Late Mississippian (Chesterian) Through Early Pennsylvanian (Atokan) Strata, Michigan Basin, U.S.A.

Shannon M. Towne

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LATE MISSISSIPPIAN (CHESTERIAN) THROUGH EARLY PENNSYLVANIAN (ATOKAN) STRATA, MICHIGAN BASIN, U.S.A.

by

Shannon M. Towne

A thesis submitted to the Graduate College in partial fulfillment of the requirements for the degree of Master of Science
Department of Geosciences
Western Michigan University
April 2013

Thesis Committee:

David A. Barnes, Ph.D., Chair
William B. Harrison III, Ph.D.
Michelle Kominz, Ph.D.
David B. Westjohn, Ph.D.
Over 2,000 linear feet of core material was analyzed to evaluate the stratigraphy and basin evolution of Carboniferous strata in the Michigan basin. Rock units were evaluated on the basis of lithofacies type, contact relationships, and existing regional geologic interpretations. The recovery of three distinct pollen and spore assemblages from core confirms the timing of deposition during the Late Mississippian Chesterian and Early-Middle Pennsylvanian Morrowan and Atokan regional stages within the Michigan basin.

The deposition of a marine carbonate succession with significant interstratified quartz sandstone occurred during the Chesterian regional stage. The Bayport interval (Bayport Limestone) is composed of seven distinct depositional lithofacies reflecting shoal-water to peritidal environments overlain at erosional contacts by tidally bedded (estuarine) quartz sandstone.

In the southern Michigan basin the Mississippian-Pennsylvanian unconformity is marked by either a coarse-grained sandstone unit or a mature paleosol overlying a karsted limestone regolith. Climate sensitive sediments indicate a transition from a predominantly arid to humid climate across the Mississippian-Pennsylvanian systemic boundary in the Michigan basin.
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Many acknowledgments are due. I would first like to thank my advisor, Dr. David Barnes, who guided me just enough to allow for independent thought and an injection of much needed ideas during the editing process. Dr. William Harrison III, for without whom none of this would be possible and Linda Harrison, who was amicable enough to aide (or allow) in the unpacking, reboxing, and re-inventorying of core and geologic material housed at the MGRRE facility. Dr. David Westjohn for providing shrewd insight into Carboniferous Michigan basin stratigraphy. Dr. Robert Ravn for providing much needed chronostratigraphic interpretations based on pollen and spore samples and excellent photomicrographs. Dr. Cortland Eble, Kentucky Geological Survey, for similarly aiding in the miospore biostratigraphy. Dr. Steve Greb, Kentucky Geological Survey, for taking time to look at cores of the geologic interval and discussions which aided in my sedimentologic and stratigraphic interpretations. Kyle Deatrick for helping in the illustration of the Bayport depositional model. The Department of Energy for providing me with a job and means to attend graduate school. My parents Jon Towne and Bobbi Martindale who provided much needed moral support. I want to especially acknowledge Carolynn DeLand-Phillips, who despite growing bored with geologic jargon and debate, entertained my muttering about the Carboniferous Michigan basin.

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CHAPTER I

INTRODUCTION

Late Mississippian strata have been extensively studied in the Illinois and Appalachian basins where outcrops are abundant and economic incentives are significant. However, in the Michigan basin Carboniferous strata occupy the shallow subsurface and they are covered by a layer of Pleistocene aged glacial deposits, preventing detailed sedimentologic and stratigraphic analysis and sampling. Examination of core material at the Michigan Geological Repository for Education (MGRRE) allows for the detailed study of Carboniferous strata in Michigan. This study applies principles of facies analysis, biostratigraphic correlation, and modern stratigraphic methods and concepts to evaluate and interpret Late Mississippian and Early Pennsylvanian Michigan basin strata.

The primary focus was placed on four continuous cores, which include a number of formal lithologic/formational units from base to top including: The Mississippian Marshall Sandstone, Michigan Formation, Bayport Limestone, and the Pennsylvanian Saginaw/Grand River Formations (Fig. 1). These lithologic units have been subjects in geologic reports and scientific literature for well over 100 years (e.g. Winchell, 1861; Lane, 1909; Newcombe, 1933; Kelly, 1936; Vugrinovich, 1984; Westjohn and Weaver, 1998), but little direct study was conducted regarding the stratigraphic relationships within and between these units, nor has detailed analysis of sedimentary facies (lithology, texture, structures, fossil content, etc.) and depositional environments been performed.

Despite previous assessments of the Late Mississippian and Early
Pennsylvanian system, the lack of detailed sedimentologic and biostratigraphic control has resulted in an incomplete understanding regarding the history of the Carboniferous in the Michigan basin, especially with regards to the Mississippian-Pennsylvanian systemic boundary. The apparent lack of Carboniferous, specifically Chesterian (upper-most Mississippian), fossils from strata in the Michigan basin is interpreted as a stratigraphic hiatus during the Chesterian Stage (Cohee, 1979; Harrell et al., 1991). Newcombe (1933) attributes the lack of Chesterian strata to tectonic uplift and erosion during the Late Mississippian, while Harrell et al. (1991) attributes the depositional hiatus to diminishing rates of tectonic subsidence.

Regional stratigraphic relationships in Late Mississippian and Pennsylvanian strata throughout the North American Midcontinent are interpreted as recording a combination of tectonic uplift (Ettensohn and Peppers, 1979; Ettensohn, 1981, 2004, 2008), high amplitude glacio-eustatic sea-level fluctuations (Smith and Read, 2001; Al-Tawil et al., 2003; Al-Tawil and Read, 2003), and climate change (Cecil, 1990; Frakes, 1992; Rankey, 1997; Smith and Read, 2000). The Late Mississippian (Chesterian Stage) directly underlies the basal Absaroka mega-sequence boundary (e.g. Sloss, 1963), which divides the Mississippian and Pennsylvanian periods in many cratonic areas globally. In the Appalachian and Illinois basins Mississippian aged marine deposits are erosionally truncated and overlain by Pennsylvanian fluvial and coal-bearing deposits (Swann, 1963; Greb and Chesnut, 1996; Kvale and Barnhill, 1994; Peppers, 1996). The unconformity is a major chronostratigraphic surface throughout the North American Midcontinent with sub and superjacent strata reflecting pronounced changes in deposition attributed to glacio-eustasy (Heckel, 1986; Ross and Ross, 1985, 1988), climate change (Cecil, 1990; Smith and Read, 2000), and the onset of the Alleghenian Orogeny (Ettensohn, 2004, 2008, 2009).
The study of pollen and spore types (palynology) is used to delineate and correlate Michigan basin strata to other units throughout the Midcontinent. Based on analysis of microflora, a Chesterian age is confirmed for the Michigan Formation and Bayport Limestone. Spore assemblages representing Chesterian, Morrowan, and Atokan regional North American Stages (Fig. 1) were confidently identified indicating Late Mississippian and Pennsylvanian strata are indeed represented in the Michigan basin, contrasting with previous interpretations (e.g. Swann, 1963, 1964; Cohee, 1979; Harrell et al., 1991). Through the use of biostratigraphy, facies analysis and modern stratigraphic approaches including the interpretation of depositional environment and climate-sensitive sediments (i.e. carbonates, evaporites, and paleosols), the Mississippian-Pennsylvanian boundary was resolved in the Michigan basin.
Figure 1. Existing Carboniferous Michigan Basin chronostratigraphy. Vertical lines represent periods of non-deposition or erosion. No Chesterian (Late Mississippian chronostratigraphic stage) fossils have been recovered prior to this study. Relative ages from Gradstein et al., 2004. Regional stages from Heckel and Clayton, 2006. Michigan basin lithostratigraphic relationships modified from Fisher et al., 1988.
Project Objectives

This study furthers the understanding of the Carboniferous Michigan basin strata by undertaking a sedimentologic and stratigraphic analysis of conventional core material believed to capture the Mississippian-Pennsylvanian sequence boundary based on existing knowledge of lithostratigraphic relationships. Due to the lack of geologic material, past research of the Michigan Carboniferous succession relied on vertically and aerially limited outcrop and quarry exposures (Winchell, 1861; Lane, 1902, 1909; Kelly, 1936; Bacon, 1971; Ciner, 1988), drill cuttings (Moser, 1963; Wanless and Shideler, 1975; Tyler, 1980; Vugrinovich, 1984), and/or borehole geophysical logs (Lilienthal, 1979; Vugrinovich, 1984; Westjohn and Weaver, 1998), which alone cannot fully characterize the sedimentology and stratigraphy of these complex units. Significant amounts of core material has since been recovered from the shallow bedrock of the southern Michigan basin and stored at the Michigan Repository for Research and Education (MGRRE), Western Michigan University. Four of these cores represent a complete stratigraphic section from the southern Michigan basin and offer a more complete understanding of Carboniferous Michigan basin stratigraphy.

The primary objectives of this study are to evaluate Carboniferous rock strata deposited within the Michigan basin using the methodology of basin analysis including: (1) lithostratigraphy and sedimentology, (2) sequence stratigraphy of lithostratigraphic units in a regional chronostratigraphic framework based on biostratigraphic analysis, and (3) the determination of the geologic controls on the deposition of rock-units within the Michigan basin. In order to meet these broad objectives, core material was: (1) analyzed for sedimentary facies characteristics and physical features including texture, lithology, bedding style, and stratal relationships,
(2) divided into rock units (facies), which were interpreted based on the evaluation of modern and ancient depositional environments described in the literature, (3) sampled for analysis of microflora (Dr. Robert Ravn, IRF Consulting Group), and this data was (4) integrated to gain insights regarding the depositional environment, regional stratigraphic relationships, and the depositional evolution of the Carboniferous Michigan basin.

**Fundamental Questions**

(1) What are the chronostratigraphic/biostratigraphic and lithostratigraphic relationships of Late Mississippian and Early Pennsylvanian Michigan basin strata?

(2) What primary depositional environments are recorded in Late Mississippian and Early Pennsylvanian Michigan basin strata?

(3) How is the Mississippian-Pennsylvanian regional sequence boundary manifested in the Michigan basin?

(4) How does Late Mississippian strata in the Michigan basin compare with coeval strata elsewhere in the midcontinent?
North American Cratonic (Sloss) “Megasequences”

Sloss (1963) divided sedimentary strata on the North American Craton into six depositional “sequences” spanning a time-frame of 10-100 million years bounded by periods of non-deposition and erosion (i.e. unconformities). A major boundary separates the Kaskaskia and Absaroka sequences (Fig. 2), approximately dividing the Mississippian and Pennsylvanian periods. Sloss (1963) noted that unconformities tend to decrease in magnitude from the cratonic arches, or basin margins, into basins and are typically transitional into nearly or completely conformable successions in the area of maximum subsidence basinward. Vail et al. (1977a,b) utilized seismic reflection data to correlate stratigraphic packages (i.e. sequences) supporting the Sloss (1963) depositional framework of the cratonic interior. Figure 2 shows the comparison between the Vail et al. (1977) relative sea-level curve and corresponding "Sloss" sequences and sequence boundaries (e.g. maximum period of erosion, "erosional maxima").

Sloss (1963) noted the complex nature of the Kaskaskia-Absaroka surface. In the Illinois basin the top of the Kaskaskia sequence is marked by cyclic "regressive deposits" composed of a heterolithic mixture of sandstone, shale, and carbonate strata (Sloss, 1963; Smith and Read, 2001; Nelson et al., 2002). The reactivation and uplift of the Transcontinental, Cincinnati, and Waverly arches may have resulted in significant erosion along the crests of these structures (Sloss, 1963; Ettensohn, 1981; Smith and Read, 2001). The Illinois basin was noted by Sloss (1963) as being prone to the influx of terrestrial detritus possibly sourced from the Canadian Shield (e.g. Potter and Siever, 1956; Sloss, 1963; Swann, 1963, 1964) and other locally "positive
elements" such as adjacent arches and domes (e.g. Transcontinental Arch) (Nelson et al., 2002). Sloss (1963) notes a decrease in the amount and thickness of the "cycles" toward the cratonic margin which transitions into hemi-pelagic shale beds deposited basinward.

While Sloss (1963) does not mention the Michigan basin in particular, this locale is noteworthy for its position within the cratonic interior and proximity to the Canadian Shield (Potter and Siever, 1956) and the above mentioned positive elements, potential source areas of clastic detritus. Carboniferous strata in the Michigan basin are predominantly siliciclastic which supports this line of reasoning.

Traditionally, the position of the Kaskaskia-Absaroka sequence boundary is placed at the contact between the Mississippian (Meramecian?) Bayport Limestone and overlying "Pennsylvanian" strata informally referred to as the "Parma" Sandstone (Lane, 1909; Newcombe, 1933; Kelly, 1936; Wanless and Shideler, 1975). Fisher et al. (1988) and Harrell et al. (1991) note the lack of reported Chesterian strata within the basin. Prior to this study, no Chesterian fossils were recovered or reported from the Michigan basin, suggesting non-deposition during the latest Mississippian Chesterian stage (Fig. 1) (Swann, 1963, 1964; Fisher et al., 1988; Harrell et al., 1991).
Figure 2. The six depositional megasequences of Sloss, showing widespread periods of deposition (color fill) and non-deposition (brown) across the North American Cratonic interior. Red-lines denote major megasequence boundaries which correspond to relative sea-level lowstands inferred by Vail et al., 1977b from seismic interpretation. A major unconformity separates the Mississippian and Pennsylvanian North American periods and represents the target interval of this study. Figure is modified from Catacosinos et al., 1990. After Sloss (1963) and Vail et al. (1977b).
The Carboniferous Michigan Basin

The Michigan basin encompasses a circular area approximately 122,000 mi² (316,978 km²) centered on the lower Peninsula of Michigan (Figs. 3 and 4) occupying portions of Wisconsin, Illinois, Indiana, and Ontario (Ells, 1979). The basin is bounded by the Kankakee and Findley arches to the southwest and southeast, the Wisconsin Dome and highlands to the northwest, and the Canadian Shield to the northeast (Cohee and Landes, 1958; Ells, 1979). Figure 3 gives a regional overview of Carboniferous-aged (Mississippian and Pennsylvanian) units, which form the dominant bedrock units of the eastern United States and in the Illinois and Appalachian basins. Paleozoic deposits preserved within the Michigan basin exhibit a downwarped, nearly symmetrically orientated synformal structure containing correlative strata, which decreases in altitude toward the central basin. The age of the subcrop increases outward from the basin center (Ells, 1979), which is approximately Clare County. Strata contained within the Michigan basin ranges from Precambrian to Pennsylvanian in age with local occurrences of Jurassic?-aged strata locally resting upon Pennsylvanian deposits (Fig. 4) (Dorr and Eschman, 1970; Ells, 1979; Fisher et al., 1988; Catacosinos et al., 1990). The thickest sedimentary rock succession is located in Clare County, where strata attain a total thickness of >4 km (Lilienthal, 1978; Ells, 1979). Proximal to the basin interior Carboniferous strata attain a maximum of 3,800 ft. (1,160 m) and represent the majority of the bedrock subcrop surface in the Lower Peninsula of Michigan (Fig. 4) (Ells, 1979). The Carboniferous dominated subcrop is covered by a layer of glacial deposits up to 1200 ft. (~400 m) in thickness, deposited during the Pleistocene Epoch (Ells, 1979).

During most of the Paleozoic, shallow marine waters inundated the North American continent preserving sediments within tectonically subsiding basins (Fig. 3).
(Dorr and Eschman, 1970). Deposition in the Paleozoic Michigan basin reflects a
tectonically quiescent environment favorable to the deposition of shallow marine,
mostly biogenic carbonates during the Early-Middle Ordovician, Early-Middle
Silurian, and the Middle Devonian times. These periods of extensive carbonate
deposition are separated by intervals of shale, sandstone, and evaporite deposition
during the Late Ordovician, Middle-Late Silurian, and are capped by mostly clastics
of the Late Devonian through the Pennsylvanian periods.

Structure

The Michigan basin is a nearly symmetrical, concave, structure preserving the
thickest accumulations of basin-fill in the basin center due to persistent, but
punctuated, Paleozoic subsidence (Howell and van der Pluijm, 1999; Dorr and
Eschman, 1970). A series of anticlines trending in a northwest-southeast orientation
are present across the basin and locally important in hydrocarbon exploration,
providing structural closure for a number of oil and gas fields (Cohee and Landes,
1958; Ells, 1979; Catacosinos et.al., 1991). Many of these structures are interpreted
as basement faults, which were periodically reactivated during the Paleozoic (Ells,
1979; Fisher et. al., 1988). Newcombe (1933) and Cohee (1979) suggest tectonic
uplift occurred during the Chesterian within the Michigan basin based on the uneven
distribution of Bayport strata.
Geophysical methods including magnetic, gravity, seismic reflection, and borehole geophysical logs have been used to interpret structures within the basin (e.g. Catacosinos et al., 1990; Howell and van der Pluijm, 1999). In the south-eastern Michigan basin, the Howell and Lucas-Monroe structures are high-angle normal faults which exhibited displacements in excess of 100 m during the Late Mississippian (Ells, 1979; Fisher et al., 1988). Periodic structural reactivation during the Late Mississippian is reported along the Lucas-Monroe fault (Fisher et al., 1988) and appears to have been modified by left lateral shear zones, interpreted by Haimson (1978) as a product of a northeast-southwest orientated compressional stress field. To the west of the Lucas-Monroe and Howell structures, deformation along the Albion-Scipio fault is inferred from a linear swath of hydrothermally dolomitized limestone, a product of upward moving, high temperature brines associated with faulting (Catacosinos et al., 1990). Folding along the Howell Anticline during the Carboniferous is suggested by Ells (1979) to account for erosion of the Michigan Formation, Bayport Limestone and overlying Pennsylvanian strata from the crest of the structure. Ells (1979) suggests periodic reactivation and erosion along the crest to explain distribution of strata adjacent to, and overlying the structure. Wanless and Shideler (1975) infer the Howell anticline stood hundreds of feet above the depositional surface before being covered by Atokan time.
Figure 4. Map displaying the prominent structural features found within and bounding the Michigan basin. Bedrock (subcrop) units increase in age toward the basin perimeter. The Howell Anticline is interpreted as being a fault block which may have been a positive feature during the Mississippian and Pennsylvanian. Studies of the Lucas-Monroe fault indicates that the structure was reactivated during the Mississippian. From Catacosinos et al. 1990.

Subsidence

Tectonic subsidence is the dominant mechanism controlling the preservation of Paleozoic strata within the Michigan basin (Cohee and Landes, 1958). The origin of the nearly symmetrical subsidence pattern is the subject of debate (refer to Fisher et
al., 1988; Catacacinos et al., 1991). Howell and van der Pluijm (1999) propose thermal contraction, associated with the 1.1 billion year old (Ba) extrusive Midcontinent Rift System, initiated basin-subsidence. They interpreted periods of narrow subsidence as being a result of load-induced crustal weakening, with intervening periods of broad basin subsidence reflecting crustal “attenuation.” Periodic eastward tilting events during the Middle-Late Ordovician and Late Devonian may reflect a tectonic signal derived from Appalachian tectonism, in particular plate-subduction along the eastern cratonic margin during the Ordovician (Howell and van der Pluijm, 1999) and during the Late Devonian and Mississippian periods as recognized by Ettensohn (2004). The flexural effects of the Taconic, Acadian, and Alleghenian orogenies the Appalachian foreland basins are described by Ettensohn (2004, 2008), but have been largely neglected in the interpretation of Carboniferous strata of the Michigan basin.

Harrell et al. (1991) suggests progressively diminishing rates of subsidence through the Mississippian Period culminated in a pronounced stratigraphic hiatus during the Late Mississippian, Chesterian Stage and infer that Pennsylvanian strata rest on Meramecian deposits in the central Michigan basin constituting a major stratigraphic hiatus or unconformity. The Mississippian is significantly thicker and more laterally extensive than the overlying Pennsylvanian strata. Isopach (thickness) maps for the Mississippian and Pennsylvanian system (Fig. 5) indicates basin centered subsidence during the Mississippian, contrasting with the lobate geometry of the Pennsylvanian system (Wanless and Shideler, 1975; Fisher et al., 1988). Erosional stripping due to widespread glacial advances and retreats occurred during the Pleistocene Epoch and represents an unconformity with a magnitude of 100s of millions of years. Despite extensive erosion around the basin margin and poor
lithologic and wireline log-control, Howell and van der Pluijm (1999) interpret an eastward tilting event during the Carboniferous. However, the rate and amount of subsidence that occurred during the Carboniferous is questionable due to the poor lateral and temporal correlation of rock strata deposited within the Michigan basin.

Figure 5. Cumulative thickness (isopach) maps of the (A) Mississippian (Kaskaskia II) and (B) Pennsylvanian (Absaroka I) sequences. From Fisher et al., 1988; Wanless and Shideler, 1975.


**Paleogeographic Setting**

During the Mississippian the Michigan basin was situated between 5 and 15° south of the paleo-equator (Fig. 6) (Scotese and McHerrow, 1990), within the Euramerican continent and occupying the landward cratonic margin of a broad intercontinental sea, covering a nearly continuous area from present day Pennsylvania to Nevada (Fig. 6) (e.g. Craig and Connor, 1979; Smith and Read, 2001). Smith and Read (2001) note the Illinois basin exhibited a 7 cm/km (0.004°) southwestward slope during the Late Mississippian (Chesterian). Paleogeographic reconstructions consistently interpret the Michigan basin as being a shallow marine embayment open to the present day south during the Mississippian (Fig. 6) (Miall and Blakey, 2008). During the Mississippian the Michigan basin was positioned ~600 km up regional dip from the southern cratonic margin and the Ouachita foredeep, which occupied portions of the present-day southern United States (Mississippi, Arkansas, Oklahoma, Louisiana, and Texas) (Smith and Read, 2001). Evidence that the Michigan basin was periodically restricted from this broad epeiric sea-way during the Mississippian period is supported by extensive evaporitic strata of the Michigan Formation (Cohee, 1979). Paleogeographic reconstructions based on the existing knowledge of interpreted paleogeography suggest the Michigan basin was subaerially exposed during the Late Mississippian (~325 Ma) (Fig. 6) resulting in non-deposition during the Late Mississippian Chesterian North American Stage (e.g. Swann, 1963, 1964; Harrell et al., 1991).
Figure 6. Global and regional paleogeographic reconstructions denoting the location of the Michigan basin (red arrow) within the North American midcontinent and the Euramerican paleocontinent. (A) The Middle Mississippian world (~345 Ma). The Michigan basin was a low-latitude shallow marine embayment within the Euramerican continent and most of the continental United States was covered by a broad intracontinental seaway. (B) Late Mississippian North America (~325 Ma) indicate nondeposition within the Michigan basin and the "docking" of Euramerica and Gondwana. Illustrations from Blakey, http://www2.nau.edu/rcb7/. Paleoequator from Miall and Blakey, 2008.

**Tectonic Setting**

The intracratonic position of the Michigan basin suggests tectonic quiescence; however Acadian and Alleghenian orogenies during the Carboniferous likely influenced deposition (e.g. Ettensohn, 2004, 2008). The Late Devonian and Carboniferous Midcontinent was heavily impacted by deformation associated with the Late Devonian Acadian, and later, the Alleghenian Orogeny during Pennsylvanian to Permian time (Ettensohn, 2004; Scotese and McKerrow, 1990). During these periods of orogeny, the Appalachian foreland basin was particularly impacted by tectonism due to proximity to the continental margin. The Taconic and Acadian orogenies occurred due to plate subduction along the eastern Appalachian plate-margin during the Ordovician and Devonian-Mississippian time respectively (Ettensohn, 2004, 2008). These periods of tectonism resulted in the deposition of large volumes of siliciclastic material in the Appalachian foreland basin (Ettensohn, 2004, 2008). These clastic "wedges," as they were referred to by Sloss (1962), are manifested most notably as progradational "deltaic" deposits which thin westward into the continental interior (Ettensohn, 2004). Plate tectonic reconstructions by Scotese and McKerrow (1990) indicate the progressive amalgamation of Euramerica
and Gondwana proceeded through the Mississippian, Pennsylvanian, and Permian periods resulting in the closure of a narrow sea-way separating the two continents. The Euramerica-Gondwana collision followed a path from the Ouachita (Southern United States) to Hercynian/Variscan (present day Nova Scotia) orogenic belts (Scotese and McKerrow, 1990; Ettensohn, 2004) and likely initiated during the Late Mississippian (Ettensohn, 2008). The timing of the collision is indicated by the structural reactivation of basement structures during the Late Mississippian (i.e. Waverly, Findlay, and Cincinnati arches) in the Appalachian basin (e.g. Ettensohn and Peppers, 1979; Ettensohn, 1981, 2008) and Illinois basin (Nelson and McBride, 1988).

Climate

The Carboniferous was a time of intense glaciations across the paleocontinent of Gondwana, which included present day South America, Africa, Madagascar, India, Antarctica, and Australia (Scotese and McKerrow, 1990). The glacial episodes correspond to global ice-house conditions (Frakes et al., 1992). Continental glaciations were most extensive over the continent of Gondwana, which contained portions of the modern continents of South America, Africa, and Australia. While the age, frequency, and extent of Late Paleozoic glaciations remains poorly constrained, glaciations intensified during the Late Mississippian (Serpukhovian, Middle Chesterian/youngest Mississippian global stage) (Smith and Read, 2000) and extended as far north as 30°S latitude during the Pennsylvanian (Frakes et al., 1992). Chronostratigraphic control of these glacial episodes is based on glacial till and dropstone deposits found within dateable strata (Frakes et al., 1992; Haq and Schutter, 2008). However, glacial events cannot be directly dated using biostratigraphic
methods, making precise correlation with midcontinent cycloths tenuous (Frakes et al., 1992).

The onset of Carboniferous glacial events is associated with global ice-house conditions which encompassed the Carboniferous and Permian periods (Frakes et al., 1992). Models of Gondwana ice-volumes are comparable to that of the Pleistocene (Crowley et al., 1991). Climate modeling performed by Crowley and Baum (1992) suggest Gondwana glaciations were associated with changes in tectonic plate locations, solar insolation, and a decrease in atmospheric CO₂. The South American plate was positioned over the southern polar region establishing a site of preferential ice accumulation (Frakes et al., 1992). Smith and Read (2000) suggest the closing of the Tethys equatorial seaway and the assembly of the super-continent of Pangaea exerted a forcing on the transition from greenhouse to icehouse conditions across the Mississippian-Pennsylvanian boundary (Fig. 7). In the Smith and Read (2000) global climate model, equatorial currents were deflected to the poles resulting in greater precipitation and an increase in ice-accumulations on the polar continent of Gondwana. This transition resulted in a steeper temperature gradient between the poles and equator resulting in the reduction in the distribution of carbonate and evaporite strata to a narrow equatorial band during the Pennsylvanian (Frakes et al., 1992). Furthermore Smith and Read (2000, 2001) suggest the onset of these glacial events occurred during the Late Mississippian (Chesterian) and constitutes a stratigraphic marker indicating a three-fold increase in frequency and amplitude of sea-level oscillations associated with Gondwana glacial events.
Figure 7. Plate tectonic reconstructions illustrating the dominant ocean-current/trade wind patterns. During the Mississippian the Equatorial sea-way was unobstructed before closing during the Early Pennsylvanian continent-continent collision (Alleghenian Orogeny). Modified from Smith and Read, 2000. After Scotese and McKerrow, 1990.
Michigan Basin Stratigraphic Framework

The Carboniferous System of the United States is divided into two periods: Mississippian (~359-318 Ma) and Pennsylvanian (~318-299 Ma) (Heckel and Clayton, 2006). These Periods or "subsystems" are divided into smaller chronostratigraphic units including Series and Stages (i.e. Fig. 1). Heckel and Clayton (2006) give an overview of recent amendments made to the Carboniferous system nomenclature to better streamline global geologic interpretation and correlation. Among the changes is the global standardization of the Mississippian and Pennsylvanian subsystem, replacing the "upper" and "lower" Carboniferous in global terminology (Heckel and Clayton, 2006). Modification of chronostratigraphic boundaries are undertaken by the International Commission on Stratigraphy (ICS).

Figure 5 displays the composite thickness of the Mississippian and Pennsylvanian systems in the Michigan basin respectively. Mississippian strata attain a thickness of ~1700 ft. (530 m) in the central part of the basin (Fig. 5a) (Ells, 1979; Fisher et al., 1988). In contrast, Pennsylvanian-aged deposits are restricted to the central Michigan basin with a maximum reported thickness of ~500 ft. (Wanless and Shideler, 1975; Fisher et al., 1988). Post depositional erosion resulted in significant erosion of previously deposited strata, especially around the basin periphery. Shale compaction studies conducted by Vugrinovich (1988) suggest a Pennsylvanian stratum was buried by up to 2600 ft. (800 m) of overburden. Furthermore, a past elevated paleogeothermal gradient is evidenced by thermally mature hydrocarbon accumulations, which cannot be explained by the present geothermal gradient of 19 °C/Km (Vugrinovich, 1989; Ma et al., 2009). While burial diagenesis is not the focus of this study, the Bayport limestone in the southern portion of the study area may have been subject to hydrothermal alteration which resulted in the physical and
chemical alteration of the parent rock (i.e. brecciation and dolomitization) (Harrison, Pers. Comm., 2012).

Figure 8. The most-recent graphical representation Carboniferous stratigraphic nomenclature and general lithologies occurring within the Michigan basin reflecting uncertainties regarding the Mississippian-Pennsylvanian systemic boundary. Modified from Catacosinos et al., 2000.

Early to Middle Paleozoic Michigan basin strata reflect deposition within a variety of distinct subtidal marine environments with limestone and dolomite being the most dominant lithotypes preserved (Cohee and Landes, 1956). During the Late Devonian to Early Carboniferous, carbonates became subordinate in an otherwise siliciclastic-dominated depositional system characterized by the deposition of
hemipelagic, fine-grained organic shale deposits of the Antrim shale that terminated carbonate deposition during the Late Devonian. Antrim Shale deposition was followed by deltaic deposition which temporarily terminated the deposition of low-energy black shales (Gutschick and Sandberg, 1991). Analysis of Late Devonian Michigan basin strata by Gutschick and Sandberg (1991) indicates deposition was heavily influenced by eustatic sea-level and Appalachian Acadian tectophases. The progradation of the Berea-Bedford deltas halted black-shale deposition and the Michigan basin became emergent during Late Devonian through earliest Mississippian (Gutschick and Sandberg, 1991). During the Early Mississippian the deposition of fine-grained shale deposits representing the Sunbury and Coldwater Shale units resumed. The overlying Marshall Sandstone, by comparison, is a relatively coarse grained marine sandstone unit (Harrell et al., 1991). Possible source terrains for detritus include the Canadian Shield to the northeast, the Wisconsin Highlands to the northwest and possibly easterly Appalachian source (Potter and Siever, 1956; Harrell et al., 1991). A northerly source area during the Early-Carboniferous is supported by a general eastward thickening of the Coldwater Shale and Marshall Sandstone (Cohee, 1979; Howell and van der Pluijm, 1999). Harrell et al. (1991) gives the Wisconsin Highlands as a possible secondary source of detritus during the deposition of upper portion of the Marshall Sandstone. The Marshall is overlain by the fine-grained Michigan Formation, Bayport Limestone, and Pennsylvanian-aged Saginaw/Grand-River Formations, which are thoroughly described later in this report. The study of regional cross-bed dip orientations by Potter and Siever (1956) was used to interpret a northeasterly Canadian Shield source for basal Pennsylvanian sandstone strata deposited in the Michigan and Illinois basins. Early interpretations by Swann (1963, 1964) proposed that Canadian Shield
sourced strata transported along a south to southwest orientated paleo-slope extending through the Michigan basin and into the eastern Illinois basin (Fig. 8). A series of northeast/southwest trending paleo-valleys filled with siliciclastic strata were interpreted by Swann (1963) as derived from the north-easterly Canadian Shield, bypassing the Michigan basin and deposited in a channelized fluvio-deltaic shoreline analogous to the Mississippi River Delta (Fig. 9). The channels have since been reinterpreted by Leetaru (2000), Smith and Read (1999, 2001), and Nelson et al. (2002) incised valleys eroded during relative sea-level low-stands and filled with a mixture of tidally deposited siliciclastic and carbonate lithologies during the subsequent marine transgression. In the Illinois basin, the source of these sediments appears to be local features including the Transcontinental and Cincinnati Arches (Nelson et al., 2002). Sequence stratigraphic methods/models are useful to explanation of the mixed carbonate and siliciclastie strata of the Chesterian Illinois basin (Smith and Read, 2001; Nelson et al., 2002) and the Appalachian basin (Al-Tawil et al., 2003; Al-Tawil and Read, 2003).

Depositional environments and detailed descriptions of rock facies is generally absent for the Mississippian section in Michigan due to the lack of significant exposures (Ells, 1979) hindering detailed stratigraphic, paleogeographic, or paleoecological interpretations. Collaborative regional mapping of the Carboniferous Midcontinent was undertaken by the USGS to place strata into regional paleotectonic sequences (i.e. Craig and Varnes, 1979; Cohee, 1979). The scheme divides Mississippian rocks into four intervals; A through D, representing the regional tectonic setting.
Figure 9. Paleogeographic and depositional interpretation of the Late-Mississippian Illinois basin and adjacent areas, including southern Michigan. Swann (1963) uses the Mississippi River Delta as an analog to account for a number of northeast-southwest trending paleo-valleys which overly carbonates in the Illinois basin. In this model/interpretation the Michigan basin is sub-aerially exposed and the site of erosion and sediment bypass. From Swann, 1963, 1964.

Interval A encompasses the Late Devonian to earliest-Mississippian-aged Bedford shale and Berea Sandstone (Eastern Michigan basin) and the Ellsworth shale (Western Michigan basin) up through the regionally extensive Sunbury Shale. In the Appalachian basin the base of the Carboniferous is similarly placed at the Sunbury shale (Eble et al., 2009). According to Cohee (1979) the interval thickens to a depth of 780 ft. in the northwest portion of the basin.

Interval B extends from the base of the Coldwater Shale (Tournasian) to the
top of the Marshall Sandstone (Osagean). Cohee (1979) interprets a pronounced eastward thickening of the Coldwater shale as an eastward source area. Interestingly, Cohee (1979) and Harrell et al. (1991) report the Marshall Sandstone as being anomalously thick in the southwestern Michigan basin. Interval B is reported thickest in the Saginaw Bay where it attains more than 1400 ft. and in the southern portion of the basin where it is 1300 ft. (including the Marshall sandstone). The interval is reported to be significantly thinner where the Marshall sandstone is overlain by the Michigan Formation in the more central portion of the basin (Cohee, 1979). McGregor (1954) notes local erosion may have impacted the Marshall Sandstone where the Michigan Formation is not present, on the western periphery of the basin.

Interval C is composed of the Michigan Formation and the Bayport Limestone (Meramecian), previously considered members of the Grand Rapids Group (Cohee, 1979). The Michigan Formation holds significant deposits of gypsum with a minable thickness ranging from 40-100 ft. (Cohee, 1979). The Michigan Formation is composed of fine-grained shale and is overlain by the cherty Bayport Limestone (Lane, 1902; Newcombe, 1933). Cohee (1979) reports interval C to be highly variable in thickness and attributes this to uplift along the Findlay and Kankakee arches which isolated the basin from neighboring depocenters (e.g. Illinois and Appalachian basins) during Meramecian time.

Due to the lack of reported Chesterian fossils, Interval D is considered to be a period of non-deposition and erosion encompassing the time between the deposition of the Bayport Limestone and the Saginaw Formation. Most geologists consider the Bayport to be Meramecian in age based on "correlation" to the St. Louis Limestone from the Mississippi Valley area (Newcombe, 1933; Anisgard and Campeau, 1961; Horowitz and Rexroad, 1972; Cohee, 1979; Harrell et al., 1991). However, the St.
Louis is now considered to be the base of the Chesterian (Nelson et al., 2002). The discontinuous nature of the Bayport is attributed by Newcombe (1933) and Cohee (1979) as being a product of post-depositional tectonic uplift. However, Vugrinovich (1984) suggests Chesterian deposition based on lithological interpretations derived from wire-line logs, core-cuttings, and comparisons of similar strata to the Illinois basin.

In order to synthesize previous observations and interpretations of Michigan basin stratigraphy, a summary of previous lithologic, stratigraphic, and depositional environment investigations of rock units positioned between the Middle Mississippian (Cohee's interval C) and the Pennsylvanian are as follows.

**Michigan Formation**

The Michigan Formation was denoted by Winchell (1861) for a succession of shale, sandstone, carbonate, and evaporitic strata exposed in Kent, Iosco, and Huron Counties that rests conformably on the Marshall Sandstone (Lilienthal, 1978; Harrell et al., 1991). A basal sandstone unit referred to by drillers as the Michigan “Stray” marks the base of the Michigan Formation (Moser, 1963; Harrell et al., 1991). This unit may be equivalent to the upper Marshall, Napoleon member in the southern Michigan basin (Moser, 1963; Harrell et al., 1991). Cohee (1979) suggests that the Marshall/Michigan contact becomes progressively younger to the southwest portion of the basin.

The Michigan Formation attains its greatest thickness in Missaukee County (>400 ft.) (Cohee, 1979; Harrell et al., 1991). The lithology of the Michigan Formation is highly heterogeneous and consists predominantly of shale and subordinate amounts of sandstone, limestone, dolostone, gypsum, and anhydrite
A number of beds within the Michigan formation have been correlated across the central basin including the "Brown Lime" and the "Triple Gypsum" (Moser, 1963; Lilienthal, 1978; Harrell et al., 1991). McGregor (1954) suggests the amount and thickness of anhydrite increases in thickness into the central Michigan basin. The “Triple Gypsum” anhydrite beds have been observed by Vugrinovich (1984) as being structurally high over hydrocarbon field associated with anticlines.

Bayport Limestone

The Bayport Limestone was designated by Lane (1899) for exposures in Huron County, near the town of Bayport for a limestone previously named the Point Au Gres Limestone by (Douglass, 1841) for limestone exposures occurring along southeastern Saginaw Bay (Harrell et al., 1991). Martin and Straight (1956) designate the type section for the Bayport as the Wallace Stone Quarry, located in Huron county, two miles west of the town of Bayport. Subsurface mapping indicates the Bayport Limestone covers an approximate area of 11,000 mi² (28,000 km²) confined to the central Michigan basin area, an area comparable to the overlying Pennsylvanian system (Ells, 1979). The thickness of the unit is reported to be highly variable ranging from 10 to 100 ft. in thickness (Lilienthal, 1978; Lasemi, 1975; Vugrinovich, 1984). The deposition of the Bayport strata is interpreted by Bacon (1971) and Vugrinovich (1984) as being the product of a marine transgression which engulfed the basin following the deposited of the comparatively restricted Michigan Formation.

Bayport strata were studied at a variety of quarry locations in the eastern and southern Michigan basin including Wallace Stone Quarry, Parma Quarry, and
Bellevue Quarry located in Huron, Jackson, and Eaton Counties, respectively. Lithologically, the Bayport is described from outcrops as a heterolithic mixture of cherty and fossiliferous limestone, dolomite, siltstone, and sandstone (Bacon, 1971; Vugrinovich, 1984; Ciner, 1988). Limestones designated the Bayport have been the subject of studies by Lasemi (1975), Tyler (1980), and Vugrinovich (1984).

Lasemi (1986) reports the presence of two facies within the Bayport: (1) a gray fossiliferous limestone and (2) a brown tan dolomite he attributes to a tidal flat origin. Bacon (1971) and Ciner (1988) interpret the Bayport as being deposited in an offshore and sabkha environments based on the observation of microbial structures, and mudcracks. Chert is common in the Bayport Limestone. Cohee (1979) suggests chert concretions within the Bayport indicate the “Bayport Sea” was enriched with silica. Lasemi et al. (2003) postulates that regional upwelling of nutrient and silica rich waters may account for extensive chert deposits within the Late Mississippian Illinois basin.

Sandstone has long been reported to occur in association with the Bayport Limestone. Lane and Seaman (1909) report an erosional contact upon which a conglomerate comprised of "pea" sized quartz grains and reworked carbonate clasts, interpreted as being the "Bayport-Parma" contact. Newcombe (1933) suggests the Bayport is interbedded with “clean” white sandstone similar to the “Parma,” but also contends the Parma and the Bayport are separated by an erosional unconformity.

Lasemi (1975) divides the Bayport into three distinct subunits from data obtained from drill cuttings and lithostratigraphically correlates them through the subsurface. The tripartite division is supported by studies of geophysical logs and drill cuttings by Tyler (1980), Vugrinovich (1984), and Ciner (1988). Ciner (1988) divides exposures observed in quarries near the towns of Bellevue (Cheney Stone
Quarry), Parma, and Bayport (Wallace Stone Quarry). On the basis of lithology, Ciner (1988) suggests the lower Bayport occurs at the Parma location and is composed of sandy dolostone. Wallace Stone Quarry is interpreted to represent the middle, open marine portion of the Bayport (Ciner, 1988) and Lasemi (1986) interpreted a deepening upward succession from that location. Subsurface mapping conducted by Vugrinovich (1984) and Westjohn and Weaver (1996, 1998) suggest an interstratification of limestone and a sandstone lithologies.

The age of the Bayport is poorly constrained. Winchell (1861) was first to document invertebrate marine macro-fauna from the Bayport (then termed the Point au Gres Limestone), to him suggesting a Chesterian age (Horowitz and Rexroad, 1972). However, Newcombe (1933) correlates the Bayport strata to the upper St. Louis Limestone and Ste. Genevieve Limestone of the Mississippi Valley. Anisgard and Campau (1961) and Horowitz and Rexroad (1972) support correlation to the Meramecian St. Louis Limestone of the Mississippi Valley based on similarities in marine fauna. Horowitz and Rexroad (1972) note the occurrence of fenestellid bryozoans, cup corals, and brachiopods from the Wallace Stone quarry and use conodont taxa to correlate Bayport strata from its type-section in Huron County to the Cheney Stone Quarry, southern Michigan. Anisgard and Campau (1963) observed a single specimen of the fusulinid genus *Paramillerella*, which they use to suggest a St. Louis (Meramecian) age for the Bayport. However, the genus ranges from Mississippian Osagean to Pennsylvanian Atokan in age (Anisgard and Campau, 1963; Vugrinovich, 1984) making any chronostratigraphic interpretations based on the previously treated macro and microfossils equivocal.
**Parma Sandstone**

Since Winchell's (1861) first description as "a white, or slightly yellowish, quartzose, glistening sandstone, containing occasional traces of terrestrial vegetation", it remains unclear whether the rock Winchell described was actually the "Parma" or a sandstone belonging to a wholly different stratigraphic unit such as the overlying Saginaw Formation (Kelly, 1936). Newcombe (1933) and Vugrinovich (1984) refer to the "Parma" as white calcareous cemented sandstone stratigraphically positioned overlying the Bayport Limestone. Newcombe (1933) reports the “Parma” contains carbonate intraclasts of underlying Bayport. Newcombe (1933) also reports the localized presence of pebble sized quartz grains and tentatively correlates the bed to the Mansville Formation of Ohio, establishing the “Parma” bed as Pennsylvanian in age.

Since its first description, the Parma Sandstone was described in a number of Michigan Geologic Survey reports and publications (Winchell, 1861; Lane, 1909; Newcombe, 1933; Kelly, 1936; Potter and Siever, 1956; Wanless and Shideler, 1975; Vugrinovich, 1984), but has not been the subject of any detailed sedimentologic or stratigraphic study. The “Parma” has historically been considered to be a basal member of the Pennsylvanian Saginaw Formation (Kelly, 1936; Cohee et. al., 1951; Wanless and Shideler, 1975). Formational status of the Parma Sandstone was eventually dropped from the Michigan basin stratigraphic nomenclature in 1964 (Wanless and Shideler, 1975). The renewed use of drill cuttings and geophysical logs from oil and gas wells enabled the mapping of the “Pennsylvanian” system by Wanless and Shideler (1975), Vugrinovich (1984), and Westjohn and Weaver (1998). These studies suggest that a sandstone unit directly overlying the Bayport Limestone referred to as the “Parma”, warrants stratigraphic distinction from both the underlying
Bayport Limestone and the overlying Saginaw Formation (Vugrinovich, 1984; Westjohn and Weaver, 1998). However, the existing stratigraphic relationships of these units are still a subject of debate due to the lack of lithostratigraphic and chronostratigraphic control.

Vugrinovich (1984) proposed that the "Parma" sandstone represents a distinct lithostratigraphic unit due to its presence across the central Michigan basin and infers a partial Mississippian age for the unit. Tyler (1980) analyzed drill cutting and geophysical logs from the Six Lakes gas storage field, Montcalm County, and identifies a series of shale, sandstone, limestone, and evaporite units directly overlying the “Parma” sandstone. Tyler’s “Cream’ limestone was denoted the “Six Lakes” Limestone member by Vugrinovich (1984), who tentatively gives a Pennsylvanian age for the limestone unit. Vugrinovich (1984) used similar tools (i.e. wireline logs and cuttings) to map these units and reconfigures the existing nomenclature by applying several new names to the subsurface deposits (e.g. Hemlock Lake Formation Six Lakes Limestone Member, Winn Member, Verne Member, etc.). The Vugrinovich (1984) nomenclature will not be followed in this study. However, much of the Vugrinovich (1984) subsurface study was only able to take into account the lithostratigraphic relationships found using well-logs. A detailed chronostratigraphic interpretation requires direct evidence (e.g. continuous core and fossil fauna/flora) to better determine conditions of deposition. Likewise, a greater study of fossils (microflora and conodonts) is required to correlate and constrain the age of deposition to other deposits located within the United States.

Saginaw and Grand-River Formations

Houghton (1836) denoted coal bearing strata the "Coal Measures" after
exposures in Eaton County, near the town of Grand Ledge. Winchell (1861) proposed a stratigraphic hierarchy for the "Coal Measures." The "Coal Measures" have long been noted for its lateral and lithologic heterogeneity (Rominger, 1876) rapidly alternating between shale, sandstone, and small beds of argillaceous carbonates (Kelly, 1936). During the late 19th and early 20th century growth of the coal mining industry prompted a series of studies by Lane (1902) who named a series of sandstone, coal, and black argillaceous limestone units within the Saginaw Formation. Newcombe (1933) denotes three members as part of the Saginaw including the Lower “Parma” Sandstone unit, the Middle Saginaw, and the Upper Woodville members. Lithologically, the Saginaw formation consists of shale, coal, micaceous sandstone, and a discontinuous argillaceous limestone referred to as the “Verne Member” (Kelly, 1936). Plant fossils *Calamites* and *Stigmaria* have been described from the Saginaw Formation (Kelly, 1936). Kelly (1936) proposed an “unconformable” relationship between the Saginaw Formation and the Grand River Formation, a predominantly sandstone unit found overlying fine-grained, carbonaceous deposits of the lower Saginaw Formation. This unconformable relationship was later supported by the study of plant macro-fossils by Arnold (1949). Kelly (1936) uses the regional occurrence of *cyclothems* (Weller, 1930) to explain the lateral heterogeneity of Pennsylvanian strata within Michigan. Shideler (1969) concluded that Pennsylvanian sandstone beds were sourced from older Paleozoic sandstones to the northeast and were deposited in a network of fluvial point-bars. Velbel and Brandt (1989) describe the Saginaw Formation as being deposited in a marginal marine to “deltaic” environment. The Saginaw Aquifer is an important source of potable water, serving a number of counties of the southern Michigan Basin, near the Capital of Lansing (Westjohn and Weaver, 1998).
At the present time, the Pennsylvanian is divided into two formational units, the lower Saginaw Formation and the "overlying" Grand River Formation; however this layer cake relationship is disputed by direct palynological correlation obtained from pollen and spores (Venable, 2006). Venable (2006) studied core material from Ingham County (Americhem core collection), Mason Township. Venable (2006) observes interstratified lithologies interpreted as deposited in distinct environments and sub-environments including channelized fluvial sandstones, shale overbank and paleosol deposits, tidally induced non-cyclic rhythmites (i.e. heterolithic bedding), and a restricted trace-fossil assemblage.

Kelly (1936) gives the age of the Verne Limestone, the only significant Pennsylvanian carbonate (Wanless and Shideler, 1975), as Early Pennsylvanian (Late Atokan) based on macrofauna. The study of coal plant material by Arnold (1949) was interpreted to indicate an Atokan age for the Saginaw Formation. Vugrinovich (1984) uses the lithologic data to infer Morrowan to Late Desmoinesian age for central basin deposits. Biostratigraphic analysis of micro-flora performed by Dr. Robert Ravn, IRF consulting, of outcrop and core material provided by Venable (2006) indicates the Saginaw and the Grand River Formations were both deposited during the Atokan North American Stage, disputing Kelly’s previous interpretation of a significant unconformity. Interstratification of generally finer-grain Saginaw “facies” and coarser-grained sandstone channel fill deposits of Grand River “facies” in cores from the Americhem site indicates that these two formations are actually a complex facies assemblage, in conformable stratigraphic contact (Venable, 2006).
CHAPTER II

DATA AND METHODS

Conventional Core and Outcrop Data

Ten cores capturing the Bayport Limestone and other Carboniferous units were used in this study. The well locations are shown in Figure 10. Eight of the ten cores were recovered as part of a geotechnical survey in preparation for a proposed Superconducting Super Collider (SCSC) located in Jackson and Ingham Counties (Fig. 10). The cores are presently housed at the Michigan Geological Repository for Research and Education (MGRRE), a core repository jointly affiliated with Western Michigan University and the Michigan Geological Survey. These cores represent the majority of the subsurface data used in this study. Of particular interest are four continuous cores (SB13, SB15, SB16, and SB18) from the site that provides data set for investigation of the Mississippian Marshall Sandstone through Saginaw Formations. In addition, two cores were recovered from strata overlying the Six Lakes Michigan “Stray” gas storage field located in Montcalm County and capture a mixed siliciclastic/carbonate interval. Representative photographs were taken noting the important attributes of a particular rock (i.e. facies) type. The slabbing and re-boxing of cores was selectively performed before the core description was undertaken to better observe the physical characteristics of the strata, as well as for curatorial purposes. Representative thin-sections were taken to evaluate and document depositional textures and microfacies within the interval, which are especially relevant for carbonate lithologies (sensu Wilson, 1975; Flugel, 2010).

Outcrop exposures located at Wallace Stone Quarry, Huron County, offer a
detailed view of the Bayport Formation in its type section. The quarry is operated by Burroughs Materials North and is managed by Eric Gardy, who provided access to the site and background information about the quarry operation. The quarry is located on a relatively low-declivity, flat topographic surface surrounded by agricultural land, ~1.5 miles (2.4 km) east of the town of Bayport, near the shore of Lake Huron. The quarry occupies a rectangular area 8,500 ft. (2,590 m) along the east-west long axis and 4,400 ft. (1,340 m) along the widest north-south transect. Two primary locations analyzed for broad changes in lithology, fossil content, and grain size were consistent with what is observed in core material encompassing the Bayport interval (this study). Locations observed include a small pit on the west side of N. Bayport Road and a larger pit located directly to east of the same road, referred to as Pit 1 and Pit 2 respectively. Pit 2 has ~20 ft. of vertical exposures and comprises of three ledges which could be traced hundreds of feet along the quarry wall. The quarry serves as a type section for the Bayport (e.g. Martin and Straight, 1956; Horowitz and Rexroad, 1972) and was used as a principal study interval for M.S. thesis projects undertaken by Bacon (1971) and Ciner (1988). While the interval observed is predominantly limestone, significant amounts of dolomite and sandstone occur as well.
Figure 10. (A) Map showing the total distribution of core material and data locations used in previous studies. This study relies heavily on the SCSC cores (southern Michigan basin) and two cores referred to as the MCGC cores located within the basin interior. Quarry locations 1-4 have been studied in previous investigations: (1) "Parma" Quarry, (2) Bellevue Quarry, (4) Wallace Stone Quarry. Site 3 represents exposures of the Pennsylvanian Saginaw/Grand River formation near the town of Grand Ledge. (B) A closeup view of the southern Michigan basin. SB16, SB15, SB18, and SB13 (green dots) capture a continuous stratigraphic succession extending from the Marshall/Michigan Formations through the Saginaw/Grand River formational "type" lithologies. Shape-file modified from Milstein, 1987. MDEQ shape-file. Structural lineaments were provided by John Esch, Michigan Office of Oil, Gas, and Minerals.

Table 1. Inventory of the name, depths, location, and aggregate core footages used in this study. All wells capture all or some of the Bayport Interval. Cores SB18, SB16, SB15, and SB13 are continuous and capture Mississippian and Pennsylvanian stratigraphic units.

<table>
<thead>
<tr>
<th>Well Name</th>
<th>Permit #</th>
<th>Drill Collar Depth (ft.) A.S.L.</th>
<th>County</th>
<th>Max. Core Depth (ft.)</th>
<th>Min. Core Depth (ft.) (Subsurface)</th>
<th>Total Linear (ft.)</th>
</tr>
</thead>
<tbody>
<tr>
<td>SL418</td>
<td>MCGC 3357200</td>
<td>994.0</td>
<td>Montcalm</td>
<td>910</td>
<td>790</td>
<td>120</td>
</tr>
<tr>
<td>SL423</td>
<td>MCGC 3357300</td>
<td>1023.0</td>
<td>Montcalm</td>
<td>902</td>
<td>802</td>
<td>100</td>
</tr>
<tr>
<td>SB16</td>
<td>NP 01135</td>
<td>950.0</td>
<td>Ingham</td>
<td>646</td>
<td>220</td>
<td>426</td>
</tr>
<tr>
<td>SB15</td>
<td>NP 07108</td>
<td>945.0</td>
<td>Ingham</td>
<td>468</td>
<td>62</td>
<td>406</td>
</tr>
<tr>
<td>SB18</td>
<td>NP 00287</td>
<td>944.0</td>
<td>Ingham</td>
<td>410</td>
<td>89</td>
<td>321</td>
</tr>
<tr>
<td>SB13</td>
<td>NP 07107</td>
<td>910.0</td>
<td>Jackson</td>
<td>232</td>
<td>62</td>
<td>170</td>
</tr>
<tr>
<td>SB11</td>
<td>NP 22627</td>
<td>910.0</td>
<td>Jackson</td>
<td>149</td>
<td>59</td>
<td>90</td>
</tr>
<tr>
<td>SB10A</td>
<td>NP 22626</td>
<td>925.0</td>
<td>Jackson</td>
<td>243</td>
<td>112</td>
<td>131</td>
</tr>
<tr>
<td>SB03</td>
<td>NP 22619</td>
<td>985.0</td>
<td>Jackson</td>
<td>207</td>
<td>64</td>
<td>143</td>
</tr>
<tr>
<td>SB01</td>
<td>NP 22617</td>
<td>945.0</td>
<td>Jackson</td>
<td>149</td>
<td>59</td>
<td>90</td>
</tr>
</tbody>
</table>
Core Description and Interpretation

Sedimentary rocks can be divided into three broad groups reflecting their composition and genesis (Boggs, 2010). These include siliciclastic, volcaniclastic, and chemical rocks, which reflect the complex interaction between physical, chemical, and biologic processes. For the purpose of this study only siliciclastic and biochemical/chemical types of rocks are present. An example graphical description of core material used in this study is displayed in Fig. 11.

Core material was analyzed and described for attributes including color, lithology, grain-size, grain-packing, fossil content, texture, sedimentary structures, and stratigraphic surfaces. Observations were made using a hand-lens, grain-size comparator chart, and binocular microscope. Where applicable, especially for paleosol intervals, color was noted using the GSA Geological Rock-Color Chart (2009). Thin-sections were analyzed with Leica polarizing light microscope. Observations were recorded on a notebook and detailed descriptions of core material are located in the appendix. Graphic core descriptions and stratigraphic time-charts displayed throughout this thesis were constructed using Adobe Illustrator and Photoshop software applications. Due to the heterolithic nature of the study interval a high resolution study was necessary, locally, on the order of inches to feet. Due to the wide range of lithologies, key differences in interpretations and classification exist between these distinct rock-types (e.g. carbonate vs. siliciclastic strata).

Carbonate rock units were analyzed on both the macro and micro-scale according to grain type, fossil content, sedimentary structures, texture, and diagenetic overprint (e.g. dolomitization, silicification, and pedogenesis). Hydrochloric acid (HCl) was used to discriminate between calcium carbonate (CaCO₃) and dolomite (Mg₆Ca₃(CO₃)₁₀(OH)₂), as the latter reacts to HCl much less vigorously than the former.
Thin-section analysis is the standard method for evaluating limestone and dolostone strata as they can reveal diagnostic textures and fossils that can be used in an environmental interpretation (Wilson, 1975). Thin-sections prepared from limestone and dolomite units were analyzed and compared to other ancient and modern examples of microfacies provided in Wilson (1975) and Flugel (2010). The amount of interstitial mud, grain types, and fossil fauna were noted when describing and interpreting carbonate facies. Limestone and dolomitic strata in this study were classified according to the Dunham (1962) classification, which emphasizes texture, mud-content, and grain packing which is typically related to the environment of deposition. Embry and Klovan (1971) modified the Dunham (1962) classification scheme to accommodate different types of binding and frame-building organisms.

Siliciclastic, or clastic, rocks were discriminated primarily on the basis of physical attributes including bedding style, grain-size, cement constituent, and to a lesser degree mineralogy. HCl and thin-sections were used to identify CaCO₃ cemented intervals which are common within sandstone rock units within the study interval. Bedding style was noted in considerable detail. Sandstones, unlike carbonate strata, commonly display primary bedding structures including planar, cross, and wavy geometries. Such primary sedimentary structures were used to interpret energy conditions (e.g. Harms, 1975, 1979). Bed-forms encountered in the observed rock units were largely interpreted based on their vertical position relative other facies in the succession (e.g. Walker, 1979). Interpretations of modern and ancient strata encountered within the scientific literature were extremely helpful when determining environments of deposition (e.g. Dalrymple et al., 1992; Dalrymple, 1992; 2010c; Dalrymple and Choi, 2007; van den Berg et al., 2007).

Features used to interpret paleosol deposits include: texture/structure, color
(high/low chroma), zones of rooting, and pedogenic nodules composed of carbonate, sulfates, and/or silica. The genetic interpretation of paleosols was largely performed using hand-specimens based on texture and structure (sensu Mack et al., 1993). Representative thin-sections were analyzed and used for interpretation (especially in carbonate lithologies). Intervals displaying characteristics of paleosols and pedogenesis were classified by the Mack et al. (1993) scheme which uses physical, textural, and mineralogical criteria to determine the “type” of paleosol.

The Mack et al. (1993) scheme uses six features for the classification of paleosols including: organic matter content, horizonation, redox conditions, in situ mineral alteration, illuviation of insoluble minerals/compounds, and the accumulation of soluble minerals. The scheme combines modern soil taxonomic classification (e.g. histosol, spodosol, oxisol, and vertisol) with soil classifications distinct to the geologic past (e.g. calcisol, gypsisol, gleysol, argillisol, and protosol) and modifiers (e.g. vertic, calcic, gypsic) can be used to denote secondary features within the interval (Mack et al., 1993). The Mack et al. (1993) scheme departs from the USDA classification scheme for modern soils by acknowledging that ancient soils display unique attributes not recognized in modern environments. Some of these soil types are common to the geologic record including calcisols. In recent years paleosol deposits have been used as proxies to infer past climate conditions (e.g. Kraus, 1999; Kahmann and Driese, 2008; Sheldon and Tabor, 2009) and are of particular relevance to Carboniferous paleo-climate models (e.g. Cecil, 1990; Rankey, 1997; Miller and Eriksson, 1999). The USDA Soil Survey (1975) was used as a guide for modern occurrences of interpreted soil observed within the study interval.
Geophysical Logs

Due to the lack of quality geophysical log data for the SCSC cores, geophysical logs including Gamma Ray, Neutron Porosity, and Bulk Density were of limited value within the context of this study. Geophysical logs were used to pick the
top of the Michigan Formation which was used as a datum for the cross-sections. The
top of the Michigan Formation was picked at the top of shale or a high density
evaporite (anhydrite). Shale in the Michigan Formation typically has a high gamma-
ray log response while anhydrite exhibits a low gamma ray response with a high
(RHOB) density reading ~2.9 g/cm³. Geophysical logs are commonly used in
subsurface mapping, but were of limited utility in this study due to cased-hole
logging required in most wells in these shallow intervals (due to regulatory protection
of underground sources of drinking water) and the small scale, heterolithic nature of
Upper Paleozoic strata in the study interval in general.

Biostratigraphic Correlation

Palynomorphs (generic term for pollen and spores) are commonly used for
regional biostratigraphic correlation. Paleopalynology, the study of ancient spore and
pollen taxa, is an important tool for correlating and constraining strata to regionally
correlative biozones (Playford and Dino, 2005) and represents the most effective
method to delineate the Mississippian-Pennsylvanian (Sub-Absaroka) boundary
within the Appalachian basin (Ettensohn and Peppers, 1979) and elsewhere in the
midcontinent. Palynologic analysis from the Pennsylvanian is especially effective in
correlating and constraining strata within the Pennsylvanian midcontinent (e.g.
Peppers, 1996). The study of ancient pollen and spore assemblages (palynology) is
widely applied to Pennsylvanian coal and carbonaceous deposits which contain
highly concentrated populations of plant microfossils and are used for local, regional,
and global palynostratigraphic correlation (Peppers, 1996; Playford and Dino, 2005).
Accordingly, the Pennsylvanian of the Eastern and Western Interior United States is
relatively well constrained due to the dominance of terrestrial and terrestrially influenced environments and an economic incentive to study such coal-bearing deposits (Peppers, 1996). Unfortunately the time-extensive (~15ma) Chesterian North American stage is poorly constrained. However, the stage-boundary separating Chesterian (Late Mississippian) and Morrowan (Early Pennsylvanian) North American Stages can be relatively easily discriminated on the basis of dominant spore types (e.g. Ettensohn and Peppers, 1979; Eble et al., 2009). The North American Regional Stage nomenclature is used for most deposits of the North American Midcontinent and likewise will be used in this study and the age of the deposits in question will be reported using North American stratigraphic stage nomenclature (e.g. Heckel and Clayton, 2005) (e.g. Fig. 1). Occasionally, Western European and global stages will be referred to due to their prevalent use in the scientific literature.

Mississippian (Chesterian North American Stage) miospore assemblages are poorly constrained within the Chesterian Stage due to lack of species and taxonomic diversity and the dominance of shallow marine carbonates present in many areas of the Midcontinent that contain few (palynomorphs) (Playford and Dino, 2005; Eble et al., 2009). Late Mississippian (Chesterian) assemblages have been distinguished in the Eastern Appalachian basin where the terrestrially influenced Mauch Chunk/Pennington clastic wedge thins into the carbonate dominated western Appalachian basin (Ettensohn and Peppers, 1979; Ettensohn, 2004). Peppers (1996) correlation scheme aids in the determination of Pennsylvanian strata, however, the correlation of known North American Mississippian microflora to Western European biozones established by Clayton et al. (1977) is complicated by the lack of a similar scheme for assemblages of the Eastern United States (Eble et al., 2009). The result is a poor stratigraphic resolution of strata deposited within the Chesterian regional stage
Core Sampling

Approximately 100 g of primarily organic rich, carbonaceous strata, was enclosed in plastic bags and sent to Global Geo Labs ltd. for processing and analysis by palynologists Dr. Robert Ravn (IRF Consulting Group). Additional samples were sent to Dr. Cortland Eble (Kentucky Geological Survey) for processing and analysis. Micro-fossils are extracted using a mixture of primarily Hydrofluoric (HF) and Hydrochloric acids (HCl) to dissolve away the fine grained SiO_2 and CaCO_3 matrix. A residue of concentrated pollen and spores specimens are placed on a glass slide for petrographic examination through transmitted light microscope. A thorough review of the methods used in extracting palynomorphs is noted by Ravn (1979). Table 2 shows the core and depths at which each sample was taken.
Table 2. An inventory of strata sampled for biostratigraphic analysis. The majority of the analysis was performed by Dr. Robert Ravn, IRF Consulting Group. Asterisk (*) marks samples which were interpreted by Dr. Cortland Eble, Kentucky Geological Survey.

<table>
<thead>
<tr>
<th>Core</th>
<th>Depth (ft.)</th>
<th>Lithostratigraphic Unit</th>
<th>Lithology (Rock Type)</th>
<th>Miospore Recovery</th>
</tr>
</thead>
<tbody>
<tr>
<td>SB16</td>
<td>618</td>
<td>Michigan/Marshall?</td>
<td>Shale</td>
<td>Y</td>
</tr>
<tr>
<td>SB16</td>
<td>598.5</td>
<td>Michigan FM</td>
<td>Interbedded sand/mud</td>
<td>Y</td>
</tr>
<tr>
<td>SB16</td>
<td>542</td>
<td>Michigan FM</td>
<td>Evaporitic Shale</td>
<td>Y</td>
</tr>
<tr>
<td>SB16</td>
<td>498</td>
<td>Bayport</td>
<td>Mudstone</td>
<td>N</td>
</tr>
<tr>
<td>SB16</td>
<td>447.75</td>
<td>Bayport</td>
<td>mudstone</td>
<td>N</td>
</tr>
<tr>
<td>SB16</td>
<td>429.5</td>
<td>Bayport</td>
<td>Interbedded sand/silt</td>
<td>N</td>
</tr>
<tr>
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<td>408.9</td>
<td>Saginaw</td>
<td>Siltstone/mudstone</td>
<td>Y</td>
</tr>
<tr>
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<td>Saginaw</td>
<td>Siltstone/mudstone</td>
<td>Y</td>
</tr>
<tr>
<td>SB16</td>
<td>228</td>
<td>Saginaw</td>
<td>Siltstone/mudstone</td>
<td>Y</td>
</tr>
<tr>
<td>SB15</td>
<td>128</td>
<td>Saginaw/Grand River</td>
<td>mudstone</td>
<td>Y</td>
</tr>
<tr>
<td>SL418</td>
<td>825</td>
<td>Bayport</td>
<td>Shale</td>
<td>Y</td>
</tr>
<tr>
<td>SL418</td>
<td>810</td>
<td>Bayport</td>
<td>Shale</td>
<td>Y</td>
</tr>
<tr>
<td>SB15*</td>
<td>238</td>
<td>Michigan FM</td>
<td>Shale</td>
<td>Y</td>
</tr>
<tr>
<td>SB15*</td>
<td>128</td>
<td>Saginaw/Grand River</td>
<td>Mudstone</td>
<td>Y</td>
</tr>
<tr>
<td>SB18*</td>
<td>154.5</td>
<td>??</td>
<td>Mudstone</td>
<td>N</td>
</tr>
<tr>
<td>SB18*</td>
<td>138</td>
<td>Saginaw/Grand River</td>
<td>Coal</td>
<td>Y</td>
</tr>
<tr>
<td>SL423*</td>
<td>820</td>
<td>Bayport</td>
<td>Shale</td>
<td>Y</td>
</tr>
<tr>
<td>SL217A*</td>
<td>1285.3</td>
<td>Michigan/Marshall?</td>
<td>Shale</td>
<td>Y</td>
</tr>
</tbody>
</table>

Sequence Stratigraphy and Stratigraphic Hierarchy

Sloss et al. (1949) was first to refer to a “sequence” as a package of rocks separated by unconformities. Sloss (1963) applied the concept to craton-wide depositional “sequences”. The concept of the “sequence” was furthered by the advent of seismic stratigraphy (i.e. Payton, 1977) with major contributions from Vail and Mitchum (1977), Mitchum et al. (1977) and Vail et al., (1977a, b). Sloss-sequences are now referred to as “megasequences” and “sequences” are reserved for strata
bounded by chronostratigraphic surfaces that can be resolved using a combination of biostratigraphy and seismic data on the scale of 1-10 million years (Vail et al., 1977b). This culminated in the construction of coastal onlap curves by tracing seismically resolvable stratal surfaces with the key assumption that resolvable seismic surfaces are chronostratigraphically significant and constitute correlative “time-lines” (Vail et al., 1977b).

Sequence Stratigraphy is a tool which integrates many sub-disciplines relating to sedimentology and stratigraphy and is used to analyze sedimentary basin-fill, assess the controls on sedimentary processes, and for better prediction of reservoir architecture in petroleum exploration (Catuneanu and others, 2009). Modern sequence stratigraphy utilizes multiple datasets (e.g. facies relationships, modern environments, biostratigraphic analysis, seismic data, and other relevant information) as a method to interpret and predict the distribution of rock strata. Inherent to sequence-stratigraphy is the observation and interpretation of facies architecture (i.e. stacking patterns) and stratigraphic surfaces that may have chronostratigraphic significance (Catuneanu, 2006). The evaluation of unconformities, sharp contact surfaces separating distinct “depositional sequences,” is a fundamental objective of this study.

Mitchum et al. (1977) defines a sequence as “a relatively conformable succession of strata bounded by unconformities or their correlative conformity.” Over the last 30 years a number of definitions regarding where to place sequence boundaries has resulted in a great deal of confusion regarding the utilization of concepts relating to sequence-stratigraphy. Embry et al. (2007) contends that the recognition of key surfaces that separate stratigraphic packages should be based on empirical data from core, outcrop, seismic, and well-logs and should not be bound to
a particular model.

The analysis of high resolution datasets such as core, outcrop, and wire-line logs have resulted in the recognition of parasequences, defined as “discrete shorter duration shallowing upward cycles bounded by a marine flooding surface or its correlative conformity” (Van Wagoner et al., 1990). Parasequences are on the order of meter thickness compared to thicker depositional sequences which may be hundreds of meters thick (Embry, 2009). Parasequence boundaries may be marked by either a flooding or erosional surface or both. These boundaries are found within the overall genetic package and should not be confused with sequence boundaries which separate unconformable facies successions (Van Wagoner, 1990). Van Wagoner et al. (1990) notes that parasequences may not be identifiable in fluvial environments due to the absence of a discernible sea-level signal. However, Greb and Martino (2005) interpret Pennsylvanian-aged Appalachian basin flooding events, on the basis of estuarine facies assemblage interbedded with an otherwise fluvially dominated environment.

Since the Van Wagoner (1990) definition the origin and identification of parasequences in the rock record is debated. Embry (2009) noted that the identification of parasequences relies on the interpretive discrimination of the Flooding Surface (FS), Maximum Flooding Surface (MFS), and Maximum Regressive Surface (MRS), which are all ill-defined and subjective. Embry (2009) argues that parasequences as defined by Van Wagoner are actually lithostratigraphic surfaces and considers the flooding surface (FS) as representing a change in lithology, and may not necessarily be a time-stratigraphic unit as defined within the realm of sequence stratigraphy. Further, Embry (2009) redefines the parasequence as discrete depositional packages bounded by a maximum regressive surface (MRS).
Important differences exist between the behaviors of clastic and carbonate systems. In shallow marine systems, fluctuations in relative sea-level may have a profound effect on the type of strata deposited. In general, carbonate systems are most productive during relative sea-level high-stands, while siliciclastic dominated strata commonly denote periods of low-stand (Van Wagoner et al., 1990; Read, 1995). The application of sequence-stratigraphic analysis to marginal-marine low-stand systems led to the recognition of *incised valleys* which can often be traced to a correlative overbank paleosol (Van Wagoner et al., 1990). Unconformities record a period of truncation and non-deposition and indicate episodes of sub-aerial or marine erosion and are often manifested as either a soil horizon or a sharp erosional surface (Van Wagoner et al., 1990).

Posamentier and James (1993) argue that sequence stratigraphy should be viewed as a "tool rather than as a template" and one should not be bound by existing models, which may not adequately explain every geologic setting. In recent years there is an ongoing attempt to streamline the methods of sequence stratigraphy and to move away from a model driven approach to one rooted in empiricism and direct observation (i.e. Embry, 2007, Catuneanu and other, 2009; Miall, 2010b). Many aspects of sequence stratigraphy are model driven (inductive) and are based on theoretical relationships rather than direct observation (deductive) (Embry et al., 2007; Miall, 2010b). Furthermore, sequence stratigraphic jargon has expanded rapidly since the sub-discipline was established in AAPG memoir 26 (Payton, 1977).

In accordance with the above theoretical principles of sequence stratigraphy, this study will be observation driven incorporating principles of facies analysis, dispositional environment interpretations, paleogeographic setting, and biostratigraphy, with the goal of predicting and interpreting the chronostratigraphic
relationships of the rock strata encountered in this study (especially in relation to the Mississippian-Pennsylvanian boundary). The principles laid out in the preceding discussion will be applied when interpreting the stratigraphic relationships between facies within an overall succession and the stratigraphic surfaces that serve as bounding discontinuities or sequence boundaries.

Data Limitations

The lack of cores and other high resolution geological data from within the Michigan basin makes it difficult to accurately determine the true lateral extent of the stratigraphic units. This is despite a number of studies utilizing geophysical log data (e.g. Lilienthal, 1978; Vugrinovich, 1984; Westjohn and Weaver, 1998). Due to the lack of sufficient log and core control in a number of locations, wireline logs give limited insights regarding the nature of deposition. The use of geophysical logs is restricted to the central Michigan basin due to poor availability of modern open-hole logs along the basin periphery. The units of interest in this study are present within the shallow subsurface in this area and have not been significant exploration targets for oil and gas. Furthermore, the presence of potable water requires the use of a steel casing to protect existing water quality within aquifer units which acts to dampen the geophysical signal (Vugrinovich, 1984). Due to the previously noted reasons, a direct evaluation of available core material is necessary to meet the broad objectives of the study.

The discontinuous nature of most Carboniferous strata in the Michigan basin makes lateral correlation of observed sequences tenuous. This may be partially due to the up-depositional dip location of the Michigan basin, which may have resulted in a
truncation of the low-stand system tract (LST) (Smith and Read, 2001) and in some cases the entire sequence. Cohee (1979) and Harrell et al. (1991) suggest low tectonic subsidence rates within the Michigan basin may have resulted in a negligible amount of accommodation space leading to periods of erosion and non-deposition. This study aims to test the current understanding of the Late Mississippian Michigan basin and possibly refine the existing Michigan basin lithostratigraphic and chronostratigraphic nomenclature. However, limited chronostratigraphic and lithologic control is present to definitively support previous interpretations.
CHAPTER III

LITHOFACIES DESCRIPTION AND PROCESS INTERPRETATION

Introduction

The main focus of this study is core material (e.g. SB18, SB16, SB15, and SB13) comprising of (in vertically ascending order): the Marshall Sandstone, fine-grained shale and gypsum deposits of the Michigan Formation, a mixture of carbonate and siliciclastic lithologies associated with the Bayport Limestone, and carbonaceous deposits of the Saginaw Formation. Bayport strata consist of a mixture of limestone, dolomite, calcareous sandstones and mudstone deposits. While not the direct subject of this study dominant Marshall Sandstone and Michigan Formation lithofacies have been noted.

Four of the eight SCSC cores capture a continuous stratigraphic section from a sandstone unit below the base of the Michigan Formation up into the Pennsylvanian Saginaw Formation. The base of these cores is locally fossiliferous, fine-to-medium-grained sandstone unit interpreted to be the upper Marshall Sandstone (Napoleon Member?) based on stratigraphic position (i.e. below the Michigan Formation). In vertical succession shale and/or dolomitic strata overlie the Marshall and are gradational upwards into the heterolithic-gypsiferous Michigan Formation. The Michigan Formation is overlain by mixed siliciclastic-carbonate strata of the Bayport Limestone. The Mississippian-Pennsylvanian boundary in the Michigan basin has traditionally been placed at the lithologic contact between the Bayport Limestone and the Parma Sandstone (Winchell, 1861; Lane, 1909; Newcombe, 1933; Kelly, 1936; Vugrinovich, 1984). While these investigations have noted the Bayport Limestone and the Parma Sandstone, no detailed documentation of texture, structures, bedding
style and vertical stacking patterns have been undertaken for the entire interval. Likewise the transition from the carbonate dominated Bayport and the overlying Pennsylvanian aged Saginaw Formation has not been thoroughly documented in either core or outcrop. The analysis and interpretation of rock units found within the Michigan Formation is considered an important step in evaluating the stratigraphy of the Late Mississippian and Early Pennsylvanian Michigan basin. The lithofacies succession presented in the following discussion approximately and generally proceeds from deepest to shallowest occurrence of these units.

**Upper Marshall and Michigan Formations**

The Michigan Formation is described in a number of Michigan basin technical reports and papers (e.g. Vugrinovich, 1984; Harrell et al., 1991; Lilienthal, 1978; Cohee, 1979; Ells, 1979) and is notable for economically significant gypsum/evaporite deposits. The Michigan Formation serves as a reference for stratigraphic position due to its apparent basin-wide distribution. Key beds that have been correlated include the “Brown Lime” and the “Triple Gypsum” (e.g. Lilienthal, 1978; Harrell et al., 1991). Fine-grained shale intervals within Michigan Formation are host to Chesterian miospores documented later in this report (Chapter 5). In most recent stratigraphic literature (Harrell et al., 1991; Westjohn and Weaver, 1998) the Michigan Formation is traditionally divided into two principal units which include: (1) the informal Michigan "Stray" sandstone unit referred to by Westjohn and Weaver (1998) as the “Upper Marshall” and (2) a heterolithic interval composed predominantly of shale with lesser amounts of carbonate, sandstone, and evaporite
units (Moser, 1963). The depositional and stratigraphic nature of “Stray” sandstone is in doubt with some workers suggesting correlation to the upper Marshall (Napoleon Member) of the southern Michigan basin (Moser, 1963; Harrell et al., 1991). The “Stray” constitutes significant natural gas accumulations, some of which were converted into significant natural gas storage reservoirs located in Montcalm and Mecosta counties.

Marshall Sandstone

General Description: The Marshall is a gray, fine-to-medium-grained, moderately well sorted, locally bioclastic, micaceous quartz sandstone and is encountered at the base of the four continuous SCSC cores (SB18, SB16, SB15, and SB13). The sandstone displays discrete intervals of crinoid/echinoderm and mollusk bioclasts toward the base of the observed core interval. These beds are often overlain by cross-bedded to planar-laminated quartz sandstone overlying the fossil beds. Toward the top of the interval the sandstone becomes progressively devoid of fossil material and exhibits a pronounced coarsening upward grain-size distribution which is capped by a pronounced zone of mottling and green-clay accumulation. A dolomitic bed commonly overlies this sandstone in three of the four cores observed consistent with observations by Moser (1963) whose study of the Michigan Formation was based on the analysis borehole-cuttings and geophysical logs. While carbonate lithologies have been observed within the interval, they are most commonly confined to the lower part of the Michigan Formation, generally lack observable biota, and rarely exceed 10 ft. in thickness.
*Preliminary Interpretation: Shoreface.* Fossil-beds (Fig. 12a) comprised of marine fauna (e.g. crinoids, bryozoans, and mollusks) reflect periods of high-energy conditions associated with a storm-dominated environment. The presence of ripple-cross laminations (HCS) (Fig. 12b) further suggests a wave dominated environment. The presence of HCS indicates mixed-energy; oscillatory flow conditions associated with environments at or below normal wave base (Duke et al., 1991). The bedforms displayed in Figure 12 suggest periods of combined fluid-flow conditions (sensu Harms, 1975). A generally coarsening upward grain-size distribution, a feature widely acknowledged from the geophysical log response (gamma ray) further indicates a prograding marine shore-face succession. Burrow mottled sandstone (Fig. 12d) is common at the top of the Marshall Sandstone and is interpreted to reflect deposition below storm-wave base and possibly indicate deepening. The overlying heterolithic, generally fine-grained, Michigan Formation commonly caps the Marshall sandstone and may mark the transition from shoreface to backshore environments. Toward the top of the formation sandstone is frequently interbedded with shale indicating a gradational contact relationship between the Marshall and the Michigan Formation in the southern Michigan basin. Fine-grained carbonate (dolomitic) units (Fig. 13a) occur most prominently at the base of the Michigan lithofacies, near this Marshall-Michigan transition.

**Michigan Formation**

*General Description:* The Marshall Sandstone is transitional from fossiliferous, fine-to coarse-grained sandstone, which grades into a heterolithic fine-grained interbedded series of shale, sand, and evaporites. The heterolithic facies of
the Michigan Formation are composed of fine-grained lenticular-bedded sandy shale, shaley sandstone, and evaporitic shale. Contorted bedding occurs throughout the interval. Occasional massive, discontinuous fine-to-medium-grained sandstone beds are present throughout the Michigan interval. The Michigan Formation displays all the variants of heterolithic bedding documented elsewhere by Reineck and Singh (1980). Flaser bedding is commonly observed where the sand-sized fraction is greater than the silt-sized component. Pyrite nodules occur commonly throughout the Michigan Formation. Evaporites including gypsum are most-commonly associated with fine-grained shale and commonly fill fractures occurring within sandstone intervals. Gypsum most commonly occurs as nodular growth forms within a fine-grained mudstone and shale matrix. No gypsum was reported in the SB13 core where the unit pinches out in the southern margin of the basin. SB16, SB18, and SB15 contain significant gypsum beds that occur in two primary locations: (1) towards the middle of the Michigan lithofacies, and (2) towards the top of the Michigan interval, directly underlying the contact between the Michigan Formation and mixed carbonate/siliciclastics of the Bayport. The middle occurrence of evaporite in the SB16 core is the thickest evaporate interval reaching ~20 ft. in thickness.

_Preliminary Interpretation: Restricted back-barrier/estuarine._ Dolomite beds devoid of fossils toward the base of the succession suggest a restricted-marine affinity. Most of the carbonate units lack observable macro-fossils possibly due to diagenetic alteration (e.g. dolomitization and dissolution). A rare brachiopod imprint is preserved in SB16 at a depth of 614 ft. directly overlying the basal Marshall Sandstone. Dolomitic units commonly display brecciation and are capped by shale reflecting a transition into lithofacies most characteristic of the Michigan Formation (Fig. 13b,c). The bulk of the Michigan Formation is dominantly shale, containing
subordinate amounts of sandstone and gypsum. Much of the formation is composed of interbedded siltstone and sandstone. Bedding style varies relative to sand and mud content from flaser bedding (fine sand), wavy (equal parts sand/silt) and lensoidal (predominantly silt size fraction) bedforms (e.g. Reineck and Singh, 1980). Heterolithic bedding is commonly associated with tide-dominated environments, including estuaries and restricted lagoons (Reineck and Singh, 1980; Dalrymple, 1992, 2010c). A restricted trace fossil assemblage is suggested from the lack of total obliteration and homogenization of the primary depositional fabric. Bioturbation is locally evident in the heterolithic facies (Fig. 13c) and are separate from syneresis cracks, which form due to fluctuations in salinity (Fig. 13c) (Shinn, 1983a). Intervals displaying low-diversity traces support a hypersaline or brackish environment of deposition not suitable for a diversity of organisms. The presence of evaporites and mudstone toward the middle and upper portions of the Michigan Formation suggests a restricted environment of deposition where a net loss of water resulted from evaporation, concentration of dissolved solids, and the eventual precipitation of sulfate minerals including Gypsum (CaSO₄*2H₂O) and anhydrite (CaSO₄) (Warren and Kendall, 1985; Kendall, 2010). Chemical precipitates in the Michigan Formation were noted by Lane (1909) who inferred a low rate of mechanical deposition and a high degree of "chemical wasting" for the Michigan Formation. The presence of distinct coarsening upward sand beds, scour and fill structures, and bedforms similar to the upper Marshall suggests lateral continuity with the shoreline environment in the lower portions of the Michigan Formation. The interbedded nature of cross-bedded fine-grained sandstone, carbonate beds, lenticular beds, a low-diversity ichnofacies, and displacive gypsum deposits suggest a tide-dominated, restricted back-barrier environment subject to intermittent storm reworking, sandstone
deposition, punctuated by periods of intensely evaporative conditions resulting in the precipitation of gypsum.

Cohee (1979) interprets the Michigan Formation as “restricted” and suggests it was "cutoff" from the neighboring Illinois or Appalachian basins during the time of deposition, which was previously interpreted to have been during the Meramecian regional stage (Harrell et al., 1991). However, biostratigraphic analysis of pollen and spore samples taken from the unit firmly indicates a Chesterian age for the formation (next chapter). The basinal thickening of the Michigan Formation is noted in the observed core material. The Michigan Formation reaches a maximum thickness of 150 ft. in the SB16 core (most basinward SCSC core) to <20 ft. of primarily fine-grained material in the SB13 (most marginal) core. A southward thinning of the Michigan formation is noted by Cohee (1979) and Harrell et al. (1991). The analysis of geophysical logs indicates the Michigan formation reaches as much as 500 ft. in thickness in the central Michigan basin (Missaukee County) (Lilienthal, 1978; Harrell et al., 1991), a possible response to ongoing basin subsidence during the time of deposition.
Figure 12. Marshall Sandstone lithofacies. (A) Skeletal (predominantly mollusk) deposit in fine-grained sandstone. (B) Ripple cross lamination (HCS?) displaying differential bed-dip directions punctuated by sharp scour surfaces. (C) Red sandstone (bottom) overlain by ripple cross laminations. (D) Burrow mottled sandstone present at the top of the marine sandstone (Marshall Sandstone).
Figure 13. Michigan Formation lithofacies. (A) Fractured dolomite unit capped by shale. The dolomite is notably present at the base of the Michigan Formation. (B) Heterolithic bedding. Note the bidirectional (herringbone?) cross stratification in the sandstone lens at the center of the photograph. (C) Heterolithic bedding. Note the syneresis cracks and isolated mollusk burrow. Trace-fossil (*Teichichnus*) in response to sedimentation. The low-diversity ichnofacies present in the Michigan Formation rarely homogenizes the primary heterolithic texture. (D) Displacive gypsum within a mudstone matrix. Such enterolithic bedforms are interpreted to have formed in an arid supratidal clastic-dominated mudflat (Cohee, 1979).

Michigan-Bayport Contact

The top of the Michigan Formation is marked by either a shale or evaporitic (gypsum/anhydrite) (Fig. 14) which is variably overlain by a carbonate, sandstone, or fine-grained siltstone/shale. The contact between the Michigan Formation and the Bayport was observed in four SCSC is consistent with Vugrinovich (1984). The transition from the heterolithic Michigan Formation to the mixed-clastic, carbonate succession of the Bayport interval is observed to be sharp (Fig. 14). In SB16, the Michigan-Bayport transition occurs at interbedded gypsum/heterolithic shale and an overlying carbonate (i.e. laminated dolomite or grainstone). The contact observed within the SB18 core occurs as red/orange stained coarse-grained sandstone directly overlying a bed of gypsum displacing a brown mudstone and contains numerous gypsum filled fractures which grade into a series of carbonate, sandstone, and mudstone beds (Fig. 14). In the SB15 core, planar bedded shale underlain by gypsum is sharply overlain by an iron stained, and oxidized coarse-grained sandstone which grades into a sandy/grainy carbonate-prone Bayport interval. The SB13 core is
located up-dip and contains a thin (<20 ft.) Michigan Formation displaying brecciated zones, possibly due to the dissolution of evaporites.

Figure 14. SB18 (340-320 ft.) core box displaying transition from a mixed/evaporite lithology grading into the carbonate prone Bayport interval. Gypsum displays enterolithic growth nodules which are interpreted as being formed in an arid back-barrier evaporitic siliciclastic mudflat environment. Bottom to top corresponds to the bottom-left to top-right of the core-box. Scale black bar ~6 in.
The Bayport Lithofacies

The lithologies that comprise the upper portion of the Mississippian section include a series of interbedded carbonate, sandstone, and mudstone lithologies referred by previous studies as the Bayport Limestone and the "Parma" (e.g. Vugrinovich, 1984). The danger in using the term “Parma” is derived from the fact that it is unknown what Winchell (1861) was actually referring to when he named the “Parma” from a small quarry located near the town of the same name. For reasons of clarity, the term Bayport will be used to refer to a mixed carbonate/clastic interval overlying the Michigan Formation including the Bayport Limestone. However, it must be noted Bayport is not exclusively composed of carbonates and includes significant amounts of quartz sandstone ranging on the scale of inches to 10s of feet in thickness and in most cores quartz sandstone forms the dominant lithology in terms of aggregate footage within the Bayport interval. Carbonates within the Bayport are composed of two primary facies (Lasemii, 1986) which have been interpreted by Bacon (1971) and Ciner (1988) from quarries of the Bayport. The lower portion of the unit comprises of a carbonate dominated interval which is typically overlain by well sorted, locally CaCO_3 cemented sandstone (Parma Sandstone?) which is consistently present in all cores investigated in this study. This stratigraphic relationship was observed by Vugrinovich (1984) in the central basin. Locally, this sandstone (~50-70 ft.) fines upward into a mudstone/paleosol before transitioning into a white, buff bioclastic limestone. The rock units within the Bayport display a range of textures and structures similar to what is reported from the Chesterian deposits Illinois basin (i.e. Bethel to Glen Dean interval; Smith and Read, 2001; Nelson et al., 2002). No comprehensive documentation of facies-type exists for the Bayport Interval. Based on detailed study of cores (specifically SB13, SB15, SB16, and SB18)
the Bayport is subdivided into seven distinct lithofacies, each described in the sections that follow.

Facies B1: Bioclastic Wackestone to Packstone

*Description:* In hand specimen Facies B1 is a white to dark gray (N9-N3) mottled to structure less fine to medium grained, bioclastic grainstone/limestone (Fig. 15a-c). The facies is most commonly well-developed/preserved in the lower-middle and upper portion of the Bayport interval. The rock fabric is composed of a poorly-sorted assortment of disarticulated skeletal bioclasts (Fig 16a-d) displaying preferred orientation along bedding (Fig. 16a), but are often found randomly orientated. Skeletal material is almost always replaced and filled with blocky calcite spar, a common occurrence in other examples of Carboniferous limestone (Wilson, 1975). Calcite cement is concentrated along grain-to-grain contacts. Peloids are common, some of which resemble bioclastic grain shapes are interpreted as "ghost" textures (e.g. Scholle and Ulmer-Scholle, 2003). Grain-packing locally ranges from wackestone to grainstone according to the Dunham (1962) classification. Silt to sand-sized bioclastic grains are composed of angular, disarticulated fossil fragments of various marine organisms including bryozoans, crinoids, ostracods, gastropods, and foraminifera. Locally, samples exhibit a calcite-spar matrix resulting in a biosparite texture. Cementation is especially prominent in the upper limestone bed encountered within the SL418, SL423, and SB18 cores. Commonly the matrix consists of micrite mud (Fig. 16a,b) and peloids (Fig. 16d). Silt-sized quartz grains are commonly disseminated throughout the matrix, also reported by Ciner (1988). Calcite (dog-tooth spar) locally fills vugs and fractures. Rugose corals occur sporadically in core
material and are observed in a continuous bed at the Wallace Stone Quarry. Pressure-solution seams (stylolites) are commonly found in this facies. In vertical section, this facies grades to and from dolomitic Facies B2. The limestone-dolomite contact occasionally appears as a series of fractures filled with green-clay.

**Interpretation: Subtidal shoal or tidal channel.** Packstone and grainstones indicate a relatively high-energy environment capable of winnowing interstitial mud due to high tidal and/or storm energy. A high energy depositional environment is also indicated by the disarticulated and abraded nature of the fossil fragments. The diversity of marine fauna including bryozoans, echinoderms, and foraminifera supports an open marine salinity (~35 ppt) (Scholle and Ulmer-Scholle, 2003; Pratt, 2010). Crinoids are particularly sensitive to changes in marine salinity and are usually absent when deviation exceeds 2 ppt from open-marine conditions (Scholle and Ulmer-Scholle, 2003). Isolated occurrences of rugose coral further indicate a marine environment of deposition. Rugose corals are widely present in Paleozoic limestone and are composed primarily of low-mg calcite which aids in the preservation of the skeletal structure (Scholle and Ulmer-Scholle, 2003). The lack of primary bedding structures is attributed to bioturbation. Figure 15a displays a burrow-mottled texture. Burrowing in subtidal marine zones exhibits a diverse behavior (MacEachern et al., 2010). In modern offshore zones, biogenic structures produced by burrowing organisms (e.g. *Callianassia sp.*) commonly result in the obliteration of primary bedding structure (Shinn et al., 1969). Similarly, trace-fossils are interpreted as being part of the *Cruziana* trace-fossil assemblage, which displays a predominantly horizontal to elongate shaped burrow path supporting a sub-tidal environment of deposition between fair weather wave base and storm weather wave base (MacEachern et al., 2010). Evidence of exposure is common upsection in vertically
adjacent, Facies B2, which is characterized by green crusts, mudcracks, intraclast breccias, and soils. Based on the lack of recognizable exposure surfaces present within this facies and the presence of an open marine biota this facies is interpreted as subtidal deposits (below mean low-tide) (sensu Shinn, 1983a) (Fig. 22a).

Figure 16a-d displays a number of microfacies types and best correspond to Wilson’s (1975) microfacies 9 and 10 based on the variable wackestone to grainstone texture (often observed within the same thin-section). This microfacies type(s) is associated with circulated, open-marine conditions at or below normal wave-base (Flugel, 2010). In modern environments carbonate accumulations displaying packstone-grainstone textures have been documented in a variety of bathymetric settings, including platform interior homoclinal carbonate ramps of the Persian Gulf and Shark Bay (Read, 1985; Purser and Seibold, 1973; Ahr, 1973). The stratigraphic position relative to the dolomitic Facies B2, supports a deeper, subtidal environment of deposition near wave-base as part of a shoaling upward facies patterns broadly described by James (1979) and Shinn (1983a) (Fig. 22b).
Figure 15. Facies B1 hand specimens. (A) Mottled bioclastic packstone, displaying a burrow-mottled texture attributed to burrowing of marine organisms (Callianassia) in the subtidal zone. (B) Wispy, mottled bioclastic packstone/grainstone. (C) Randomly distributed ostracod bioclasts within a micritic matrix. (D) Stylolitic packstone-grainstone, devoid of primary bedding structure presumably due to bioturbation. Scale in cm.

Figure 16. Facies B1 microfacies. (A) Bioclastic grainstone composed of fine-sand sized bryozoan spines (By), echinoderms fragments, and foraminifera. (B) Bioturbated bioclastic packstone composed of brachiopods, foraminifera, mollusks, and echinoderm fragments in a silty mudstone matrix. Note that skeletal material has undergone textural inversion to calcite spar cement. (C) Wackestone-Packstone displaying a cross-sectional view of an echinoderm spine with sagittal sections of foraminifera (Fo) bioclasts scattered throughout the silt-sized peloidal matrix. (D) Bioclastic wackestone-packstone with abundant ostracod tests, foraminifera, and calcite-spar cement constituent. All samples in PPL.
Facies B2: Laminated Dolomite

Description: Facies B2 in the Bayport is a grayish orange to yellowish brown (10YR 7/4-10YR 4/2), laminated, vuggy, argillaceous dolomite that displays a notable lack of fossil material (i.e. bioclasts) (Fig. 17a,c). The matrix is composed of silt-sized dolomite rhombs with common sucrosic texture (Fig. 18b). Sand-sized quartz grains are commonly found distributed both randomly and along distinct laminar horizons. Laminations grade from horizontal to a disrupted/clotted texture. A localized brecciated fabric occurs commonly in conjunction with fracture filling green (10G 4/2) clay. Sand and mud-filled downward narrowing fissures (mud cracks?) (Fig. 19a) are found interbedded with thin beds of fine quartz sandstone and green-clay crusts (Fig. 17c). Brecciated intraclast-rich deposits (Fig. 19b) are common at the top of the lithofacies. Vuggy zones are widely observed in core material (Fig. 17c). Vugs are commonly spherical to elongate in shape, may be open or filled with calcite or anhydrite (Fig. 17d) and commonly occur along horizontal to slightly-inclined bedding planes. Local fractures, breccias, and crusts often filled with a greenish residuum are observed in the facies. Both bedded and nodular occurrences of chert are present. Bedded chert nodules are common within the laminated dolomite (Fig. 17a). The top of the facies occurs as either a brecciated or fractured dolomite sharply overlain by poorly sorted granule-pebble sized chert and carbonate intraclasts, which often mark the transition to Facies B4 and B5 (Fig. 19b).

Interpretation: Intertidal-supratidal (peritidal) tidal-flat complex. Microbial laminations, desiccation features, birdseye fenestrae, keystone vugs, and soil crusts are common in the intertidal to supratidal zone environment of carbonate depositional settings (Shinn, 1968; Shinn, 1983a). These features are common in Facies B2. Laminites are interpreted to of microbial origin and were most likely formed from
photosynthetic cyanobacteria and contain similar attributes to those described by Riding (2000). Microbial mats and laminations are most common in the supratidal environment (Fig. 22a) where desiccated mud-chips are reworked during storms resulting in "flat pebble" conglomerates (Shinn, 1983a) which is commonly observed within the Bayport (Fig. 19b). These features are common along with mudcracks throughout laminated, dolomitic units of the Bayport. Mudcracks and other desiccation features (Fig. 17a,d) are interpreted as the direct product of sub-aerial exposure, a common feature of the intertidal zone. Mudcracks occur as vertically orientated cavities which are subsequently filled with material from the overlying deposits (Shinn, 1983a). Elongated, ovoid vugs are interpreted as being fenestrae which are synonymous with birds-eyes, shrinkage pores, and loferites (Shinn, 1983b). Fenestrae are indicative of microbialite deposited in the supratidal zone of deposition (Fig. 22a) within carbonate tidal flats. Fenestrae occur as ovoid to spherical, 1-to-3 mm vugs filled with calcite and anhydrite which protects the geometry from subsequent burial compaction (Shinn, 1968; Shinn, 1983a,b). These structures are attributed to the decay of organic material resulting in degassing, which are later filled by calcite or anhydrite (Shinn, 1968; Shinn, 1983b). Keystone vugs, by contrast, are spherical cavities produced in association with the swash zone (Shinn, 1983a,b). Interbedded dolomite and silt to sand sized quartz grains indicate an environment proximal to a source of erogenous clastic material which may have been wind-blown or washed into the depositional environment.

Interbedded quartz sand within the Bayport was interpreted by Ciner (1988) as the product of eolian dune deposits analogous to sands deposited along the southeast coast of Qatar Peninsula in the Persian Gulf, reported by Shinn (1973). Along the Qatar Peninsula eolian sand is blown from the northwest (locally known as
the Shamal) into sheltered, carbonate dominated environments forming thin lenses of quartz sand, which are preserved as wedges between layers of carbonates due to changes in relative sea-level (Shinn, 1973). In the Bayport, Ciner (1988) notes well rounded sand-grains and angular silt-sized quartz, which support eolian deposition. However, the base of sandstone beds are often erosive (Fig. 23a,b) and contain carbonate lithoclasts disseminated throughout the basal portions of sandstone beds (Fig. 18c). Other surfaces display evidence of exposure at the base of sand-beds. These features are not explained exclusively by wind abrasion. While wind-blown sand may be present, the majority of the quartz sandstone was water-lain based on features and primary sedimentary structures discussed later.

Facies B2 in thin-section consists of laminated quartz silt/sand and carbonate mud corresponding with microfacies 21 (Fig. 18d) (i.e. Wilson, 1975; Flugel, 2010). Occurrences of non-laminated, sucrosic microtexture (Fig. 18b) support deposition in an environment subjected to hypersaline conditions (e.g. tidal ponds, Flugel, 2010). Laminations, birds-eye vugs, intraclasts, sand-filled mudcracks, green crusts, and brecciated horizons are good evidence of intertidal to supratidal zone of deposition consistent with processes occurring in modern tidal-flats of the Persian Gulf (Kendall and Skipwith, 1969), Andros Island, (Shinn, 1968; Shinn., 1983a,b) as well as other ancient stratigraphic examples (e.g. James, 1979; Shinn, 1983a), including previous interpretations of the Bayport Limestone by Bacon (1971) and Ciner (1988).

Chert within the Bayport occurs along bedding planes (Fig. 17a) and filling fractures (Fig. 21a). Knauth (1979) suggests that nodular chert forms in the meteoric-marine mixing zone commonly found in coastal systems where biogenic opal-A is readily dissolved in connate water undersaturated, with respect to calcite and aragonite. Similarly, chertification is widely noted in coeval Late Mississippian
Illinois basin deposits. Lasemi et al. (2003) interprets chert found in the Illinois basin as a product of regional upwelling of silica rich waters which coincided with a shallowing of the basin. Regional upwelling may account for the suggestion by Cohee (1979) that the “Bayport Sea” was “enriched with silica.” However, at this time it is difficult to determine the true nature of the chert without more detailed chemical analysis.

Sharp contact surfaces occur within the Bayport separating carbonate from the overlying siliciclastic dominated lithofacies (Facies B4-B6). This is well displayed in the two examples shown in Figure 23a,b. The contact surfaces are characterized by a sharp contact between carbonate and overlying quartz sandstone comprising of a poorly-sorted mixture of carbonate rip-ups clasts and pebble-sized chert clasts observed directly on top of limestone in a poorly sorted dolomitic/silty matrix. These sharp, erosional surfaces are interpreted as being deposited in a high energy environment of deposition characterized by scour and erosion. Carbonate and chert lithoclasts were likely derived from the underlying carbonate and display marine-fossils. These surfaces appear to be localized. The lack of fluvial facies may be attributed to reworking and erosion during the subsequent marine transgression, as observed in other environments in previous studies (Zaitlin et al., 1994; Smith and Read, 2001). A similar interpretation was applied to Chesterian deposits of the Illinois basin by Smith and Read (2001) and Nelson et al. (2002).
Figure 17. Facies B2 hand specimen.  

(C) Smooth laminations in dolomite displaying bedded chert.  

(B) Intraclasts in dolomite.  

(C) Dolomite within an intertidal and supratidal, tidal-flat succession sharply overlain by a green crust and intraclasts at the top. This feature is interpreted as an exposure surface.  

(D) Coarse-grained sandstone grading in laminated, vuggy and intraclastic dolostone displaying distinct laminations interpreted as microbial structures.  

(D) Interpreted soil crust, a product of sub-aerial exposure. Calcite-filled vugs are present at the bottom of the sample.  

Scale in cm.
Figure 18. Facies B2 microfacies. (A) Laminated (Lm) texture with silt-sized quartz grains distributed throughout a vuggy, sucrosic dolomite matrix. The clotted texture is possibly due to sub-aerial exposure. Vugs are denoted by blue-epoxy. (B) A close-up of the previous sample displaying a highly porous and permeable "sucrosic" dolomite texture. (C) Brecciated horizon composed of 1-5 mm carbonate intraclasts in silt to fine-grained sandstone matrix. (D) A domally laminated "crypt-algal" deposit.

Microbialites in the Rock Record

Microbialites, as described by Riding (2000), are most commonly classified on the basis of macrofabric and texture. Dolomitic laminites exhibiting flat, domal, and clotted textures have been attributed to cyanobacteria (Riding, 2000). Riding (2000) favors the term microbialite to describe a variety of ancient textures, referred to as “algal mats” in older literature. Cyanobacteria, previously referred to as blue-green algae, comprise a broad group of photosynthetic bacteria commonly found in shallow marine and lake sediments (Riding, 2000). Recent studies indicate blue-green algae are actually eukaryotes, not bacteria, while prokaryotic cyanobacteria are genetically similar to chloroplasts (Riding, 2000). Microbial processes have been widely recognized in modern sediments. Bathurst (1966) attributes micrite envelopes and boring structures to algae. Chafetz (1986) gives evidence to support bacterial carbonate precipitation in the formation of peloids and Robbins and Blackwelder (1992) observed that cyanobacteria play an important role in the precipitation of CaCO₃ in the water column resulting in the formation of whitings. It appears that the high evolutionary variability of cyanobacterial communities can result in their occurrence in most surficial Earth environments.

Cyanobacteria, as well as other species of bacteria, operate through a number of processes including substrate secretion, the trapping and binding of grains, and
carbonate precipitation (Riding, 2000). Cyanobacteria often produce a mucilage or
sheath, referred to as an extracellular polymeric substance (EPS), which provides a
substrate for microbial mat growth and enables the trapping of fine to coarse grained
particles (Riding, 2000; Andres and Reid, 2006). The precipitation of carbonate can
result in the calcification of both the cyanobacteria and mucilage (Riding, 2000).
This process of calcification is governed by environmental and biological factors
including ocean water chemistry and the creation of alkaline conditions associated
with uptake of \( \text{CO}_2 \) and \( \text{HCO}_3^- \) during photosynthesis (Pentecost, 2000).

Stromatolites became subordinate to filter-feeding and shelled organisms
during the Early-Paleozoic (Riding, 2000; Scholle and Ulmer-Scholle, 2003). Modern
examples of stromatolite communities have been observed in the highly saline Shark
Bay Western Australia and off the coast of Lee Stocking Island, Bahamas, where tidal
energies are exceptionally high and ecological competition is scarce (Riding, 2000).
Ginsburg (1955) reported that cyanobacteria are the primary stromatolite forming
organism occupying the tidal flats of Andros Island, Bahamas. Logan (1961) uses
sediment trapping cyanobacteria to explain the formation of stromatolites in Shark
Bay. Andres and Reid (2006) promote the hypothesis that sedimentation rate is the
dominant factor controlling stromatolite growth near Highborne Cay, Bahamas.
Riding (2000) contends that calcification and cementation potential is the overarching
control on the preservation of stromatolites/microbialites in the geologic record.
Based on modern occurrences, microbialites favor a stressed hypersaline or high
energy environment of deposition (e.g. Shark Bay, the Khor al Bazam, and Lee
Stocking Island), though this may have varied through geologic time due to
environmental controls (Riding, 2000).

Kendall and Skipwith (1969) observe modern laminated “algal mats” in the
Khor al Bazam saline lagoon, Abu Dhabi, and observe distinct textural belts which grade, in a landward direction, from smooth laminations overlying quartzose sand and anhydrite, a crinkled and disrupted zone, a polygonal zone defined by desiccation cracks, and the cinder zone consisting of a black pustular texture. Kendall and Skipwith (1969) conclude that the growth and distribution of these “algal” structures are defined by the length of subaerial exposure and salinity of the water.

**Microbialites in the Bayport**

The examples presented as part of Facies B2 are consistent with both ancient and modern trapping and binding microbialite behavior. Cyanobacterial lamination structures are often interpreted as a product of discrete episodes of sediment accretion and mat growth. Bedded quartz sand observed within Facies B2 may support a regular episodic growth of cyanobacterial mats (sensu Riding, 2000). Clotted and disrupted textures are common to Facies B2 and have been described and interpreted as “algal” mats from the Bayport by Bacon (1971) and Ciner (1988) from exposures at Wallace Stone Quarry. Based on the modern environments where stromatolites are found (e.g. restricted hypersaline and/or high energy tidal environments) such as Hamlin pool at Shark Bay or the Khor al Bazam (Persian Gulf), a restricted intertidal environment of deposition is inferred for Facies B2. Microbial structures are often encased with quartz sandstone indicating these features may be controlled by either the local or regional influx of terrestrially derived siliciclastic material similar to what is proposed by Andres and Reid (2006).
Figure 19. Facies B2 core specimens continued, displaying a number of textures associated with microbial mats. **(A)** Silicified laminated structures within an otherwise laminated dolomite with significant amounts of quartz sand bedding. Mud-cracks and chert nodules are also common to Facies B2. **(B)** Flat-pebble-breccia composed of dolomite clasts overlain by poorly sorted coarse-grained quartz sandstone. These structures are likely the result of storm reworking of desiccated microbialite "chips" (Sensu Shinn, 1983a). Silicification of microbial mats is a post-depositional process.

**Dolomitization**

Primary carbonate sediment is composed of a mixture of aragonite, calcite, high magnesium calcite, and minor dolomite (Tucker and Wright, 1990). Dolomite, (Ca,Mg(CO3)2, is composed of equal parts calcium and magnesium ions and exhibits a higher density than calcium carbonate (CaCO3) (Scholle and Ulmer-Scholle, 2003). The process of dolomitization occurs through a number of primary depositional (pentcontemporaneous) and post-depositional (diagenetic) mechanisms. While the direct precipitation of dolomite is recorded in highly saline environments (e.g. the Coorong, South Australia), most dolomite is believed to be secondary in nature (Tucker and Wright, 1990). The alteration of calcium-carbonate to dolomite requires solution enriched with magnesium ions (Mg2+) for either the direct precipitation of dolomite or the replacement of previously deposited limestone (CaCO3) deposits (Tucker and Wright, 1990).

The creation of dolomite in modern environments alone does not explain the widespread occurrence of dolomite in the geologic record. Post-depositional burial dolomitization formed under conditions of high temperature and pressure may result
in a disproportionate occurrence of dolomite in the geologic record (Tucker and Wright, 1990). Dolomitic crusts and surfaces are commonly observed association with microbialites, which form within the intertidal to supratidal zone, and suggest periodic sub-aerial exposure during low-tide in supratidal environments semi-arid to arid environments of the Persian Gulf and Western Australia (Shinn, 1983a). Arid Sabkha environments of the Trucial Coast, Arabian Gulf, in particular are widely seen as good modern analogs to explain the occurrence of dolomite in ancient strata (Tucker and Wright, 1990). However, the widespread dolomitization of ancient strata cannot be fully explained by modern processes and the failure to replicate the process of dolomitization under low-temperature laboratory conditions has exacerbated the problem (Tucker and Wright, 1990).

A number of models are available to explain the occurrence of dolomitic strata either related to primary deposition or post-depositional processes. In modern environments, dolomite is commonly found associated with microbial structures and evaporites such as anhydrite and gypsum (Tucker and Wright, 1990; Bontognali et al., 2010). The common occurrence of dolomite below calcium-rich evaporitic strata in the geologic record supports the seepage-reflux model which relies on downward percolating water enriched with magnesium ions (Tucker and Wright, 1990). Hsu and Schneider (1973) analyzed excavated pit walls along Abu Dhabi Island to test three hydrologic models which can explain the creation of dolomite: the seepage reflux, capillary action, and evaporative pumping models. The observation of vertically orientated hydraulic gradients was used to validate the evaporative pumping model. Hsu and Schneider (1973) differentiate between the processes of capillary draw and “evaporative pumping,” the latter of which is caused by a decrease in pore-pressure attributed to the evaporation of intra-formational fluids resulting in the upward
mobility of water-table brines enriched with respect to magnesium ($Mg^{2+}$).

Microbial mediation is noted by Bontognali et al. (2010) as the only pathway for dolomitization in surficial environments. Bontognali et al. (2010) gives two hypotheses for the formation of dolomite based on extensive geochemical analysis of coastal sabkha environments of the Arabian Gulf: (1) dolomite formed directly in the supratidal zone, and (2) dolomite continues to form within the mats during and after burial. Regardless of the exact timing of dolomitization, the occurrence of dolomite associated with laminated, fenestral, and desiccation features in association with tan-brown dolostone Facies B2 within the Bayport Interval supports an upper-intertidal to supratidal environment, similar to structures reported in the modern Trucial Coast and Shark Bay (e.g. Shinn, 1983a; ).

**Facies B3a: Carbonate Breccia**

*Description:* Breccias and conglomerates (Fig. 20a,b) are composed of carbonate clasts ranging from 1-to-5 cm in diameter and are locally present near the top of various the carbonate-dominated facies. Textures include both clast and matrix-supported conglomerate-breccias (Fig. 20a,b). The matrix consists predominantly of fine-grained dolomitic mudstone. The orientation and size of clasts is generally random with increasing clast size towards the top of the brecciated horizon. Green-clay residue commonly occupies interstitial space between carbonate clasts. Fine-grained silty laminations are often present, especially near the top of the deposits. In thin-section, this facies displays a nodular texture (Fig. 21b,c). Circular to ovoid “glaebules” surrounded by calcite spar cement is observed throughout the section (Fig. 22a-c). Fossiliferous material can be discerned where not obliterated by
post-depositional alteration (e.g. Fig. 22c). This facies caps many sub-meter carbonate beds. In thin-section this facies displays a nodular texture (Fig. 22b,c) with calcium carbonate spar, and, rarely, fibrous chalcedony (Fig. 22a) cements form rinds surrounding the discrete pisoliths. Nodular chert is also common throughout this facies. Silicification and chertification occur along discrete seams and inside relatively large (>10 cm diameter) cavities is common and consistent with previous descriptions of the Bayport (Bacon, 1971; Ciner, 1988). A notable zone of extensive silicification in limestone is observed capping the Bayport in core SB18, which is then overlain by a thick soil profile (Fig. 22d, Fig. 33a-d).

**Interpretation: Karst/Caliche.** Breccias observed in this study may have been produced in one of two ways: (1) the influx of meteoric water, which results in the dissolution and alteration of underlying evaporite and/or carbonate material and collapse or, (2) repeated periods of exposure and desiccation. Soils are often observed in modern carbonate deposits at subaerial exposure surfaces, but have a poor preservation potential in the geologic record. Underlying caliche and karst deposits are more likely preserved (Estaban and Klappa, 1983) but may cap carbonate succession from inches to feet in thickness. Similarly described mudstone deposits have been reported in the Chesterian-aged Illinois basin (Glen Dean to Tar Springs Interval) by Smith and Read (2001) and in the Appalachian basin by (Al-Tawil et al., 2003; Al-Tawil and Read, 2003).

Brecciation, dissolution, disrupted laminations, pisoliths, and the rare occurrence of evaporate minerals common in Facies 3a indicate periods of sub-aerial exposure and pedogenesis in a semi-arid environment (Estaban and Klappa, 1983). Karst results from a diagenetic process resulting in the dissolution of calcium carbonate irregular surface landscape, subsurface fissures and caves, re-precipitation
products (speleothems and globulites), and collapse structures from the net loss of underlying carbonates (Estaban and Klappa, 1983). Calcrete/caliche microtextures observed in Fig. 22a-c are the result of periodic wetting and drying (sensu Estaban and Klappa, 1983; Flugel, 2010). Caliche or calcrites may form either directly from pedogenesis and soil formation or cementation in the phreatic zone (Flugel, 2010). Evidence for the presence calcrete include: (1) vadose pisolites/glaebules surrounded by calcite "rinds", (2) obliteration and inversion of primary texture/structure to calcite spar, and (3) association with other facies/features resulting from subaerial exposure (e.g. paleosols and silica crusts) (Estaban and Klappa, 1983). Carbonate units observed in this study display evidence for diagenetic modification in a subaerial environment.

Figure 20a,b display hand specimens of the topmost portion of an interpreted caliche deposit. These brecciated horizons are often overlain by a mature soil profile and is interpreted as a substantial period of exposure (e.g. Fig. 31). The precipitation of calcium carbonate in arid environments commonly results in early-lithification and the occlusion of porosity (Estaban and Klappa, 1983). Caliche or calcrete is described as a white fine grained calcium carbonate which may form as an alteration product of pre-existing subtidal deposits (Estaban and Klappa, 1983). Likewise Figure21c indicates textural inversion and obliteration of fabrics and constituents interpreted as subtidal deposits (upper left quadrant) including foraminifera which are common to Facies B1. Chalcedony is reported to form from the replacement of evaporite minerals by Folk and Pittman (1971) and may indicate arid-sabkha environments. Silicification at exposure surfaces is reported by Retallack (1991) and commonly occurs in association with unconformities (Retallack, 1997). Silicification requires an available source of silica (Knauth, 1979). Knauth (1979) outlines a simple
geochemical model which may explain the genesis of nodular chert in the meteoric-marine mixing zone. A similar interpretation was given by Ciner (1988) to explain the occurrence of chert in the Bayport based on occurrence of pore filling, fibrous, microcrystalline silica which can be represented by a botryoidal texture within the carbonate host rock (Fig. 21a). Breccias have been reported from Bellevue quarry, Eaton County. Parker and Taylor (1971) report the occurrence of a 1-5 ft. thick breccia layer with intragranular space filled with pyrite and calcite. These deposits are often overlain by unaltered limestone beds suggesting that brecciation may be a product of post-depositional dissolution and collapse of a pre-existing limestone deposit, though a hydrothermal origin cannot be discounted (Harrison, Pers. Comm., 2012). The close vertical proximity of soils and brecciated deposits are likely the product of exposure and are interpreted to represent a significant drawdown in relative sea-level compared with depositional conditions of the underlying Facies B1 and B2.

Facies B3b: Blocky Mudstone

Description: Facies B3b consists mostly of pale green (5G 7/2-5G 5/2) to grayish olive (10Y 4/2), fine-grained silty mudstone/claystone with local occurrences of carbonate, evaporite, and silica nodules ranging from 1-8 cm in diameter. The structure of mudstone layers is typically heavily fractured and lineated and displays irregular, columnar to platy partings (Fig. 20c,d). This lithology is found interbedded with both sandstone and carbonate. Facies B3b also occurs interbedded with sandstone Facies B4 and B5. Beds most commonly occur as thin (<1 in) beds capping carbonate deposits (Facies B2) throughout the Bayport interval.
Interpretation: Vertic paleosols. Facies B3b is characteristic of ancient soils. The drab green to olive gray coloration is interpreted as a product of gleying associated with the reduction of iron (Fe$^{3+}$ to Fe$^{2+}$) possibly due to prolonged submergence below the water table (Retallack, 1991; Mack et al., 1993) or in the subsurface in a process referred to as burial gleization by Retallack (1997). Calcic and nodular horizons are interpreted as a product of an arid/semi-arid environment favorable to the accumulation of highly soluble calcium carbonate and evaporite minerals. The most pronounced nodular zone occurs in the SB18 core and directly overlies a fine grained argillic clay interval. These beds are often interbedded with fine-to-medium grained quartz sandstone which indicates proximity to either a marine or terrestrial source of clastic material. The presence of these distinct horizons within the Bayport is further suggestive of periodic subaerial exposure of both siliciclastic and carbonate lithologies. The number of interpreted exposure horizons suggests the system was located within a shoreline environment sensitive to changes in local sea-level and/or sediment supply. These green horizons often cap carbonate units and possibly indicate a decrease in relative sea-level.
Figure 20. Facies B3a and b. Caliche breccia and blocky mudrock paleosol. (A) Limestone breccia. (B) Limestone breccia overlying a prominent evaporite bed. Interpreted as the product of post-depositional dissolution and collapse. Note green-clay filling the matrix surrounding the carbonate clasts. (C) Green, lineated mudstone displaying with prominent carbonate clasts. (D) Green-mudstone displaying evaporite nodules and iron staining. Scale in cm.
Figure 21. Facies B3 photomicrographs showing caliche breccia and blocky mudrock-paleosol. (A) Pisolitic deposit with coated carbonate pisoliths surrounded by pore-filling fibrous silica (chalcedony) (XPL). Such deposits are a common feature of ancient exposure horizons (e.g. Folk and Pittman, 1971). (B) Caliche deposit with pisoliths attributed to the shrink-swell (wetting and drying) in the vadose zone. Wetting and drying in the soil horizon results in the creation of glaebules and the precipitation of calcite spar between within the newly created intergranular crevice (Esteban and Klappa, 1983). (C) Highly mature caliche deposit with coated grains, pisoids (Pi) and carbonate glaebules enclosed by calcite-spar rinds. This example resembles a mature caliche deposit and is associated with the systemic boundary. Upper left quadrant contains the relatively unaltered foraminiferan packstone-grainstone parent material. The sample is associated with a major exposure surface. (D) Dark red/maroon vertic texture paleosol from Figure 33b. Note the opaque fracture filling iron oxide (FeO$_3$).
Figure 22. (A) A generalized depiction of evaporative processes with respect to the dominant tidal zones of deposition; Marine (subtidal zone), intertidal zone, supratidal, and land. These zones exhibit distinct facies distribution (From Shinn, 1983a). (B) An idealized shoaling (shallowing) upward carbonate facies succession commonly observed in ancient peritidal carbonate deposits. The facies succession begins with a coarse-grained deposit (A) overlain by fossiliferous limestone (B) then grade into laminated (“crypt-algal”) deposits punctuated by exposure on the intertidal-supratidal zone of deposition (C-E) (Modified from James, 1979).

Shallow Marine Carbonate Depositional Environments

The distribution of rock facies and diagnostic sedimentary structures found within modern shallow marine, paralic, carbonate systems including Andros Island, the Persian Gulf, and Shark Bay have been thoroughly documented in the literature (e.g. Shinn, 1983a,b; Shinn, 1968; Kendall and Skipwith, 1969; Logan and Cebulski, 1970; Alsharhan and Kendall, 2003). The recognition of diagnostic structures such as birdseye-fenestrae, microbial laminations, mudcracks, intraclasts, and soils horizons aid in the interpretation of ancient carbonate strata (discussed in the previous chapter (Fig. 22b, Shinn, 1983a). Ancient carbonate tidal-flat successions are commonly observed in the geologic record and often exhibit a shallowing (shoaling) upward facies motif (Fig. 22b). The successions often transition from deeper subtidal and intertidal to evaporites, soils, and/or dissolution/subaerial exposure related structures at the top (e.g. Fig. 22, e.g. James, 1979; Shinn, 1983a). In the supratidal zone, climate is an especially important control on the character of exposed sediments (Shinn, 1983a). The arid-subtropical Persian Gulf and portions of the humid-tropical/sub-tropical Great Bahama Bank (GBB) serve as useful modern
environments and display common facies textures and successions, which are observed within ancient strata (Alsharhan and Kendall, 2003; James, 1979; Shinn, 1983a).

Tides play a large role in depositional processes in shallow marine environments. Tidal systems are divided into three zones (Fig. 22a) on the basis of tidal regime, each marked by diagnostic sedimentary features (e.g. Shinn, 1983a): (1) supratidal, or the zone of exposure is the site of pedogenesis, karst dissolution, and evaporite production, (2) the intertidal zone, which is partially exposed during low tides and often dominated by microbial communities, and (3) the subtidal zone, which is composed of a sublittoral environment characterized by bioclastic debris including skeletal sands, ooids, and peloids and is rarely exposed to subaerial alteration (Shinn, 1983a). Paleozoic tidal flat deposits have been widely recognized in the geologic record and constitute important resource exploration targets due to a clear stratigraphic trapping mechanism (i.e. impermeable evaporites) which often pinch out up-dip of the underlying permeable reservoir rock (i.e. fenestral grainstone) (Shinn, 1983a).

The supratidal zone contains a number of diagnostic sedimentary structures including mudcracks, crypt-algal/microbial laminites, intraclasts, birdseye-fenestrae, and evidence of pedogenesis (Shinn, 1983a,b). In arid climates, evaporites such as gypsum and anhydrite are precipitated often directly above a water table supersaturated with calcium, magnesium, and sulfate ions. Desiccation features, including mudcracks, are diagnostic of a supratidal environment and have been widely reported in both ancient and modern strata (Shinn, 1983a; James, 1979). Intraclast and flat pebble conglomerate deposits may indicate recurrent storm events (Shinn, 1983a). Tide dominated environments often preserve distinctive, rhythmically
laminated bedding structures referred to as tidal laminites, which are documented to occur in both carbonate and siliciclastic settings (Ginsburg, 1975). Early diagenetic precipitation of sulfates and occurs in arid environments such as the Trucial coast (southern Persian Gulf, Abu Dhabi) coastal sabkhas associated with the above facies. In comparison, humid environments, such as the Bahaman Platform, receive substantially more rainfall resulting in the lack of soluble evaporites. In these environments exposures are characterized by extensive dissolution cavities referred to as karst (Shinn, 1983a). A number of climate dependent processes occur in supratidal tidal flat environments including dolomitization and evaporite precipitation in arid-climates and soils and karst dissolution in humid environments (Shinn, 1983a). In arid environments dolomitic surface crusts are created by capillary action and/or evaporative pumping, which transport water table brines with high Mg⁺ concentration upward, altering (i.e. dolomitizing) the limestone host deposit (Tucker and Wright, 1990). Evaporites, including gypsum and anhydrite, form readily in arid environments including the Persian Gulf, but are not preserved in Andros Island and humid areas of the modern Persian Gulf due to the high humidity and rainfall (Shinn, 1983). The presences of evaporites are perhaps the most diagnostic of ancient sabkha succession.

The intertidal zone lies between mean high and low tides. The onlapping (transgressive) tidal flat deposits of Andros Island contain numerous tidal channels and burrowing organisms have thoroughly mixed the sediment eliminating sedimentary structures (Shinn et al., 1969; James, 2010). In contrast, the offlapping (regressive) deposits of the Persian Gulf reside in an arid setting, limiting the effects of bioturbation on the upper-intertidal portion of the offshore system (Shinn, 1983; James, 2010). Microbial laminites and birds-eye fenestrae structures are most
commonly found associated with upper-intertidal to supratidal zone of deposition (Shinn, 1983a,b).

Subtidal environments fall below the influence of tides and may comprise of shoals, lagoons, and/or subtidal portions of tidal channels. Along the coast of Abu Dhabi (i.e. Trucial Coast) a complex system of barrier islands composed of skeletal and coralline sandstone grade into lagoons, microbial mats, and a coastal sabkha environments in a landward trajectory (Alsharhan and Kendall, 2003).

Facies B4: Planar to Cross-Bedded Quartz Sandstone

*Description:* The facies is composed of moderately poor-to-moderately well-sorted, subangular to rounded, fine to medium-grained quartzose sandstone with horizontal (planar) to low angle (15-25°) cross-bedding. Bedsets are commonly planar and horizontal to high-angle (>25°) stratification. This facies is commonly cemented with calcite and consists of predominantly sub-angular to sub-rounded sand-sized quartz grains (Fig. 26a-d). The base of this facies is typically marked by a poorly sorted lag deposit composed of chert and carbonate lithoclasts ranging from 6 in. to 1 ft. in thickness. Other occurrences (SL423) appear to be gradational with the underlying carbonate unit (Facies B3). Figure 24c shows cross-strata sharply transitional to horizontal laminations. Figure 24b displays beds with consistent dip angles separated by reactivation surfaces (Fig. 24b,c). In core the bedforms resemble tabular cross-strata. However, in outcrop (i.e. Wallace Stone Quarry) bedforms appear to have a sigmoidal configuration with the overlying erosional surface indicating trough cross-stratification. The challenges in discriminating between tabular and trough cross-stratification was noted by Harms (1975), who indicates
trough-cross stratification is best identified in a horizontal section, transverse to the principle flow-direction. The laminations display, sub-mm, mud/silt partings with thick (~0.6 in., 1.5 cm) and thin (~0.04 in., ~0.1 cm) layers. So-called “double mud” drape structures are common (Fig. 24a,c). Repetitive (rhythmic?) thick and thin laminations are best preserved in planar bedded deposits (Fig. 24a). Re-activation surfaces are common and are often concordant with the bedding plane upon which bedforms displaying contrasting dip orientations rest (Fig. 24b), but may be erosional as well (Fig. 24c). Facies B4 transitions into Facies B5 moving upward in the vertical succession. Locally, calcium carbonate cements are common. Carbonate cement concentration tends to occur in highest proportions toward the base of the sandstone units where a deposit of medium to coarse-sand containing lithified carbonate rip-up clasts and chert fragments overlies Facies B2. In these basal zones subtle isopachous rim cements enclose quartz grains (Fig. 26b). The underlying carbonate beds are regularly fractured with local calcium carbonate, chert, and rarely sulfide minerals fill including pyrite (as is the case in SB15, 238 ft) (Fig. 23a,b).

**Interpretation: Estuarine tidal channels.** The presence of preserved sedimentary structures including planar and cross-laminations indicates an environment not favorable to bioturbating organisms. Trough cross-stratification most commonly occurs in association with down current migrating dunes (i.e. sand-waves) while planar laminations indicate greater shear stress (current velocity) compared to cross-strata (Harms, 1975). The presence of <1 mm mud-drapes orientated in a repetitive “thick-thin” occurrence is interpreted to record alternating periods of high, maximum tidal current velocity and low energy slackwater conditions. These features are common in tide dominated environments (Dalrymple, 1992, 2010c). Figure 24b shows evidence of bidirectional currents and accompanying
reactivation surfaces, further supporting a tidal environment of deposition (Dalrymple, 1992, 2010c). However, as noted earlier, trough cross-stratification may yield a similar bidirectional configuration when observed perpendicular to the original flow direction (Harms, 1975). Reactivation surfaces, which are often marked by a discrete silt-layer, which have been reported in tidal estuarine environments under conditions of changing flow velocity (e.g. Klein, 1970; Dalrymple, 1992, 2010c). The presence of elongate sandbodies with primary sedimentary structures preserved either from bioturbation or soft-sediment deformation indicates high energy environment of deposition. The presence early-marine cements are suggested by isopachous rims enclosing quartz grains which “float” within a carbonate matrix (Fig. 26b) tentatively indicating syndepositional cementation in a marine environment.

Rhythmic laminations associated with tide-dominated environments have been reported from Precambrian deposits (Bose et al., 1997) and from the Carboniferous mid-continent (Archer, 1998). Tidal rythmites may be laterally or vertically accreted in a tidal channels or offshore sandbars (Mazumder and Arima, 2005; Bose et al., 1997). The absence of exposure surfaces, which may be represented by root traces, oxidized horizons, and soils, indicates a predominantly subtidal environment of deposition for Facies B4. In vertical section Facies B4 is commonly associated with FaciesB5, which is interpreted to represent a more landward, lower energy, depositional environment (see below).
Figure 23. The contact between carbonate and clastic dominated lithofacies B2 and B4. (A) Chert and carbonate intraclasts overlying a brecciated and fractured dolomite deposit. The contacts grades into cross-bedded quartz sandstone of Facies B4. (B) A similar contact in the SL418 core (~70 mi away from sample A). Red arrow denotes the transition to quartz sandstone of Facies B4. Scale in cm.
Figure 24. Facies B4 hand specimens. Cross-bedded quartz sandstone. (A) Planar bedded calcite cemented quartz sandstone displaying bimodal “repetitive” thick (medium-grained sand) and fine-grained thinly bedded mud-draped laminae. Small-scale trough cross bedding is present at the base. (B) Fine-to-medium quartz sandstone displaying prominent reactivation surfaces (red arrows) resulting from changes in current speed and direction. (C) An example of the same facies displaying a prominent reactivation surface (red arrow). Note the rapid transition from inclined (cross) bedding to planar bedding and the tabular nature of the bed-contact. Scale in cm.

Facies B5: “Heterolithic” Wispy Bedded Quartz Sandstone

Description: Wavy to wispy heterolithic bedded sandstone is found in all cores analyzed and represents the most common siliciclastic-dominated facies. Grain-size is heterolithic and alternates between fine–to-medium-grained sand. Bedforms are silt dominated "wispy" bed partings often displaying a wavy ripple-like geometry in fine-grained white sandstone. Locally, Facies B5 displays mottling and contorted bedding. Irregular laminae may be inclined at angles exceeding $45^\circ$. Notably, this facies is most commonly non-calcareous and is typically interbedded with microbial laminitic structures common in Facies B2. Bedding is often disrupted and displays irregular laminations. Brecciated dolomite and silica clasts (flat pebble conglomerates) of laminated dolomite were observed in the SB16 and SL423 core.

Interpretation: Intertidal sand-mudflat. “Heterolithic” bedding is typically associated with tide-dominated depositional environments (Reineck and Singh, 1980; Dalrymple, 1992, 2010c). Van Straaten (1961) observed a distinct landward grain-size fining trend in North Sea coastal tidal flats. Heterolithic bedding is defined based on the proportion of mud to sand grain-sizes and include flaser, wavy, and lenticular
bedding (Reineck and Singh, 1980). The presence of mud indicates a low-energy environment influenced by intermittent periods of suspension deposition supporting a more landward depositional environment compared to Facies B4. Wispy bedding indicates alternating periods of high and low energy conditions depositing a mixture of fine-to-medium-grained sand and mud. Silty wisps are interpreted as deposits formed during slack-tide condition resulting in the suspension-settling of fine-grained mud and clay out of the water column. Occasional vertical traces indicate burrowing; a common occurrence on modern intertidal flats (Dalrymple, 1992, 2010c). Contorted bedding due to soft sediment deformation may explain the heterolithic nature of the wispy bedforms. Soft sediment deformation is associated with the intertidal zone of modern tidal flats (Reineck and Singh, 1980). Microbial structures similar to Facies B2 commonly occur interbedded within Facies B5 further suggesting an intertidal to supratidal environment of deposition.
Figure 25. Facies B5 hand specimens. Wispy bedded fine-grained sandstone-interpreted as clastic tidal flat deposits. (A) Wispy-bedded medium-grained quartz sandstone. (B) Interbedded medium sandstone and silty draped lamina. (C) Wavy to lenticular bedded fine quartz sandstone/siltstone display mud-draped ripple-like lamina. (D) A sub-vertically orientated trace-fossil. Scale in cm.

Figure 26. Facies B4 and B5 microfacies. (A) Well sorted and finely laminated horizontally bedded fine grained quartz sandstone. (B) A carbonate (calcite) cemented quartz sandstone. Red arrows denote isopachous rimming cements, possibly a product of early-marine cementation. (C) Quartz sandstone containing bryozoan fossils and isolated mudclast fragments (intraclasts). Red arrow marks a pore (blue epoxy) that was probably previously calcite cement (i.e. dissolution porosity). (D) Wispy bedded Facies B5. Red arrows denote mud-drapes and are often overlain by relatively coarser grain-size.
Tidal Processes and Deposition

Tidal systems and processes have been treated in a number of publications covering both siliciclastic and carbonate dominated environments (e.g. Shinn et al., 1969; Shinn, 1983a; Ginsburg, 1975; Dalrymple, 1992, 2010c; Kvale, 2012). Two primary models explain the occurrence of tides, the *tidal equilibrium model* and the *dynamic tidal model* (Kvale, 2012). The tidal equilibrium model describes tides as a function of the Earth-Moon orbital system which results in the presence of two oppositely orientated tidal centers (tidal bulges) on the Earth corresponding to a point closest to and furthest from the Moon and resulting in flood (rising) and ebb (falling) tides corresponding with the rotation of the Earth about its axis (de Boer et al., 1989). The dynamic tidal model is more complex, taking into account the gravitational effect of both the sun and moon relative to the Earth (Kvale, 2012). In the previously discussed model, tides attain their maximum amplitude when the Earth, Sun, and Moon are aligned 180° relative to one another and are weakest when they are aligned perpendicular (90°) (Kvale, 2012). Tidal currents by nature operate on daily, monthly, and yearly scales (Kvale, 2012). Key indicators of ancient tidal periodicity lie in their physical structure and laminae thickness, features referred to as *tidal rhythmites* (Williams, 1991; Dalrymple, 1992, 2010c). Each day, a given position of the earth undergoes two complete ebb and flood cycles, which under ideal condition results in the deposition of alternating couplets of fine and coarse grained deposits (Fig. 27a). Tidal rhythmites result from the deposition of a coarser grained layer during the flood tidal current, while during the period of slackwater, when ebb and flow tides are in equilibrium, a suspension deposit formed during a period of slackwater at high and low water when tidal currents are in equilibrium followed by deposition of a thinner, coarse bed during the subordinate tidal current, which is
followed by another period of slackwater and the deposition of a mud layer from suspension (Dalrymple, 1992, 2010c). Other features that may be associated with tidal current dominated environments include thick-thin lamina pairs, reactivation surfaces, and oppositely dipping beds (herringbone cross stratification). Taken in total, these features are important indicators of siliciclastic tide dominated environments (Dalrymple, 1992, 2010c).

Figure 27a shows interpreted siliciclastic-dominated tidal-rhythmites. Although these features have been reported in carbonate strata, they are not typically as well preserved (Davis et al., 1998; Dalrymple, 1992, 2010c; Kvale, 2012). Williams (1991) notes that vertically accreted, mixed sand/silt strata have the best potential for the preservation as tidal rhythmtes in Precambrian examples from southern Australia. Modern occurrences of tidal rhythmtes in siliciclastic strata have been documented in the Bay of Fundy, Nova Scotia by Dalrymple et al. (1991) and from the Bay of Mont St. Michel, northwestern France, by Tessier (1993).

A statistical model based on laminae thickness was constructed by De Boer (1989) to test for diurnal rhythmicity. Quantitative studies using this procedure of recording the number and thickness of laminae have been used to infer tidal periodicities for ancient Carboniferous Midcontinent strata (Archer, 1998; Greb and Archer, 1998; Kvale, 2012). However, Greb and Archer (1998) note that determining the tidal hierarchy based on laminae thickness is complicated due to incomplete preservation-potential of daily cycles. Greb and Archer (1998) suggest an annual scale of preservation is most likely for ancient tidal deposits. Dalrymple (2010c) notes that even known tidal deposits "fail the test" for tidal cyclicity. However, the presence of abundant laminations, heterolithic bedding, double mud-drapes, and thick-thin lamina pairs, as well as the presence of these structures and facies within
the overall succession supports deposition under a tide-dominated energy regime (sensu Dalrymple, 2010c).

Estuarine facies models based on tidal channels within the Bay of Fundy and the Bay of Mont. St. Michel estuaries have been proposed as a modern analogue for Carboniferous Midcontinent strata (Greb and Archer, 1998; Archer, 1998; Smith and Read, 2001; Greb and Martino, 2005). Tessier et al. (1995) compare examples of tidal Rhythmites from the Midcontinent USA to modern examples deposited at the Mont St. Michel Bay, France. The typical tidal flat succession displays a fining-upward grain-size motif, from coarser-grained subtidal tidal channel deposits upwards into finer-grained, lower energy intertidal deposits, and capped by the supratidal or terrestrial facies (Dalrymple, 1992, 2010c). The fining upward grain-size trend is a response to tidal channel switching and meandering, an autocyclic process similar to meandering fluvial channels and carbonate dominated examples (Dalrymple, 1992, 2010c). Despite the occurrence of open-coast tidal-flats in the modern (e.g. Georgia Coast and North Sea Coasts) regressive tidal flats are rarely preserved in the geologic record due to the high potential for erosion. Dalrymple et al. (1992) suggest that
Figure 27. Tidal depositional processes associated with quartz sandstone intervals. (A) A model illustrating the deposition of a tidal rhythmite within a flood-dominated depositional environment. Tides consist of two components: the advancing (landward) flood-tide and the receding (seaward) ebb tide. Slackwater occurs twice a day (diurnal tidal cycle) resulting in the deposition of a mud layer by suspension settling. Mud layers may be amalgamated if the subordinate current (in this case the ebb) is too weak to deposit sand. From Dalrymple, 2010c. (B) An example of calcareous quartz sandstone displaying a series of interbedded bundled thick/thin mud-drapes interpreted as being a product of tidal energy. An interpreted tidal rhythmite from the Bayport interval. (C) An example of a reactivation surface created under a bidirectional flow regime. (D) Model explaining the creation of a reactivation surfaces within tide dominated estuarine environments. From Clifton, 1982. After Klein, 1970.
estuarine environments have a high-preservation potential due to their common occurrence within incised valleys, which are fluvially formed by fluvial erosion. Further, siliciclastic tidal-flat facies are common in estuarine environments, including the Bay of Fundy, which form during periods of rising relative sea-level and estuarine valley filling resulting a fining upward, progradational vertical stacking pattern (Dalrymple et al., 1992; Dalrymple, 1992, 2010c; Tessier et al., 2006) and interpreted within the Bayport.

Facies B6: Mudclast Breccia and Climbing-Ripple-Laminated Sandstone

*Description:* Facies B6 occurs as two distinct sub-facies: (1) mudclast breccias overlain by (2) a bed displaying climbing ripple laminations. This facies was observed only in SL423, and is Chesterian in age on the basis of microflora (next chapter). In SL423 the lithofacies is overlain by evaporitic shale (Facies B7). The base of the facies occurs as coarse-grained sandstone grading into a pronounced mudclast breccias. The breccia extends ~5 ft. upward where it is overlain by climbing-ripple bedforms that fine upward into mudstone/shale (Facies B7). Vertical burrows occur at the base of the facies, within heterolithic sandstone of Facies B5. The breccia deposits are composed of angular to rounded 1 to 4cm mudelasts randomly distributed within a coarse grained sandstone matrix. Reddish orange (10R 6/6) iron staining is commonly present throughout this facies. The mudclast prone deposits grade into a ~5 ft. bed of climbing ripple laminations of variable inclinations (5-30°) that display prominent erosional surfaces and micro-faults with bed offsets. The most diagnostic occurrence of climbing ripple lamination (Type A) is shown in
(Fig. 28b) and occurs at the top of the facies. The interval grades into an evaporitic shale of Facies B7.

Interpretation: Cutbank collapse channel lag deposit. Mudstone lithoclasts are interpreted to represent a period of erosion and scour possibly associated with the base of high energy fluvial channel deposit (Miall, 2010a). Mudclasts are commonly observed within the Pennsylvanian Saginaw Formation, where they are characteristically found at the base of coarse-grained sandstone deposits, interpreted as channel lags produced by high energy scouring possibly along the cutbank of a fluvial channel (Venable, 2006; Venable et al., 2010). Climbing ripple laminations have been reported in a number of depositional environments including subaqueous delta levees (Reineck and Singh, 1980), fluvial deposits (Ashley et al., 1982), and within the fluvial-tidal transition zone associated with estuarine environments (Lanier and Tessier, 1998; van den Berg et al., 2007). Climbing ripples form under conditions of unidirectional current velocities and high concentrations of suspended sediment, which contributes to sediment fallout and the depositional of silt layers (Ashley et al., 1982). Climbing ripples are divided into two subtypes; Type (A) and (B) (Ashley et al., 1982). Type (B) climbing ripples show both the lee and stoss side of the ripple, while only the lee side is preserved in the “A- type” due to erosion (Reineck and Singh, 1980; Ashley et al., 1982). Factors resulting in the creation type A (erosional-stoss) and B (depositional-stoss) climbing ripples are documented by flume experiments conducted by Ashley et al. (1982). Lanier and Tessier (1998) note that climbing ripples have been observed within fluvial-tidal transition zone at the headward portion of the Mont Saint Michel Bay macrotidal estuary, northwest France. The fluvial-tidal transition is the subject of a number of recent studies (e.g. Dalrymple and Choi, 2007; van den Berg et al., 2007) and is influenced by the
mixture of tidal and fluvial energy regime, producing sedimentary structures due to the interaction of contrasting flow conditions (van den Berg et al., 2007). The laterally discontinuous occurrence of these deposits, along with the vertical association relative to the underlying tidal deposits and the overlying evaporitic mudflat, Facies B7, suggest a channel deposit, possibly of mixed fluvial/tidal origin. In the neighboring Illinois basin, rare occurrences of fluvial deposits occur at the base of incised valleys which formed during sea-level low-stands (Smith and Read, 2001; Nelson et al., 2002). Nelson et al. (2002) notes that sequence boundaries are marked by erosion (i.e. incised valleys) or caliche/breccia paleosols, a similar scenario observed in Bayport strata within the Michigan basin.
Figure 28. Mudstone breccia and associated climbing ripple laminations (Facies B6 and B7; SL423) (A) Isolated mudclast within a coarse grained quartz sandstone matrix interpreted as channel collapse lag deposits. (B) Climbing ripple lamina displaying a pronounced truncation surface eroding the stoss side of the ripple formed under conditions of high aggradation (Type A ripple) (red arrow). (C) Horizontally bedded mudstone/shale deposit. (D) Enterolithic anhydrite mudstone displaying interpreted fenestral texture associated with the decay of organic matter. Scale in cm.

Facies B7: Evaporitic Shale and Nodular Anhydrite

*Description:* Mudstone/shale of Facies B7 is present only in the SL418 and SL423 cores where it is locally interbedded with thin beds (~3 in) of gypsum and anhydrite. Associated shales are commonly horizontally bedded, highly fissile, and fracture along horizontal bedding planes. Localized occurrences of ripple bedforms are observed. Nodular anhydrite occurs in one core, SL418 (806.5-803.7 ft.), Montcalm County. Nodules are typically 3-5 mm in diameter and are surrounded by interstitial clay/mudstone deposits. This occurrence of anhydrite overlies mudstone/shale (817-806.5 ft.) and is surrounded by a hemispherical/laminated dolomite displaying horizontally orientated fenestrae filled with anhydrite. The anhydrite deposit is associated with a highly brecciated carbonate (Fig. 20b) (Facies B3) which grades into massive white fossiliferous limestone of Facies B1 (Fig. 15b). Green clay is found throughout the anhydrite filling fractures. Locally, mudstones display elongate cavities, concentrated along bedding planes filled with a mixture of anhydrite/gypsum, calcite, and silica.

*Interpretation:* *Arid supratidal mudflat.* Mudstone/shale is interpreted as
suspension deposits resulting from low-energy conditions at the site of deposition. The shale is devoid of fossils and evaporitic intervals within the shale interval indicate a supratidal environment of deposition, similar to the Michigan Formation, although Facies B7 lacks sandstone with wave-dominated structures indicating a back-barrier environment for the Michigan. Evaporites such as gypsum often form in hypersaline pools within the supratidal zone where the rate of evaporation is greater than the rate of water recharge (Kendall, 2010). Gypsum is typically converted to anhydrite as a result of elevated temperature and dehydration during burial. Evaporites are highly soluble and extremely susceptible to dissolution making them rare to nonexistent in humid environments where low pH (slightly acidic) meteoric waters percolate into the subsurface dissolving evaporite minerals (Warren and Kendall, 1985). The presence of evaporites including gypsum and anhydrite are commonly formed above the water table under semi-arid to arid conditions. The most prominent example in recent deposits is observed in the Trucial Coast located in the Arabian Gulf (Shinn, 1983a).

Evaporite minerals including gypsum and anhydrite have been thoroughly studied in the sabkhas of the Persian Gulf (Kendall, 2010) where they occur in broad landward facies belts (Alsharhan and Kendall, 2003). In arid coastal sabkha environments, such as the Trucial Coast, Abu Dhabi, the diagenetic replacement of gypsum with anhydrite results in the growth of displacive nodular and enterolithic structures within a pre-existing matrix (Warren and Kendall, 1985; Kendall, 2010) (e.g. Fig. 28d). Coastal sabkhas tend to be found parallel and landward of the shoreline and often cap ancient shoaling (shallowing) upward peritidal carbonate successions (Warren and Kendall, 1985; James, 1979; Shinn, 1983a) as is the case in the Bayport. The rare occurrence of evaporites within the Bayport indicates: (1)
localized deposition or preservation possibly due to climate fluctuation between arid and humid conditions (e.g. Rankey, 1997), and/or (2) poor preservation due to the influx of meteoric water under post-burial conditions. The latter scenario is consistent for what is reported from Late Mississippian strata of the Midcontinent by Cecil (1990).

Dissolution breccias forming key surfaces may further indicate relatively humid depositional conditions predominating within the Bayport interval. However, the paleosol deposits of Facies B3 indicate periodic aridity. In the observed example, the anhydrite caps a carbonaceous mudstone/shale deposit which occur at the top of a carbonate/clastic dominated fining upward succession composed of Facies B1 through B5. For the previously mentioned reasons, as well as its position relative to other facies (e.g. Facies B1 and B6), it is interpreted that deposition occurred within an arid supratidal mud-flat subject to marine flooding, possibly dissected by channel deposits of Facies B6. Vugrinovich (1984) and Tyler (1980) note an anhydrite bed overlying the “Bayport-Parma” interval, which is capped by the “Six Lakes Limestone.” Anhydrite is observed in cuttings and interpreted from geophysical borehole data. Tyler (1980) reports anhydrite from the Six-Lakes gas-storage field, Montcalm County. Similar to what is reported by Tyler (1980), the anhydrite is observed in core SL418 in association with an overlying limestone bed referred to as the “Cream” Limestone by Tyler (1980) and the Six-Lakes Limestone Member by Vugrinovich (1984). Vugrinovich (1984) interpreted the limestone as Pennsylvanian in age.
Table 3. Summary of facies observed within the mixed-siliciclastic/carbonate Bayport interval overlying the restricted Michigan Formation.

<table>
<thead>
<tr>
<th>Facies/Lithology</th>
<th>Occurrence</th>
<th>Color</th>
<th>Grain-type and texture</th>
<th>Sedimentary Structures</th>
<th>Biota</th>
<th>Interpretation</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>B1 Limestone</strong></td>
<td>Identified in all cores observed within the Bayport Interval</td>
<td>Buff white with black organic wisps</td>
<td>Packstone- Grainstone. Heavily bioturbated and Stylolitic with sand sized disarticulated bioclasts within a carbonate mud (micrite/pelsparite) matrix</td>
<td>Stylolites, wispy bedding, and rare original low-angle laminations</td>
<td>Open Marine; Echinoderms, bryozoans, brachiopods, horn corals, ostracods, and foraminifera</td>
<td>Subtidal grain shoal or tidal channel</td>
</tr>
<tr>
<td><strong>B2 Dolomite</strong></td>
<td>Occurs interbedded with Facies B1 and B4</td>
<td>Tan-brown</td>
<td>Composed of silt-sized dolomite mud, peloids, and quartz sandstone</td>
<td>Laminated, birdseye fenestrae, intraclasts, and mud-cracks</td>
<td>Microbial mats and rare gastropods</td>
<td>Restricted intertidal microbial mud-flat</td>
</tr>
<tr>
<td><strong>B3 Caliche-breccia and nodular, blocky mudrock</strong></td>
<td>Occurs in all cores observed, most commonly at the top of individual meter-scale carbonate beds</td>
<td>Mudrocks are locally green, red, and/or gray. White carbonate clasts/nodules</td>
<td>Conglomerate/ breccias commonly contain silicified nodules. Both clast and matrix supported arrangement.</td>
<td>Occasional reduction haloes, and shrink-swell lineaments (peds)</td>
<td>Reduction haloes may indicate rooting by terrestrial vegetation</td>
<td>Exposure surfaces and incipient pedogenesis; Karst, collapse breccias, and reduced and vertic (soil) horizons</td>
</tr>
<tr>
<td><strong>B4 Cross-Bedded quartz sandstone</strong></td>
<td>Occurs in all cores observed and marked by carbonate rip-up clasts at the base.</td>
<td>White to light yellow</td>
<td>Fine-Medium grained carbonate cemented quartz sandstone. Isopachous cements at the carbonate/sandstone contact</td>
<td>High-angle cross and planar laminations, reactivation surfaces, double mud-drapes occasionally in a rhythmic thick-thin, arrangement</td>
<td>Rare brachiopod and unidentified silicified fossil? beds</td>
<td>Subtidal Estuarine channels and offshore sandbars; Coastal marine,</td>
</tr>
<tr>
<td><strong>B5 Heterolithic Sandstone</strong></td>
<td>Occurs in all cores observed</td>
<td>White to light yellow</td>
<td>Fine-medium grained subrounded to subangular sandstone</td>
<td>Wavy (laser?) bedding</td>
<td>Locally bioturbated by Cruziana ichnofacies.</td>
<td>Alternating low and high energy regime. Intertidal mudflats</td>
</tr>
</tbody>
</table>
The Bayport at Wallace Stone Quarry

The type locale for the Bayport Limestone is the Wallace Stone Quarry (WSQ), Huron County, MI. The westward portions of the quarry were analyzed and attributes of the rock units were observed. The exposures occur just beneath the surface in an area in open quarry pits which were excavated over the course of the mines’ history. Sedimentary facies in the Quarry are similar to those observed in core. Previous studies of the Bayport were conducted strictly in the WSQ which is recognized as the type locale for the Bayport Limestone (Horowitz and Rexroad, 1972). The WSQ does not expose a lower boundary of the Bayport and is directly overlain thinly (~1 m) by Pleistocene and modern soil deposits.

The Bayport Formation exposures at Wallace Stone Quarry were the subject of M.S. thesis by Bacon (1971) and Ciner (1988). Bacon (1971) described a stratigraphic succession at the quarry including chert nodule-rich limestone, ‘algal’ mats, desiccation features, and marine limestone beds suggesting deposition within a sabkha to offshore marine environment. Bacon (1971) divided the depositional system into three zones: (1) a limestone-dominated lagoon zone composed of skeletal sands, (2) quartzose sand deposited in a beach, and (3) a dolomitic zone
corresponding to upper, middle, and lower zones of a hypothetical sabkha environment. The lower unit is composed predominantly of dolomitic and desiccated mat structures interpreted to have been deposited in the upper-intertidal to supratidal environment of deposition (Bacon, 1971).
Bacon (1971) noted the heterogeneous nature of the mat structures which may be chertified, finely bedded, and often topped by black “algal coal” deposits. Upward in the section, the dolomitized mats grade into a bed referred by Bacon (1971) as the “middle carbonates.” This zone is floored by chert beds which grade into “algal” laminites, sandy carbonates, and marine limestone (Bacon, 1971). The presence of quartz sand was noted by Bacon (1971) who attributed it to frequent storm activity which washed the grains onto the upper intertidal zone. Bioturbation is reported in this zone indicating a gradual transition from a restricted environment (dolomitic strata) to a more open marine environment of the lower intertidal zone. The presence of chert in particular was noted by Bacon (1971) to occur in two forms: (1) chertified “algal” laminated structures and (2) an ovular, nodular form. Chert nodules are reported to constitute up to 30% skeletal material, which is also noted by Ciner (1988). Further, Bacon (1971) notes the presence of chert nodules associated with green residuum which fill downward pointing tubules displaying preferential silica replacement. The green color is referred by Bacon (1971) as glauconite and suggests the tube structures are the product of roots associated with sea-grasses. However, grasses did not start to encroach into the subtidal zone until the Late Cretaceous.
before rapidly colonizing marine environments during the Miocene (Brasier, 1975). These structures are interpreted as being burrows and silica may have preferentially replaced the burrow-fill fabric, which is common in the subtidal zone consistent with the carbonate grainstone to packstone deposits. Bacon (1971) notes that the “upper” limestone unit is composed of ‘branching’ corals, crinoids stalks, pelecypod, and brachiopod valves. A zone of coral is noted from the Wallace Stone Quarry and is interpreted to have been deposited below wave-base due to the lack of sorting (Bacon, 1971).

Observations of the exposures at Wallace Stone Quarry are consistent with facies observed in core, reported previously in this chapter. The base of the succession is a mixture of siliciclastic and dolomitic strata. Stromatolitic textures, fenestrae, flat pebble breccias/conglomerates within the dolomitic zones indicate a restricted environment of deposition subject to the influx of terrestrially derived quartz sandstone. Cross-bedded quartz sandstone is striking and has previously been ignored in previous studies of the Quarry (e.g. Bacon, 1971; Ciner, 1988; Lasemi, 1986). The interbedded carbonate (dolomitic) and quartz sandstone is similar to what was observed in core material. The heterolithic dolomitic/siliciclastic facies is overlain by a thick (~10 ft.) grainstone-to-packstone carbonate deposit consisting of 0.5-1 ft. chert nodules and vertical tubules filled with green-clay residuum. Thin shale beds can be disseminated within the quarry walls. A quarry-wide rugose (horn) coral bed and directly overlies a similarly extensive shale bed. The presence of similar facies tracts present throughout south, central, and western Michigan supports a basin-wide distribution of the Bayport interval as suggested from the study of wireline logs by Lasemi (1975), Vugrinovich (1984), and Westjohn and Weaver (1998).
The Mississippian-Pennsylvanian systemic boundary was observed in five of the SSC cores from the southern Michigan basin (SB03, SB10A, SB13, SB15, SB16, and SB18). The Bayport interval is observed to be disconformably overlain by either a mudstone (paleosol) at a karsted limestone regolith or very-coarse micaceous quartz sandstone (Facies S2, see Saginaw Formation Facies below) that forms an erosional contact/surface (e.g. Fig. 30). These facies can either overly the sandstone or carbonate facies (B4, B2) of the Bayport making wireline log correlations of the boundary difficult. The sharp transition from a mixed carbonate-clastic interval to a depositional system almost completely devoid of carbonates in all of the cores analyzed. Figure 30 shows the contact observed within the SB13 core. In the SB15 core quartz “pebbles” are observed disseminated throughout a very-coarse grained sandstone matrix suggestive of conditions of bed-load dominated deposition within a fluvial channel. Associated carbonaceous shale and plant debris suggests a terrestrially dominated depositional system. The latter deposits are darker in color, contain siderite nodules, and display plant impressions along bedding planes. Carbonaceous debris and indicators of pedogenesis (e.g. clay horizons, slickensides, nodules, and carbonate breccias) strongly suggest a terrestrially dominated environment of deposition and strongly contrasts with the mixed clastic/carbonate succession occurring within the Bayport Limestone. The transition from the Bayport Limestone to the overlying Saginaw Formation is disconformable.
Figure 30. Interpreted Mississippian-Pennsylvanian systemic boundary in core. Dashed line at 92 ft. displays a sharp transition from a mixed sandstone, dolomite, and white limestone truncated by organic mudclasts overlain by coarse-grained sandstone interpreted fluvial in origin. An erosional surface overlying a white, buff, marine limestone which is sharply transitional into mudclasts within a coarse sandstone matrix, an indication of erosional origin.
Figure 31. Caliche/calcrete and brecciated limestone regolith interpreted as the Mississippian-Pennsylvanian systemic boundary. (A) A limestone displaying brecciation at the base, silica crusts and nodules. This limestone displays vadose textures (e.g. Fig. 21c). (B) A carbonate breccia containing disseminated green clay, the likely product of karst-dissolution. Both of these samples are overlain by a succession of mudrocks interpreted as paleosols.

SB18 Mudrock Paleosol Succession

The SB18 core is noteworthy due to a heavily altered marine limestone overlain by a prominent mudrock succession from 178-~100 ft. The limestone is heavily diagenetically altered displaying pisolites and remnant foraminifera grains. The limestone is increasingly fractured and dominated upsection that contained by clay-fill (Fig. 31). The color of the clay fill varies from moderate yellowish green (10GY 5/2) to dark yellowish green (10GY 4/4). A similar relationship is observed in SB10A (Fig. 31b). Directly overlying limestone is a reddish-orange (10R 6/6), fractured to blocky mudstone interval (Fig. 32b). Polished/slickensided surfaces occur along fracture partings. There are also unstained gray siltstone matrix in these units. In thin-section the red-mudstone contains fractures filled with opaque iron-oxide mixed with a siltstone-mudstone matrix (Fig. 21d) also containing circular iron nodules. The red fractured mustone is overlain sharply by an interval of light-medium gray (N7-N5) horizontally bedded claystone (Fig. 21d). The “bedding” displays layers separated by distinct horizons. This is then overlain by a dark reddish brown (10R 3/4) nodular horizon. Upsection is a bed exhibiting a variegated grayish-purple (5P 4/2) matrix with greenish-yellow circular downward directed mottles (10Y 6/6)
occurs. This deposit is a calcareous, nodular, mudrock (Fig. 33d) distinct compared to samples lower in the section. Nodules are composed predominantly of silica, though softer gypsum nodules are also present. Gypsum is apparently more commonly replaced by silica (e.g. Wilson and Pittman, 1978). Intense mottling was observed in the section overlying the nodules. Upward the higher chroma, red facies grades into low-chroma white, gray, and black (N9-N1), carbonaceous, and sideritic mudstone and coal deposits. Plant microfossils recovered from depth 138 ft. indicate an Atokan age of deposition (Eble, Pers. Comm., 2012).

The previously described mudrock succession in the SB18 core contains similar attributes of ancient soils reported in the literature (e.g. Retallack, 1988, 1991; Mack et al., 1993; Kraus, 1999) including those from the late Chesterian aged Pennington Formation located in eastern Kentucky reported by Kahmann and Driese (2008). The underlying carbonate strata displays heavy diagenetic alteration consistent with meteoric digenesis. A similar relationship is observed in the Michigan basin in the SB18 core. The core preserves ~150 ft. of facies corresponding to the Bayport interval, which is capped by a prominent buff marine carbonate bed displaying features indicative of exposure (Fig. 30a). The caliche textures and extensive silicification in the limestone support a predominately semi-arid environment (Estaban and Klappa, 1983; Retallack, 1997). The overlying breccia deposit likely represents subaerial exposure processes related to related to the influx of relatively low pH waters resulting in dissolution. The limestone is overlain by a prominent bed of red mudstone which is interpreted as a vertic oxisol. Reddening is likely due to the oxidation of iron-oxides in the soil horizon. Burial reddening occurs as well (Retallack, 1997), however the presence of the deposits within the stratigraphic succession is suggestive of a lateritic origin, supporting surface
weathering. Vertic shrink and swell features (Fig. 32b) indicate periodic wetting and drying resulting in the expansion and contraction of clay minerals (e.g. Mack et al., 1993; Retallack, 1997). Gleyed intervals displaying horizonation may represent the illuviation of clay materials through the soil column, a process associated with the creation of an argillaceous horizon and the formation of argilosols (Fig. 32c). The low-chroma drab-gray color indicates a reducing environment, likely associated with hydric conditions. The overlying sample contains a nodular evaporite and carbonate horizons displaying distinctive mottling (reduction haloes), vertic fractures, and nodules of carbonate, evaporite, and silica (Fig. 33d) associated with calcisols, gypsisols, and vertisols (Mack et al., 1993). Nodules are interpreted to represent the precipitation of sulfates and carbonate either due to downward mobility of dissolved ions in the vadose zone or the upward movement of water-table brines due to capillary draw (Mack et al., 1993). The presence of nodular beds overlying the relatively low-permeable clay-bed suggests a perched water table resulting in the accumulation of water heavily saturated with dissolved solids. Such horizons may indicate periodic changes in the water table related to eustasy (Miller and Eriksson, 1999), climate (Cecil, 1990; Kahman and Driese, 2008) and/or localized water-ponding (sensu Beuthin and Blake, 2002). In the SB18 core this paleosol grades into carbonaceous shales, rooted clay horizons, and coal of Facies S2. The underlying green paleosols (Facies B3b) found within the Bayport displays similarly lineated vertic textures, though not as pronounced, either due to a shorter duration of pedogenesis or to the lack of preservation.

Overall, the presence of distinct soil horizons, structures, and zones of rooting strongly support a major ancient soil profile and possible significant hiatal surface. This surface is interpreted as the manifestation of the Mississippian-Pennsylvanian
Kaskaskia-Absaroka systemic boundary and provides evidence of a significant base-level fall following the deposition of the dominantly marine influenced, mixed-siliciclastic/carbonate Bayport interval. The stratigraphic distribution of climate sensitive features indicate an overall arid textural overprint (e.g. caliche, lateritic-vertic-oxisols, and gypsic/calcic horizons). Overlying gleyed soils and carbonaceous deposits formed in a persistently wet environment. This climate dislocation supports the occurrence of the Late Mississippian-Early Pennsylvanian sequence/systemic boundary (e.g. Cecil, 1990; Kahmann and Driese, 2008).

The age and climatic implications of these strata are consistent with regional interpretations of climate change (e.g. Cecil, 1990; Cecil et al., 1985) and eustasy (e.g. Ross and Ross, 1988; Haq and Schutter, 2008) associated with the Mississippian-Pennsylvanian boundary. Regionally, this boundary is transitional into a predominately humid-terrestrial deposit (e.g. coals, siderite, and rooted underclays), similar to what is reported from Late Mississippian aged coeval Appalachian basin strata (i.e. Pennington Formation) (Kahmann and Driese, 2008), consistent with observation made from the Michigan basin.

Paleosols are commonly found associated with unconformities and reflect landscape stability (Kraus, 1999). Driese et al. (1994) noted the presence of extensive paleosol development overlying a marine limestone within the Pennington Formation. Furthermore the SB18 core is located proximal to the Lucas Fault structure, possibly indicating a tectonic control on the formation and/or preservation of the paleosol deposit. A similar relationship was reported in eastern Kentucky by Ettensohn and Peppers (1979). In general, the formation of paleosols at unconformites reflect an allogenic forcing (e.g. Kraus, 1999).
Figure 32. A representation of the mudrock paleosol succession observed in the SB18 core. The deposit is sharply transitional from a brecciated carbonate unit to a redbed paleosol displaying a number of soil horizons and textures. High chroma deposit generally grade into gleyed carbonaceous and coal bearing strata of the Saginaw Formation. This soil profile is interpreted to represent changing climate conditions associated with the Mississippian-Pennsylvanian systemic boundary (red line).
Paleosols have interpretative value and may be used to reveal chemical, physical, and biological conditions of past environments (Retallack, 1991; Kraus, 1999). The recognition of distinctive syndepositional and post-depositional textures, structure, and rock types are fundamental to interpreting ancient soil horizons (sensu Retallack, 1997). In general, paleosols form on stable landscapes not subject to periods of widespread erosion (Kraus, 1999). While predominantly siliciclastic, paleosols have been associated with underlying karst limestone regoliths (Retallack, 1997). Retallack (1988) notes three main features which distinguish paleosols from other rock-types including: root traces, soil horizons, and soil-structure. Root traces are perhaps the most distinguishable attribute of certain soil horizons (Retallack, 1988).

While the identification and interpretation of paleosols is ongoing in the scientific literature, paleosols have practical significance for the interpretation of relative sea-level and may have allostratigraphic significance (Kraus, 1999). The identification of ancient soils is extremely important in sequence stratigraphy, where they represent a relative fall in sea-level and may denote sequence boundaries (i.e. van Wagoner et al., 1990). Furthermore, paleosols are extremely significant for paleoclimatic interpretations (e.g. Cecil, 1990; Kahmann and Driese, 2008; Kraus, 1999).

Laterites are characterized by "surface crusts," possibly due to precipitation from groundwater and commonly form under alternating conditions of high and low rainfall (i.e. wet and dry climate conditions) and form on a wide range of rock-types (Retallack, 1997). Lateritic soils are predominately composed of hematite (Fe₂O₃) with lesser amounts of gibbsite (Al₂(OH)₃) and kaolinite (Al₄(Si₄O₁₀)(OH)₈). They are commonly associated with oxisols (Retallack, 1997). Hematite nodules and
opaque iron-oxide filled fractures are clearly present in thin-section at 175 ft. (SB18) (Fig. 33b), further suggesting a laterite deposit for the vertic-oxisol deposit formed by pedogenic processes associated with a Bayport limestone regolith at 178 ft. Further analysis utilizing x-ray diffraction (XRD) would confirm the mineralogy of this deposit.

The mudrock deposit described above is interpreted to represent a pronounced period of subaerial exposure, resulting in pedogenesis and soil formation on the exposed Bayport Limestone. In general, paleosols of this type and "maturity" occur on stable landscapes developed over millions of years forming heavily cemented calcrete, silcrete, laterite, and bauxite deposits (Kraus, 1999; Rettallack, 1997). The deposition of this paleosol indicates a pronounced fall in relative sea-level representing a sequence boundary. The vertic/oxic and argillaceous paleosols are transitional into drab (gleyed) mudstone and coal deposits (35a,b), strata, consistent with much more humid and disoxic conditions during the deposition of the Pennsylvanian Saginaw Formation (e.g. Lane, 1902; Kelly, 1936; Venable, 2006).
Figure 33. Dominant pedotypes overlying the Bayport interval. (A) A limestone regolith displays clay-filled dissolution fractures attributed to the influx of meteoric water during sub-aerial exposure. (B) Maroon-red, mudstone with a fractured, slickensided texture interpreted as a vertic oxisol. The cross sectional view displayed at the bottom of the picture identified as Fe₂O₃ filled fractures and possible rhizoliths. (C) Gleyed mudstone with horizontal clay coated partings (cutans) interpreted as argillisol. (D) High chroma, mottled, vertic texture with silica/gypsum nodules.
Figure 34. Views of prominent soil-textures in core. (A) Bedding plane of a high chroma, mottled texture with small rhizoliths? Gray mottling (i.e. reduction haloes) is interpreted to be the product of plant rhizoliths (e.g. Retallack, 1997). (B) Bedding plane view of a variegated mottled yellow/dark purple calcareous vertic mudrock. Red arrow marks the distinctive yellow/green mottling indicative of disoxic pedogenic conditions. (C) Polished (slickensided) surfaces of an interpreted vertic oxisol deposit.

Figure 35. SB18 carbonaceous paleosol deposits of the Saginaw Formation. (A) Leached claystone (seat-clay) displaying organic matter traces, evidence of rooting. (B) Dark silty carbon-rich (not coal grade) mottled carbonaceous deposit composed of a mixture of silt and organic (plant) material.
A terrestrial to marginal marine environment of deposition has long been recognized for the Pennsylvanian Saginaw Formation in the Michigan basin. In a M.S. thesis focused on study of shallow bedrock core material in Mason, MI (Fig. 10) near the type section of the Saginaw and Grand River formations at Grand Ledge, MI, Venable (2006) used the alluvial model to explain the lateral heterogeneity of Atokan aged units (verified by palynology) located in core and exposures at Grand Ledge. Venable (2006) suggested that generally fine-grained Saginaw and coarse-grained sandstone-dominated Grand River strata are facies of one another. These distinct Saginaw facies are roughly synchronous and not separated by a chronostratigraphically significant hiatus, as contended by Kelly (1936). The lithologies encountered within Atokan portions of SB16, SB18, and SB15 are consistent with observations of the Saginaw Formation by Venable (2006) and Venable et al. (2010).

Facies S1: Coarse-Grained Quartz Sandstone

Description: Yellow-to-white, massive-to-cross-bedded, medium-to-very coarse-grained quartz sandstone is present in strata at a sharp contact overlying the Bayport interval in 4 of the 8 cores analyzed from the SSC site. Approximately 10-20 cm brecciated intervals contain angular to rounded mudstone and/or siderite clasts ranging 1-3 cm in diameter are commonly present at the base and transition into cross-bedded sandstone. These mudstone breccia deposits are similar to the mudclast breccias of Facies B6 within the Bayport. However, the breccias of Facies S1 do not grade into either a carbonate or evaporitic shale and these deposits are interbedded
with gray to black mudrock and carbonaceous shale (Facies S2). Noteworthy is sandstone consisting of pebble sized quartz clasts disseminated within a medium-to-coarse-grained matrix (Fig. 36a). This pebble deposit is found exclusively in the SB15 core where it forms a thick succession, ~100 ft. thick, and capped with shale in a distinct fining upward trend. Organic remains, pyrite, siderite, and contorted bedding structures occur commonly in this facies. Chert pebble sandstone directly overlying the Bayport facies is unique to the SB15 core indicating a localized occurrence of these coarse-grained deposits. In SB13 and SB16 this facies occurs as medium-coarse sandstone without the occurrence of chert pebbles, consistent with what was observed by Venable (2006) from Atokan-aged deposits found near Grand Ledge.

**Interpretation: Channel Fill.** This facies is interpreted as a terrestrial-marginal marine channel/bay-fill deposit. The presence of siderite rip-up clasts, organic material and a close association with the mudstone/shale Facies S2 is strongly suggestive of a terrestrial/marginal marine environment of deposition. This facies occurs most prominently in SB16, SB15, and SB13. Mudclast breccias commonly form the base of channel deposits due to the erosion/collapse of partially lithified over bank deposits (Facies S2) and localized occurrences of pebble sized chert fragments within discrete fining upward (3-4 cm) bedsets indicating waning flow conditions and bedload deposition under tractive transport conditions common in braided rivers. These types of deposits are similar to what is found in a number of modern fluvial environments (e.g. the Amazon River Basin). Regionally, similar, likely coeval strata are found at the base of the Pennsylvanian system including the Caseyville Formation (Illinois basin), the Pottsville (Sharon Conglomerate member, Ohio), and the Lee (Kentucky) and Pocahontas Formations (Western Virginia) of the
Appalachian basin (Peppers, 1996). Wispy laminations are often present in these deposits and may indicate the periodic extension of tidal energy into a terrestrially dominated alluvial plain during periods of low river discharge. The occurrence of heterolithic bedding exhibiting a low-diversity estuarine trace-fossil assemblage (Greb and Chesnut, 1996; Greb and Martino, 2004) is also consistent with facies in the Saginaw Formation in Michigan. Facies models for Pennsylvanian strata from the fluvial-estuarine transition of the Appalachian basin are discussed in detail by Greb and Martino (2004) and have been applied to Atokan aged Michigan basin deposits by Venable (2006).

Heterolithic intervals are present in deposits overlying Facies S1 and have been reported in Pennsylvanian (Atokan) deposits (Venable, 2006). The thickest occurrence of these deposits occurs in the SB15 core, the western most SSC core, where the deposit is approximately 120 ft. and consists of multiple discrete fining-upward successions capped by Facies S2. In this core, the clastic/carbonate (Bayport) succession is uncharacteristically thin (~50 ft.). A similar relationship occurs in SB13 where an erosional surface separates a white fossiliferous marine limestone (Facies B1) and Facies B4 and B5 from coarse-grained sandstone (Facies S1) (e.g. Fig. 30). Mud-clasts are noted at the contact between the silty sandstone and the coarse-grained sandstone unit that also displays quartz (chert) pebbles. In the SB16 core the sandstone is thin (<10 ft.), while the mixed clastic/carbonate interval is thick (~100 ft.), and is represented as a pebbly contorted-bedded quartz sandstone displaying oxidized horizons 6-7 ft. thick, thin mudstone paleosol deposits, and heterolithic intervals. Coarse-grained sandstone is interpreted to be laterally equivalent to paleosol and overbank deposits (Facies S2) encountered in the eastward SB18 and SB10A cores.
Figure 36. Facies S1 hand specimens. Coarse-grained sandstone deposited as a fluvial/tidal channel fill. (A) Coarse grained sandstone with pebble sized clasts overlaying a scour surface with large 4-5 cm diameter mudclast at the base overlying fine-grained sandstone. (B) Mudclast Breccia. Interpreted cutbank deposit. (C) Mudclast found within a medium to coarse-grained sandstone matrix similar strata described in facies B6. Mudclasts often form the base of this facies succession. Palynomorphs from strata underlying from pictures B and C clearly indicate a Pennsylvanian (late Morrowan) age (Ravn, Pers. Comm., 2012; Eble, Pers. Comm. 2012).

Facies S2: Gray Mudstone, Carbonaceous Shale, and Coal

Description: Light gray-to-black (N8-N1), silty shale displaying red/orange concretions make up the bulk of Facies S2. Facies S1 occurs overlying the mixed carbonate/siliciclastic deposits of Facies B1-B6. Chalky white claystone beds displaying traces with of organic matter (Fig. 35a, 37c) occur interbedded with the carbonaceous shale in the SB18 and SB16 cores. White (N9) claystone displays contorted bedding and vertically orientated organic matter traces (Fig. 35a) and are generally overlain by carbonaceous shales or coal. The latter can be discerned by a vitreous appearance. Coal seams range in thickness from 5-12 in. Yellowish-orange (10YR 6/2-10YR 6/6) staining is locally common and dull red/orange (10 R 3/4-5R 4/6) nodules are commonly found along bedding planes within the dark carbonaceous shale deposits (Fig. 37). Plant material and small disarticulated fossils (Lingula sp.) are occasionally present along shale bedding planes.

Interpretation: Floodplain and paleosol deposits. Carbonaceous shale and coals are interpreted as marginal marine to terrestrial swamp deposits. Dark
carbonaceous deposits containing plant material along bedding planes (Fig. 37c) and indicates a terrestrial-marginal marine environment of deposition. Mud-filled vertically orientated tubules as root traces (e.g. Fig. 37c) are common. Rettalack (1997) notes that root macropores are commonly filled by overlying sediment prior to deposition (Retallack, 1997). Drab red, siderite (FeCO₃) concretions are consistent with previous analysis of the Saginaw Formation (Venable, 2006). Locally, siderite occurs as massive (~6 in) beds. Siderite is commonly found in modern freshwater swamp sediments (Postma, 1982) and in brackish marginal-marine deltaic environments (Bhattacharya and Walker, 1992). Mozley (1989) reports that the purity of siderite diminishes within the marine environment resulting from extensive Mg and Ca substitution into the crystal structure. Geochemical analysis of the siderite concretions observed in the SB16 core should be undertaken to better document the true nature of the deposits. The carbonaceous and siderite deposits grade vertically to and from gleyed horizons displaying root traces locally overlain by thin (2-5 in) bituminous coal seams. Coal is assumed to indicate the presence of a histic epipedon, an organic rich soil layer which accumulates at or near the land-surface is classified as histosols (Mack et al. 1993; Retallack, 1997). Histosols form in reducing environments under conditions of poor drainage with the local water table at or near the land surface for prolonged periods (Mack et al., 1993). Siderite nodules and traces of terrestrial plant material indicates deposition within brackish to freshwater floodplain supporting an alluvial environment of deposition. Often sandstone Facies S1 underlies, or is interbedded with Facies S2 as is the case in the SB15 core.
Figure 37. Facies S2 hand specimens. (A) A core with interbedded histolic paleosols (gray/gleyed), carbonaceous shale (darker color), and coal (labeled) interpreted as alternate between subaerial soil and brackish floodplain deposits. (B) Close-up of Carbonaceous shale with interbedded dark red/orange nodules of siderite. (C) Root traces, note vertically orientated organic matter fill. (D) A leaf imprint found along a bedding plane in carbonaceous shale, leaf litter.

Facies S3: Heterolithic Bedded Sandstone/Shale

_Description:_ Facies S3 is a fine-to-medium-grained, locally finely laminated, quartz sandstone/shale displaying a similar bedding style as the Michigan Formation. This facies was most notably observed in the SB15 core where the coarse sandstone Facies S1 grades into heterolithic bedding of Facies S3. The bedding appears to be composed of a heterolithic mixture of fine-coarse-grained sand and siltstone layers, appear rhythmically bedded, and occur in a number of variants including flaser, wavy, and lenticular bedding styles depending on the sand/silt ratio (e.g. Fig. 38a-c). Sandstone intervals display thin laminations and cross-bedding, similar to what is reported in Facies B4 of the Bayport interval. The facies contains disseminated iron (likely siderite) and pyrite nodules. These beds either grade into carbonaceous shale of Facies S2 or are sharply truncated by a mudclast conglomerate which grades into Facies S1.

_Interpretation: Estuarine Channel Fill._ Heterolithic bedding, as noted earlier in the text, is a product of rapidly changing current velocity and commonly forms in restricted tide-dominated environments (sensu Reineck and Singh, 1980). Heterolithic bedding is reported from the Pennsylvanian (Atokan) Appalachian basin by Greb and
Martino (2005) who interpret a tidal influence at the headward portion of an estuary in a zone referred to as the fluvial-tidal transition zone (sensu Dalrymple and Choi, 2007; van den Berg et al., 2007). Venable (2006) presented a similar interpretation for deposits from the Saginaw Formation (Atokan) Michigan basin. The fluvial-tidal transition zone is influenced by a number of marine and non-marine (i.e. fluvial) processes which influence the distribution of grain-size and formation of primary sedimentary structures (e.g. Dalrymple and Choi, 2007). The presence of heterolithic bedding and thick-thin laminae (double-mud drapes) suggests a tidal influence (sensu Dalrymple, 1992; 2010c). The presence of low diversity, small circular to horizontally elongate burrows present in this facies was noted by Venable (2006) who interpreted a brackish depositional environment for the Facies S3.
Figure 38. Heterolithic Facies S3 hand specimens. (A) Burrow mottled mudstone displaying a low-diversity ichnofacies. (B) Wavy cross-bedding in medium grained quartz sandstone. Note the occurrence of amalgamated “double” mud-draped laminae. (C) An example of heterolithic flaser bedding. (D) Another example of flaser bedding located upsection sharply overlain by a mudclast breccia (green line).
Table 4. Summary of lithofacies representing the Pennsylvanian Saginaw and Grand River Formations.

<table>
<thead>
<tr>
<th>Facies/ Lithology</th>
<th>Occurrence</th>
<th>Color</th>
<th>Texture/ Grain-type</th>
<th>Sedimentary Structures</th>
<th>Biota</th>
<th>Interpretation</th>
</tr>
</thead>
<tbody>
<tr>
<td>S1</td>
<td>Variable, Locally overlays Bayport facies</td>
<td>White to yellow locally red-orange</td>
<td>Poorly sorted with local occurrence of “quartz pebbles” (SB15, SB03)</td>
<td>Locally cross-bedded. Basal mudclasts.</td>
<td>Carbonaceous. Possible root traces</td>
<td>High energy, bedload dominated fluvial channel deposits.</td>
</tr>
<tr>
<td>Quartz Sandstone</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>S2</td>
<td>Commonly occurs above the Bayport facies</td>
<td>Gray-black</td>
<td>Fine grained, siltstone/mudstone, locally slickesided</td>
<td>Siderite nodules are common along bedding planes. Vertically orientated organic clasts</td>
<td>Root traces common and leaf imprints</td>
<td>Marginal marine Swamp</td>
</tr>
<tr>
<td>Interbedded Carbonaceous shale, blocky mudrock, and coal deposits</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>S3</td>
<td>Commonly observed within the Saginaw lithofacies</td>
<td>Black and white</td>
<td>Coarse to fine/silty sandstone</td>
<td>Wispy-bedded mixture of quartz sandstone and siltstone/shale flasers. Locally oxidized and pyritized.Localized tidal bedding (Reineck and Singh, 1980)</td>
<td>Carbonaceous material. Brackish trace-fossil fauna</td>
<td>Mixed flow regime, high and low-energy deposits. Mixed fluvial-tidal/estuarine deposits?</td>
</tr>
<tr>
<td>Heterolithic sand/siltstone</td>
<td></td>
<td></td>
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</tbody>
</table>
Incised Valley-Fill Deposits

The recognition of incised valley deposits has become increasingly important in recent years (e.g. Boyd et al., 2007; Zaitlin et al., 1994). Incised valleys are erosional landforms, which form by fluvial erosion during periods of relative sea-level lowstands (Zaitlin et al., 1994; Boyd et al., 2006). The most prominent recent example is the Mississippi River System, which was eroded during last glacial maximum (~20 Ka) (Van Wagoner et al., 1990). Zaitlin et al. (1994) notes that the presence of incised valleys may be the only evidence of lowstand (LST) to early transgressive (TST) deposition within shallow-marine ramp settings. Incised valleys may be several miles to tens of miles in width and commonly form in response to changing base-level (Van Wagoner et al., 1990). Van Wagoner et al. (1990) describes two distinct periods of evolution in the formation of incised valley-fill deposits due to allocyclic mechanisms: (1) erosion and channelizing of sediment to a lowstand shoreline and (2) filling of the valley during a subsequent sea-level rise or transgression. When sea-level decreases below the shelf, incision may proceed across the entire area into a network of submarine canyons at the shelf-break (Van Wagoner et al., 1990; Zaitlin et al., 1994). These valleys may be subsequently drowned and subject to marine processes during periods of high-stand. These processes result in a number of depositional environments and systems associated with incised valley deposits including deltas, tidal-flat, and estuarine systems (Van Wagoner et al., 1990; Dalrymple et al., 1992). Within incised valley fill deposits marine tidal strata transition landward into a system of coarse-grained fluvial deposits which may then transition laterally along depositional strike into sub-aerial exposure surfaces including paleosols (Van Wagoner et al., 1990).

Incised valley fill deposits are observed in Late Mississippian strata within the
Michigan basin and locally at the Mississippian-Pennsylvanian sequence boundary, consistent with the above regional depositional models. Incised valley-fill deposits are likewise reported in the Chesterian Illinois basin, where they are interpreted to mark sequence boundaries (e.g. Smith and Read, 2001; Nelson et al., 2002). Two forms of incised valley-fill deposits are recognized in the Michigan basin including, inter-system boundaries, which form the boundaries between carbonate and siliciclastic lithofacies of the Bayport. These valley-fill deposits comprise tidal deposits which often grade upwards into paleosols and carbonates. These boundaries may reflect pronounced periods of relative sea-level low-stand and are of key allostratigraphic significance (Zaitlin et al., 1994; Smith and Read, 2001). The presence of incised valley fill deposits within the Late Mississippian system is consistent with what is reported by Leetaru (2000), Smith and Read (2001), and Nelson (2002) in the neighboring Illinois basin. Incised valley-fill deposits are recognized throughout the Late Mississippian Illinois basin, and have been interpreted as being the product of low-stand deposition, along with paleosols that form high frequency sequence boundaries (Smith and Read, 2001; Nelson et al., 2002).

Incised valley-fill successions have been observed throughout the Carboniferous Midcontinent (Archer et al., 1994; Martino, 2004; Aitken and Flint, 1994), associated with the Mississippian-Pennsylvanian boundary in the Illinois basin (Kvale and Barnhill, 1994). In the Illinois basin the basal Pennsylvanian system is marked by a series of incised valleys directly overlying Mississippian strata (Kvale and Barnhill, 1994). These valleys range from a kilometer to 30 km in width and ~150 m in depth (Kvale and Barnhill, 1994). A similar relationship is observed in cores capturing the Mississippian-Pennsylvanian systemic boundary in Michigan...
Erosion of the valleys due to erosional scour acts to increase accommodation space and thus increases preservation potential in the geologic record (Zaitlin et al., 1994; Boyd et al., 2006). These drowned valleys may subsequently become estuaries and are affected by marine processes (e.g. tide and wave) especially down-dip at the marine basin margin with fluvial discharge at its landward head (Dalrymple et al., 1992). Estuarine environments constitute a wide range of depositional sub environments and processes which are not yet fully understood (i.e. Dalrymple and Choi, 2007; van den berg et al., 2007) which may have hindered ancient interpretations of estuarine strata. Valley incision within the Bayport interval is evidenced by a shoaling carbonate succession overlain by either a karst remnant, green clay residuum (soil) truncated by fine-medium tidally bedded CaCO$_3$ cemented sandstone displaying carbonate and chert rip-up clasts (e.g. Fig. 23a,b).

Facies Distribution

All of the conventional cores analyzed in this study contain a succession of repeating clastic and carbonate strata ranging from ~50 to 140 ft. in thickness, which occupy a stratigraphic position between the underlying, evaporitic Michigan Formation and superjacent terrestrial deposits of the Saginaw/Grand River Formations (Facies S1-S3). The thickest observed occurrence of Bayport strata occurs in the SB18 core where the succession approaches 150 ft. in thickness while in the adjacent SB15 core the Bayport is ~65 ft. thick (Fig. 39). This disparity is interpreted as being the product of an incised valley fill succession preserved in the
SB15 core, which apparently truncated much of the Bayport Interval in the core.

Limestone units encountered in the study consist of high-energy packstone to grainstones (Facies B1), which display a dominantly open marine fossil assemblage and that is transitional into finely laminated and mudcracked microbialites (Facies B2). These successions are punctuated above by subaerial exposure surfaces. Sub-meter carbonate packages are topped by pronounced subaerial exposure horizons.

In other locations, including SB18, a ~22 ft. thick carbonate succession is overlain by a breccia/paleosol deposit (Facies B3). However, these paleosol deposits are not commonly preserved resulting in hiatal surfaces. These surfaces may be due to erosion during relative sea-level low-stands and valley incision (Smith and Read, 2001) and represent a pronounced dislocation of facies type compared to the underlying carbonate/clastic intervals.

Three distinct carbonate facies were observed within the Bayport interval and are inter-bedded with siliciclastics, sandstone and mudstone deposits. Carbonate packages range from ~2-10 ft. in thickness. Each core observed contains carbonate packages and some component of an idealized “shallowing upward” facies geometry grading from subtidal, intertidal, and finally supratidal deposits up-section. Subaerial exposure processes including desiccation, karsting and incipient soil formation progressively alter the tops of each package. Diagenetic processes including dolomitization and silicification are likewise most prevalent at the top of these carbonate packages. It should be noted that this is the idealized facies patterns and significant deviations from the ideal commonly occurs. Such discrepancies may be due low of preservation potential of these lowstand systems tract deposits similar to relationships reported in the Illinois basin by Smith and Read (2001).

The base of siliciclastic Facies B4 is commonly marked by erosional surfaces
(Fig. 23a,b) (Facies B4). The cross-bedded quartz sandstone units display characteristics similar to many modern and ancient tidal deposits (e.g. reactivation surfaces, mud-drapes, and rhythmic thick-thin laminations). Facies B4 is commonly topped by a buff white, altered marine limestone typically ~10-15 ft. in thickness as observed in core from SL418, SL423, SL18, SB13 and SB11. Vugrinovich (1984) refers to this unit as the “Six Lakes Limestone” (after the “Cream limestone” described by Tyler (1980)) and mapped its distribution using cuttings and geophysical logs. The age of this unit is poorly constrained, but was interpreted by Vugrinovich (1984) as Pennsylvanian in age with little evidence. Overlying this limestone in the SB18 and SB10A cores, is a red and/or olive-gray vertic mudstone (Facies B3), which grades into carbonaceous shale and coal deposits (Facies S2) of the Saginaw Formation. Facies S2 is inferred to be Pennsylvanian in age on the basis of widespread Pennsylvanian coal in the Illinois and Appalachian basin. Cores SB16, SB15 and SB13 lack a paleosol deposit and contain coarse-grained sandstone unit (Facies S1) directly overlying either fine-grained calcareous sandstone (e.g. SB16, SB15) or a carbonate (e.g. SB13, SB03) of the Bayport.

Vertical-facies-stacking patterns and key stratal surfaces have critical predictive value and are important in developing a relevant depositional interpretation (sensu Walker, 1979). High resolution sequence stratigraphic analysis is used as a tool to integrate depositional environments and key stratigraphic surfaces (Van Wagoner, 1990; Catuneanu, 2006) in detailed core and wire-line log studies. Key surfaces observed in this study are interpreted to occur within the heterolithic Bayport interval. The intraformational boundaries are considered to be parasequences due to the fact that the packages are bounded by exposure (e.g. paleosol) without significant dislocation of constituent facies. The topmost contact of the Bayport interval below
the Saginaw Formation is interpreted to represent a much more significant disconformity consisting of a marine/restricted marine facies assemblage (Facies B1-B5) overlain by extensive paleosol or a coarse-grained fluvial sandstone of the Saginaw Formation. Figure 39 indicates the variation in stratal relationships interpreted for the Bayport-Saginaw (Mississippian-Pennsylvanian systemic? boundary) observed in core. Figure 40 shows an example of the cyclic stacking pattern/sequence stratigraphic surfaces (green lines) present within the Bayport interval. The evaporitic/dolomitic shale overlying carbonate prone quartz sandstone is best represented in SL418 and SL423. The shale was noted by Tyler (1980) and Vugrinovich (1984), who refers to it as the "Hemlock Lake Member". This facies of the Bayport may represent a return to restricted, low-energy conditions similar to the subjacent Michigan Formation. Evaporitic shale is overlain by a white limestone, referred to as the Six-Lakes Limestone member (Vugrinovich, 1984) and defined as (Facies B1). This cap limestone is locally present in the SCSC cores (SB18, SB10, and SB01) where it apparently survived fluvial erosion along the basin margin. The position of these cores adjacent to the Lucas-Monroe fault-block suggests a possible tectonic control on preservation. In cores SB13, SB15, SB16, and SB03 fluvial sandstone directly overlie Facies B1-B3. This contact is sharp and marked by an erosion surface below terrestrially dominated facies of the Saginaw Formation. Based on these facies interpretation, this contact is the Bayport-Saginaw boundary. A similar relationship is observed in the Illinois basin by Kvale and Barnhill (1994) who interpreted an incised valley deposit at the Mississippian-Pennsylvanian boundary.
Figure 39. Facies successions observed in the SB18 and SB15 cores. The Bayport has radically different thicknesses interpreted to have resulted from valley incision and local preservation of interfluvial soil horizons. The SB18 core is located adjacent to the Lucas Fault while the SB15 core is located ~16 mi to the west. In the SB15 core the Bayport interval is truncated by very coarse-grained fluvial sandstone (Facies S2) while the Bayport is overlain by a paleosol. Bayport Facies Succession

As noted in the introduction, cyclothems are believed to be the product of changes in accommodation space controlled by a glacio-eustasy (Heckel, 1986; Miall, 2010b) and possibly crustal flexure (Miall, 2010b). The Kansas “type” cyclothem is carbonate dominated and is interpreted to represent a relatively “pure” eustatic signal (Wanless, 1962). In contrast, the Appalachian basin accommodation was predominantly influenced by tectonic loading and flexural response of the lithosphere, attributed to Acadian and Alleghenian orogenic events (Miall, 2010b). The observed succession in the Michigan basin is comparable to the Illinois basin described by Smith and Read (2001) and Nelson et al. (2002). However, a number of differences are apparent including the lack of a well-developed open-marine facies including the fossiliferous-marine shale units which have been interpreted as transgressive deposits by Heckel (1984) and Smith and Read (2001). However, the notable lack of these marine shale units may be explained by the regional up-dip location of the Michigan basin resulting in a high degree of topset erosion. Smith and Read (2001) note that due to the updip location of the Illinois basin, highstand carbonate deposits were subjected to extensive erosion and sub-aerial modification during times of relative sea-level lowstands. Observations of Michigan basin strata are consistent with this previously mentioned interpretation.
Figure 40. An example facies succession observed within the SB18 core. The succession shares similar attributes to midcontinent cyclothem (e.g. Heckel, 1984). Basal sandstone units grade upward into carbonates overlain by paleosols (Facies B3) sharply overlain by tidally bedded sandstone intervals (Facies B4 and B5).
Four cores (SB13, SB15, SB16, and SB18) were used to discriminate the Mississippian-Pennsylvanian boundary in the Michigan basin. Figure 41 displays a cross-section displaying the four cores as well as two more basinal cores. The SB18 core is positioned directly over the Lucas fault and displays extensive paleosol development overlying the Bayport Limestone. In adjacent cores fluvial sandstones are interpreted to erosionally truncate the upper portion of the Bayport Interval. The Mississippian-Pennsylvanian systemic boundary is interpreted in the cross-section (Fig. 41). The age of the strata was determined through the use of sampling and analysis of pollen and spores (next chapter).
Figure 41. An interpretative cross-section (A-A’) constructed using described core-sections from the SCSC cores located in southern Michigan and two partial intervals (with geophysical logs) capturing the Bayport interval located in a more basinal setting to the north in Northern Montcalm County. Red tabs denote the depths at which biostratigraphic samples were taken. The Mississippian-Pennsylvanian unconformity (red line) was discriminated based on facies analysis (Chapter 3) and approximately constrained using biostratigraphy. The Mississippian-Pennsylvanian boundary is manifested as either an erosional surface overlain by coarse sandstone and shale and underlain by either a carbonate (Facies B1, B3) or a medium-fine grained sandstone unit (Facies B4, B5). The SB18 core shows no sign of fluvial incision. In this core the Bayport Interval is capped by a buff/white limestone upon which rests a paleosol succession, which is then overlain by mixed carbonate-siliciclastic strata of the Bayport. This relationship is similar to what is observed in the basinal SL418 and SL423 cores in the central Michigan Basin. Green lines denote high frequency sequence boundaries within the Bayport Limestone.

Depositional Environments

Models taking a sequence stratigraphic approach have the best chance of predicting the deposition of sedimentary rock types of ancient carbonate deposits. Realistic constraints must be placed on depositional models which synthesizes tectonics setting, sea-level cycles, paleo-geography, depositional topography, climate, and diagenesis. Outcrop locals may be used to better understand the stratigraphic unit in question, but due to the lack of these exposures in most circumstances, the study of analogous modern environments may be the only option for understanding the lateral distribution of facies. Studies of ancient carbonate shelf deposits including the Illinois and Permian Basins, USA, have supported the sequence stratigraphic approach and the use of modern analogs for predictive utility (Leetaru, 2000; Smith and Read,
2001; Grammer et al., 2004). It is important to note that no stratigraphic model can fully account for the heterolithic nature of the units in question. This section is an attempt to integrate modern and inferred ancient depositional systems into a realistic depositional model for the Bayport interval.

**Carbonate Dominated Environments**

Shallow marine carbonate deposition is fundamentally different from that of siliciclastic deposits which are derived from extra-basinal terrains. In contrast, carbonates are largely dependent on in-situ sediment production referred to as the "carbonate factory" (Hanford and Loucks, 1993). The precipitation of carbonate is governed by the liberation of CO$_2$ from the water resulting in the oxidation of bicarbonate to carbonate (Wilson, 1975). Mechanisms that result in an increased rate of carbonate precipitation include increased temperature, high rates of evaporation, marine upwelling, bacterial decay and photosynthesis (Wilson, 1975). The transport of carbonate grains is controlled (similar to siliciclastic deposits) by the threshold of tractive transport and the velocity at which grains come to rest (i.e. settling velocity), but bedforms produced by wave and tidal activity are much less likely to be preserved compared to their siliciclastic counterparts (Scoffin, 1987). Likewise, grain-size may be controlled by biologic controls (i.e. organism’s shell-size) rather than directly by the current velocity (Scoffin, 1987).

Carbonate rocks occur as a mixture of calcite and dolomite and are extremely sensitive to sea-level, climate, and diagenesis (Wilson, 1975; Tucker and Wright, 1990). Most, but not all, carbonate strata preserved in the geologic record is derived from marine organisms, which are most productive in warm, clear, and circulated
water located within 30° latitude of the equator (Wilson, 1975). This is due to the prevalence of photozoan organisms, which are most productive within the photic zone extending to between 80 and 100 m under conditions of low turbidity (Jones, 2010).

In general, carbonate deposits are predominantly intrabasinal and are transported short distances from the growth site (Wilson, 1975). These deposits occur within a variety of shallow marine bathymetric settings including homoclinal and distally steepened ramps, rimmed shelves dominated by an abrupt slope break, and isolated platforms such as the Bahamas (Read, 1985; Ahr, 1973). Ahr (1973) was first to acknowledge the gently dipping (<1°) Persian Gulf as a modern homoclinal ramp. Carbonate ramps are commonly interpreted for Mississippian aged, epicratonic carbonate deposits of the St. Genevieve, Newman, and Greenbriar Limestone of the eastern Midcontinent (Smith and Read, 2001; Nelson et al., 2002). Smith and Read (2001) describe a regional carbonate ramp extending from Virginia to New Mexico during the Late Mississippian. Modern examples of carbonate ramps (including the Persian Gulf and the Shark Bay homoclines) dip at low angles (<1°) and are characterized by a number of distinct facies belts, which are governed by a combination of wind and marine circulation patterns and regional location (Purser and Seibold, 1973; Logan and Cebulski, 1970). These facies zones include tide and storm dominated coastal marine environments characterized by high energy skeletal grainstone facies, which commonly extend into low-energy, fine-grained fossiliferous shale deposits in a more distal (i.e. deeper) environment (Ahr, 1973; Read, 1985).

Carbonate producing organisms are extremely sensitive to the influx of siliciclastic strata (Wilson, 1975). Scoffin (1987) notes that calcium-carbonate secreting organisms may be inhibited by the influx of terrestrial derived elastic
sediment through a number of mechanisms: (1) diluting existing carbonate deposits, (2) resultant in unfavorable substrates for benthic organisms, (3) increase in turbidity reducing light penetration to the sea-floor, and (4) tendency to be associated with a reduction of salinity, which inhibits carbonate precipitation. Carbonate grains and laminated dolomite deposits constitute the majority of the carbonate strata observed in the Bayport interval and are interpreted to represent conditions intermittently favorable to marine organisms. The heterolithic nature of these deposits, presence of microbial structures, and the widespread occurrence of sub-aerial exposure surfaces within the Bayport strongly suggests carbonates within the Bayport Interval were deposited within the subtidal to supratidal zones.

Shallow marine carbonate deposits, including tidal flats, have been recognized in a variety of low-latitude coastal environments including the Great Bahama Bank (e.g. Andros Island), the Persian Gulf, and Shark Bay, Western Australia (Shinn, 1983a). Carbonate tidal flats most commonly occur in low-declivity ramp environments sheltered from the impacts of waves (Shinn, 1983a) and represent the shallow water interface between land and sea. Tidal flats are extremely susceptible to small fluctuations in sea-level and climate regime which may result in pronounced subaerial exposure surfaces. The impact of diagenesis is dependent on a number of factors including sediment composition, the hydrology of interstitial fluids, as well as physio-chemical processes that are depositional and post-depositional (diagenetic) in origin (Scoffin, 1987).

Packages of carbonate strata comprising marine fauna (e.g. Facies B1) are interpreted to reflect biological, chemical, and physical conditions suitable for the precipitation of carbonate rocks similar to what is found in a number of modern carbonate dominated depositional environments and settings (e.g. Ahr, 1973; Wilson,
1975; Read, 1985). This study compares depositional features observed within the Bayport interval to modern tidal flats and shallow marine (sublittoral) settings in a number of carbonate and clastic dominated environments including the Persian Gulf, Shark Bay, Bay of Fundy, and the Bay of Mont. St. Michel. The study of ancient analogs, particularly the intercontinental Persian Gulf and Shark Bay (Western Australia) provides insight into the depositional processes and facies architecture observed in the Bayport Limestone and related strata.

The Arabian Gulf

The Persian Gulf is a semi-enclosed, semi restricted foreland basin that is host to a number of carbonate dominated marine and restricted-marine depositional environments (e.g. Kendall and Skipwith, 1969; Purser and Seibold, 1973; Alsharhan and Kendall, 2003). The modern environments along the coast of the United Arab Emirates (i.e. Trucial Coast) in particular, have been widely studied and are used as a modern analog in the interpretation of ancient strata (sensu Alsharhan and Kendall, 2003; Shinn, 1983a).

The Bayport Limestone was compared to arid region carbonates of the Persian Gulf by previous workers (Bacon, 1971; Ciner, 1988). Deposition within the 500 km long and 60 km wide Persian Gulf occurs in a wide but restricted embayment connected to the Gulf of Oman and the Indian Ocean by the narrow straits of Hormuz, which is ~80 km in width (Purser and Seibold, 1973). Deposition is primarily controlled by bathymetry and regional climate patterns such as humidity and wind-direction (Purser and Seibold, 1973). The Persian Gulf is separated into two geological provinces by the axis of the basin: (1) the tectonically stable Arabian Foreland, and (2) the tectonically active Iranian fold-belt, which is present along the
northern coast of the Gulf area and provides a source of terrestrially derived detritus from estuaries and marshes located along the northwestern margin of the restricted embayment (Purser and Seibold, 1973).

The Persian Gulf has an asymmetrical bathymetric profile and consists of a gently northeast sloping (35 cm/km) Arabian side, which is abruptly transitional to a steeper basin profile (175 cm/km) along the Iranian coast and the Zagros fold-belt (Purser and Seibold, 1973). Ahr (1973) and Read (1985) classify the southern part of Persian Gulf as a gently sloping homoclinal ramp consisting of a shoreward sabkha, dominated by microbial ("algal") boundstone, which grades into skeletal and oolitic grainstone/packstone facies that are transitional, downdip, into lower energy pelagic mudstones.

The Persian Gulf is located between 24° and 30° N latitude and the climate is characterized as arid to sub-tropical. Coastal areas of the Persian Gulf are dominated by seasonal fluctuations in water temperature and salinities (Purser and Seibold, 1973; Alsharhan and Kendall, 2003). Salinity ranges from 40 ppt along the deeper portions of the Iranian coast to >100 ppt along the southern Saudi Arabian Coast (Purser and Seibold, 1973). Consequently local salinity, nutrient availability, and temperature are heavily impacted by an anticlockwise, gyreng current, though with little impact on sediment transport (Purser and Seibold, 1973). Average seasonal temperatures range from 45-50 °C in the summer to as low as 0 °C during winter months and the area receives little annual rainfall (<5 cm) resulting in an environment conducive to the formation and preservation of carbonate and evaporite deposits (Purser and Seibold, 1973; Alsharhan and Kendall, 2003).

The climate of the Persian Gulf varies along the coast and local variations in climate have a profound impact on the distribution of sedimentary environments. The
Gulf of Qatar lacks well developed evaporite and microbial laminations compared to the Trucial Coast due to higher rainfall (Shinn, 1983a). Purser and Seibold (1973) note the north/northwest "Shamal" winds result in the stimulation of wave and surface currents which transport terrigenous clastics material sourced from the Zagros Mountains into the southeastern marine realm, including the Trucial Coast (Abu Dhabi). Tidal currents trend along the axis of the gulf and attain bottom current velocities of >0.6 m/s in coastal channels, however, the coastal tidal range is relatively small, ranging between 0.5 and 1.5 m (Purser and Seibold, 1973).

Along the southern Gulf (Trucial Coast), a wide variety of depositional structures and textures are represented (Fig. 42). In offshore marine environments not subject to subaerial processes reefs and oolitic/skeletal shoals dominate (Fig. 42). Locally, these open marine shoal and reef facies grade into lower energy lagoons (i.e. Khor al Bazam, Fig. 42; Kendall and Skipwith, 1969). Subtidal facies grade into “algal” mats and sabkha environments in intertidal and supratidal coastal environments respectively (Shinn, 1983a; Purser and Seibold, 1973; Kendall and Skipwith, 1969; Alsharhan and Kendall, 2003). These sabkha environments have been widely studied and are the basis for many depositional models for ancient carbonate successions.

Models for carbonate sedimentation in the southern Persian Gulf have been applied to the Bayport Limestone by Bacon (1971) for exposures at Wallace Stone Quarry, Huron Co., MI. Observations in this study using primarily core material are consistent with the work of Bacon (1971). A predominantly arid environment of deposition is supported by the presence of dolomite, microbial (“algal”) laminations, mudcracks, and the occurrence of gypsum/anhydrite. While the carbonate facies can be explained using the Trucial Coast Analog (Fig. 42) the interbedded siliciclastic
facies observed in the interval cannot be explained solely by looking at this carbonate dominated depositional environment.

Figure 42. Dominant depositional environments of the coast of Abu Dhabi (United Arab Emirates). The system is characterized by a number of facies belts which grade from offshore reefs and coralline/oolitic sands into lower energy skeletal and pelletal lagoonal sands, and into "algal" mats and evaporite prone supratidal sabkha environment in a landward trajectory. The figure highlights the lateral adjacency of a variety of carbonate-dominated environments. From Alsharhan and Kendall, 2003. After Kendall and Skipwith, 1969.

Tide-Dominated Coastal Siliciclastic Environments

While no one modern depositional environment fully represents the broad range of facies present within the Bayport interval, the study of modern environments
remains critical to the understanding of ancient strata. The Bayport, as observed in core, is interpreted to represent a number of depositional environments ranging from carbonate tidal flats, siliciclastic tidal channel deposits, and other more terrestrially dominated systems. Interbedded sandstone, limestone, dolomite and mudstone within the Bayport interval suggest variable and cyclic depositional conditions. A number of modern coastal environments are influenced by tidal currents including tidal flats, estuaries and deltas (Dalrymple, 1992, 2010c). In shallow marine environments tidal range is divided into microtidal (0-2 m), mesotidal (2-4 m), and macrotidal (>4 m) (Dalrymple, 1992). Tidal dominance is dependent on the relative energy of the tides compared to other processes such as storm and wave generated currents as well as overall geomorphology and bathymetry (Dalrymple, 1992).

Processes observed in modern clastic dominated tidal depositional systems (e.g. Bay of Mont St. Michel, Bay of Fundy, the North Sea coast) have been widely studied (e.g. Dalrymple, 1992, 2010; Reineck and Sigh, 1980; van Straaten, 1961; Nyandwi, 1998). These environments exhibit widely diverging morphological and evolutionary attributes (Dalrymple, 1992, 2010c). Processes documented in tide dominated environments are effectively used to explain primary sedimentary structures and facies distribution observed in ancient Carboniferous strata of the North American Midcontinent (Tessier et al., 1995; Archer, 1998; Archer and Greb, 2012), including the Illinois and Appalachian basins (Greb and Archer, 1998; Archer and Greb, 2012; Smith and Read, 2001).

Macrotidal environments, such as the Bay of Fundy and the Bay of Mont St. Michel, are the focus of much recent published literature (Tessier, 1993; Dalrymple, 1992; Dalrymple et al., 1992; Dalrymple and Choi, 2007; Dalrymple, 2010c). Accommodation space formed by fluvial erosion during low-stand conditions is the
dominant mechanism for the preservation of estuarine strata in the geologic record (Dalrymple et al., 1992). By definition, tide dominated estuaries are transgressive in nature (Dalrymple et al., 1992) and tend to (though not always, e.g. Dalrymple and Choi, 2007) form in incised valley features, many of which were fluvially eroded during the last glacial maxima (e.g. Bay of Mont St. Michel, Tessier et al., 2006).

**Estuaries**

Estuaries are unique environments, which record a complex interplay of physical, chemical and biological processes. Prichard (1967) defines an estuary as "a semi-enclosed coastal body of water which has a free connection with the open sea and within which sea water is measurably diluted with fresh water derived from land drainage." However, Pritchard's definition has proven unsatisfactory for stratigraphic interpretation and a definition based on relative coastal energy (i.e. wave vs. tide vs. river) is most appropriate in the geologic sense (Dalrymple et al., 1992). Dalrymple et al. (1992) define an estuary as "a drowned valley that receives sediment from a mixture of tide, wave and fluvial processes and extends to the landward limit of tidal energy and to the seaward limit of coastal facies at its mouth." This definition was modified by Dalrymple and Choi (2007) to encompass abandoned delta plains undergoing transgression which are not incised valleys. Estuaries, by definition, form during sea-level rise which are typically controlled by allogenic mechanism, including glacio-eustasy, which makes them potentially important in sequence stratigraphic studies (Dalrymple et al., 1992; Boyd et al., 2006).
Wave Dominated Estuaries

A wide range of processes are recognized in modern estuarine environments. Dalrymple et al. (1992) use a tripartite classification scheme to divide estuaries on the basis of relative strength of tide, wave, and fluvial energy, and further discriminate wave and tide dominated forms. Wave dominated estuaries are semi-enclosed by barrier bars forming a restricted low-energy, back-barrier bay, which receives input at its landward head via fluvial drainage and terrestrial run-off acts to dilute the salinity of the estuarine bay (Dalrymple et al., 1992). Wave dominated estuaries are characterized by a pronounced "tripartite" grain-size distribution (i.e. coarse-fine-coarse) ranging from coarse grained barrier bars grading into low-energy muddy estuarine deposits, to coarse-grained fluvial deposits orientated toward the headward (i.e. landward) portions of the estuary (Dalrymple et al., 1992).

Tide Dominated Estuaries

In contrast to wave-dominated estuaries, tide-dominated estuaries display a characteristic open funnel shape resulting in exceptionally high tidal current velocities (e.g. Bay of Mont St. Michel, 2m/s) and tidal amplitudes (Dalrymple et al., 1992) due to the cross-sectional decrease in area towards the landward head of the estuary (Dalrymple and Choi, 2007). Tide dominated estuaries consists of an erosional coast transitioning into tidal bars (sand waves) before transitional with a fluvial-tidal transition zone at the estuary head (Dalrymple et al., 1992) (Fig. 42). Tide dominated estuaries lack a well-developed low-energy, interior basin and low energy shale deposits due to the funneling of tidal energy in a landward direction resulting in a landward grain-size fining trend and the development of coastal mudflats (Fig. 42) (Dalrymple et al., 1992). Modern examples of tide dominated
estuaries include the Bay of Fundy and the Bay of Mont St. Michel, which will be discussed later in the Chapter.

Fluvially dominated coastal environments are considered to be progradational and thus deltaic in origin, but estuarine environments also exhibit progradation during the last stages of infilling (Dalrymple et al., 1992; Tessier et al., 2006). While both tide dominated deltas and estuaries have been documented, the differences separating them are yet to be fully understood (e.g. van den Berg, 2007; Dalrymple and Choi, 2007). In general, tide dominated deltas exhibit coarsening upward successions in which delta front sands are reworked by tides and prograde over fine-grained prodelta deposits (Dalrymple and Choi, 2007). These relationships contrast with estuaries and tidal flat systems, which exhibit a generally fining-upward trends and progradation of landward tidal flats (Dalrymple et al., 1992). Sharp scour surfaces overlain by tidally influenced quartz sandstone (Facies B4), which generally fines upward into a heterolithic Facies B5 within the Bayport interval suggests a progradational geometry associated with estuarine tidal channels (e.g. Dalyrmple et al., 1992).

The facies area representing the fluvial-tidal transition zone (Fig. 43) of tide dominated settings is the subject of a number of modern studies (e.g. van den Berg et al., 2007; Dalrymple and Choi, 2007; Tessier and Lanier, 1998) and represents a unique depositional environment. In this setting sedimentary structures are influenced by the complex interaction of tidal and fluvial energy regimes (e.g. climbing ripples; Tessier, 2006). Tessier and Lanier (1998) compared modern occurrences of climbing ripple beds with examples from Pennsylvanian aged strata deposited in the Midcontinent by Chesnut and Greb (1996). Greb and Martino (2004) suggest that deposition within the fluvio-tidal transition was common during the Carboniferous in the Appalachian basin. Similar structures, such as climbing ripples have been
observed within the Bayport Interval (e.g. Facies B6).

Figure 43. A generalized model of a tide-dominated estuary. (A) An Idealized representation of grain-size and sedimentary structures within the subtidal-supratidal siliciclastic tidal-flat, Bay of Fundy. Figures B and C shows the type and relative energy strength and the relative grain-size affecting the cross-section of a tide-dominated estuary respectively. The estuary displays the characteristic tripartite grain-size distribution (C) grading from coarse grained subtidal sandbars grading to into Grain size displays a relatively coarse grain-size at the estuary mouth due to high wave energy, a decrease in grain-size and energy in the middle portion, and an increase in grain-size due to the confluence of tidal and fluvial energy at the estuary mouth. Subtidal and intertidal sands are capped by upper intertidal and supratidal muds and salt marshes). From Dalrymple and Choi, 2007; Modified from Dalrymple, 1992.
Bay of Fundy

Perhaps the most widely studied, modern tide-dominated estuarine environment is the Bay of Fundy. The Bay of Fundy (Dalrymple et al., 1992; Dalrymple and Choi, 2007) is located within an embayment along the Atlantic coast of Nova Scotia. This inlet is structurally bounded by grabens that result in an elongate, southward facing estuary (Dalrymple et al., 1992). The landward head of the estuary is the confluence of the Salmon River, which feeds the estuarine Cobequid bay, at the landward head of the estuary. The Bay of Fundy is macrotidal estuaries with diurnal differences in mean high and low tides of ~14-16 m (Klein, 1970; Dalrymple, 1992, 2010c). This estuary contains a number depositional sub-environments including a high energy mouth, tidal flats bordering the inner coast, and a high energy “funnel,” grading into the fluvial-tidal transition which displays a characteristic straight, meandering, straight channel morphology (Dalrymple et al., 1992; Dalrymple, 1992, 2010c).

Wave energy dissipates landward along the axis of the estuary resulting in a pronounced increase in tidal activity (Dalrymple et al., 1992). The Bay receives sediment input from both tide and wave energy. Klein (1970) reports that tidal velocity reaches 1.1 m/s in the Minas Basin, a predominantly intertidal interestuarine bay. Sedimentation and the formation of sedimentary structures (i.e. estuarine sandbars and reactivation surfaces) are governed by opposite orientated ebb and flood tidal currents.

The Bay of Fundy is the most extensively studied example of a modern, tide-dominated estuary in temperate latitudes. As such, use of Bay of Fundy depositional models as an analogue to climatically dissimilar, Paleozoic estuarine deposits maybe inappropriate. However, the lack of well documented, low-latitude tropical/arid and
tide dominated estuarine environments makes the study of physical processes operating in modern settings essential to the interpretation of ancient sedimentary structures, in the Bayport interval.

**The Mont Saint Michel Bay (MSMB)**

The macrotidal Mont Saint Michel Bay (MSMB) (or Bay of Mont St. Michel) occupies a 500 km² coastal embayment along the Atlantic coast of northwest France (Tessier et al., 2006). The MSMB displays a similar morphology and facies distribution as compared to the Bay of Fundy (Figs. 42 and 43) (Dalrymple et al., 1992).

Tidal energy within the MSMB varies from 5-6 m during neap condition to 10-11 m during spring tides and is sheltered from waves by the Chausey Archipelago (Fig. 44; Tessier et al., 2006). Tidal currents are strongest inside tidal channels located in the eastward head of the estuary where they can be up to 3 m/s and dissipate to the westward portion of the embayment (Tessier et al., 2006). Minor rivers including the Couesnon, Sèe, and Sèlune discharge fine grained sediment to the MSMB with relatively insignificant water and sediment discharge rates (8-15 m³/s) (Tessier et al., 2006). This results in predominantly marine sediment source composed of a mixture of clastic and (50%) bioclastic carbonates (Tessier et al., 2006). The bioclastic component, which is unusually high for temperate environments, is composed of mollusk shells and other calcareous micro-organisms which form subtidal shoals orientated parallel to the coast (L’Homer and Caline, 1990). This contrasts with the perpendicularly orientated intertidal sandbars located in the mouth of the estuary (L’Homer and Caline, 1990). In the upper tidal zone laminated carbonate mud occur resulting from the erosion and maceration of
bioclastic material (L’Homer and Caline, 1990).

Estuaries have proven to be especially important to concepts relating to sequence stratigraphy (e.g. Boyd, 2006; Dalrymple et al., 1992). The MSMB, like other modern estuaries, is located within an incised channel system created during the last glacial maximum and progradational estuarine fill was deposited during the Holocene sea-level transgression (11,700-6,500 yrBp) (Tessier et al., 2006). The MSMB displays a number of environments including the estuary mouth flanked by sandy shoreline environments, tidal channels, creeks, and mudflats (Tessier et al., 2006). Barriers composed of shelly, bioclastic debris occupy offshore barriers to the southwest and east of the inlet and are highly subject to storm reworking and destruction. Tessier et al. (2006) used seismic reflection profiles and facies mapping and report that nearly all of the progradational estuarine fill occurred during the Holocene transgression until 11700-6500 yrBp (Fig. 44) representing the transition from the transgressive systems tract (TST) into the highstand system tract (HST).
Figure 44. (A) The location of the tide-dominated Mont-Saint Michel (MSM) Bay estuary located along the Atlantic coast of Northwest France. (B) The depositional environments and conditions with and adjacent to the Mont-Saint Michel (MSM) Bay which is a tide dominated estuary. Salt marshes (sm) dominate the landward coastal section. Fine-grained salt-marshes grade into coarser grained clastics along the coast. The hatch marks denotes the boundary of Holocene transgressive fill and the outer hard-rock setting. From Tessier (2006). (C) Satellite view of the Bay of Mont Saint Michel from Google Earth.

Ancient Analogue

Carbonate tidal flats are common in the geologic record (e.g. James, 1979). These deposits typically form shallowing upward facies mosaics with basal open marine shoals or lagoons that grade into the peritidal facies (intertidal and supratidal) in vertical section (Fig. 22). The shallowing upward carbonate model proposed by James (1979) is commonly observed in shallow marine carbonate environments and represents a facies succession similar to the carbonate facies present within the Bayport interval. Figure 22 shows the model presented by James (1979) for a carbonate dominated succession commonly observed in ancient environments.

Carbonate peritidal facies have been recognized in Mississippian deposits throughout the mid-continent (Wilson, 1975). During the Early-Late Meramecian (Middle Mississippian), the Illinois basin was transitional from dominantly distal-marine cool water, heterozoan dominated, mud mounds to a warm, shallow-water water photozoan dominated assemblages (Lasemi et al., 2003). Lasemi et al. (2003) contends that the transition was tectonically driven and associated with the reactivation of the Reelfoot rift. As a result diminishing rates of subsidence and a shallowing of the Illinois basin occurred before deposition of a series of mixed
carbonate-clastic sequences beginning during the Early-Middle Chesterian (Smith and Read, 2001). Similar deposits strata are reported in the Appalachian basin by Al-Tawil et al. (2003) and Al-Tawil and Read (2003).

Mixed Carbonate/Clastic Depositional Models

Late Mississippian Midcontinent strata contains a diversity of marine and non-marine facies assembled in a "cyclic" manner (Swann, 1963; Smith and Read, 2001) possibly reflecting a global climate transition from a predominantly arid (Greenhouse) conditions into the Pennsylvanian icehouse conditions (Smith, 1995; Smith and Read, 2000). Smith and Read (2001) and Al-Tawil and Smith (2003) contend that climate-induced, high frequency (fourth order) cycles, were the predominant control on deposition during the Late Mississippian. Smith and Read (2001) were able to correlate paleosols into adjacent sandstone-filled channels, which constitute sequence boundaries. Similarly Al-Tawil et al. (2003) and Al-Tawil and Read (2003) were able to correlate paleosols across much of the Appalachian basin.

The Late-Mississippian Illinois basin is notable for the occurrence of a mixed-carbonate and siliciclastic depositional system composed of facies similar to those encountered within the Bayport interval of the Michigan basin. Green and red mudrock paleosols and caliche-breccias generally cap dolomitized skeletal and oolitic pack-stones in the Chesterian Illinois basin (Smith and Read, 2001; Nelson, 2002). The later Mississippian is dominated by pronounced disconformity surfaces composed of caliche and coarse-grained chert deposits overlying erosion surfaces (Smith et al., 1995), similar to those reported in Chapter 3 from the Bayport. Tidally bedded quartz sandstone in the Illinois basin is generally confined to the lower
portions of the cycles and directly overlies a disconformity surface (Smith, 1995; Smith and Read, 2001).

The cyclic nature of deposition in the Late Mississippian Illinois basin was noted by Sloss (1963), Swann (1963, 1964), and Smith and Read (1999, 2001) among others. The realization that tidal depositional systems, including estuaries, may be more common in the geologic record than previously recognized is in part based on the prevalence of estuarine environments along modern highstand, coastal environments (Dalrymple et al., 1992). This has resulted in considerable research of modern tide-dominated environments and sedimentary features and succession, which may be applied to the interpretation of ancient strata. The construction of a ternary classification scheme proposed by Dalrymple et al. (1992) relating the relative amount of wave, tidal, and fluvial energy has proved useful in this endeavor. Since a clear facies model of wave and tide dominated estuaries were proposed by Dalrymple et al. (1992) (Fig. 43) many examples of ancient strata have been reinterpreted as representing estuarine environments (Boyd et al., 2006).

Illinois Basin: Bethel to Glen Dean Interval

Sequence Stratigraphic analysis of the Chesterian Illinois basin was undertaken by Leetaru (2000), Smith and Read (1999, 2001), and Nelson et al. (2002). Noteworthy occurrences of northeast-southwest trending paleovalley features have historically been interpreted as deltaic (e.g. Swann, 1963, 1964). However, Smith and Read (2001) interpret a series of incised valleys filled with a mixture of siliciclastics and marine carbonates. Smith and Read (2000, 2001) interpret these erosional features as a product of high frequency (fourth and fifth order) eustatic sea-
level changes, a possible response to the onset of Gondwanan glaciations, similar in nature to other coeval rock units throughout the world (i.e. Caspian Region, Smith and Read, 2001).

The incised valleys may be up to 75 m in depth and occupy the stratigraphic position between the Bethel and the Glen Dean, deposited during the Middle Chesterian North American Stage (Nelson et al., 2002; Smith and Read, 2001). The fluvial deposits within these valleys are poorly preserved and valley fill is typically tidally bedded quartz sand and estuarine deposits of the transgressive systems tract (Nelson et al., 2002). The underlying St. Genevieve Limestone is composed of a series of oolitic shoals and is marked by a significant exposure surface overlain by paleovalley deposits between the St. Genevieve and superjacent Bethel Formation (Smith and Read, 2001; Nelson et al., 2002). Figure 45 displays the Smith and Read (2001) depositional model for the Middle Chesterian Bethel to Glen Dean Interval. Transgressive deposits are overlain by a “shallowing” upward carbonate facies succession, which is commonly capped by siliciclastic tidal-flat and terrestrial paleosol deposits. These paleosols can be laterally traced into correlative incised valleys (Nelson et al., 2002; Smith and Read, 2001). Due to the up-dip location of the Illinois basin low-stand deposits (i.e. paleosol and fluvial facies) are poorly preserved due to erosion (Nelson et al., 2002; Smith and Read, 2001).

Smith and Read (2001) divide the interval into a series of basin-wide sequences denoted by paleosols and/or incised valley deposits. These same authors suggest that pentcontemporaneous faulting occurred during this time which resulted in differential subsidence rates, particularly in proximity to the Cincinnati Arch. Smith and Read (2001) use the depth of incision to estimate the magnitude of sea-level fall and have created a sea-level chart for the Illinois Basin Interval.
Figure 45. A depositional model depicting the Late Mississippian (Early-Middle Chesterian) Glen-Dean to Tar Springs Interval, Illinois basin. The system is composed of a mixture of carbonate and siliciclastic strata. Except for the widespread presence of fossiliferous shale the lithologies and depositional environments are interpreted to be similar to those observed and interpreted for the Bayport interval within the Michigan basin (this study). From Smith and Read, 1999, 2001.

This study documents similar rock successions within the Michigan basin, which are often marked by erosion and the development of paleosols (e.g. caliche, breccia). A prominent sandstone interval, referred to by Vugrinovich (1984) as the "Parma" sandstone displays characteristics of incision including the presence of a sharp, scoured surface overlying carbonate rocks. This sandstone is well preserved in the central basin SL423 and SL418 cores and consists of tidally bedded calcareous quartz sandstone (Facies B4 and B5). These sandstone units erosionally overlie a
Bayport Interval Integrated Depositional Model

A depositional model for the mixed siliciclastic/carbonate Bayport interval was constructed based on the local and regional occurrences of diagnostic lithofacies and is shown in Figure 46. Carbonate Facies (B1-B3) of the Bayport were deposited in coastal environments similar to those interpreted by Bacon (1971) and in a similar facies association as observed for carbonate strata of the Trucial Coast (Fig. 42). Siliciclastic facies are interpreted as in incised valley-fill deposits based on the presence of pronounced basal erosion, the presence of tidal rhythmite and heterolithic bedding, and the presence of microbialites within the siliciclastic facies. These facies were deposited within a tidally influenced, embayed coastal environment, likely a major estuary in the North American epicratonic seaway.

The occurrence of microbial laminations, birdseye fenestrae, and evaporite minerals, particularly anhydrite, strongly support a peritidal depositional environment for the Bayport. Inter-bedded white/buff limestone and brown, dolomitic strata occur throughout the study interval. The contact between these units is often sharp and is associated with fractures, vugs, intraclasts, and green crusts. These dolomites are then overlain by erosional, scoured surfaces, a mudstone, or a caliche/breccia. Furthermore, the observations made are similar to previous descriptions and interpretations of the Bayport Limestone in outcrop (particularly at Wallace Stone Quarry) by Bacon (1971) and Ciner (1988) who interpret a supratidal sabkha to sub-littoral (subtidal) environment of deposition for the Bayport. This study interprets, in a tide-dominated, estuarine environment, for sandstones of the Bay (sensu Dalrymple
et al., 1992). Siliciclastic intervals typically comprise of fining upward siliciclastic successions, which are topped by open-marine limestone of Facies B1 and restricted dolostone of Facies B2.

Figure 46. An idealized facies model for the Bayport interval. Subtidal carbonate shoal and sand bodies of Facies B1 grade in a landward trajectory to laminated dolomite (Facies B3), tidal sand-flats of facies B4, mixed flats of Facies B5, channel Facies B6, evaporitic low-energy mudflat deposits of Facies B7.
CHAPTER IV

PALYNOLOGY

Introduction

Stratigraphic correlation using pollen and spores has proven suitable for Carboniferous aged, particularly Pennsylvanian strata (Playford and Dino, 2005). Palynology is useful for chronostratigraphic interpretation. The SB16 core was the focus of pollen sampling due to the large vertical recovery (>550 ft.) in this core including several stratigraphic units and also because the more basinward position relative to other SCSC cores. Analysis and chronostratigraphic interpretation were performed by Dr. Robert Ravn, IRF consulting group.

Useful pollen and spores were recovered from two principle intervals: (1) the Mississippian Michigan Formation and (2) coal bearing strata of the Pennsylvanian Saginaw Formation. Miospore recovery from a third shale interval, in the SL418 and SL423 cores establishes a Chesterian age for the Bayport interval. Age determinations of characteristic pollen and spore assemblages/specimens are constrained by similar biota within the Carboniferous elsewhere Midcontinent. The documentation of pollen and spore types within Carboniferous strata represents an increase in understanding of the Mississippian-Pennsylvanian Michigan basin and the stratigraphic units therein. The biostratigraphic interpretations presented here indicate that the Chesterian stage is represented in the Michigan basin, a significant departure from the existing stratigraphic nomenclature (e.g. Catacosinos et al., 2000; Fisher et al., 1988).
**Chronostratigraphic Framework**

The Carboniferous system is formally divided into two periods, the Mississippian and Pennsylvanian (Heckel and Clayton, 2006). These periods are divided into smaller time-units including series and stages (e.g. Fig. 47). In the U.S Midcontinent, the Mississippian Period is divided, from oldest to youngest, into the Kinderhookian, Osagean, Meramecian, and Chesterian stages. The overlying Pennsylvanian is divided into the Morrowan, Atokan, Desmoinesian, Missourian, and Virgilian stages (Fig. 47). Stratigraphic Stages are commonly defined by rock-units and the fossils they contain (Peppers, 1996) and are generally correlated using biostratigraphy (i.e. macrofossils and/or microfossils). During the Carboniferous land-plants dominated large portions of the Midcontinent and Western Europe and presently Carboniferous and Permian strata are best correlated using spore and pollen specimens (Playford and Dino, 2005). Ettensohn and Peppers (1979) use plant microfossils to delineate the Mississippian-Pennsylvanian boundary in the eastern Appalachian basin. The chronostratigraphic interpretations presented here are based on the study of pollen and spores undertaken by Dr. Robert Ravn, IRF Consulting Group, and secondarily Dr. Cortland Eble of the Kentucky Geological Survey.
Figure 47. Current Carboniferous Michigan Basin chronostratigraphy. Vertical lines represent periods of non-deposition or erosion. No Chesterian (Late/Upper Mississippian) fossils have been recovered prior to this study. Relative ages are from Gradstein et al., 2004. Regional stages are from Heckel and Clayton, 2006. Michigan basin lithostratigraphic relationships modified from Fisher et al., 1988. Time (Ma) based on global stages from Gradstein et al., 2004.
Carboniferous Plant Biota

Carboniferous land plants have been observed in plant impressions and as plant mega and microfossils. Carboniferous palynomorphs and miospore (pollen or spore <200 µm in diameter) assemblages primarily originated from lower vascular plants including lycopods, horsetails, and ferns (Kosanke, 1969; Playford and Dino, 2005). Pennsylvanian terrestrial vegetation has been divided into five broad groups including lycopods, ferns, seed ferns, sphenopsids, and cordaites (Peppers, 1996). During the Carboniferous, particularly the Pennsylvanian period, a humid-tropical climate predominated (Cecil, 1990). Hydrophytic plants adapted to wetland conditions include lycopod trees whose fossil plant genus includes *Lepidophloios*, *Lepidodendron*, and *Paralycopodites* (Peppers, 1996). Lycopsids, genus *Lycospora*, were especially adapted to wetland conditions and have been widely observed in Pennsylvanian coal flora while tree ferns and seed ferns (*Sigillaria, Psaronius*, and *Medullosa*) grew in more arid, sandy substrates (Peppers, 1996). During the Early Pennsylvanian, lycopods were the dominant plant group with tree ferns more prevalent and eventually dominant during the Late Pennsylvanian (Desmoinesian, Missourian, and Virgilian regional stages; Peppers, 1996).

Classification of Ancient Pollen and Spores

A Miospore is a generic term for a pollen or spore less than 2 mm in length (Kosanke, 1969). Spore specimens are divided according to morphology and are separated into three basic divisions based on observed symmetry and include: (1) bilateral and monolete (*Apiculatasporites*), (2) radial/trilete (*Apiculatisporites*), and alete (Ravn, 1986; Kosanke, 1969). Kosanke (1969) reports that radial/trilete spores
were the most abundant spore-type occurring within Carboniferous strata. Bilateral spores referred to as the fern genus *Laevigatosporites* are also relatively common in Pennsylvanian aged assemblages (Kosanke, 1969). Alete forms lack a central aperture and are comparatively less common during the Carboniferous (Kosanke, 1969).

**Carboniferous Biozones**

Late-Paleozoic biozonal schemes of the Appalachian basin include both megafloral and microfloral assemblages (Eble et al., 2009). The relative age of a species or assemblage is stratigraphically constrained by recording the first and last occurrence of the single or group of specimens (e.g. Peppers, 1996). Based on this criteria individual pollen and spore genera and species are placed within a spore assemblage zone which is then used for biostratigraphic correlation based on similar regional and global distribution patterns (e.g. Peppers, 1996). Mississippian midcontinent spore assemblages of the Pennington and Mauch Chunk Formations lack clear correlation to the discrete Western Europe stages. Unfortunately, the lack of clear spore assemblages discourages direct comparison to Europe (Eble et al. 2009). For a recent summary of regional and local biozones of the Pennsylvanian Midcontinent see Peppers (1996) and Eble et al. (2009).

**Preservation**

The preservation of pollen and spores requires a highly reducing environment due to the rapid decomposition of organic matter in the presence of oxygen (i.e.
oxidizing/atmospheric conditions). Anoxic wetland environments such as swamps and bogs are most suitable for the preservation of pollen and spores and the formation of peats which were eventually transformed into coal under burial conditions. During the Pennsylvanian, Eastern North America experienced a predominantly humid climate clearly evidenced by organic-rich coal and shale deposits (Cecil, 1990). In the Saginaw Formation, these coal and carbonaceous deposits are commonly observed and display plant imprints along bedding surfaces along with occasional, rooted paleosol deposits (e.g. Fig. 37).

Michigan Basin Samples

Pollen spore taken from core SB16 represent the basis for stratigraphic correlation and age interpretation of the strata within the Michigan-Saginaw Formation interval. The vertical extent of the SB16 core enables a clear vertical profile of palynoflora from the Michigan basin. The core yielded three diverse species assemblages which conform to Chesterian, Morrowan, and Atokan North American stages (Ravn, Pers. Comm., 2012) upsection. A Chesterian miospore assemblage has not been recorded in the Michigan basin and a palynological analysis of Morrowan strata is sparse to non-existent in the literature. The Atokan assemblage observed in this study is directly comparable with samples obtained in shallow subsurface bedrock core (Americhem site) and exposures at Grand Ledge reported by Venable (2006). As briefly mentioned earlier, the Chesterian miospore assemblages obtained from Midcontinent strata lack a definite correlation to European spore
assemblages but can be broadly correlated with other occurrences of Chesterian miospore assemblages throughout the midcontinent region, with an albeit lower resolution (Eble et al., 2009). The Chesterian Assemblage was recovered from interbedded silt/shale lithofacies in the Michigan Formation.

The SB16 core was chosen as a biostratigraphic reference (Fig. 41) because of its basinward location relative to the other SCSC cores, the presence of the most complete stratigraphic section, and the thickest occurrence of Michigan Formation strata. The fine-grained, low-energy, terrestrially influenced strata of the Michigan Formation were sampled in two locations which yielded a spore assemblage of Chesterian age. Three attempts were made to acquire useful from green clay horizons within the mixed siliciclastic/carbonate Bayport Interval directly overlying the Michigan “type” lithology, but no dateable samples were recovered. Overlying the Bayport in SB16 is a thick gray-black bed of carbonaceous shale from the overlying Saginaw, which provided the most diverse species assemblage. Four other cores (SB15, SB18, SL423, and SL418) provided microflora assemblages which could be dated. Six of the samples were analyzed by Dr. Cortland Eble of the Kentucky Geological Survey (Table 5).

The SB16 core was thoroughly sampled and “yielded diverse and well-preserved samples of pollen and spores ranging from Late Mississippian to Early Pennsylvanian age” (Ravn, Pers. Comm., 2012). The specimens displayed in Fig. 49 document three distinct miospore assemblages representing Chesterian, Morrowan, and Atokan North American depositional stages with no evidence of post-depositional reworking (Ravn, Pers. Comm., 2012). The analysis of the Chesterian miospore assemblage is noteworthy due to the rare occurrence of terrestrial vegetation in the Appalachian basin because of marine inundation during deposition
of the Greenbriar and Newman Limestone formations (Eble et al., 2009).

The following is an adaptation and stratigraphic interpretation based on a detailed analysis performed by Dr. Ravn, who supplied photomicrographs and preliminary chronostratigraphic interpretations in a report (Appendix B). The interpretations were made using what is known of ancient Chesterian, Morrowan, and Atokan spore and pollen assemblages from the North American Midcontinent reported by Ettensohn and Peppers (1979), Ravn and Fitzgerald (1982), Ravn (1986), Peppers (1996), and Eble et al. (2009).

Table 5. Summary of results of the palynological analysis conducted by Dr. Robert Ravn (SB16 core) and Dr. Cortland Eble (the remaining 6 samples). Asterisk (*) marks the samples described by Dr. Cortland Eble, Kentucky Geological Survey.

<table>
<thead>
<tr>
<th>Core</th>
<th>Depth (ft.)</th>
<th>Lithostratigraphic Unit</th>
<th>Age (N. American Stage)</th>
</tr>
</thead>
<tbody>
<tr>
<td>SB16</td>
<td>618</td>
<td>Michigan/Marshall?</td>
<td>Chesterian</td>
</tr>
<tr>
<td>SB16</td>
<td>598.5</td>
<td>Michigan FM</td>
<td>Chesterian</td>
</tr>
<tr>
<td>SB16</td>
<td>542</td>
<td>Michigan FM</td>
<td>Chesterian</td>
</tr>
<tr>
<td>SB16</td>
<td>498</td>
<td>Bayport</td>
<td>N/A</td>
</tr>
<tr>
<td>SB16</td>
<td>447.75</td>
<td>Bayport</td>
<td>N/A</td>
</tr>
<tr>
<td>SB16</td>
<td>429.5</td>
<td>Bayport</td>
<td>N/A</td>
</tr>
<tr>
<td>SB16</td>
<td>408.9</td>
<td>Saginaw</td>
<td>Morrowan</td>
</tr>
<tr>
<td>SB16</td>
<td>280.4</td>
<td>Saginaw</td>
<td>Atokan</td>
</tr>
<tr>
<td>SB16</td>
<td>228</td>
<td>Saginaw</td>
<td>Atokan</td>
</tr>
<tr>
<td>SB15</td>
<td>129</td>
<td>Saginaw/Grand River</td>
<td>Atokan?</td>
</tr>
<tr>
<td>SL418</td>
<td>825</td>
<td>Bayport</td>
<td>Late-Chesterian</td>
</tr>
<tr>
<td>SL418</td>
<td>810</td>
<td>Bayport</td>
<td>Late-Chesterian</td>
</tr>
<tr>
<td>SB15*</td>
<td>248</td>
<td>Michigan FM</td>
<td>Chesterian</td>
</tr>
<tr>
<td>SB15*</td>
<td>129</td>
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</tr>
<tr>
<td>SB18*</td>
<td>154.5</td>
<td>??</td>
<td>N/A</td>
</tr>
<tr>
<td>SB18*</td>
<td>138</td>
<td>Saginaw/Grand River</td>
<td>Atokan</td>
</tr>
<tr>
<td>SL423*</td>
<td>820</td>
<td>Bayport</td>
<td>Late-Chesterian</td>
</tr>
<tr>
<td>SL-217A*</td>
<td>1285.3</td>
<td>Marshall/Michigan?</td>
<td>Late-Chesterian</td>
</tr>
</tbody>
</table>
Figure 48. An example stratigraphic section from central Ingham County. The SB16 core (left) represents a continuous coring of the strata. The core on the right displays a partial core and wireline log suite from Montcalm County 72.6 mi to the northwest. Biostratigraphic samples were taken and denoted by the red tabs. The biostratigraphic analysis of the SB16 core constrains the Parma-Bayport interval to between the Chesterian (latest Mississippian) and Morrowan (earliest Pennsylvanian) regional stages. Interpretation of spores recovered from SL418 suggests a later Chesterian age for the Bayport Interval in the central Michigan basin. (Ravn, Pers. Comm., 2012)

Chesterian Miospore Assemblage

A Chesterian miospore assemblage was recovered from the Michigan Formation from depths 618 ft., 598.5 ft., and 542 ft. These samples *Spelaeotriletes triangulus*, *Neoraistrickia* sp. cf., *N. loganii*, and a number of forms of the genus *Vallatisporites* sp. (Fig. 48, pictures 1-8). These species and genera are interpreted by Ravn (Pers. Comm., 2012) as Chesterian age and are directly comparable to the Pennington Formation of Eastern Kentucky reported by Ettensohn and Peppers (1979). The Chesterian miospore specimens possess low taxonomic diversity (Ravn, Pers. Comm., 2012), attributed by Playford and Dino (2005) as a less evolutionarily developed assemblage present during the Mississippian compared to the Pennsylvanian Period. Depositional environments play a large role in the accumulation and preservation of palynomorphs and predominantly shallow marine depositional systems generally lack microflora as in the Chesterian Greenbriar and Newman Limestone Formations within the Appalachian basin (Eble et al., 2009). Eble et al. (2006) note that significant ecological changes occur between Mississippian and Pennsylvanian time, reflecting a predominantly arid environment compared to humid species common during the Pennsylvanian.
Samples from shale beds (Facies B7) within the SL418 core from depths 810 and 825 ft. are reported by Dr. Ravn as highly unusual and a radical departure from the miospore population reported in the SB16 core. The palynomorphs are reported as the genus *Retusotriletes*, which are present in Mississippian and Devonian strata (Ravn, Pers. Comm., 2012). However, the presence of *Lycospora pusilla*, a species prevalent in Upper Mississippian through Middle Pennsylvanian strata, is interpreted to represent a Late Chesterian age of deposition for the SL418 shales and the absence of *Lycospora pusilla*, from the SB16 samples is interpreted as indicating an earlier Chesterian age (Ravn, Pers. Comm., 2012). Based on the analysis of pollen and spores the Michigan Formation and the overlying Bayport Interval were likely deposited during the Chesterian Regional Stage. The previous statement is also supported by regional lithologic comparisons of the interval (i.e. the presence of carbonates and evaporites) which suggest deposition under an arid climate regime consistant with observations of Late Mississippian strata (Cecil, 1990).

**Morrowan Miospore Assemblage**

The Morrowan palynomorph assemblage was taken from depth 408.9 ft. of the SB16 core, at the base of a thick (~150 ft.) mudstone/shale bed directly overlying a sandstone bed interpreted as part of the Bayport. The mudstone is interpreted to overlay an erosional surface above a basal ~10 ft. bed of coarse-grained, iron-stained, quartz cemented sandstone. The mudstone grades upwards from a light gray color, with sulfide flakes along bedding planes, to a dark gray to black sideritic carbonaceous shale (Facies S2).

Figure 49 displays key species from these samples. The sample contains *Dictyotriletes bireticulatus, Densosporites anulatus, Densosporites irregularis,*
Radiizonates striatus, Cristatisporites indignabundus, Savitrisporites concavus, and Savitrisporites nux (Fig. 48, picture 9-14; Ravn, Pers. Comm., 2012). Ravn (1986) notes that the base of the Morrowan is characterized by the genera Lycospora, Densosporites, and Radiizonates. The species found in the upper parts of the SB16 core compare favorably to those identified from coal beds found within the Caseyville Formation of Iowa, Western Interior Basin reported by Ravn and Fitzgerald (1982) and Ravn (1986). Based on the presence of these spores, the sample indicates a Morrowan age for the fine-grained carbonaceous strata overlying the mixed carbonate-siliciclastic interval (Ravn, Pers. Comm., 2012).
Figure 49. Selected photomicrographs of Carboniferous miospores from the sampled SB16 core intervals. (1) Vallatisporites sp. (598.5'), (2) Vallatisporites sp. (598.5'), (3) Vallatisporites sp. (598.5'), (4) Vallatisporites sp. (598.5'), (5) Retusotriletes sp. (598.5'), (6) Neoraistrickia loganii, (Winslow) Coleman & Clayton 1987 (598.5'), (7) Calamospora sp. cf. C. liquida Kosanke 1950 (598.5'), (8) Spelaeotriiletes triangulus, Neves & Owens 1966 (598.5'), (9) Densosporites anulatus (Loose) Schopf et al. 1944 (408.9'), (10) Densosporites irregularis, Hacquebard & Barss 1957 (408.9'), (11) Cristatisporites indignabundus, (Loose) Potonié & Kremp 1954 (408.9'), (12) Radiizonates striatus, (Knox) Staplin & Jansonius 1964 (408.9'), (13) Laevigatosporites desmoinesensis, (Wilson & Coe) Schopf et al. 1944 (408.9'), (14) Savitrisporites concavus, Marshall & Smith 1965 (408.9'), (15) Grumosisporites varioreticulatus, (Neves) Smith & Butterworth 1967 (280.4'), (16) Lycospora pusilla (Ibrahim) Schopf et al. 1944 (280.4'), (17) Lycospora noctuina (Butterworth & Williams 1958) (280.4'), (18) Ahrensispories guerickei (Horst) Potonié & Kremp 1954 (280.4'), (19) Lophotriiletes sp. (280.4'), (20) Pilosisporites aculeolatus (Kosanke) Ravn 1986 (280.4'), (21) Spackmanites sp. of Ravn & Fitzgerald 1982 (280.4'), (22) Savitrisporites nux (Butterworth & Williams) Sullivan 1964 (280.4'), (23) Bellisporites nitidus (Horst) Sullivan 1964 (280.4'), (24) Cordylosporites papillatus (Naumova) Playford & Satterthwaite 1985 (280.4'), (25) Florinites pumicosus (Ibrahim) Laveine 1965 (280.4').


Atokan Miospore Assemblage

Ravn (Pers. Comm., 2012) describes an Atokan age for samples at 280.4 ft. and 228 ft. from the SB16 core and at 128 ft. of the SB15 core. An Atokan age of deposition is supported by the occurrence of Savitrisporites concavus, Savitrisporites nux, Grumosisporites varioreticulatus, and Densosporites anulatus. This spore assemblage is comparable to other occurrences of the Atokan stage in the Western Interior Basin reported by Ravn (1986) and from the Illinois basin by Peppers (1996). The presence of Cordylosporites papillatus which is characteristic of Mississippian and Morrowan strata of Iowa, is interpreted as Atokan, based on the occurrence of
species described in sample 280.4 ft. (Ravn, Pers. Comm., 2012) containing spores characteristic of the Atokan North American Stage. At 280.4 ft. the SB16 core yielded specimens including *Vestispora costata* which is indicative of an Atokan age and an isolated occurrence of *Bellisporites nitidus* which is reported to reflect the proximity to the Morrowan-Atokan boundary (Ravn, Pers. Comm., 2012). Notably the species *Lycospora noctuina* and *Lycospora pusilla* are unique to samples 228 ft. and 280.4 ft. (aside from those extracted from two samples studied from SL418), further indicating a Pennsylvanian (Atokan) age. These specimens are comparable to those analyzed by Dr. Ravn and reported by Venable (2006) from Americhem cores and the Saginaw/Grand River Formation outcrops near the town of Grand-Ledge. The Americhem cores are in close proximity (~2 mi) from the SB16 core. Furthermore, these samples were taken from a subsurface depth similar to those of the Americhem cores reported by Venable (2006).

Discussion

Midcontinent Biostratigraphy

Pollen and spore types may differ latitudinally as well as temporally (Playford and Dino, 2005). Peppers (1996) was able to demonstrate key similarities between miospores species recovered from Pennsylvanian strata throughout the Midcontinent to Western Europe and Russia (Donets Basin). Correlation using macro- and microfossils has been undertaken to correlate facies within midcontinent (Western Interior, Illinois, and Appalachian basins) cyclothemic successions, suggesting that individual beds could be correlated for “many miles” (Peppers, 1996). However
debate still exists regarding the depositional controls on midcontinent cyclothems despite such widespread correlation (Miall, 2010b). Many have argued for a glacio-eustatic control (Wanless and Shepard, 1936; Heckel, 1986; Ross and Ross, 1988). Peppers (1996) suggests that the coal “underclay” paleosol may represent the longest time within the cyclothem interval.

Most of the early palynological correlation schemes were based on those derived from Western Europe (e.g. Clayton et al., 1977), which is considered to be “standard” for regional biostratigraphic correlation (Peppers, 1996). However, this scheme has not been generally used for strata deposited within the Illinois and Appalachian basin (Peppers, 1996; Eble et al., 2009). In European nomenclature the Namurian stage is divided into three substages: Namurian A (Early Chesterian), Namurian B (Early Morrowan), and Namurian C (Late Morrowan). Ettensohn and Peppers (1979) use palynology to distinguish between Mississippian coals of the Pennington from overlying Pennsylvanian strata and indicate the presence of a distinct change in spore taxa across the Mississippian-Pennsylvanian boundary, possibly indicating an unconformity. Ravn and Fitzgerald (1982) and Ravn (1986) document Pennsylvanian miospore assemblages from strata located in the Mississippi Valley Region in Iowa and suggest strong correlation with established European biozones. Carboniferous stratigraphic correlation is best constrained for deposits located in Western Europe and is best represented in Clayton et al. (1977), who divided miospores into a number of biostratigraphic assemblages resulting in a much more precise age correlation for Pennsylvanian biota compared to their Mississippian counterparts (e.g. Peppers, 1996). This is indicated in the Michigan basin by the occurrence of two distinct and diverse Pennsylvanian spore assemblages, Atokan and Morrowan, from the intervals sampled. As noted earlier the Chesterian lacks the
taxonomic diversity of the Pennsylvanian system, displaying relatively few genera and species. These characteristics are comparable to other assemblages observed throughout the Midcontinent United States (Ravn, Pers. Comm., 2012). It is beyond the scope of this study to undertake a global comparison of Michigan basin assemblages reported in this study to those of Western Europe.

The greater taxonomic diversity of pollen and spores during the Pennsylvanian period is generally attributed to more humid climate conditions, contrasting with dominantly arid conditions characterizing the Mississippian period (Eble et al., 2006; Cecil, 1990). Pollen and spores have been studied in detail from Midcontinent strata (Iowa) by Ravn and Fitzgerald (1982) and Ravn (1986) who suggest correlation of Midcontinent microflora to the Illinois basin. Peppers (1996) uses the Illinois basin as a stratotype for regional comparison. Peppers (1996) was able to correlate midcontinent Pennsylvanian spore assemblages found within formational units within the Illinois basin to those found in Western Europe and the Russian Donets Basin based on histograms comparing the distribution of species type and relative abundance through the Pennsylvanian regional stages. As a result, the Pennsylvanian period is relatively well correlated to the European continent but the Mississippian is far less understood, presumably due to the lack of widespread terrestrial vegetation and markedly different depositional environments (i.e. more marine; Eble et al., 2009).

Palynological study of the Michigan basin is scarce to nonexistent. Richardson (2006) uses the occurrence of pollen and spores from the Lower Mississippian Coldwater Formation and the overlying Marshall Formation to suggest a Kinderhookian and Osagian age for the respective units based on comparison to known Western European Miospore Assemblages. The analysis of Pennsylvanian
spores from the Saginaw Formation was reported by Venable (2006), with the assistance of Dr. Robert Ravn, who reports an Atokan age for carbonaceous material found in the outcrops of the Grand River Formation at Grand Ledge and subsurface core material located ~20 miles to the southeast, as part of the Americhem industrial remediation site. The present study represents the most complete record of pollen and spore taxa within the Carboniferous Michigan basin, especially with respect to the Chesterian and Morrowan depositional stages.

The Chesterian palynomorph assemblages reported here in from the Michigan Basin exhibit a low taxonomic diversity (Ravn, Pers. Comm., 2012) similar to what is reported from coeval strata of the Midcontinent and Europe (Playford and Dino, 2005). In contrast, Pennsylvanian spore assemblages in Michigan are comparatively more “cosmopolitan” with regard to taxonomic diversity observed in spore assemblages present in Pennsylvanian coal-bearing strata (Playford and Dino, 2005). The lack of diversity in Mississippian-aged fauna recovered within strata deposited in the eastern United States may be a function of paleo-geography depositional environments (i.e. greater marine affinity), and possibly climate (Eble et al., 2006; Eble et al., 2009). The most thorough delineation of the Mississippian-Pennsylvanian boundary was performed on the Pennington formation, Eastern Kentucky by Ettensohn and Peppers (1979).

**Mississippian-Pennsylvanian Boundary**

As noted in the introduction the Mississippian-Pennsylvanian boundary is recognized as a significant regional time-stratigraphic boundary (Sloss, 1963), an apparent result of non-deposition across much of the eastern midcontinent (Peppers, 1996). In the past, considerable debate has revolved around the relationship between
the Mississippian and Pennsylvanian strata in time and space (Ettensohn and Peppers, 1979; Ettensohn, 1980). While some investigations have proposed a conformable relationship (Horne et al., 1974), others have been used to support an unconformity surface. Ettensohn and Peppers (1979) analyzed Mississippian aged coal beds forming parts of the Pennington Formation and conclude that a major boundary exists separating Namurian A and Namurian B times (i.e. Chesterian and Morrowan) on the basis of a marked extinction event corresponding to the Mississippian-Pennsylvanian boundary. Peppers (1996) reports that coal beds in the Illinois basin (southeastern Illinois, Brown County) found within paleo-valleys lying at the base of the Pennsylvanian system are likely Early Pennsylvanian in age. Likewise, shale/coal beds resting above the unconformity in Ohio have been determined to be Namurian B (Early Pennsylvanian) in age (Peppers, 1996). Comparisons of conodonts recovered from the Wayside member of the Pennsylvanian Caseyville Formation with those from the underlying Grove Church Formation have been the subject of debate in the literature. Rexroad and Merrill (1979) report taxonomic similarities of conodonts found within the two rock units, suggesting a gradational contact across the boundary. However, Weibel and Norby (1990) indicate taxonomic differences between conodonts found in the two sites suggesting an unconformity does in fact separate Mississippian and Pennsylvanian strata. Peppers (1996) suggest a Namurian B (Early Pennsylvanian) age for the Wayside Shale based on palynomorph taxa.

Stratigraphic Implication

No documented account of Chesterian palynoflora exists in strata from the Michigan basin. A revised stratigraphic framework for the Late Mississippian
Michigan basin is presented in Figure 50 and is based primarily on the analysis of miospores by Dr. Robert Ravn. The documentation of Chesterian aged pollen and spores significantly modify the existing understanding of Mississippian aged strata deposited within the Michigan basin, in particular the Michigan and the Bayport intervals. Based on a Chesterian age for pollen and spores from these formations a significant revision of the existing stratigraphic relationships in of the Michigan basin is indicated.

The Michigan Formation was extensively sampled due to abundant, dark colored, fine-grained deposits interpreted as terrestrially influenced due to the presence of evaporites, oxidation horizons and an estuarine trace-fossil assemblage. The Michigan Formation has significant preservation of pollen and spore assemblages as indicated by the high degree of sample recovery. The relative age of the Michigan Formation is clearly Chesterian (Late Mississippian). The analysis of pollen and spore samples from a shale bed overlying Vugrinovich's "Bayport-Parma" interval has likewise been dated as Mississippian by Dr. Eble, Kentucky Geological Survey.

In conjunction with facies analysis and regional comparison to coeval Mississippian strata in the Illinois and Appalachian basins, these relationships, in total, strongly indicate a Chesterian age (Late Mississippian) for the mixed carbonate-clastic interval above the Michigan Formation, referred to as the "Bayport Limestone" and the informal "Parma Sandstone" in previous work. The overlying strata composed predominantly of fine-grained carbonaceous debris-rich shale established as Morrowan in the SB16 core. In this core mudstone grades into a mixed sand and mudstone/shale succession-lithologically similar to strata documented by Venable (2006) as Atokan in age from core material and outcrop at Grand Ledge
The study of miospores indicates that deposition in Michigan basin was ongoing during the Chesterian North American stage. A significant unconformity does not separate Middle Mississippian (Meramecian) and Early Pennsylvanian strata (Morrowan) at the locales analyzed, as proposed by Swann (1963) and reiterated by Fisher et al., (1988) and Harrell et al. (1991). These previous interpretations were based on a lack of documented Chesterian fossils and the overreliance on inter-regional lithostratigraphic correlations. Furthermore, the Bayport Limestone is constrained to a Chesterian to Morrowan age of deposition. The detailed analysis of facies (previous chapters) suggests deposition under an arid/sub-humid climate consistent with other occurrences of Chesterian strata present in the Illinois and Appalachian basins.

The biostratigraphic data generated in the course of this study was used to constrain the age of strata and the Mississippian-Pennsylvanian boundary within. Due to the lack of appropriate material (i.e. shale and mudstone), an accurate location of the Mississippian-Pennsylvanian sequence boundary is difficult to determine using biostratigraphy alone and a detailed facies analysis was performed to more clearly define the position of the Kaskaskia-Absaroka systemic boundary (Chapter 3). A conformable biostratigraphic succession recording Chesterian, Morrowan, and Atokan North American stages indicates a lack of a chronostratigraphically significant stratigraphic break in Michigan and, contrary to previous interpretations by Swann (1963) and Harrell et al. (1991) deposition was ongoing in the Michigan Basin during the Chesterian Regional Stage. A Morrowan age was determined for carbonaceous shale deposits consistent with the Saginaw Formation overlying the heterolithic mixed carbonate/clastic Bayport Interval. This constrains the deposition (also with the aid of Dr. Ravn).
of the Bayport Interval to between the Chesterian (Late Mississippian) and Morrowan (Early Pennsylvanian). Spores recovered from a shale bed from the SL418 and SL423 cores further supports a Late Chesterian age for the Bayport interval. In the Illinois basin, Chesterian strata is composed of heterolithic, mixed siliciclastic and carbonate lithologies deposited during the Late Mississippian, below the regional Mississippian-Pennsylvanian unconformity. The Chesterian Illinois basin stratigraphy compares favorably to the Bayport interval. Pennsylvanian, Morrowan strata comprises mainly of carbonaceous shale and fluvio-estuarine, channel-fill sandstone units interpreted as a comparatively humid environment deposits, consistent with regional climatic interpretation of Early Pennsylvanian strata throughout the Midcontinent (e.g. Cecil, 1990).
Figure 50. The revised nomenclature for the Late Mississippian (Chesterian) and lower Pennsylvanian Michigan basin strata. The age of the Bayport and Michigan Formation was determined to be Chesterian in age based on the analysis of Pollen and Spores by Dr. Robert Ravn. Similar palynological analysis presented by Venable (2006) was performed by Dr. Robert Ravn for Saginaw Formation and Grand River strata from core and outcrop indicating an Atokan age for both units. The nature of post-Atokan stratigraphy of the Michigan basin is currently the subject of study. Nomenclature from Heckel and Clayton, 2006. Time (Ma) based on global stages from Gradstein et al., 2004.
CHAPTER V

CONCLUSION

The Mississippian-Pennsylvanian Unconformity

The Mississippian-Pennsylvanian unconformity in the Appalachian basin was of particular interest to a number of workers (e.g. Horn et al., 1974; Ettensohn and Peppers, 1979). Most early researchers favored a tabular erosion model. In the tabular erosion model, a large unconformity on the magnitude of millions of years is believed to separate the Mississippian (locally the Pennington Formation) from the overlying Pennsylvanian strata (Lee and Breathitt Formations) in the Appalachian basin.

During the 1970s, contradictive interpretations were made regarding the tabular erosion model in the Appalachian basin, particularly from outcrops in eastern Kentucky. Horn et al. (1974) used facies models and interpretive cross-sections to demonstrate depositional continuity between the Mississippian Newman Limestone, Pennington Formation shales, and the overlying Pennsylvanian Lee Formation. Horn et al. (1974) further postulated that the three units are part of a facies mosaic based on the lateral gradation and inter-fingering of the units observed in outcrops exposed along the then recently constructed Interstate Highway 64. The so-called Lee-Newman Barrier-Shoreline model of Horn et al. (1974) suggested that the Mississippian aged Newman Limestone represents restricted carbonate shoals which were depositionally enclosed by “red and green shales” of the Pennington formation. Orthoquartzite pebble deposits of the Lee Formation were interpreted to represent high-energy barrier islands and channel deposits, which grade into the back-barrier environments composed of sideritic, coal-bearing Breathitt Formation. Due to the interest in the relatively new and innovative, stratigraphic techniques inherent in the
facies-model approach and the interpretation of depositional environments, these ideas regarding the Mississippian-Pennsylvanian boundary were not immediately rejected, though they proved to be controversial (i.e. Ettensohn and Peppers, 1979; Ettensohn, 1980).

Ettensohn and Peppers (1979) demonstrated, through the study of palynology, that a systemic boundary is present and European stages (Namurian A and B European Stages corresponding to Late Chesterian and Morrowan North American Stages) and the laterally restricted occurrence of the Late Chesterian Pennington Formation was attributed to extensive post-depositional erosion (Ettensohn and Peppers, 1979). Further Ettensohn and Peppers (1979) document the occurrence of local syndepositional tectonism along the Waverly Arch and proposed a tectonostratigraphic model to account for the preservation of the Pennington Formation in the study area. Through the biostratigraphic analysis by Ettensohn and Peppers, 1979 the Horn et al. (1974) contemporaneous depositional model was refuted based on key differences in palynomorph assemblages and correlation to their European counterparts.

The Late Mississippian Pennington Formation is the subject of a number of studies. The Pennington Formation is composed of a series of paleosols, siltstone, sandstone, and limestone (Ettensohn and Peppers, 1979; Kahmann and Driese, 2008). Ettensohn and Peppers (1979) report that the Pennington Formation is overlain by a major unconformity upon which rests Pennsylvanian aged coals of the Breathitt group. This unconformity may be conformable in the eastern Appalachian basin due to increased subsidence in that area (Ettensohn, 2004; Greb et al., 2009).

Elsewhere in the eastern Appalachian basin, in western Kentucky, the Mississippian-Pennsylvanian boundary is an unconformable contact between the
Middle Mississippian Newman Limestone and the Pennsylvanian Breathitt Group. In Ohio the systemic boundary occurs between the Maxville Limestone and the overlying Pottsville Formation (Greb et al., 2009). Barnhill and Kvale (1994) note the occurrence of incised valleys marking the Mississippian Pennsylvanian systemic boundary in the Illinois basin. These observations are consistent with those reported in this study in the Southern Michigan basin.

The Chesterian Illinois Basin

Chester deposits were referred to by Sloss (1963) and Swann (1963, 1964) as the Pope Supergroup. Sloss (1963) noted that the Eastern Interior basin (Illinois Basin) was highly influenced by the influx of terrigenous clastic material during the Early-Middle Chesterian. This part of the section is characterized by Swann (1963), who postulates the occurrence of a “Michigan paleoriver” (Fig. 9) which carried sediment from the Canadian Shield into the Illinois basin, bypassing the Michigan basin. This supported work by Potter and Siever (1956), whose interpretation was partially based on Michigan basin deposits. However, Nelson et al. (2002) suggests the Transcontinental Arch (to the north and west) was the primary source of quartz sandstone into the Illinois basin. The analysis of pollen and spores suggests Chesterian strata are indeed present in the Michigan basin refuting Swann’s generalized interpretation of regional deposition during Carboniferous.

During the Meramecian regional stage (Late Visean in European Nomenclature) deep water, low-diversity mud mound accumulations (Walsortian) have been widely reported in the Illinois basin as well as in parts of Western Europe (Ahr et al., 2003). The overlying St. Louis Limestone lies below the base of the Chesterian (Nelson et al., 2002). The St. Louis is a cherty “sublithographic”
limestone containing economic deposits of gypsum (McGregor, 1954; Nelson et al., 2002). During the Chesterian the Illinois basin experienced a pronounced shallowing to a peritidal facies. Lasemi et al., (2003) suggests regional upwelling of nutrient and silica-rich water along the Reelfoot rift, southern Illinois, resulted in significant accumulation of chert deposits. Chesterian aged Illinois basin limestone units have been interpreted by Lasemi et al. (2003) as forming in response to shoaling. This regional shallowing of the basin was related to tectonic uplift resulting transition from a deeper water heterozoan community to a peritidal setting dominated by marine and tidal flat biota (Lasemi et al., 2003).

The limestone facies in the Bayport interval are similar to the Chesterian-aged mixed siliciclastic/carbonate strata of the Illinois basin reported by Swann (1963, 1964) and Smith and Read (2001). Of particular interest is the sharp transition between the Middle Mississippian (Meramecian) St. Louis Limestone and the overlying St. Genevieve Limestone/Beaver Bend to Tar Springs Interval representing a shallow marine, mixed carbonate and clastic dominated succession. Skeletal, oolitic grainstones and laminated dolomites appear to have been deposited in shallow marine, paralic setting bounded by exposure surfaces (Smith and Read, 2001). Quartzose siliciclastics are present in paleo-valleys, which are complexly distributed throughout the Illinois basin. Clastic deposits are primarily confined to the lower portions of each sequence and interpreted to display estuarine tidal filling (Smith and Read, 2001). This time relationship is common in modern estuaries such as the Bay of Mont St. Michel, which underwent incision during the last glacial maximum and subsequent filling during the transgressive system tract (TST; e.g. Tessier, 2006). In the Chesterian Illinois basin Smith and Read (2001) suggest the rapid transition in facies is related to Gondwanan glaciations, which produced high frequency global
sea-level fluctuations (cyclothems). Tidal deposits within the Illinois basin have been noted by Leetaru (2000) and Nelson et al. (2002). The Late Mississippian Michigan basin shows many similar relationships compared to the Illinois basin. The Mississippian-Pennsylvanian boundary is considered by Kvale and Barnhill (1994) as an incised-valley filled with fluvial sandstone of the Caseyville Formation and correlative overbank coal and paleosol deposits. Incised valleys have been widely reported in Midcontinent strata during the Late Mississippian (Smith and Read, 2001), at the Mississippian Pennsylvanian boundary (Kvale and Barnhill, 1994), and into the Pennsylvanian (Aiken and Flint, 1994).

Geologic Controls on Deposition

A number of factors are used to explain the stratigraphy of the Carboniferous Midcontinent, including glacio-eustasy, climate-change, and tectonic uplift. Long-term climate fluctuation can be explained by continental drift and plate-tectonic reconstructions (e.g. Scotese and McKerrow, 1990; Frakes et al., 1992; Cecil, 1990). Cecil (1990) interprets Appalachian basin strata as being deposited under rapidly alternating climate conditions. During the Mississippian, Euramerica drifted northward from tropical climate belts during the Osagean (Tournasian), into arid latitudes during the Late Mississippian (Meramecian to Chesterian), and again into tropical latitudes during the Pennsylvanian (Cecil, 1990). Changes in insolation patterns associated with Milankovich-driven orbital cycles are widely interpreted from Carboniferous Midcontinent strata (e.g. Heckel, 1986; Smith and Read, 2001; Al-Tawil et al., 2003; Al-Tawil and Read, 2003). Previous investigations of the tectonic vs. eustatic/climate controls remain contradictory and constitute a subject of
vigorous debate (Ettensohn, 2004, 2008; Miall, 2010b; Miller and Eriksson, 1999; Beuthin and Blake, 2002).

Tectonics

Sloss (1963) attributed the presence of regionally correlative unconformity surfaces to periods of tectonic uplift. This observation is consistent with regional reactivation of basement faults within the Appalachian (Ettensohn, 1981) and Illinois basins (Mcbride and Nelson, 1988; Smith and Read, 2001), and possibly indicates the inception of the Alleghenian and Ouachita orogenies (Ettensohn, 2008). During the Middle Chesterian the Appalachian Foreland was filled by carbonates representing the Greenbriar, Newman, and Slade units (Ettensohn, 2004). The carbonates are overlain by a marginal marine clastic "wedge" composed of the Pennington and Mauch Chunk Formations (Ettensohn, 2004, 2009).

Locally, the Howell Anticline and the Lucas Monroe Fault are believed to have exhibited significant displacements during the Late-Mississippian (Fisher et al., 1988). The Lucas Fault is a significant geologic feature in the southern Michigan basin. Figure 51 displays a significant thinning of the heterolithic Michigan Formation in a west to east gamma-ray cross-section across Eaton and Ingham counties, supporting the presence of a paleobathymetric feature overlying the Lucas fault. This feature was reported by Ells (1979) and Fisher et al. (1988). There may be two explanations for the abrupt truncation of the Michigan Formation: (1) The area overlying the Lucas Fault represents a depositional high and the Michigan Formation was not deposited, or, (2) an episode of syndepositional tectonic reactivation. The latter may explain the preservation of the SB18 paleosol, which is located directly on
the mapped Lucas structure. The nearby Howell Anticline was interpreted by Wanless and Shideler (1975) as a monadock representing a depositional high. The analysis of core and wireline logs supports syndepositional tectonics during the Chesterian stage (likely Early to Middle) in the Michigan basin.

Ettensohn (2004) divides the tectonic evolution of the Appalachian region into distinct tectophases which exerted substantial control on the type and style of basin fill. The Acadian Orogeny resulted from the subduction of microcontinents along the active eastern and southern margins of the Euramerican paleocontinent and culminated with the Alleghenian orogeny and continent-continent transpression (Ettensohn, 2004). During the Late Devonian through Mississippian the Acadian Orogeny resulted in the deposition of fine-grained clastic dominated strata throughout the eastern midcontinent (Ettensohn, 2004). A period of tectonic quiescence during the Middle Mississippian resulted in the accumulation of carbonate strata (Greenbriar, Newman, and Slade units) in the Appalachian basin (Ettensohn, 2009). Tectonic activity resumed during the Late Mississippian concomitant with the Ouachita orogeny along the southern margin of Euramerica (Ettensohn, 2008). In the Appalachian and Illinois basins this tectonic activity is thought to correspond with the reactivation of basement faults associated with the onset of the Alleghenian orogeny (Ettensohn and Peppers, 1979; Ettensohn, 1981). Ettensohn (1981) reports that the reactivation of faults within the Kentucky River Fault zone resulted in the progradation of the Pennington-MauchChunk clastic wedge westward into the Appalachian Basin (Ettensohn, 2004). Regional tectonism continued into the Pennsylvanian which resulted in the deposition of large amounts of “deltaic” terrigenous clastic material across the eastern Midcontinent (Ettensohn, 2004, 2008).

The continent-continent collision between Euramerica and Gondwana
(Alleghenian Orogeny) resulted in the eventual amalgamation of the supercontinent of Pangaea (Ettensohn, 2004, 2008). The movement of the tectonic epicenters associated with a transpressional regime (i.e. Euramerica and Gondwana) heavily influenced the distribution and timing of deposition along the axis of the Appalachian basin (Ettensohn, 2004). Wanless (1962) notes that eustasy, not tectonics may be the primary control on accommodation space in the tectonically quiescent continental interior resulting in rhythmic deposition associated with cyclothems. The Illinois basin is considered to be intermediate between these two extremes and was influenced by both rapid changes in eustatic sea-level association with Gondwanan glaciations as well as tectonics and differential subsidence patterns depending on the proximity to active fault structures undergoing uplift (Smith and Read, 2001, 2000).

Siliciclastic rocks are derived from pre-existing source terrains and strongly reflect the local and/or regional tectonic setting (Miall, 1992, 2010a; Dalrymple, 2010b). During the Paleozoic, orogenic activity resulted in the progradation of a number of clastic "wedges" into the Appalachian foreland basin from uplifted source terrains located along the continental margin through a network of fluvial "deltaic systems" and paralic environments (Ettensohn, 2004). During the Late Mississippian the mixed carbonate-clastic Pennington wedge prograded into the carbonate dominated Appalachian basin (Ettensohn, 2004, 2009). The mixed carbonate/siliciclastic Bayport interval may be an extension of this regional influx of siliciclastic material into the Midcontinent region.
Figure 51. Top left box shows the location of observed core material while the top right box shows the location of borehole penetrations located in the black box. The underlying east-west structural cross-section indicates a subtle thickening of the heterolithic Michigan formation (mixed shale, sandstone, carbonate, and evaporite). Truncation of these units is evident over the Lucas Fault structure (east-most well displayed in the cross-section).

**Eustasy**

The Late Mississippian is considered to be a time of glacio-eustasy. Smith and Read (1999, 2000, and 2001) note that the Late Mississippian Illinois basin was controlled by glacio-eustasy, and shares similar attributes with the midcontinent Pennsylvanian cyclothem (sensu Heckel, 1984). Al Tawil et al. (2003) and Al Tawil and Read (2003) discuss the effects of eustasy on the tectonically active Late Mississippian Appalachian basin and conclude that glacio-eustasy overprints the tectonic signal. Late Mississippian strata deposited in these basins are a mixture of shallow marine carbonates, shales, siliciclastics, and terrestrially derived paleosols and rare coals (Smith and Read, 2001; Al-Tawil and Read, 2003).

Pronounced fluctuations in local sea-level across the midcontinent have been used to suggest an allogetic (global) control on deposition during the Pennsylvanian. Glacio-eustasy is widely regarded as having exerted a large control on base-level, which impacted low-latitude sedimentary basins (Miall, 2010b). Midcontinent cyclothsems are explained by changes in glacio-eustasy (Wanless and Shepard, 1936; Heckel, 1986). Fluctuations in global sea-level are considered responsible for the rapid changes in depositional facies, alternating between marine and non-marine
depositional environments, which, in some cases, can be correlated for hundreds of miles (Heckel, 1986; Peppers, 1996). Small scale sea-level fluctuations on the order of ~400 Ky have likewise been interpreted from Late Mississippian strata (Smith and Read, 2001; Al-Tawil et al., 2003). The construction of global onlap and sea-level curves for the Carboniferous was attempted by Ross and Ross (1988), Smith and Read (2001), and Haq and Schutter (2008) indicate a pronounced increase in the frequency and intensity of sea-level change during the Chesterian Stage. Late Mississippian Midcontinent deposits are considered to be primarily regressive in nature reflecting a mixture of open marine carbonates and terrestrially derived siliciclastic strata (Sloss, 1963; Ross and Ross, 1988). The Ross and Ross (1988) and Haq and Schutter (2008) global onlap and sea-level curves (Fig. 52) suggests that a pronounced sea-level drawdown occurred at the Mississippian-Pennsylvanian boundary.

This study supports regional similarities between the Illinois and Appalachian basins. The vertical stacking pattern of facies within the Bayport support rapid changes in base-level. Smith and Read (2001) identify a number of sequences bounded by incised valleys and their correlative paleosol deposits within the Illinois basin. A similar relationship is reported to exist for the Bayport interval. The presence of discrete alternating packages of carbonate and siliciclastic strata is interpreted to result from rising and falling eustatic sea-level. The global sea-level curve constructed for the Carboniferous system by Ross and Ross (1988). Smith and Read (2000) suggest the onset of high-frequency sea-level fluctuations corresponds to the onset of Gondwanan glaciations. A global sea-level curve constructed by Haq and Schutter (2008) (Fig. 52) shows repetitive fluctuations in "global" sea-level.
Figure 52. Global onlap and deviations in global sea-level. A sharp increase in the frequency of deposits occurs approximately at the Meramecian-Chesterian boundary. Carboniferous glaciations occurred periodically during the lower-Mississippian before expanding over the continent of Gondwana during the Late-Mississippian (Chesterian, Serpukhovian) and Pennsylvanian periods and are thought to be responsible for cyclic (cyclothemic) deposition across the Midcontinent. Modified from Haq and Schutter, 2008.

Climate Change

The Carboniferous was a period of transition from a greenhouse to icehouse global climate (i.e. Fig. 7) (Frakes et al., 1992; Smith and Read, 2000). The
differences in deposition between carbonate and siliciclastic strata have been widely acknowledged (e.g. Dalrymple, 2010b; Cecil, 1990) and the analysis of climate sensitive sediments within the Carboniferous Midcontinent supports fluctuating, wet and dry climate observed in Late Mississippian and Pennsylvanian strata throughout the Midcontinent (Cecil, 1990; Rankey, 1997; Miller and Eriksson, 1999; Beuthin and Blake, 2002; Kahmann and Driese, 2008). In addition to tectonics, climate is also acknowledged as a driving mechanism for the liberation and delivery of clastic material into a basin (e.g. Cecil, 1990). According to Cecil (1990), the highest amounts of siliciclastic input throughout the Carboniferous Midcontinent is associated with alternating periods of wet and dry conditions resulting in the highest weathering potential of a source-rock. Biochemical rocks such as carbonates and evaporites tend to accumulate under persistently arid conditions while coal/histosols form under persistently humid conditions (Cecil, 1990; Rankey, 1997). The mapping and correlation of the global latitudinal distribution of climate-sensitive strata through time was undertaken by Frakes et al. (1992) who uses these features as a proxy for global climate. Cecil (1990) contends that climate change is a function of long-term (1-100 Ma), intermediate (100-400 Ky), and short-term (10 Ky to instantaneous) periodicities associated with long-term tectonism and relatively short-term climate oscillations attributed to Milankovich orbital cycles. Milankovich cyclicity is used to explain Pleistocene climate change and is supported by a number of independent proxies (Hays et al., 1976; Frakes et al., 1992). The Pleistocene example of Milankovich cyclicity was used by Goldhammer et al. (1987, 1990) to model the deposition of the Triassic Italian Alps. This rationale has been observed and applied to Carboniferous cyclothemic depositional successions throughout the midcontinent (e.g. Heckel, 1986; Cecil, 1990; Rankey, 1997). However, considerable debate exists
regarding the relationship between climate and eustatic sea-level influenced the midcontinent cyclothem (e.g. Rankey, 1997; Miller and Eriksson, 1999; Beuthin and Blake, 2002). The widespread occurrence of evaporite minerals within the Michigan Formation, deposited during the Chesterian Stage, indicates semi-arid to arid climate (Cohee, 1979) in the Michigan basin at this time. The overlying Bayport Limestone likewise contains features that may indicate a semi-arid/arid climate (Ciner, 1988) including carbonate and evaporitic lithologies.

The deposition of significant amounts of sandstone within the Bayport may be a response to alternating wet/dry climate during the Late Mississippian (Cecil, 1990; Kahmann and Driese, 2008). In the cores observed siliciclastics dominated intervals occur intermittently throughout and commonly overlain by open marine limestone units. These sandstones are interpreted as tide-dominated deposits based on sedimentary structures (e.g. mud-drapes, thick-thin laminations, reactivation surfaces, and interpreted tidal rhythmites) observed within the sandstone packages (sensu Dalrymple, 2010c). The presence of heterolithic, mixed sand and silt, in a generally fining upward succession further supports a tide dominated depositional setting. These sandstone features incised valley deposits due to the generally erosive nature of the contacts with underlying carbonate package.

Figure 53 is a paleoclimate curve for the Carboniferous Appalachian basin. Cecil (1990) and indicates the onset of high frequency climate change during the Late Mississippian (Chesterian Stage) based on the observation of climate sensitive sediments (i.e. carbonates, evaporites, and coal). Similar analysis of Late-Mississippian Appalachian basin strata by Miller and Eriksson (1999) supports Cecil's (1990) climate model.
Kahmann and Driese (2008) undertook a textural and geochemical analysis of paleosols within the Pennington Formation in an effort to calculate the Mean Annual Precipitation (MAP) values. These workers support an alternating wet and dry climate during the Late Chesterian associated with monsoonal conditions and high-frequency climate fluctuations in association with the onset of Gondwanan glaciations. The climate cyclicity resulted from expansion and contraction of the
Paleo-Intertropical Convergence Zone (ITCZ).

The presence of paleosol deposits in the SB18 core offers the best evidence for past climate conditions within the Mississippian-Pennsylvanian systemic boundary of the Michigan basin. The marine limestone bed underlying the paleosol succession displays characteristic glaebules/caliche structures, interpreted to have formed due to the shrinking and swelling of the carbonate host material associated with vadose conditions (e.g. Estaban and Klappa, 1983). The presence of a vertic oxisol (i.e. red bed) at the top of the subaerially altered limestone further indicates an alternating wet and dry climate (Retallack, 1997). An increase in preserved carbonaceous deposits was observed upsection (e.g. coal and carbonaceous shale) indicating a reducing, more humid terrestrially influenced environment of deposition (Cecil, 1990; Rankey, 1997). Cecil (1990) favors Milankovich driven climate cycles. Miller and Eriksson (1999) suggest changes in global sea-level may represent a local climate forcing. However, Beuthin and Blake (2002) propose that interpreted climate change from the sedimentary record could be a product of the depositional environment and localized conditions such as water ponding. The coupling influence of eustasy on climate or vice-versa remains a point of debate (e.g. Rankey, 1997; Miller and Eriksson, 1999; Beuthin and Blake, 2002; Kahmann and Driese, 2008) and a developing avenue of research.

During the Late Mississippian the climate of the Appalachian region was transitional between semi-arid to seasonally wet and dry climate as indicated by the deposition of carbonates, evaporites, paleosols, and the periodic influx of siliciclastic sediments (Cecil, 1990). In contrast, the Pennsylvanian system is predominantly composed of terrestrially derived siliciclastic and carbonaceous shale/coal deposits interpreted to be the product of a tropical rainy to wet-dry climate (Cecil, 1990). The
transition from semiarid/arid to more humid conditions occurred approximately at the Mississippian-Pennsylvanian boundary (Cecil et al., 1985; Cecil, 1990; Witke, 1990) suggesting regional climate change (Donaldson, 1985; Cecil et al. 1985; Cecil, 1990). Many researchers have sought to explain the relationship between sea-level fluctuations associated with repetitive cyclothem deposition during the Late-Mississippian and Pennsylvanian midcontinent strata (e.g. Wanless and Shepard, 1936; Cecil et al., 1985; Cecil, 1990; Rankey, 1997; Miller and Eriksson, 1999; Beuthin and Blake, 2002).

In the Michigan basin, an arid climate during the Mississippian is supported by economically significant evaporite deposits (gypsum and anhydrite) of the Michigan Formation. Evaporitic strata are reported to overly Bayport strata in the central basin (Tyler, 1980; Vugrinovich, 1984). Furthermore, previous interpretations of the “algal” and desiccation structures reported within the Bayport Limestone further indicate an arid environment of deposition (Cohee, 1979; Bacon, 1971; Ciner, 1988). In contrast, a preliminary review of the literature regarding Pennsylvanian deposits within the Michigan basin suggest a humid marginal-marine to terrestrially dominated strata based on the occurrence plant traces (Winchell, 1861), brackish water fauna (Lingula sp.; Lane, 1902), descriptions of paleosol and terrestrial fluvial deposits (Kelly, 1936; Venable, 2006), terrestrial plant derived pollen and spores (Venable, 2006), and the notable lack of carbonate deposits observed in this study, though a thin bed of limestone referred as the Verne Member was reported in previous studies of the Saginaw Formation (e.g. Lane, 1902; Kelly, 1936; Wanless and Shideler, 1975; Vugrinovich, 1984).
Michigan Basin Stratigraphy

The documentation of key pollen and spore types indicate the Michigan Formation is Chesterian in age, contrasting with previous interpretations suggesting a Meramecian age (Catacosinos et al., 2001; Harrell et al., 1991). The late Chesterian age of the Bayport interval is further supported by deposition of carbonates, tidally influenced siliciclastics, evaporites, and paleosols which are consistent with coeval deposits reported in the Illinois basin (i.e. Smith and Read, 2001; Nelson et al., 2002) and the Appalachian basin (i.e. Al-Tawil et al., 2003; Al-Tawil and Read, 2003). Changes in glacio-eustasy are the common explanation for rapid lithologic changes present in Carboniferous aged Midcontinent strata (e.g. Heckel, 1986; Smith and Read, 2001). However, regional climate change has also been documented in Carboniferous strata (Cecil, 1990; Rankey, 1997; Miller and Eriksson, 1999; Kahmann and Driese, 2008) and the relationship between eustasy and climate is not well understood (Rankey, 1997).

Historically, the Mississippian-Pennsylvanian boundary in the Michigan basin was placed at the contact between the Bayport Limestone and the "Parma" Sandstone (Kelly, 1936; Newcombe, 1933; Wanless and Shideler, 1975). Vugrinovich (1984) suggests that the "Parma" unit may straddle the Mississippian and Pennsylvanian periods in the central Michigan basin, based on cuttings, wireline log data and comparison to Illinois basin strata. However, the validity of using the name "Parma" is questionable due to the lack of a well-defined type-section. It is proposed that the name "Parma" should, at most, be relegated to member status within the Bayport Formation (new stratigraphic term).
The Bayport Formation is a series of interbedded sandstone, carbonate, and paleosols stratigraphically positioned between the Michigan Formation and the Saginaw Formation. These discrete packages typically overlie cross-bedded quartz sandstone, which fines upward before passing into the deeper water carbonate facies. The carbonate units rarely exceed 5 ft. in thickness in southern Michigan and are severely overprinted by diagenetic alteration associated with sub-aerial exposure including paleosols and karst-dissolution structures. Smith and Read (2000, 2001) describe the Late Mississippian as a regionally significant time-marker due to the onset of rapidly fluctuating global base-level conditions, particularly within the Illinois and Appalachian basins. Similar relationships appear to be present within the Bayport Formation. The paleogeographic position, which is up regional dip from other cratonic interior basins, including the Illinois and Western Interior basins, and inferred diminished rate of subsidence of the Michigan basin compared to the Illinois, Appalachian, and other basins of the Midcontinent, may indicate a significant amount of lost time within the Bayport.

This study documents the existence of Chesterian strata in the Michigan basin previously considered to be either eroded or not deposited. Tyler (1980) and Vugrinovich (1984) describe the occurrence of a Limestone bed overlying the "Parma Sandstone," referred to as the Six-Lakes Limestone. This bed was encountered in the analysis of core material and is interpreted to be Chesterian in age based on the prevalence of marine biota and biostratigraphic dating of the underlying shale bed. In the SB18 a white-buff limestone overlain by a series of paleosols is interpreted as the Mississippian-Pennsylvanian systemic boundary within the Michigan basin. This
Limestone is best preserved in proximity to the Lucas Fault, suggesting, fault-related control on the preservation of the Bayport unit. In core material adjacent to the Lucas structure fluvial erosion has eroded the limestone unit (e.g. SB13, SB15, and SB16). The preservation of the Bayport, which was noted by Newcombe (1933) as discontinuous in nature and may be governed by tectonic factors (syndepositional faulting), which may have controlled fluvial erosion associated with the Mississippian-Pennsylvanian systemic boundary in the southern Michigan basin.

The Parma Sandstone is the source of much debate in Michigan basin geological literature. Since it was first described by Winchell (1861), the Parma is noted for its discontinuous nature. Later scientists place the Parma at the base of the Pennsylvanian system (Kelly, 1936; Newcombe, 1933; Wanless and Shideler, 1975). Vugrinovich (1984) describes the Parma Sandstone as overlying the carbonate facies of the Bayport Limestone. However, this study disputes a significant depositional break or unconformity between the two units and suggests that limestone, dolomite, shale, and quartz sandstones are in facies relationship with one another. The sandstones within the Bayport are interpreted as tidal channel/shoreline deposits and are observed to be intercalated with limestone units similar to what was reported by Vugrinovich (1984) and Westjohn and Weaver (1996, 1998), from the study of geophysical logs in the central Michigan basin. Lithologically, the Parma may deserve member status, though, the sandstone units may be expected to be diachronous based on their depositional nature (i.e. channels). The discontinuous nature of the Bayport may also be related to complex facies relationships, the distribution of which may have been governed by glacio-eustasy and climate, similar
to what is interpreted from the Late Mississippian Illinois basin (Smith and Read, 2000; Cecil, 1990), as well as autocyclic processes inherent to the depositional system such as shoaling and the lateral migration of channel features. Based on the above discussion it is recommended the Mississippian-aged mixed carbonate-siliciclastic interval of the Bayport Limestone and, informally, the Parma Sandstone, be elevated to *Bayport Formation* in the Michigan basin stratigraphic nomenclature.

**Conclusions**

• The Chesterian aged Bayport Formation (proposed stratigraphic term) comprises a mixed carbonate/siliciclastic succession deposited in an arid, paralic shore-line environment influenced by high frequency, glacial interglacial cyclicity.

• Significant facies boundaries within the Bayport formation are marked by carbonate exposure horizons, paleosols, and sharp erosional surfaces overlain by tidally bedded quartz sandstone. Such discontinuities have been observed throughout the Bayport.

• The interpretation of climate sensitive strata (evaporites) indicate a predominantly arid climate during deposition of the Michigan Formation, an increasingly humid environment through the mixed carbonate/siliciclastic Bayport Formation, and a predominantly humid climate superjacent strata of the Saginaw Formation, based on the presence of coal and abundant carbonaceous organic material and the absence of evaporites and carbonates.

• Biostratigraphic analysis indicates the Michigan and Bayport Formations
were deposited during the Chesterian North American Regional Stage, while carbonaceous-rich Saginaw Formation lithofacies are Morrowan and Atokan in age.

- The Mississippian-Pennsylvanian systemic boundary is marked by either a paleosol deposit grading from textures indicating an arid environment into a carbonaceous deposits during the Pennsylvanian, or an incised valley fill erosionally truncating the Bayport.

- The Lucas Fault structure was likely active during the Late Mississippian (Chesterian) based on the analysis of core material and well logs. Paleosols appear to be preferentially preserved along the up-thrown portions of the structure. Syndepositional reactivation of the fault is indicated by a thickening of the Michigan Formation observed using wireline logs from the area.
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APPENDIX A

CORE DESCRIPTIONS
### LEGEND

<table>
<thead>
<tr>
<th>Bedding/Erosional Features</th>
<th>Biota/Biogenic</th>
<th>Diagenetic</th>
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<tr>
<td>Planar Laminated</td>
<td>Undifferentiated Fragments</td>
<td>Calcite Cement</td>
</tr>
<tr>
<td>Cross-bedded</td>
<td>Echinoderm</td>
<td>Moldic Vug(s)</td>
</tr>
<tr>
<td>Wispy</td>
<td>Brachiopod</td>
<td>Brecciated Zone</td>
</tr>
<tr>
<td>Horizontal</td>
<td>Bryozoa</td>
<td>Fractured</td>
</tr>
<tr>
<td>Undulatory/wavy</td>
<td>Microbially Laminated</td>
<td>Chert Nodule</td>
</tr>
<tr>
<td>Lenticular Bedding</td>
<td>Burrowed</td>
<td>Evaporite Nodules</td>
</tr>
<tr>
<td>Contorted Bedded</td>
<td>Solitary Horn Coral</td>
<td>Styloitic</td>
</tr>
<tr>
<td>Lag Deposit</td>
<td>Ostracod</td>
<td>Pyrite Concretion</td>
</tr>
<tr>
<td>Carbonate Clasts</td>
<td>Root Traces</td>
<td>Siderite Nodule</td>
</tr>
<tr>
<td>Organic/mud Clasts</td>
<td>Plant remains</td>
<td>Horizonation</td>
</tr>
<tr>
<td>Fenestrae</td>
<td></td>
<td>Vertic Fractures</td>
</tr>
<tr>
<td>Mud-crack</td>
<td></td>
<td>Slickensided</td>
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</tbody>
</table>

#### Lithology

- Marine Limestone
- Carbonaceous Shale
- Dolomite
- Evaporite
- Sandstone
- Coal
- Mudrock

#### Data Point/Interpretation

- P: Recovered Palynological sample
- X: No Palynomorph recovery
- Sequence Boundary
- MegaSequence Boundary
APPENDIX B
PALYNOMORF INVENTORY

<table>
<thead>
<tr>
<th>MISSISSIPPIAN</th>
<th>PENNSYLVANIAN</th>
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<tr>
<td>Chesterian</td>
<td>microfossils</td>
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<tr>
<td></td>
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<tr>
<td></td>
<td>barren of</td>
</tr>
<tr>
<td>Morrowan</td>
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</tr>
<tr>
<td>Atokan</td>
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### SB16

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<th>598.5 ft.</th>
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<th>408.9 ft.</th>
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| PENNSYLVANIAN | MISSISSIPPIAN |
| Atokan | Chesterian |
| 618 ft. | 128 ft. |

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