to figure 9C, is marked by a lenticular condensed phosphorite (B). C) Lower Silurian Irondequoit Limestone/Rochester Shale contact in western New York. This sharp facies offset is marked by a glauconitic skeletal limestone bed (D) overlain by a 2 meter-thick bioherm that is draped by the lower Rochester Shale. E) Upper Ordovician Sulphur Well/Stamping Ground member contact within the Lexington Limestone of central Kentucky. The sparry skeletal grainstone-rudstones of the Sulphur Well contain multiple closely spaced ferric crust hardgrounds that become clustered near the contact with the Stamping Ground. F) In places small (10 to 20 centimeter-thick) bryozoan-algal bioherms are formed on the uppermost hardground of the Sulphur Well and are draped by shales and argillaceous limestones of the Stamping Ground.
tectonic conditions condensed phosphorites in the basin center laterally transition toward the clastic margin of the foreland basin into ironstones. Thick, widespread, and complex ironstones cap regressive sandstone successions forming combined thin TSTs and maximum starvation surfaces. Alternatively, TST’s in on the siliciclastic margin may be represented by heavily winnowed and/or glauconite cemented thin sandstone lags. The forced regression surface, which marks the contact between the HST and FSST, may be associated with time-averaged thin skeletal lag beds in up-ramp areas on the carbonate side of the basin and phosphate nodules beds (precursor beds of Brett and Baird, 1996) toward the basin center (Fig. 11). However, this discontinuity surface disappears toward the clastic side of the basin and is commonly only marked by a sharp facies dislocation. Both the maximum starvation and forced regression surfaces are interpreted to form during periods of rapid sea level change. The maximum starvation surface appears to represent a much more extended period of sediment starvation as the discontinuity surfaces and condensed beds that mark it are much more complex than those of the forced regression surface. The hierarchy of discontinuity surfaces and condensed beds presented here is in good agreement with that proposed by Garrison et al. (1987) who recognized increasing complexity of cemented, mineralized, encrusted, and reworked beds within a Cretaceous passive margin succession.

Summary

The sections above describe stratal patterns, discontinuity development and biotic changes as responses to shut-down of siliciclastic influx and carbonate production, commonly in response to rapid sea level rise. Sequence stratigraphic models of carbonate platforms call for cessation of carbonate production in response to: 1) light stress (drop below the euphotic zone), 2)
eutrophication in response to upwelling, 3) oxygen stress, and 4) cool-water inhibiting carbonate precipitation (Schlager, 1998; Bosence, 2004). Many of these processes are applicable to modern reefs and continental shelf platforms, where hermatypic corals rely on zooxanthellae, which are extremely sensitive to changes in light, nutrients, water temperature, and oxygenation of water. This type of symbiosis is a relatively new invention; yet “drowned” carbonate platforms are relatively common in the geologic record.

Development of cemented layers and their exposure as hard substrates or early diagenetic mineralization requires very low net sedimentation rates and/or enhanced seafloor erosion. In terms of sequence stratigraphy, this is the critical connection in the present context: hardgrounds and other discontinuity surfaces mark prolonged reduction of sediment accumulation and thus should be especially characteristic of transgressions. However, the Phanerozoic record of mixed carbonate-siliciclastic successions does not demonstrate a one-to-one relationship between well developed cyclicity and well developed discontinuities. Many cyclic stratigraphic successions do not contain recognizable discontinuity surfaces, especially not those of shorter duration. However, concretions are more commonly reported than hardgrounds.

At present it is not fully clear why hardgrounds are much more common at certain times, such as the Late Ordovician and Middle Jurassic, than others, but it does appear to be the case that the expression of discontinuity surfaces of relatively short duration is a function of oceanic conditions during certain periods in Earth’s history and not simply related to cyclicity. Based on published accounts, hardgrounds are most common, widespread, and highly modified during the Ordovician and Jurassic, the peak periods of calcite seas, and least common during times of aragonite oceans such as Carboniferous and Neogene (Sandberg, 1983; Wilkinson, 1986). Calcite ocean intervals are characterized by lowered seawater Mg/Ca ratios, which favor very
early dissolution of aragonite and precipitation of interstitial calcite cements (Stanley and Hardie, 1998; Wilson and Palmer, 2004). However, even during times of calcite seas, other factors may be at play.

In many modern subtropical passive margin carbonate platforms the rate of carbonate sediment productions is known to balance or exceed normal rates of glacio-eustatic sea level rise. Hence, reduction of carbonate production, although commonly linked with transgression, is less pronounced and involves other processes that may curtail carbonate production. It has been noted that many limestones formed in Paleozoic and perhaps early Mesozoic epicontinental seas are not composed of coralgal type sediments (Taylor and Allison, 1998). These limestones are largely composed of varying quantities of pelmatozoan ossicles, brachiopods, bryozoans, and fragments of trilobite with little or no algal or coral component. Thus, they appear more closely related to modern “cool water” carbonates whose production rates are much lower. Whether this character indicates truly cool water production or simply different suites of common carbonate platform organisms through time, it is likely that production rates were indeed lower than in healthy, warm water carbonate platforms at present, and thus more readily exceeded by rapid sea level rise. Moreover, the common occurrence of organic-rich muds above hardgrounds suggests that incursions of dysoxic, turbid and/or nutrient rich water could have been related to temporary curtailment of skeletal production. Hence, a propensity for carbonate factory shut-down is probably more pronounced in mixed carbonate-siliciclastic systems (common in well studied areas of both Upper Ordovician and Middle Jurassic) than in pure carbonate buildups. More extreme starvation was associated with seafloor erosion and produced reworked hardgrounds and concretions. Storm-generated currents were probably ubiquitous in these environments and
FIGURE 14. Generalized hierarchical diagram of discontinuity surfaces and condensed beds across a foreland basin profile. The duration assigned to individual discontinuities and condensed beds is estimated based increasing complexity and sedimentology. Time, plotted on the Y-axis, is on a logarithmic scale. Note the abundance of indicators of sediment starvation preserved on the carbonate side of the foreland basin.
served to remove sediments unless they accumulated relatively rapidly. Hence, erosion may be a consequence of sediment starvation in areas of relatively low carbonate production.

The tendency toward hardground and discontinuity surface production may also be a matter of water depth and environmental energy. Marshall and Aston (1980), in a comparative study of three Middle Jurassic hardgrounds from eastern England, suggest that the intensity of early lithification was related to the depositional environment, with high energy environments contributing maximum circulation of sea water resulting in the most highly lithified hardgrounds.

**Conclusions**

Recognition of a hierarchy of increasingly complex discontinuity surfaces (Fig. 14) contributes to an evolving view of cratonic sequence stratigraphy, which gives insight into relative change in epeiric seas throughout the Phanerozoic. The basic units of discontinuity surfaces are the simple firmground and hardground. If sedimentation does not resume these surfaces are repeatedly modified through the competing forces of bioerosion and encrustation and develop into composite hardgrounds, typical of small-scale (5th-order) cycles. Composite hardgrounds are often associated with monomictic conglomerates (possibly a signature of earthquake activity) and ferric and phosphatic crusts, which also indicate extensive periods of low sedimentation and sea floor modification. Further modification during extended periods of extremely low and episodic sedimentation result in formation of polymictic intraformational conglomerates characteristic of the carbonate margin of foreland basins and epeiric carbonate platforms. Polymictic intraformational conglomerates transition down-ramp into phosphorite beds where carbonate production is typically very low. During the waning phases of orogeny, phosphorites in foreland basins may grade up-ramp into ironstones, otherwise grading into
glauconitic or fossiliferous clean sandstones. Whereas polymictic intraformational conglomerates form at maximum starvation surfaces capping the limestone-rich TST, it is typically not possible to differentiate a separate maximum starvation surface from the TST in highly condensed phosphorites and ironstones (or their clean sandstone counterpart). Hardgrounds, reworked concretions, phosphatic beds and other indicators of condensation are most typical of 3rd order late TST successions, but may also occur at discontinuity surfaces associated with lesser flooding surfaces and forced regression surfaces.

Establishing the relationship between discontinuity surfaces and sequence stratigraphy reinforces the importance of outcrop based studies. Similarities between facies successions and major discontinuity surfaces on passive margins, where depositional sequences are best documented, and cratonic successions reinforces the argument for the applicability of sequence stratigraphy to the craton. However, because passive margin successions display an apparent lack of widespread short-duration discontinuity surfaces (i.e. composite hardgrounds) and greater sediment thicknesses between major discontinuity surfaces, recognizing the differences between these two tectonically distinct provinces is equally important in establishing the geologic history and completeness of that history for a given area.

Acknowledgements

The authors are indebted to our many colleagues for fruitful discussions that greatly helped refine the ideas presented here. In particular, we would like to thank Steve Donovan and Tim Palmer for organizing the special session on hard substrates through time at the 2004 Geological Society of America annual meeting, at which gave impetus for compiling of these ideas. Great thanks are also expressed to Brian Pratt and Chris Holmden for compiling this volume and for
their great patience. Acknowledgement is also made to the donors of the American Chemical Society Petroleum Research Fund for partial support of this research.
References


Brett, C.E., Turner, A.H., McLaughlin, P.I., Over, J. and Storrs, G., 2003, Middle-Upper Devonian (Givetian-Famennian) bone/conodont beds from central Kentucky: reworking and
event condensation in the distal Acadian foreland basin: Proceedings of the 15th annual
Senckenberg Conference Corrier Forschungsinstitut, Senckenberg. 125-139.


Keith, B. and Wickstrom, L., 1993, Trenton Limestone: the karst that wasn’t there, or was it?: In Wilson, J.L., Yurewicz, D.A., Fritz, R.D. Eds., Paleokarst Related Hydrocarbon Reservoirs, SEPM Core Workshop 18, 167-179.


Macquaker, J.H.S, Taylor, K.G., Young, T.P., Curtis, C.D., 1996, Sedimentological and geochemical controls on ooidal ironstone and ‘bone-bed’ formation and some comments on


Palmer, T.J., Wilson, M.A., 1990, Growth of ferruginous oncoliths in the Bajocian (Middle Jurassic) of Europe. Terra Nova 2, 142-147.


Pope, M.C., Read, J. F., 1997, High-Resolution surface and subsurface sequence stratigraphy of the Middle to Late Ordovician (late Mohawkian-Cincinnatian) foreland basin rocks, Kentucky and Virginia. AAPG Bulletin 81, 1866-1893.


Raiswell, R., 1976, The microbiological formation of carbonate concretions in the Upper Lias of NE England: Chemical Geology 18, 227-244.


CHAPTER 6

A Unified Sequence Stratigraphic Model for Foreland Basins: Advances from Analysis of Mixed Carbonate-Siliciclastic Successions

Patrick I. McLaughlin* and Carlton E. Brett

Department of Geology, University of Cincinnati, Cincinnati, OH 45221

(anticipated submission to Journal of Geology; June, 2006)
ABSTRACT

A unified model of sequence stratigraphy for foreland basins is proposed based on integration of a standard siliciclastic-based model with recent advances in the understanding of mixed carbonate-siliciclastic successions. The ideas governing the latter model are largely derived from comparative analyses of Paleozoic synorogenic deposits in eastern North America (associated with the Taconic, Acadian, and Alleghenian orogenies). Critical to development of this model was: 1) the recognition of repeating stratigraphic patterns common to all foreland basin systems studied, 2) detailed analysis of mixed carbonate-siliciclastic strata, which yield the most sensitive record of environmental change within foreland basin systems and provide a critical connection between the well studied pure siliciclastic and pure carbonate end members; and 3) the assumption that load induced subsidence rates and sedimentation rates on the siliciclastic-dominated margin of the foreland basin are much greater than on the carbonate-dominated margin. Key aspects of this model are: A) differential development of transgressive systems tracts as relatively thin lags on the siliciclastic margin and as thicker and relatively clean skeletal limestones on the carbonate margin; B) identification of a sharp surface of maximum sediment starvation (MSS) particularly well developed on the carbonate margin as a mineralized composite hardground surface, associated with maximal rates of sea level rise and drowning of the carbonate factory, as opposed to a more subtle maximum flooding surface (MFS) in overlying muddy deposits and in some cases separated by up to several meters from the MSS; and C) identification of a sharp forced regression surface (FRS), and in some cases, an overlying condensed shell rich bed (“precursor bed”) at the base of D) distinctly shallowing upward falling stage (regressive) deposits, typified by progradation of coarser sediments on the siliciclastic margin and heavily reworked skeletal sands on the carbonate margin.
INTRODUCTION

A great deal of new sequence stratigraphic research and theory has been presented in the nearly two decades since the release of the seminal publication on sequence stratigraphy in 1988, SEPM Special Volume 42 “Sea Level Changes: An Integrated Approach”. Much of the progress in siliciclastic sequence stratigraphy is summarized in Catuneanu (2002). However, no such review is available for advancements in mixed carbonate-siliciclastic successions. The following manuscript not only provides a synthesis of current models of siliciclastic sequence stratigraphy, but also discusses the interrelatedness of carbonate and siliciclastic systems in the context of foreland basin settings.

Mixed carbonate-siliciclastic successions are the most sensitive indicators of changes in sea level, sediment supply, climate, and tectonics within the foreland basin system (Saylor, 2003), as they are poised at the interface between the allochthonous prograding clastic wedge and the autochthonous carbonate factory and are built upon the relatively stable foundation of the craton. By contrast, pure siliciclastic successions on the collisional side of the foreland basin are almost completely allochthonous, subject to autocyclic processes in near shore areas because of very high sedimentation rates (i.e. delta lobe switching), and rest upon the wobbly footings of the continental margin (e.g. see Sloss (1988) for subsidence histories of North American continental margins, basins and arches). Therefore, it is ironic that today siliciclastic sequence stratigraphy represents the primary mode of foreland basin analysis and that concepts of mixed carbonate-siliciclastic sequence stratigraphy are poorly understood and hence not widely applied (see discussion of misapplication of sequence stratigraphic models in cratonic successions in Sloss (1996) and McLaughlin and Brett (in review)).
In the following a modified model of foreland basin sequence stratigraphy is presented that proposes a common response to relative sea level change across the entire foreland basin system, including both its siliciclastic and carbonate end members. As with any model, the ideas presented here are only generalities and it is expected that depositional history will vary somewhat both temporally and spatially within and between foreland basins. That said, the depositional patterns of foreland basins described below are well documented throughout the middle Paleozoic of eastern North America, representing deposition during three consecutive orogenic phases. Further, preliminary analysis of several other geographic regions and time periods, coupled with review of the literature suggests these patterns are typical of underfilled foreland basin systems.

**Historical Perspective of Foreland Basin Sequence Stratigraphy**

Over the past century stratigraphic paradigms have had a cyclicity all their own. In North America the century began with Ulrich’s (1911) ideas of layer-cake stratigraphy, principally derived from analysis of Upper Ordovician mixed carbonate-siliciclastic successions of eastern North America. Largely in agreement with the basic tenets of layer-cake stratigraphy were Udden’s (1912) ideas on cyclic sedimentation in Pennsylvanian age heterolithic strata of Illinois. The term cyclothem was introduced by Wanless and Weller (1932) to describe these cyclic packages of marine and non-marine strata in the Pennsylvanian deposits of the Midcontinent and more fully explored ideas about their genesis. A great deal of the cyclothem literature spans the 1930’s to the mid-1960s, a substantial portion of this literature focused on unraveling the origin of cyclothems (Weller, 1964). While the idea of stratigraphic (mega)sequences was getting a boost from the studies of Sloss (1969; though Newberry (1874) recognized these “circles of
sedimentation nearly a century before) during the middle of the century, high-resolution stratigraphic studies were suffering a major set back. Recognition of the complexity of many modern coastal settings (especially carbonate environments of the Bahamas) led many researchers to embrace a facies mosaic model of the stratigraphic record. This view was in line with the denigration of E.O. Ulrich for his “misguided” ideas on the continuity of strata proposed early in the twentieth century. With the advent of seismic stratigraphy focus shifted away from foreland basins to passive margins, primarily siliciclastic systems (see papers in Payton et al., 1977). Out of seismic stratigraphy sequence stratigraphy was created (Posementier et al., 1988). Sedimentary studies soon returned to the craton as efforts for an outcrop analog to siliciclastic passive margin successions was sought (Van Wagoner et al., 1990). This initially resulted in a great resurgence of interest in cratonic stratigraphy (e.g. Witzke et al., 1996). During the last decade advances in siliciclastic sequence stratigraphy have continued to be forthcoming (see discussion below). However, work on mixed carbonate-siliciclastic successions has been largely stagnant. Pennsylvanian and Permian stratigraphers continue to work on relating cyclothem stratigraphy to sequence stratigraphy, but many view this is an isolated problem, one that does not clearly relate to cyclicity or the lack thereof during other portions of the Phanerozoic. In part this disconnect comes from both a poor record of glaciation during other portions of the Phanerozoic (excluding the Cenozoic) and an overemphasis on the “ideal cyclothem” (*sensu* Weller and Wanless, 1939) as a base line for recognizing cyclicity in mixed carbonate-siliciclastic strata (see discussion in Wilkinson et al., 2003).
Evolving view of glacio-eustasy

Over the last few decades the evidence for substantial glaciations in the geologic past has improved beyond the traditional end Ordovician, Pennsylvanian, Permian, and late Cenozoic view. For example glacial deposits are now suspected to reach back some ten to fifteen million years from the end Ordovician into the early Late Ordovician and extend forward into several parts of the Silurian and end Devonian (Frakes et al., 1992). In addition, several studies that demonstrate globally synchronous facies changes during portions of the Cretaceous (Gale et al. 2002; Sandulli and Raspini, 2004) challenge the notion that only massive continental glaciation like that of the Pleistocene can produce eustatic fluctuations (also see discussion in Miller et al., 2005).

The early sequence stratigraphic models considered eustasy as the primary driving component in sequence development at any scale (i.e. Vail et al., 1977; Haq et al., 1987; Posementier et al., 1988). This sparked major debates in the professional community about the ability to reliably identify the relative influence of eustasy and tectonics (e.g. Christie-Blick et al., 1990; Miall, 1992; Vail et al., 1991), which resulted in an SEPM special volume (Dennison and Ettensohn, 1991) on the topic. The overwhelming consensus of authors in this volume was that Milankovitch-scale (4\textsuperscript{th}, 5\textsuperscript{th}, 6\textsuperscript{th}-order) and slightly larger (3\textsuperscript{rd}-order) cycles/sequences were governed primarily by eustasy and that localized tectonic effects were identifiable within sub-regional studies. It was also recognized that region and global tectonic processes occur over a much longer time period than typical outcrop-scale sequences (also see Miller et al., 2005 for discussion) and that they too have a distinctive cyclic nature, though discernibly different than that of eustatic sequences. Some authors have quickly dismissed any attempt to disentangle the relative effects of eustasy and tectonics (e.g. Van Wagoner et al., 1990) and began referring to
“relative sea level”, a combination of eustatic fluctuation and subsidence/uplift. This trend continues to be dominant in sequence stratigraphic literature today.

Existing Models

Siliciclastic Sequence Stratigraphy

Siliciclastic successions have received more intense sequence stratigraphic study than any other major lithologic group; as a result siliciclastic sequence stratigraphic models are the most highly resolved. As noted above, the study of siliciclastic sequence stratigraphy got its start primarily in seismic studies of the Exxon research group. The basic concepts of siliciclastic sequence stratigraphy are presented in manuscripts and entire volumes on the subject by Posementier et al. (1988), Van Wagoner et al. (1990), Van Wagoner and Bertram (1995), and Posementier and Allen (1993). All but Posementier et al. (1988) focused specifically on siliciclastic sequence stratigraphy in foreland basin successions. These studies provide a readily accessible outcrop analog in the Cretaceous deposits of the western United States that has been visited by thousands of geologists and students. Further important contributions to siliciclastic sequence stratigraphy of foreland basins from the Cretaceous deal directly with the record of relative sea level fall (e.g. Plint, 1988, Hadley and Elliot, 1993; Fitzsimmons and Johnson, 2000; Plint and Nummedal, 2000; Posementier and Morris, 2000) and have aided in resolving inconsistencies associated with earlier models. These modifications sparked some debates; however, they are largely resolved (see excellent review of Catuneanu (2002)). This important contribution to the sequence stratigraphic literature places an emphasis on the continuity of various siliciclastic stratigraphic models (such as genetic stratigraphy, transgressive-regressive cycles, as well as sequence stratigraphy). Thus, the discussion of siliciclastic sequence
stratigraphy in foreland basins presented below largely follows the summary of ideas presented by Catuneanu (2002).

**Carbonate Sequence Stratigraphy**

A considerable amount of literature has also been directed toward carbonate sequence stratigraphy. The modern synthesis of carbonate sequence stratigraphy is largely the result of seismic studies of passive margin successions (e.g. Kendall and Schlager, 1981; Sarg, 1988; Kerans and Tinker, 1997; Schlager, 1999) and outcrop and core studies of Cretaceous deposits of the Tethyan passive margin now exposed in the Alps and surrounding uplifts (e.g. Hunt and Tucker, 1993; Föllmi et al., 1994). Additionally, a large proportion of the sequence stratigraphic literature is dedicated to open ocean carbonate systems such as atolls, guyots, and seamounts (e.g. Schlager, 1998). A review of this literature emphasizes to the cratonic worker that the realm of the open ocean represents a very different carbonate depositional setting than that with which he/she is acquainted. Subsidence and carbonate production rates in many cases are an order of magnitude higher on the passive margin than they are in epicontinental seas. High subsidence rates in particular, when combined with eustatic rise often result in drowning of the carbonate factory for millions of years (Schlager, 1999; Föllmi et al., 1994). Such long term drowning unconformities are virtually unknown in cratonic successions. Rather, subaerial exposure for millions or even tens of millions of years is much more commonly the case in the cratonic carbonate record (Sloss, 1963; Foos, 1996). Because of these extreme differences a thorough treatment of carbonate sequence stratigraphy on the craton/foreland basins has not been attempted in the same manner as that of siliciclastic sequence stratigraphy. However, Tucker et al. (1993) provide a sequence stratigraphic model for homoclinal carbonate ramps that is the
nearest approximation. Their concepts are integrated in part in the following discussion of mixed carbonate-siliciclastic successions. The discussion of mixed systems in connection to siliciclastic sequence stratigraphy will hopefully provide the intermediate step required to more fully understand carbonate sequence stratigraphy on the craton.

**Mixed Carbonate-Siliciclastic Sequence Stratigraphy**

Analysis of mixed carbonate-siliciclastic strata has received the least amount of consideration in the recent sequence stratigraphic literature of foreland basins. Further, much of the literature on the subject involves direct application of models derived from analysis of siliciclastic- or carbonate-dominated successions in passive margin conditions (e.g. Holmes and Christie-Blick, 1993; Holland, 1993; Pope and Read, 1997; Choi and Simo, 1998; Holland and Patzkowsky, 1998). These studies signify that there is a general lack of recognition that mixed carbonate-siliciclastic successions are indeed a mixture of these two end members and having been deposited in foreland basins have very different tectonic histories than passive margin successions. Notable exceptions include work on the late Proterozoic of Namibia (Saylor, 2003), Jurassic and lower Cretaceous of India (Fürsich et al., 1991; Fürsich et al., 1992; Fürsich and Pandey, 2003), and the lower Cretaceous of Alberta, Canada (Banerjee and Kidwell, 1991). Of course the literature of Upper Mississippian-Permian cyclothems has for nearly a century considered the geologic phenomena that generate mixed carbonate-siliciclastic sequences in foreland basins (though typically using a different nomenclature). Unfortunately, the principles derived from study of cyclothems are little applied to mixed systems outside the late Paleozoic. A notable exception can be found in the more recent stratigraphic literature of late Tertiary and Quaternary strata of New Zealand (e.g. Naish and Kamp, 1997). Both of these time periods, the
late Paleozoic and late Cenozoic, experienced waxing and waning of large continental ice sheets that created eustatic fluctuations on the order of several 10s to over 100 meters resulting in very compartmentalized facies packages that follow a general stacking pattern (the cyclothem). Failure to recognize similar facies successions and a lack of evidence for similarly extensive glaciations has resulted in a failure of workers in mixed systems outside of these time periods to apply these concepts. Further, an important contribution to the literature of mixed carbonate-siliciclastic sequence stratigraphy by Baum and Vail (1988) considering the Paleogene strata of the Gulf Coastal Plain has been largely overlooked by many researchers. The following discussion incorporates these sequence stratigraphic concepts with those derived from analysis of Upper Ordovician (Holland et al., 2001; McLaughlin et al., 2004, Brett et al., 2003; Brett et al., 2004; McLaughlin and Brett, in review), Lower Silurian (Brett et al., 1990; Goodman and Brett, 1994; Brett et al., 1998), and Middle Devonian (Ver Straeten and Brett, 2000, Brett and Baird, 1996) mixed carbonate-siliciclastic foreland basin strata in eastern North America. Read (1998) considered the middle part of the Paleozoic transitional between icehouse and greenhouse states. Thus, the development of systems tracts and the degree of facies offset at sequence surfaces is expected to vary slightly from those of typical cyclothems, but aspects of the facies succession should be similar.

THE MOTIF OF SYSTEMS TRACTS ACROSS A FORELAND BASIN PROFILE

Systems tracts vary predictably across foreland basin profiles. Of course, each foreland basins has unique characteristics which form in response to geologic age; paleogeography; type, angle, and size of collision; etc. However, they share many gross similarities of sedimentation pattern (Fig. 1) that allow for genesis of a general model of foreland basin sequence stratigraphy.
FIGURE 1. Foreland basin profile for the Upper Cretaceous Sevier Basin showing general thickness patterns and interpreted subsidence histories (Modified from Kaufman, 1985).
FIGURE 2. Schematic cross section of Lower Silurian strata deposited across New York State during the end Taconic and Salinic orogenies. Note that a complete foreland basin profile is available through several consecutive depositional sequences. Modified from Brett et al., 1998.
FIGURE 3. The general motif of depositional sequences across the foreland basin profile.
For the sake of discussion the foreland basin is divided here into the siliciclastic margin, basin center, and carbonate margin (Fig. 2, 3). These divisions represent geographic as well as general lithologic domains. The siliciclastic margin is located on the collisional side of the foreland basin and thus is subject to high levels of tectonic activity and is in close proximity to high relief areas and volcanics. The siliciclastic margin is idealized as a gently inclined ramp, although with localized areas of moderate to high relief (e.g. delta front slopes), having high rates of terrigenous sedimentation (primarily siliciclastic sand-size grains and larger), and high subsidence rates. The basin center is a relatively deep, low relief area, marked by moderate subsidence rates, low to moderate terrigenous sedimentation rates (primarily clay and silt), low oxygen levels, and is mostly below the depth of the photic zone (hence carbonate production by benthic invertebrates is minimal). On the craton side of the foreland basin the carbonate margin is typified by a gently inclined ramp, low subsidence and siliciclastic sedimentation rates (as adjacent areas of exposure are generally low lying and may be mantled by previously deposited carbonates), and primarily autochthonous carbonate sedimentation. The following section provides discussion of variations in these general sedimentation patterns resulting from changes in relative sea level. The discussion of each systems tract proceeds from the better studied siliciclastic margin, across the basin center, and onto the more tectonically stable carbonate margin. An emphasis is placed on the interrelatedness of patterns of sedimentation in the various parts of the foreland basin.

Lowstand Systems Tract

On the siliciclastic margin the lowstand systems tract (LST) is “bounded by the subaerial unconformity and its marine correlative conformity at the base, and by the maximum regressive
FIGURE 4. General profiles for the siliciclastic and carbonate margins of forelands during initial sea level rise (Lowstand Systems Tract).
surface at the top. It forms during the early stage of base level rise when the rate of rise is
outpaced by the sedimentation rate (case of normal regression)” (Catuneanu, 2002; Fig. 4). The
maximum regressive surface (Helland-Hansen and Martinsen, 1996) is defined by the T-R curve,
separating prograding from retrograding strata, and is generally conformable (Catuneanu, 2002).
The LST is typically composed of a slightly progradational to aggradational succession of
argillaceous sandstones and mudstones on the siliciclastic margin (Fig. 5D).

In the basin center the lowstand is marked by relatively low levels of siliciclastic input and a
fining upward succession. During the lowstand the basin center is near its shallowest while
siliciclastic input is waning. As a result the LST in may be marked by heavily bioturbated silts
and muds or a brief return to mud only deposition. Depending on latitude, the LST near the
craton side of the basin center may also contain distal calcareous tempestites. In this setting
slowing sedimentation rates may be marked by closely spaced concretion horizons.

On the carbonate margin of the foreland basin lowstand may be marked by aggradation of
shoals during initial sea level rise (sensu Tucker et al., 1993). This build-up of shoals results in
formation of relatively deep and extensive lagoons in near shore areas. For example multiple
horizons in the Upper Ordovician Lexington Limestone in Kentucky and lateral equivalent
Nashville Group in Tennessee contain a stromatoporoid carbonate wackestone facies that is rich
in ostracods and commonly contains interbeds of organic rich shale enclosing the carbonaceous
remains of dasycladacean green algae (Fig. 5A). Similar deposits are reported from the Lower
Silurian Gasport Dolostone in New York State (LoDuca and Brett, 1997) and its lateral
equivalent the McKenzie Formation in Pennsylvania (Fig. 5C; Brett et al., 1990). The fauna
suggests a shallow restricted environment. However, these deposits commonly show
aggradational stacking of several beds totaling more than two meters in thickness (up to 25 m in
FIGURE 5. General foreland basin facies of the lowstand systems tract. A) Lagoon and shoal facies of the LST on the carbonate margin of the Taconic Foreland Basin in central Kentucky (Devils Hollow Member of the Lexington Limestone). B) Mid-ramp facies of the LST on the carbonate margin of the Taconic Foreland Basin in northern Kentucky (Point Pleasant member of the Lexington Limestone and overlying Fulton and Brent sub-members of the Kope Formation). C) Close-up of lagoonal LST facies from the Lower Silurian Salinic Foreland Basin of central Pennsylvania (McKenzie Formation). The rhythmically bedded marls alternate with organic-rich shales containing an abundance of ostracodes and carbonized dasycladacean green algae. D) LST facies of the clastic margin of the Upper Mississippian Alleghenian Foreland Basin of southeastern Indiana (Sample Formation).
the case of the McKenzie). This facies is always sharply overlain by cross-bedded calcarenites (shoal deposits; transgressive ravinement surface) occasionally containing reworked stromatoporoids as a basal lag (e.g. Strodes Creek Member of the Lexington Limestone; Taha McLaughlin, 2006). In the mid-ramp this period of initial sea level rise, accompanied by a cut-back in siliciclastic input on the opposing margin, is marked by formation of clean skeletal grainstone-rudstones that alternate with thin intervals of argillaceous limestones and thin shales. This pattern is well expressed in the uppermost member of the Lexington Limestone (Point Pleasant), which contains at its base two 5th-order cycles with this motif (Fig. 5B; McLaughlin and Brett, in review). Unlike their counterparts on the siliciclastic margin, faunal gradient analysis of these slightly argillaceous deposits suggests deepening upward (McLaughlin and Brett, in review). As noted above, the relatively low subsidence and sedimentation rates on the carbonate margin largely preclude normal regression (i.e. progradation) resulting in vertical facies changes that more closely record relative sea level fluctuations (relative to the basin center or siliciclastic margin). Thus, an interval containing an aggradational to deepening upward faunal assemblage, but with high clay content relative to the overlying strata defines the LST on the carbonate margin.

**Transgressive Ravinement Surface**

The transgressive ravinement surface is a diachronous erosion surfaces formed by the landward migration of the high-energy shoreface zone. The transgressive ravinement surface is only found updip of the most basinward position of the upper shoreface during the preceeding regressive phase (falling stage). The transgressive ravinement surface on the siliciclastic side of foreland basins has been described as a sharp erosion surface (Fig. 5D). In many cases it
removes a few meters to greater than ten meters of strata (Catuneanu, 2002). Such high levels of incision often result in cannibalization of the sequence boundary. The transgressive ravinement surface on the carbonate margin of the foreland basin is also highly erosive, especially where it overlies clay-rich facies. The transgressive ravinement surface may be less erosive on the carbonate margin than on the siliciclastic margin because of the binding effects of early cementation. Still, transgressive ravinement surfaces with local relief in excess of two meters are known. For example, the transgressive ravinement surface within the upper Devils Hollow Member of the Lexington Limestone forms broad channels near Frankfort, Kentucky and has been observed not only to remove the underlying lagoonal facies but also to cut down across the sequence boundary into the underlying falling stage deposits (Fig. 5A; McLaughlin and Brett, in prep).

**Transgressive Systems Tract (TST)**

The transgressive systems tract is bounded below by the maximum regressive surface and above by the maximum flooding surface in siliciclastic systems (Catuneanu, 2002). On the siliciclastic margin of the foreland basin the transgressive systems tract forms when the rate of increase in accommodation outpaces sedimentation rate. Consequently the transgressive systems tract is typified by abundant and thick estuary deposits (Fig. 6). This nearshore trapping of sediments results in retrogradational stacking and a generally fining upward succession in shallow portions of the ramp. In many cases the foreland basin may become starved of terrigenous input altogether during formation of the TST (Fig. 6). Sediment starvation in many cases leads to authigenic mineralization and formation of glauconitic, hematitic, and/or phosphatic transgressive lags in up-ramp areas that directly overly the transgressive ravinement
FIGURE 6. General profiles for the siliciclastic and carbonate margins of forelands during rapid sea level rise (Transgressive Systems Tract).
surface (e.g. Brett et al., 1998; Fig. 7E). This drop in terrigenous input and cleaning of the water column may also result in formation of widely traceable thin shell concentrations (Van Wagoner et al., 1990; Banerjee and Kidwell, 1991; Brett and Baird, 1996).

TSTs down-ramp become increasingly condensed and are typically thinnest in the basin center. These deposits are typically marked by thin, closely spaced, distal storm beds and shelly stringers associated with concretions in the outer ramp (Brett and Baird, 1996). Depending on age and water chemistry this portion of the ramp may also contain limestones formed from calcareous ooze (e.g. Elder et al., 1994). In the more anoxic settings of the basin center the TST may only be marked only by a thin lag of pyrite and other refractory material such as phosphatized skeletal grains, bone, and chert (Brett et al., 1990; Brett and Baird, 1996).

The TST thickens on the carbonate side of the basin and is divisible into two distinct phases (early and late). The ETST is bound by argillaceous deposits of the LST below and the maximum starvation surface above. The ETST is commonly dominated by widespread, often hardground-rich, highly amalgamated shell beds with little to no clay content (Fig. 7A; McLaughlin and Brett, in review). These shell beds are thickest in mid ramp areas, thinning up dip into heavily winnowed (typically phosphatic) shoals that eventually interfinger with shallow lagoonal carbonate wackestones and fenestral micrites. Hardgrounds are not only abundant stratigraphically, but also show broad areal distributions from shoal facies to distal storm beds. Patch reefs and bioherms are common components of ETST deposited in tropical settings (Fig. 7B; e.g. Lower Silurian Irondequoit Limestone and Gasport Dolostone, Fig. 7D; Goodman and Brett, 1994: Edgecliff Member of the Lower Devonian Onondaga Limestone; Ver Straeten and Brett, 2000). The early phase of the transgressive systems tract is separated from the later phase (LTST) by the maximum starvation surface (MSS). The late TST (LTST; early highstand of
FIGURE 7. General foreland basin facies of the early and late systems tracts. A) Close-up of ETST-LTST contact (maximum starvation surface) in inner-ramp facies of the carbonate margin of the Taconic Foreland Basin in central Kentucky (Sulphur Well and Stamping Ground members of the Lexington Limestone). Note that the ETST contains multiple hardgrounds near its top (marked by red dashed lines) and a small iron mineralized bryozoan-algal bioherm (adjacent to hammer head). B) Poorly bedded biohermal facies of the ETST (LST absent locally) from the Lower Silurian Salinic Foreland basin in western New York State (Niagara Falls Member of the Goat Island Dolostone). C) Outer-ramp, tightly stacked, calcisiltites and lesser grainstones of the ETST (Edgecliff Member of the Onondaga Formation) sharply overlain by the LTST interbedded organic-rich shales, calcisiltites and thin packstones (Union Spring Member of the Marcellus Shale). D) Mid-ramp facies of the ETST in relation to overlying systems tracts deposited on the carbonate-margin of the Salinic Foreland Basin in western New York (Irondeqoit through Gasport formations; represents two 4th-order depositional sequences). The ETST contains skeletal grainstone-rudstone that becomes increasingly enriched in glauconite near the top and is capped by a thrombolitic bioherm. 
Brett et al., 1990) records a retrogradational to aggraded biofacies trend and like the LST contains higher clay content than the ETST (Fig. 7A, C; e.g. McLaughlin and Brett, in review). This higher clay content of the LTST signals renewed progradation on the siliciclastic margin of the foreland basin. The LTST is typically thin and may contain hardgrounds, although typically not in the abundance found in the ETST. The upper boundary of the LTST is the maximum flooding surface (MFS).

**Maximum Starvation and Maximum Flooding Surfaces**

The maximum starvation surface (MSS) is believed to form during the most rapid rate of relative sea level rise (Baum and Vail, 1988) when sedimentation largely ceases across the entire foreland basin (Fig. 6). The maximum starvation surface has typically not been recognized (or designated as such) on the siliciclastic margin of the foreland basin (but, see Goodman and Brett, 1994). Several conditions hinder the recognition of the MSS on the siliciclastic margin: 1) authigenic mineralization formed during the MSS may not be different from that developed during the ETST, 2) sediment starvation may preclude the development of an ETST, and 3) strata of the LTST may not contain biostratigraphic resolution or other sedimentary indicators to accurately pinpoint the MFS (thus reference to a maximum flooding zone, likely equivalent to the LTST). However, in some cases transgressive lags of the ETST are capped by authigenic crusts (MSS) and overlain by shelly mudstones of the LTST. The maximum flooding surface (MFS) on the siliciclastic margin of the foreland basin marks the level between retrograding and aggrading to prograding strata, separating the LTST from the overlying HST (Catuneanu, 2002).

In the foreland basin center the maximum starvation surface is commonly manifest as a phosphate mineralized surface that is not easily differentiated from the phosphatic lag deposit of
the ETST. Further, because of very low sedimentation rates in the basin center during this period the MSS and MFS may be merged.

On the carbonate margin the difference between the MSS and the MFS is most pronounced. Skeletal grainstone-rudstones of the ETST are often overlain by more argillaceous facies of the LTST across a sharp contact (the MSS). Within many successions such contacts are made additionally apparent by the presence of hardgrounds, which commonly occur as a cluster near the top of the ETST. The uppermost of these discontinuity surfaces is typically the most complex, suggesting a long period of nondeposition and modification (Baum and Vail, 1988; Fürsich et al., 1992; McLaughlin and Brett, in review). These “drowning unconformities” are most likely the result of temperature and/or oxygen stress induced by the rapid rise in relative sea level. The MSS is overlain by the argillaceous LTST, which is capped by the deepest water facies of the sequence; this contact marks the MFS. In a down-ramp direction, as noted, the MFS may eventually merge with the MSS as the intervening strata become increasingly condensed. Maximum starvation marks the period of most rapid sea level rise and consequently this is when the formation of bioherms on the carbonate margin is most prevalent (Tucker et al., 1993). Although bioherms may show roots as low as the preceding FSST, they often expand slightly and interfingering with ETST skeletal grainstone-rudstones and show their greatest vertical accretion (lacking interfingering with surrounding deposits) on top of the ETST, coincident with formation of the MSS (Brett et al., 1990; Ver Straeten and Brett, 2000). Occasionally, these bioherms develop an uppermost hardground after the reef has become dormant. Tall bioherms, such as pinnacle reefs, are also subject to destruction through current action and large blocks are commonly found within surrounding highstand deposits (Watts and Riding, 2000).
FIGURE 8. General profiles for the siliciclastic and carbonate margins of forelands during sea level standstill and initial fall (Highstand Systems Tract).
**Highstand Systems Tract (HST)**

On the siliciclastic side of the foreland basin the HST is marked by a coarsening upward succession indicative of progradation. It rests on the MFS and is bound above by the forced regression surface (Catuneanu, 2002). Sedimentation rates may be comparatively low during the initial HST as relative sea level is still rising and approaching still stand; though, it is now being outpaced by sedimentation rate, thus progradation is occurring (Fig. 8). The HST on the siliciclastic-dominated mid-ramp is typically composed of shales and siltstones with an increasing percentage of sand upward (Fig. 9E). Many models would suggest that deposition of the HST ends at maximum relative sea level (Catuneanu, 2002). However, evidence from the carbonate margin of the foreland basin suggests that deposition of the HST continues during initial relative sea level fall (see discussion below).

In the basin center the HST is marked by thick deposits of dark, typically organic-rich shales with an upward increasing concentration of thin siltstones (Fig. 9D; typically storm beds or distal turbidites; e.g. Brett and Baird, 1996); in this settings the HST is slightly progradational. Sediment bypass delivers clays and silts down ramp as turbidity currents to the basin center. Oxygen content in the basin center is typically at its lowest during the initial part of the HST; coupled with the increasingly high amounts of clay influx, this interval is commonly characterized by high levels of organic carbon burial (Bohacs, 1998). K-bentonites are also very common in these deposits as a result of generally low energy and the possibility of rapid burial by distal turbiditic mud flows (Ver Straeten, 2004).

The HST on the carbonate side of the foreland basin is typically marked by argillaceous carbonates interbedded with thin shales and a shallowing upward faunal assemblage (Fig. 9A-C). As mentioned earlier, because of the low sedimentation rates normal regression is unusual on the
FIGURE 9. General foreland basin facies of the highstand systems tract. A-C) Inner- to outer-ramp HST facies from the carbonate margin of the Taconic Foreland Basin in central Kentucky (Stamping Ground, Bromley, and Brannon members of the Lexington Limestone, respectively). A) Inner ramp amalgamated and locally cross-bedded argillaceous calcarenites with occasional shale beds. B) Mid-ramp nodular wackestones and packstones with irregular interbedded shales and shale partings. Contrast this argillaceous facies against the overlying white, sparry grainstones of the subsequent LST and ETST (Peaks Mill member). C) Outer-ramp rhythmically interbedded shales and marls with lesser calcisiltites of the HST. Note the sharp erosional contact with the overlying argillaceous calcarenites of the FSST. D) Basin center thick organic-rich black shales of the HST. Note the presence of a K-bentonite near the base of the succession (Indian Castle Shale, Taconic Foreland Basin, New York). E) Thick interval of silty olive-gray shales and lesser siltstones of the HST overlying ironstones of the LTST and ETST from the Lower Silurian Salinic Foreland Basin of central Pennsylvania (Williamson Shale and Westmoreland Ironstone equivalents).
carbonate margin of foreland basins. Yet, argillaceous limestones and interbedded shales that make up the HST show distinctive stratigraphic trends in faunal composition, taphonomic state, and sedimentary structures indicative of shallowing (e.g. Taha McLaughlin, 2006). These patterns suggest that the HST is deposited during still stand and initial fall of relative sea level. Obrution deposits (rapidly smothered benthic assemblages) are most common in the HST. The burial layer of many obrution deposits in this setting is clay-rich. The widespread nature, taphonomic characteristics of the preserved organisms, and distinctive silt-free clay fabrics of these deposits suggest rapid rain-out of transported clays from hyperpycnal flows (Obrien et al., 2001). This pulsed influx of smothering clays may help to explain the relatively low carbonate sedimentation rates during the HST on the carbonate margin. Hardgrounds are typically lacking from the HST, but occasionally become more common in up-ramp facies where winnowing reduces net sedimentation rates.

**Forced Regression Surface (FRS)**

The FRS forms during the most rapid rate of sea level fall (Fitzsimmons and Johnson, 2000). In the mid-ramp on the siliciclastic side of the foreland basin the FRS is typically coincident with the regressive ravinement surface, however further offshore it may be preceded by closely stacked storm beds. These beds contain “precursor” gutter cast horizons that have been interpreted to mark this rapid fall preceding regressive ravinement (Plint and Nummedal, 2000).

Toward the basin center the forced regression surface may be marked only by a thin condensed bed. This “precursor bed” (*sensu* Brett and Baird, 1996) is typically manifest as a reworked concretion or phosphatic pebble bed overlying a subtle erosion surface at a major facies dislocation. Rapid lowering of base level (falling inflection point) results in a short period
of disequilibrium where increased current energy leads to bypass and erosion of clays and fine silts. A lag of phosphatic material (such as bone) acts as the nucleation site for additional phosphate to accrete to form nodules (sensu Follmi et al., 1994). Sedimentation resumes when coarser clastic sediments eventually migrate out into the basin center.

On the carbonate margin of the foreland basin the FRS is typically highly erosive and ranges in morphology from channeled to planar (McLaughlin and Brett, in prep.). In up-ramp areas the FRS is typically planar and commonly removes up to a several meters of the underlying HST. In some cases a thin, condensed skeletal grainstone-rudstone bed, similar to facies of the ETST forms a “precursor bed” on the carbonate margin.

**Falling Stage Systems Tract (FSST)**

On the siliciclastic margin of the foreland basin the falling stage systems tract records a period of base level fall and is marked by deep incision of fluvial systems and subsequent build-out of large delta systems (Fig. 10). In this setting the FSST is manifest as a thick series of downstepping, shingled shoreface deposits dominated by argillaceous sandstones (Plint and Nummedal, 2000; Fig. 11D, E). These represent some of the thickest deposits of the entire depositional sequence (if they are not totally removed under the sequence boundary).

In the basin center thick packages of siliciclastics deposited during the FSST are typically dominated by combinations of silt and clay that are often highly bioturbated with a notable drop in organic carbon content (Fig. 11C, D). In steep sided foreland basins the FSST in the basin center may be marked by thick turbidite successions (e.g. proximal portions of the Taconic peripheral foreland basin axis).
FIGURE 10. General profiles for the siliciclastic and carbonate margins of forelands during rapid sea level fall (Falling Stage Systems Tract).
On the carbonate ramp the FSST is dominated by a series of thin, down stepping, shingled shoreface deposits, typically an order of magnitude thinner than their siliciclastic counterparts (McLaughlin and Brett, in prep.). The FSST in the most up-ramp areas of the carbonate margin may be composed of a thin interval of lagoonal to supratidal deposits dominated by green mud cracked shale containing a restricted fauna. Down-ramp argillaceous calcarenites form shoal deposits that down-step in a seaward direction and record long-distance regression (sensu Posementier and Morris, 2000; Fig. 11A). More distal deposits are formed of stacked storm beds of argillaceous calcarenite representing material sourced from the shoals (Fig. 11B). The FSST on the carbonate margin is also typically a time of low diversity faunas. However, in distal parts of the ramp, away form the influence of the prograding shoals, the FSST may have a very high diversity, shallow water fauna containing small patch reefs and biostroms (e.g. Brett et al., 1990 “Willowvale” facies).

**Sequence Boundary**

The sequence boundary is marked by subaerial erosion in proximal areas, submarine erosion in slightly more distal areas, and the correlative conformity in down-ramp areas which marks a switch from rapid progradation to aggradation in shallow marine settings and is coincident with the end of forced regression (Posementier et al., 1988; Catuneanu, 2002). Type 2 sequence boundaries form where subsidence rate outpaces absolute sea level fall (Posementier et al., 1988). Type 2 sequence boundaries occasionally occur on the siliciclastic margin of foreland basins, but are rare on the carbonate margin because of the relatively low rates of subsidence.

On the siliciclastic side of the foreland basin the sequence boundary may be marked by a transition from braided to meandering stream deposits (Van Wagoner et al., 1990). Near shore a
FIGURE 11. General foreland basin facies of the falling stage systems tract. A) Carbonate margin FSST containing cross-bedded argillaceous dolo-calcarenite, sharply overlain by LST herringbone cross-bedded dolo-grainstone-rudstone (DeCew Dolostone overlain by the Gothic Hill Member of the Gasport Dolostone, Salinic Foreland Basin, western New York). B) Distal outer-ramp of the Upper Ordovician Taconic Foreland Basin carbonate margin showing calcisiltite ribbon limestones (FSST; upper Dolgeville Formation) overlain by a compact fine-grained grainstone (ETST; Stueben Formation equivalent) that is capped by a phosphate pebble bed (maximum starvation surface; thruway unconformity of Baird and Brett, 2002). C) Basin center FSST (lower Centerfield Formation) of the Acadian Foreland Basin in central New York State. Note the sharp contact between the heavily bioturbated silty mudstones of the FSST and the organic-rich black shales of the HST (Butternut Shale). D) Sharp contact between the interbedded silty mudstones and siltstones of the HST (Blue Hill Shale Member of the Carlisle Shale) and the heavily bioturbated muddy sandstones of the FSST (Codell Sandstone Member of the Carlisle Shale) in the basin center to outer-ramp of the siliciclastic margin of the Upper Cretaceous Sevier Basin in central Colorado. E) Alternating light gray and maroon bands in the HST (Henley Beds), sharply overlain by semitabular, thick bedded sandstones and thin interbedded shales of the FSST (Farmers Member) on the outer-ramp of the siliciclastic margin of the Acadian Foreland Basin in eastern Kentucky.
switch from red, heavily oxidized paleosols to green, reduced paleosols marks a rising water table, which form part of the LST. These deposits are often partially to fully removed by the effects of shoreface erosion, which form the transgressive ravinement surface. On deeper part of the siliciclastic ramp the correlative conformity is typically only marked by the shift from progradational to aggradational stacking of parasequences (Catuneanu, 2002).

In the basin center the sequence boundary (correlative conformity) is recorded by a reduction in siliciclastic grain size, typically marked by a pronounced switch to mud deposition (LST; Brett and Baird, 1996).

On the carbonate ramp the sequence boundary may be marked by a karst surface and a switch from deposition of supratidal desiccation-cracked mudstones of the FSST to deeper lagoonal deposits of the LST. In subtidal portions of the carbonate margin the sequence boundary is marked by a sharp change from argillaceous calcarenites of the FSST to clean skeletal grainstones and rudstones of the basal LST. In some successions the sequence boundary is marked by a marine hardground suggesting a period of non-deposition during temporary reorganization of the carbonate factory in response to increasing accommodation (McLaughlin and Brett, in review).

**DISCUSSION**

**Continuity and disparity of facies change across the foreland basin**

The transgressive-regressive curve is slightly out of phase on the two sides of the foreland basin (Fig. 12). While during initial sea level rise normal regression occurs on the siliciclastic margin (Catuneanu, 2002), aggradation and retrogradation occur on the carbonate margin (orange shaded area-1 in Fig. 12). Once sedimentation is shut down on the siliciclastic margin
FIGURE 12. Summary diagram of relationship between contemporaneous depositional processes on the siliciclastic and carbonate margins of a foreland basin.
high subsidence rates result in more rapid deepening than on the carbonate margin (yellow shaded area-2 in Fig. 12). Because of the high sedimentation rate the siliciclastic side of the foreland basin reaches maximum water depth soon after sea level rise begins to slow (just following the rising inflection point, purple shaded area-3 in Fig. 12) when sedimentation rate exceeds the rate of relative sea level rise and normal regression resumes (green shaded area-4 in Fig. 12). Because of the high sedimentation rates the rate of water depth shallowing on the siliciclastic margin will greatly outpace that of the carbonate margin.

**Combined effects of high subsidence and highly fluctuating sedimentation rate on the magnitude of facies change on the siliciclastic margin**

If the creation of near shore accommodation is such that subtidal sedimentation rates on the siliciclastic margin of the foreland basin drop to zero, then the magnitude of both shallowing and deepening on the siliciclastic margin will be greater than that on the carbonate margin. High subsidence rates on the siliciclastic margin of the foreland basin have an additive effect on eustatic rise and a negative effect on eustatic fall. At the same time eustatic rise has a negative effect on subtidal sedimentation and eustatic fall has a positive effect on sedimentation rate. Thus, water depth increase (deepening) during the TST will exceed eustatic rise by the amount of subsidence that occurs during that time interval. Because sedimentation rate is increasing throughout much of the HST and FSST it is more difficult to predict the magnitude of shallowing in excess of eustatic fall. However, as long as sedimentation rate exceeds the subsidence rate shallowing will exceed eustatic fall. Given that both low subsidence and low and relatively uniform sedimentation occur on the carbonate margin, facies on that side of the foreland basin
more closely represent eustasy and thus the magnitude of facies variation would be less on that side of the foreland basin than on the siliciclastic margin.

**Relation of mixed carbonate-siliciclastic sequences between icehouse and transitional climate states**

Sequence stratigraphic patterns of Permo-Carboniferous Midcontinent cyclothems (*sensu* Heckel, 1994) deposited during icehouse climate states bear many similarities to middle Paleozoic sequences deposited during transitional climate states (form here on referred to as transitional cyclothems; Fig. 13). Similarly, sequences developed during transitional climate states yield insight into the construction of systems tracts that may not be available in icehouse cyclothems because of the greater level of condensation resulting from greater amplitude, and therefore higher rates, of relative sea level fluctuation. Icehouse cyclothems frequently contain regionally persistent paleosol horizons, making identification of the sequence boundary very straightforward. These paleosols typically are situated between strongly regressive calcarenites and oolites and overlying transgressive coals, marine shales and condensed limestones. The primary difference in the transitional cyclothems is that exposure surfaces are not as widely distributed, yet where present, separate regressive from transgressive limestones. Because of the slower rate of sea level rise carbonate deposition is more continuous, as recorded in the LST and TST of the transitional period; whereas in the icehouse cyclothem the transgressive limestone is highly condensed. For the same reason maximum starvation and maximum flooding surfaces are easier to differentiate in the transitional cyclothems. Surprisingly, the falling stage argillaceous calcarenites, oolites, and paleosols of the icehouse cyclothem commonly occupy a greater proportion of the sequence (greater than one-half) than they do in the transitional sequence (less
FIGURE 13. Comparison of the motif of transitional cyclothsems and icehouse cyclothsems. (Icehouse cyclothem after Heckel, 1994)
than one-third). The greater asymmetry of the icehouse cyclothem may relate to continental-scale glacial processes (i.e. rapid melting reported from Pleistocene glacial deposits), whereas processes that may have a more symmetry pattern such as storage/release of freshwater on land, thermal expansion/contraction of seawater, and alpine glaciation, which are suggested as mechanisms for generating cyclicity during transitional climate states (Miller et al., 2005). Further, small-scale cyclicity is typically present within the systems tracts of transitional climate states, whereas they are generally lacking in icehouse cyclothems (Felton and Heckel, 1996).

CONCLUSIONS

In conclusion, the siliciclastic and carbonate margins of the foreland basin are intricately intertwined with expansion and growth of one side coming at the expense of the other as a result of fluctuations in relative sea level. This process of expansion and contraction of the opposing sides of the foreland basin creates a vertical succession of genetically related facies (systems tracts), the variation in argillaceous content on the carbonate margin providing clues as to the timing of progradation and retrogradation on the siliciclastic margin of the foreland basin. The characteristics of these 5 systems include: 1) The LST, which forms during the initial period of sea level rise and is composed of a mixed interval of clean limestones and thin argillaceous strata of on the carbonate margin, condensed muds in the basin center, and a progradational to aggradational heterolithic succession of silty mudstones, siltstones and lesser sandstones on the siliciclastic margin. 2) The ETST, which forms as the rate of sea level rise increases. During this period the carbonate margin expands to its maximum. Formation of widespread, clean, skeletal sand sheets and bioherms forming occurs on the carbonate margin while the basin center and much of the siliciclastic margin experience sediment starvation and occasionally authigenic
mineralization or colonization by benthic invertebrates as a result of sequestering of siliciclastics in near shore traps of the siliciclastic margin. This systems tract is capped by the MSS, which forms during the most rapid rate of sea level rise. It is during this period that all sedimentation may cease across the foreland basin, although bioherms may try to keep up initially. 3) As sea level rise begins to slow the siliciclastic margin again begins to expand as near shore traps are over filled and progradation resumes marking the HST on the siliciclastic margin. The basin center receives a large amount of mud at this time as well. On the carbonate margin this slowing of sea level rise permits recolonization of the ramp by shelly faunas mixed with a fresh influx of clays, forming the LTST. Maximum relative sea level is marked by the maximum flooding surface on the carbonate margin of the foreland basin. 4) HST on the carbonate margin begins as sea level still stand and initial relative sea level fall results in aggradation to progradation. During this period of sea level fall fine grained siliciclastics are transported across the basin center to the carbonate margin increasing turbidity, which ultimately results in reduced carbonate production rates. The most rapid seaward migration of the shoreline on either side of the foreland basin occurs during the most rapid rate of sea level fall and formation of the forced regression surface. 5) Progradation is at its highest on both sides of the foreland basin during formation of the FSST. During this period the siliciclastic margin is most fully expanded marked by progradation of a coarse clastic wedge. The regional nature of these shallowing trends coupled with local coarse wedges indicates that the progradation is a response to, rather than the cause of shallowing.
ACKNOWLEDGEMENTS

PIM would like to recognize the University of Cincinnati for support of doctoral studies through a University Distinguished Graduate Fellowship as well as the Department of Geology for continuing financial support. Acknowledgement is also made to the donors of the American Chemical Society Petroleum Research Fund for partial support of this research.
REFERENCES


HOLMES, A.E. AND CHRISTIE-BLICK, N., 1993, Origin of sedimentary cycles in mixed carbonate-siliciclastic systems, an example from the Canning Basin, Western Australia, in


**Biography**

Patrick Ian McLaughlin was born on March 24, 1973 in the till plains of central Illinois to parents of mixed Irish and German decent. Pat spent most of his childhood in Peoria; later he would find out that this is the type area of Pennsylvanian cyclothems, the literature of which played a key role in his doctoral research. His early interests were in carpentry and landscaping and remain as favored hobbies today. Pat’s only exposure to rocks as a child was the blurred images of road cuts he would glimpse through the car window during his families occasional vacations to St. Louis, Missouri. Earth history became his primary interest after taking introductory geology courses from Cheryl Emmerson at Illinois Central College in the spring and early summer of 1996. He went on to be influenced by the enthusiasm of professors David Malone and Jed Day at Illinois State University. Field camp in the Black Hills of South Dakota and Big Horns of eastern Wyoming during the summer of 1999 gave Pat the first opportunity to utilize his newly attained geologic knowledge in intensive field-based projects. This was the beginning of his love of fieldwork. Pat joined the Department of Geology at the University of Cincinnati shortly thereafter as a masters student. His research project, under Dr. Carlton Brett focused on widespread soft-sediment deformation horizons within the Upper Ordovician Lexington Limestone in Kentucky and southwestern Ohio. This research was aided by access to a series of new, very large road cuts north of Frankfort, Kentucky, which also played a primary role in his doctoral research. Pat’s current interests are in global correlation, integrating the sequence stratigraphic model he has presented here with isotope geochemistry and absolute dating. He will begin his first fulltime teaching position as a sabbatical replacement at Bucknell University in the fall of 2006.