64 ANNUAL

TRI-STATE GEOLOGICAL FIELD CONFERENCE

GUIDEBOOK

Pleistocene, Mississippian, & Devonian Stratigraphy of the Burlington, Iowa, Area

October 12-13, 2002

Iowa Geological Survey **Guidebook Series 23**

Cover photograph: Exposures of Pleistocene Peoria Loess and Illinoian Till overlie Mississippian Keokuk Fm limestones at the Cessford Construction Co. Nelson Quarry; Field Trip Stop 4.

64th Annual Tri-State Geological Field Conference

Pleistocene, Mississippian, & Devonian Stratigraphy of the Burlington, Iowa, Area

Hosted by the Iowa Geological Survey

prepared and led by

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INTRODUCTION TO THE FIELD TRIP

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Field Trip leader Brian Witzke contemplates an exposure of Warsaw Fm at the Orba & Johnson barge terminal road, north of Keokuk, a trip stop on the Sunday Morning Field Trip

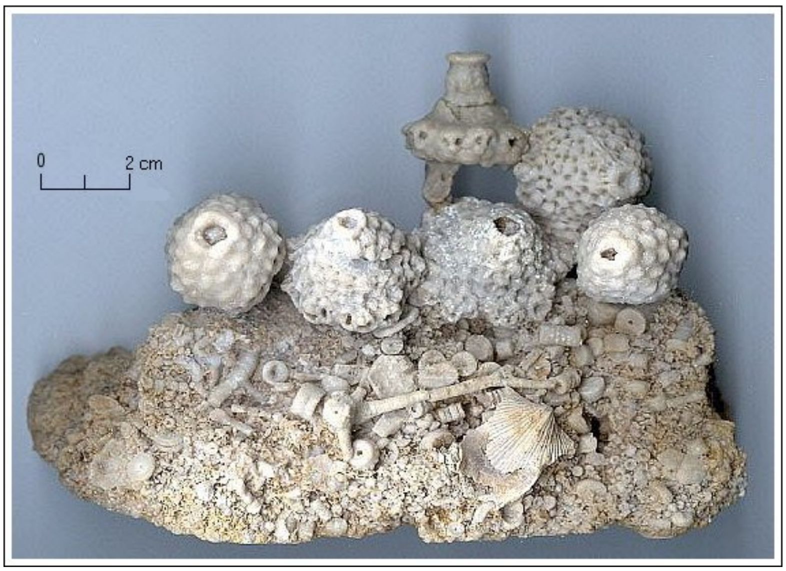
The Mississippian System was historically proposed for the succession of strata exposed in the Mississippi River Valley between Burlington, Iowa, and St. Louis, Missouri. The bedrock exposures in and around the City of Burlington comprise part of the historic "body statotype" on which the concept of the Mississippian System was defined and based. Until recently, the Mississippian was primarily a North American chronostratigraphic label, roughly synonymous with the Lower Carboniferous of the Old World. Recently the International Union of Geological Sciences and the International Commission on Stratigraphy officially subdivided the Carboniferous System into upper (Pennsylvanian) and lower (Mississippian) subsystems. As such, the Mississippian now has global meaning and application as a major chronostratigraphic subdivision of geologic time, and the strata exposed in the Burlington area constitutes a significant historic reference. The 64th Tri-State field trip can be viewed as an international pilgrimage to the classic Mississippian succession in this type area. We will see the entire Mississippian section of the region, including the McCraney, Prospect Hill, Wassonville, Burlington, Keokuk, Warsaw, Sorora, and St. Louis formations, as well as underlying Upper Devonian English River Siltstone and Maple Mill Shale, and an overlying Pennsylvanian sandstone channel fill and Pleistocene strata.

The Pleistocene Illinoian ice advanced south across most of northern Illinois and westward into the southeastern-most portion of Iowa. The Illinoian ice overrode Paleozoic bedrock and earlier pre-Illinoian glacial deposits that were emplaced by a series of ice sheets that advanced into Iowa from the north and northwest between about 2,500,000 and 500,000 years ago. This field trip will visit exposures of Illinoian glacial till and a possible pre-Illinoian till. Additionally, we will see a lacustrine deposit that fills a bedrock channel (possibly an Illinoian ice-marginal lake), a well-developed Sangamon Geosol, and a thick, section-capping loess sequence.

The trip will also see examples of karst development in Mississippian carbonate rocks, the many faces of the Mississippi River floodplain, and geode development in the Warsaw Formation.

After the Tri-State banquet we will hear Forest Gahn (University of Michigan Museum of Paleontology and Burlington native) discuss the spectacular Burlington crinoids. These crinoids were popular even before the geology of the region was understood. In his 1895 publication on the Geology of Des Moines County, Charles Rollin Keyes commented that *"Burlington crinoids are known throughout the world as objects of surprising beauty. They are sought for and highly prized everywhere. In consequence, a great deal of attention has been directed to the consideration of the fossils in the rocks, rather than the rocks themselves."*

The field trip organizers and other staff of the Iowa Geological Survey welcome all trip participants from Wisconsin, Illinois, Missouri, and elsewhere to Iowa. We hope that you and all of the Iowa participants will enjoy an informative, enjoyable, and safe 64th Annual Tri-State Geological Field Conference.



A cluster of Burlington Formation Crinoids found and prepared by Carl Cook. This picture appeared on the cover of the October 2000 MAPS publication and displays three *Uperocrinus pyriformis* (at each end and at top) and three *Physetocrinus ventricosus* (identifications by Rick Poropat).

Photo and information from: http://www.lakeneosho.org/

DEVELOPMENT OF THE MISSISSIPPI RIVER IN SOUTHEAST IOWA

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INTRODUCTION

The configuration of tributaries and the position of the Mississippi River have changed numerous times throughout the Quaternary. Certain aspects of the history of the Mississippi River are poorly understood, but many studies have speculated on the Mississippi River positions and its response to the glacial advances of the Quaternary.

PRE-ILLINOIAN

Prior to the Illinoian ice advance, the Mississippi River may have been diverted several times by Pre-Illinoian events, but this history is difficult to sort out. It is known that the Mississippi River did not follow its present course south of Clinton, Iowa, prior to the Illinoian (Udden, 1899; Trowbridge, 1959; Anderson, 1968). Instead, it turned to the southeast, following the Meredosia Channel between the modern Mississippi and Rock River valleys in Illinois (see Figure 1). It then flowed into the Princeton Bedrock Channel and joined with the present Illinois Valley in the vicinity of Hennepin, Illinois. At that time, no valley existed between Rock Island and Muscatine, and the present course of the Mississippi River below Muscatine was occupied by the ancestral Cedar, Iowa, Skunk and Des Moines valleys (Udden, 1899; Horberg, 1950; Hansen, 1973).

Multiple Pre-Wisconsin drainage lines exist in southeast Iowa as evidenced by deep valleys cut into the bedrock surface. The most prominent of these is the Cleona Channel, which trends southwesterly across the western part of Muscatine County and joins the Udden Channel in Louisa County (Hansen, 1972). Both of these valleys contain Pre-Illinoian tills and multiple buried valley fills (Bettis, 1994; Bettis and Autin, 1997). The Udden Channel was buried by the Illinoian glacial advance. The Cleona Channel was ice-marginal during the Illinoian and carried the diverted Mississippi River flow. Neither channel is occupied by a large stream today.

ILLINOIAN

During the advance of the Illinoian age Lake Michigan Lobe, ice blocked the Princeton Channel and a large lake, Glacial Lake Moline, formed upstream in what is now the Green River Lowland of Rock Island County, Illinois (Anderson, 1968). The Mississippi River was diverted from its interglacial course in the Princeton Bedrock Valley of Illinois to the Cleona Channel between Dixon and Wilton in eastern Iowa (Anderson, 1968; Bettis and Glenister, 1987; Bettis and Autin, 1997). The Illinoian glacier advanced from the northeast out of the Lake Michigan Basin and moved across Illinois into Iowa. During the early Illinoian, ice covered what is currently the Mississippi channel from Clinton to Fort Madison, Iowa. The westernmost margin of the advance reached a width of approximately 4 miles (6.5 km) from the Mississippi River near Muscatine and up to 20 miles (32 km) from the river just north of Burlington (Kay and Graham, 1943). At the maximum extent of the Illinoian glaciation, ice blocked the lower reaches of the Iowa and Cedar valleys and also the Wapsipinicon River Valley farther to the north.

Diversion channels developed between the Maquoketa and Wapsipinicon valleys (Goose Lake Channel) and the Wapsipinicon and Cedar valleys (Cleona Channel) (Anderson, 1968).

Several outlets for Lake Moline developed as the Illinoian ice continued to advance, but were subsequently buried (Anderson, 1968). Portions of Edwards Valley, Copperas Creek, and Andalusia Gorge of the present Mississippi Valley provided outlets for the ice-dammed lake. Nott (1981) and Hallberg et al. (1980) have also shown that portions of Crooked Creek, the Skunk River, Cedar Creek, and Sugar Creek (collectively known as the Leverett Channel of Schoewe, 1921) carried meltwater to the south from the Illinoian front.

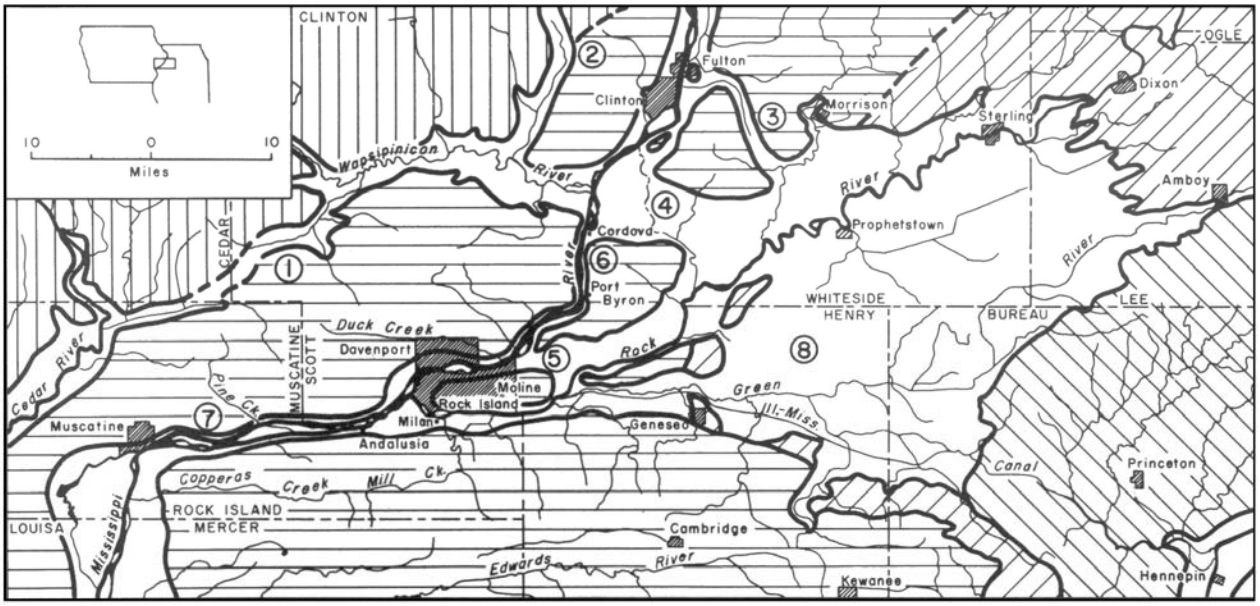

Figure 1. Glacial geology and drainage features of the Rock Island area. Vertical ruling: Pre-Illinoian Drift with thick cover of loess; horizontal ruling: Illinoian drift with thick cover of loess; NE-SW ruling; Shelbyville drift; NW-SE ruling; Bloomington and younger drift. Lowlands and channels: 1=Cleona Channel; 2=Goose Lake Channel; 3=Cattail Channel; 4=Meredosia Channel; 5-Pleasant Valley; 6-Port Byron Gorge; 7-Andalusia Gorge; 8-Green River Lowland. Modified from Anderson (1968), Fig. 1.

SANGAMON

At the end of the Illinoian glaciation, the Mississippi River returned to its course through the Princeton Bedrock Channel and down the present Illinois River Valley (Anderson, 1968). Downcutting, headward extension, and drainage network development continued from the end of the Illinoian stage to approximately 55,000 years ago to form the modern configuration of the major tributary valleys in Iowa. The Sangamon Geosol also developed on the stable uplands and valley surfaces at this time.

LATE WISCONSINAN

The last major episode influencing the drainages in southeastern Iowa occurred approximately 21,000 years ago. Glacial ice of the Lake Michigan Lobe again blocked the Ancient Mississippi in the Princeton Channel in Illinois and formed Glacial Lake Milan in the Green River Lowland of northeastern Illinois (Schaffer, 1954; Anderson, 1968). Lake Milan eventually deepened to the elevation of a low divide and drained through the Andalusia Gorge located between Rock Island and Muscatine. The Mississippi River changed its course to the present route through the Andalusia Gorge and south through the Ancestral

Iowa/Cedar Valley to St. Louis, where it joins the pre-Wisconsinan course of the Mississippi Valley (Hobbs, 1990; Bettis, 1997).

Between 21,000 to 11,000 B.P., the Mississippi Valley was aggraded by outwash from glaciers in the northern part of the Mississippi Basin and the Lake Michigan Basin. During this time, outwash accumulated in the Green River Lowland and extended into the Andalusia Gorge (the modern Mississippi River valley), which was carrying meltwater and lake discharge. The Andalusia was cut down to about its present level at this time and linked the Ancient Mississippi above Rock Island with the Ancient Iowa valley below Muscatine (Anderson, 1968). Following the retreat of the ice and the cessation of meltwater and outwash discharge into the Green River Lowland, the Mississippi River continued to flow through the Andalusia Gorge and the modern valley configuration was established.

In the area between Rock Island and Burlington the aggradation was interrupted by down-cutting events around 13,000 and 10,500 B.P. These down-cutting events resulted in the development of two late glacial terraces: the Savanna Terrace (an older, higher level) and the Kingston Terrace (younger and lower level). These terraces have previously been referred to as the Mankato and Late Mankato terraces (Trowbridge, 1954; Edmund and Anderson, 1968).

The Savanna Terrace remnants are recognized along the length of the Mississippi Valley from the St. Croix River to the Meramac River south of St. Louis (Hobbs et al., 1990). Radiocarbon dates of terrace sediments indicate that the Savanna Terrace was accumulating 17,000 years ago (Bettis and Hallberg, 1985). Stratigraphic relationships with other radiocarbon dated alluvial deposits indicate that the Savanna Terrace deposition ended between 12,000 and 11,500 years ago.

Floods occurred when meltwater from glaciers in Minnesota and Wisconsin was discharged into the Mississippi Valley. Floods originating from the Lake Superior Basin carried distinctive reddish brown clay derived from the Upper Keweenawan Fon du Lac Formation, whereas floodwaters from other sources did not. Most of the fill comprising the Savanna Terrace consists of gray and grayish brown silt, loam and sand that is a combination of non-Superior-source flood sediment and locally-derived alluvium (Bettis, 1997).

With the development of the Savanna Terrace, the Mississippi River had downcut to within a few meters of its present floodplain level by 11,000 years ago. Outwash then accumulated, and the Mississippi floodplain aggraded until about 10,500 B.P. when glacial meltwater was no longer directly discharged into the Upper Mississippi Valley. Around 10,500 B.P. another down-cutting episode isolated the latest glacial floodplain and formed the Kingston Terrace. The discharge of large volumes of largely sediment-free meltwater as a result of the opening of the southern outlet of Glacial Lake Agassiz may have been the cause. Glacial Lake Agassiz was located in the Red River Lowland in west-central Minnesota and the eastern Dakotas. This glacial lake drained into the Upper Mississippi via River Warren in the Minnesota Valley between about 11,000 and 9,500 years ago (Matsch, 1983) and marked the final episode of glacier-related discharge into the valley.

HOLOCENE

Following the last glacial down-cutting event, the Mississippi River underwent a change in channel pattern from a braided pattern to an island braided pattern, which has continued through the Holocene. The Mississippi floodplain level has been at nearly the same elevation since about 11,000 years ago, but several shifts in channel position have occurred. During the early Holocene (before 6,000 B.P.) the main channel of the Mississippi was just east of the bluff line between Burlington and the village of Kingston, Iowa. Alluvial fans began to build out across the early Holocene floodplain along the bluff Line approximately 9,500 years ago. The Mississippi channel belt had shifted eastward to a position three to four miles east of the western bluff line in the vicinity of Kingston by 6,000 B.P. Alluvial fans continued to prograde, burying portions of the early Holocene channel belt and remained active through the middle Holocene. These alluvial fans stabilized during the late Holocene about 2,500 years ago.

LANDFORM SEDIMENT ASSEMBLAGES IN SOUTHEAST IOWA

The Late Wisconsinan and Holocene alluvial fills of the Upper Mississippi Valley are defined by landform sediment assemblages (LSA) of Bettis et al. (1996). Each LSA is a set of discontinuous geologic units. Figure 2 is a generalized map showing the LSAs of southeast Iowa. Brief unit descriptions for LSAs in southeast Iowa are listed below (summarized from Bettis et al., 1996):

Gilead Terrace (GILEAD)

The Gilead Terrace is the highest unit in the Mississippi Valley that is underlain by alluvium. Its distribution is not documented due to difficulties mapping loess-mantled surfaces in the Upper Mississippi Valley. The Gilead Terrace is mantled by thick Peoria Silt. Only remnants of this terrace protruding into the valley were mapped in the project area. The Gilead Terrace typically occurs as low relief areas elevated greater than 20 feet (6 m) above the Savanna Terrace.

Cuivre Terrace (CUIVRE)

The Cuivre Terrace is also mantled by Peoria Silt, but these loess deposits are thinner than those of the Gilead Terrace. The mantle of silt and sand may interfinger with Cuivre Terrace deposits and grade downward into alluvium ranging in age from approximately 25,000 to 17,000 BP. The alluvial deposits associated with the Cuivre Terrace accumulated before those associated with the Savanna Terrace. The Cuivre Terrace typically occupies relatively flat areas at elevations 15 to 20 feet (4.5 to 6 m) above the Savanna Terrace.

Savanna (SAVAN)

The Savanna Terrace represents the highest terrace remnants in the Mississippi Valley without loess cover. The Savanna Terrace formed from approximately 17,000 to 12,000 years ago (Bettis and Hallberg, 1985). Near the Iowa/Minnesota border, the Savanna Terrace is approximately 20m (66 ft) above the Mississippi River floodplain. This decreases to near 10m (33 ft) south of Quincy, Illinois. The gradient of the terrace in the main valley is the highest of any Wisconsinan or Holocene surface. Very little of the Savanna Terrace remains intact today, partly due to extensive modification by human activities (urban development, agricultural activity, quarrying, etc.).

Kingston (KINGS)

The Kingston Terrace consists of streamlined, sandy terrace remnants elevated 3-5m (10-16 ft) above the Mississippi floodplain. Terrace remnants are associated with a now-buried paleochannel system with channels several times broader than the historic Mississippi channel. The gradient is less than that of the Savanna Terrace, but greater than the Holocene floodplain gradient; therefore, the height of the Kingston Terrace above the Mississippi floodplain decreases down valley. Younger Mississippi River abandoned channel areas typically separate the Kingston Terrace remnants from the valley wall.

Kingston Terrace deposits are greater than 10m thick and consist of valleytrain outwash deposited between about 12,000 and 10,400 years ago. Silty, loamy, and Superior Basin-source reddish brown silty clay deposits are present in some of the overflow channels as well as in all the deeper Mississippi paleochannels associated with the Kingston. The reddish brown silty clay sediments accumulated during the last Superior Basin overflow events into the Upper Mississippi Valley between 9,800 and 9,500 B.P. The sandy fluvial deposits of the Kingston Terrace are mantled with thin (usually less than 2m/6.5ft thick), eolian sand comprising sand sheets and low dunes. The Kingston Terrace is underlain by trough cross-bedded and planar-bedded sand and pebbly sand. These deposits are usually less pebbly than deposits associated with the Savanna Terrace.

Fan/Colluvial Slope (FANCO)

The Fan/Colluvial Slope LSA includes alluvial fans and colluvial slopes along valley margins. These surfaces are above the Mississippi River floodplain and bury the Savanna and Kingston terraces. FANCO has interfingering relationships with EMHOL, YAZOO, and TRIFA LSAs. FANCO deposits are stratified silty, loamy, clayey, sandy, and pebbly sand alluvium derived from erosion in tributary valleys and from the valley wall. The thickness ranges from 3 to 15m (10 to 49 ft), and these deposits contain several upward-fining sequences with paleosols. FANCO deposits accumulated between 9,000 and 2,500 B.P.

Early to Middle Holocene Channel Belt (EMHOL)

The Early to Middle Holocene Channel Belt encompasses low-relief, slightly undulating, poorly drained, linear to broadly arcuate surfaces on the Mississippi floodplain, which mark the location of Mississippi paleochannel positions and associated islands during the Early and Middle Holocene. High stage Mississippi floodwaters overtop portions of this LSA where unrestricted by artificial levees. North of the Quad Cities most of this LSA is flooded by pools of the lock and dam system. EMHOL is located on both sides of the floodplain and is most extensive in the wide valley reaches south of the Quad Cities. EMHOL is inset below the Savanna and Kingston terraces, can either cut or be cut out by the Yazoo Meander Belt LSA, and is buried by the Mississippi Levee LSA. The Late Holocene Channel Belt and Tributary Fan LSAs cut out the Early to Middle Holocene Channel Belt LSA.

The Early Middle Holocene Channel Belt deposits consist of a variable thickness of loamy, silty clay loam and clay loam overbank alluvium grading downward to sandy loam, sand, and pebbly sand in-channel deposits. The fine-grained deposits mantle most of the LSA and range in thickness from about 1.5 meters (5 ft) on swells to over 6 meters (20 ft) in abandoned channel areas. The oldest portions of the LSA are underlain by Superior-source reddish brown silty clay slackwater sediments that were deposited between 9,600 and 9,200 B.P. Buried soils (formed during periods of low Mississippi River flood frequency) are common in this LSA and many wetland areas are underlain by peat, muck or fine-grained lacustrine sediments that may contain well-preserved paleoenvironmental records (from pollen, plant macrofossils, insects, ostracods, fish, etc.). Deposits range in age from 10,400 to about 4,500 B.P. Thin younger deposits may overlap portions of this LSA.

The Late Holocene Channel Belt (LAHOL)

The Late Holocene Channel Belt encompasses low relief, slightly to moderately undulating, poorly drained, broadly arcuate surfaces where sloughs and abandoned channels are abundant on the Mississippi floodplain. LAHOL deposits consist of variable thicknesses of loamy, silty clay loam, and clay loam alluvium overlying sand and pebbly sand in-channel and sand ridge deposits. The fine-grained deposits range from 1-2.5 meters (3.3-8.2 ft) in thickness and mantle most of the LSA but are thin to absent on ridges. Deposits of this LSA accumulated between 4,500 B.P. and the Historic period.

The Yazoo Meander Belt (YAZOO)

The Yazoo Meander Belt consists of low-relief, undulating to slightly undulating, arcuate surfaces associated with tributary and anabranch stream channel belts on the Mississippi floodplain. Deposits consist of a complex mosaic of criss-crossing channels and associated scroll bars, natural levees, and abandoned channels (which may contain marshes and small oxbow lakes). Although no complete record is present at any one locality, these deposits span the Holocene.

Mississippi Levee (LEVEE)

The Missississippi Levee LSA occurs in isolated areas of the Upper Mississippi Valley in areas with relatively stable river banks. Deposits typically consist of planar and trough cross-bedded loam, silty clay loam, sand, and pebbly sand. Deposits are sandier near the river and coarsen in the distributary channels. Chronologic control is limited, but it suggests that these deposits began accumulating near 7,000 B.P. and continued to the Late Holocene in some areas. LEVEE deposits may bury and interfinger with deposits of the EMHOL, YAZOO, and FANCO LSAs. The Late Holocene Channel Belt may either cut or or interfinger with LEVEE and is truncated by ISLAN. LEVEE does not occur with the Savanna Terrace, Kingston Terrace, or Tributary Fan LSAs.

Tributary Fan (TRIFA)

The Tributary Fan LSA includes all deposits and landforms related to major tributary rivers on the Mississippi floodplain and the lower reaches of tributary valleys including abandoned channels, oxbow lakes, scroll-bar complexes, natural levees, crevasse splays, and floodbasins. These surfaces are slightly elevated above the Mississippi floodplain and are inset below the Savanna and Kingston terraces. TRIFA can can interfinger with, bury or truncate all other LSAs except YAZOO, LEVEE and ISLAN. Deposits are highly variable and typically consist of two to four meters of fine-grained alluvium grading downward to sand and pebbly sand alluvium. On natural levees and splays the fine-grained mantle is thinner or absent, whereas abandoned channels may have thick fine-grained fills and include zones of detrital organics, peat or muck.

Island (ISLAN)

The Island LSA includes all landforms and deposits on historic islands in the Mississippi Valley. The ISLAN deposits consist of variable thicknesses of fine-grained alluvium overlying sandy and pebbly sand alluvium. Fine-grained deposits may be lacking on sandy ridges and recently formed bars on the island margins. South of the Quad Cities, post-settlement alluvium mantles all surfaces in this LSA, reaching thicknesses of two meters or more along the margins of the islands (thickness of the post sediment alluvium is more variable north of the Quad Cities ranging from >1m to a few cm). Deposits in the ISLAN LSA range from about 5,000 B.P. on large islands north of the Quad Cities to less than 3,500 B.P. on large islands south of the Quad Cities.

Unknown Category (UNKNOWN)

Several areas are mapped as “unknown”. Reasons for this designation include landforms that are not distinguishable on air photos and cannot be field checked, as well as landforms that have been destroyed and cannot be attributed to a different LSA.

Post-Settlement Alluvium (PSA)

Post-Settlement Alluvium usually occurs on other LSAs described above and is therefore not a discrete LSA. This LSA was established primarily for the purposes of locating cultural resources, as its thickness may help determine appropriate techniques. PSA is divided into “Pre-Lock and Dam PSA” and “Post-Lock and Dam PSA”, but the distinction is not always easy to make. Pre-Lock and Dam PSA is located primarily on the Island LSA south of the Quad Cities and consists of silty, loamy, and sandy, horizontally bedded, noncalcareous alluvium. It is usually less than 0.5m (1.5 ft) thick and buries the early Historic land surface. Post-Lock and Dam PSA occurs on many LSAs, especially in the lower elevation areas. These deposits consist of horizontally bedded, sand, silt, and loam that is calcareous and is thickest between the artificial levees. Post-Lock and Dam PSA is lighter colored and much thicker than Pre-Lock and Dam PSA.

TERRACES IN THE VICINITY OF CRAPO PARK

The primary terraces near Crapo Park are the Savanna Terrace and the Early Middle Holocene Channel Belt. The Early Middle Holocene Channel Belt comprises the floodplain east of the river up to approximately 550’ (168m) elevation. The Savanna Terrace extends from the 550’ (168m) contour interval to the eastern edge of the floodplain. Other LSAs in the area include Island, the Late Holocene Channel Belt, and the Kingston Terrace. The Island LSA includes the area between the river channels and the Late Holocene Channel Belt is in a thin band on the east margin of the river. Kingston Terrace deposits are in patchy areas within the Early and Middle Holocene Channel Belt.

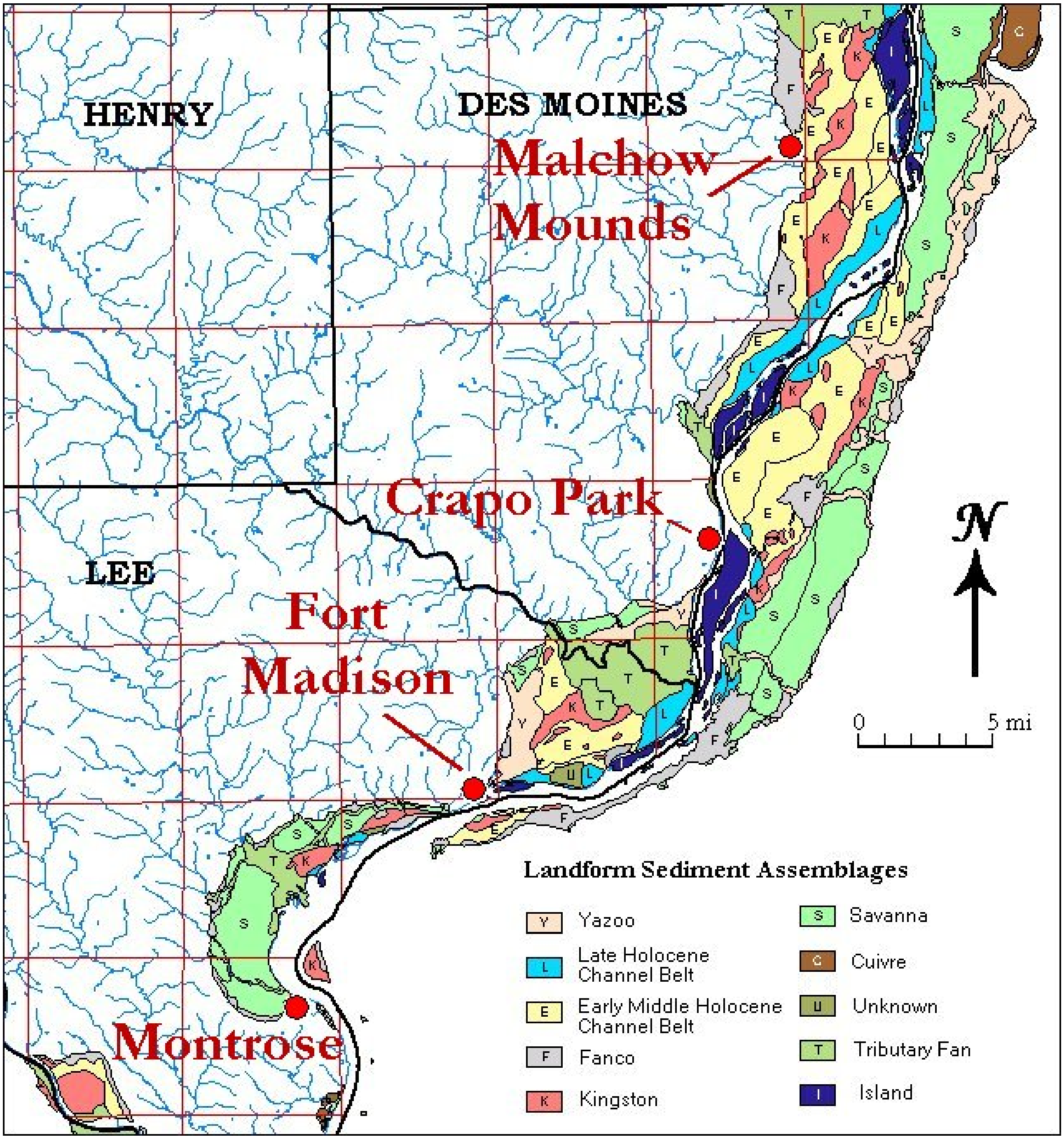

Figure 2. Landform Sediment Assemblages of the Upper Mississippi River Valley (Modified from Bettis et al., 1996)

TERRACES IN THE VICINITY OF MALCHOW MOUNDS

LSAs near Malchow Mounds include the Early to Middle Channel Belt, the Kingston Terrace, Island deposits, the Early to Middle Holocene Channel Belt, the Savanna Terrace, and the Yazoo Meander Belt. The Early to Middle Holocene Channel Belt extends from the western bluffline to the river and has a few areas of Kinston Terrace present within it. Island deposits are present in the middle of the river channel and the Late Holocene Channel Belt borders the river on the east side. The predominant deposits east of the river are Savanna Terrace deposits. A band of the Yazoo Meander Belt makes up the eastern boundary of the floodplain.

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PRE-WISCONSINAN STRATIGRAPHY IN SOUTHEAST IOWA

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INTRODUCTION

Nelson Quarry provides a unique look at an Illinoian till exposure. The Illinoian age till was first described by Leverett, and the current stratigraphy in Iowa was defined by Hallberg (1980). The majority of this work was based on cores, as exposures are not common. The Illinoian till advanced into Iowa from the northeast and has a limited distribution in the state.

ILLINOIAN

Historical Studies

Leverett (1898a, 1899) was the first to study the Illinoian glacial deposits in southeast Iowa. Early studies included detailed observations of the Pleistocene deposits of the "Illinoian Lobe" and led to the conclusion that the ice had advanced into Iowa from the northeast across Illinois. The term Illinoian stage was first used to include both the interval of glaciation and the drift sheet. Leverett (1898a, 1898b, 1899) also identified weathering zones marking the upper and lower boundaries of the Illinoian and termed these the Yarmouth and Sangamon soils, respectively.

Subsequent researchers have made few changes to the original concept of the Illinoian stage as defined by Leverett (1899). The most significant change was the subdivision into three substages (Willman et al., 1963; Frye et al., 1964; Willman and Frye, 1970). These time-stratigraphic units were based on end moraine positions and mineralogy and were originally termed (oldest to youngest) the Payson, Jacksonville, and Buffalo Hart substages. The rock-stratigraphic concepts by Willman and Frye in the 1960's for the Quaternary were formalized in 1970 with the naming of several till members of the Glasford Formation. (Lineback, 1979)

Work during the 1950's and 1960's focused on characterizing the mineralogy and petrography. Multiple till units were identified and related to the substages of the Illinoian glaciation. Leverett's Illinoian till was renamed by Leighton and Willman (1950) as the Payson substage, and later modified by Willman et al. (1963) to include the Petersburg Silt. In 1964, it was suggested (Frye et al.) that the terms Mendon Till and Payson till be abandoned due to a lack of Illinoian age tills as far west as had been previously mapped. Instead, the name Liman was proposed for the oldest substages of the Illinoian which encompassed the time of deposition of the Petersburg Silt, the Mendon Till, and other Illinoian deposits stratigraphically below the Jacksonville Till.

The Pleistocene deposits and stratigraphy of Illinois were more formally classified by Willman and Frye in 1970, and the oldest Illinoian age till was named the Kellerville Till Member of the Glasford Formation. Lineback (1979) later divided the Kellerville into two members: the upper (unnamed Till A) and lower. However, data presented by Wickham (1980) does not provide enough evidence to support the subdivision of the Kellerville Till into a lower part and an upper part. The study suggested instead that unnamed till member A is a lateral variation of the type Kellerville Till Member, as unnamed till member A was not found in superposition with the lower part of the Kellerville Till.

Hallberg (1980) defined the formal stratigraphic nomenclature in Iowa. The only Glasford Formation till present in Iowa is the Kellerville Till Member which was described from roadcut exposures in western Illinois (Willman and Frye, 1970). The type section is located in southeastern Adams County, Illinois, approximately 100 km (62 miles) south of Burlington, Iowa. The Kellerville Till Member replaces the terms Mendon Till (Frye et al., 1964; Frye et al., 1969) and Payson Till (Leighton and Willman, 1950; Wanless, 1957).

Source Area

The Illinoian ice sheet is the only ice advance into Iowa believed to have a northeastern source area. Based on the eastern-derived erratics and the prominent west-facing terminal moraines, it was determined that the ice had moved from the east across Illinois and into Iowa, leaving drift widely exposed in Illinois, Indiana, and Ohio (Leverett, 1898a; 1899). The Illinoian till was deposited by the advancing Lake Michigan Lobe which moved across western Illinois into Iowa from the northeast (Leverett, 1899; Wickham, 1980). The Lake Michigan Lobe incorporated Paleozoic bedrock materials from the Lake Michigan Basin which are distinguished by both the clay mineralogy of the matrix and the pebbles and clasts in the Illinoian deposits (Lineback, 1980; Wickham, 1980). Till fabrics and glacial landform orientations also provide evidence of a northeastern source area for these tills (Lineback, 1979). The differing provenances between the Illinoian and Pre-Illinoian tills (which moved into Iowa from north-northeast) creates identifiable and distinguishing physical and mineralogical characteristics. Most specifically, the Kellerville Till Member is differentiated from the Pre-Illinoian tills by the relatively high illite content, high dolomite content and the abundance of Pennsylvanian lithologies in the very coarse sand through cobble size particles. Physical stratigraphy, pebble lithology, and quantitative values of clay mineralogy, particle-size distribution, matrix carbonates, and sand-fraction lithology have been used to correlate the Illinoian deposits (Hallberg et al., 1980).

Distribution and Thickness

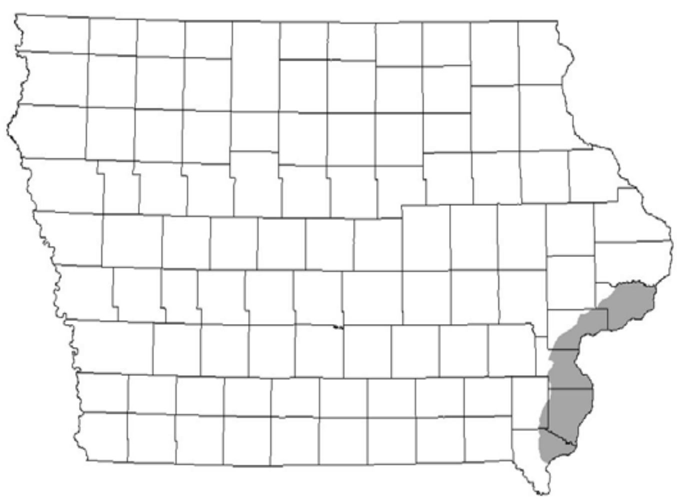

Figure 1. Map showing the distribution of Illinoian till in southeast Iowa.

As shown in Figure 1, the Lake Michigan Lobe of Illinoian ice did not reach far into Iowa, only advancing into the most southeastern portion of the state. Leverett (1898a) originally defined the terminal boundary, and his distribution map in Iowa has only been slightly modified since (Kay and Graham, 1943; Ruhe, 1969).

The Illinoian deposits in southeast Iowa extend along the western edge of the Mississippi River from just south of Fort Madison northward to near the mouth of the Wapsipinicon River at the boundary between Scott and Clinton Counties (Leverett, 1898a, Kay and Graham, 1943; Ruhe, 1969). The distance west from the Mississippi River varies, ranging from approximately 6.5 km (4 miles) near Muscatine to 32 km (20 miles) north of Burlington. Illinoian deposits have been identified in Lee (NE ¼), Des Moines (all but NW corner), Henry (only SE corner), Louisa (eastern 2/3), Muscatine (all but NW ¼), and Scott Counties (southern 2/3).

PRE-WISCONSINAN STRATIGRAPHY IN SOUTHEAST IOWA

Three Pre-Wisconsin till formations, each containing several members, are present in southeastern Iowa: the Illinoian age Glasford Formation and the Pre-Illinoian Wolf Creek and Alburnett formations. Paleosols are associated with each of the till members. These formations are not all present at any one individual section or core, but all are found at some locality in southeastern Iowa.

The basic stratigraphy of southeast Iowa consists of the following:

- Wisconsinan Loess
- Sangamon and Late-Sangamon Paleosols
- Glasford Formation
  - Kellerville Till Member
    - Superglacial Facies
    - Subglacial/Basal Till Facies
- Yarmouth Paleosol
- Wolf Creek Formation
  - Hickory Hills Till Member
  - Dysart Paleosol
  - Aurora Till Member
  - Franklin Paleosol
  - Winthrop Till Member
  - Westburg Paleosol
- Alburnett Formation
  - Undifferentiated Members

CHARACTERISTICS OF PRE-WISCONSINAN TILLS

The Pre-Illinoian deposits (Alburnett and Wolf Creek formations) of southeast Iowa have the same general characteristics as described by Hallberg (1980) in east-central Iowa and are differentiated in the same manner. As seen in other areas, the Wolf Creek and Alburnett formations represent basal tills (uniform, dense, overconsolidated), and are separated based on clay mineralogy. The Alburnett Formation represents the oldest identified glacial deposits in southeast Iowa and consists of several undifferentiated members. However, only one Alburnett Formation till has been recognized at any particular site in southeast Iowa. The Wolf Creek Formation consists of three members (youngest to oldest): Hickory Hills, Aurora and Winthrop. These deposits show essentially the same stratigraphy and characteristics as elsewhere in Iowa. The Kellerville Till Member and associated sediments of the Glasford Formation are the only Illinoian age materials identified in the area.

Distinction between Illinoian and Pre-Illinoian Materials

The distinction between Illinoian and Pre-Illinoian tills is based primarily on matrix carbonate data, clay mineralogy, and sand-fraction lithology. The contrasting mineralogy is a result of the Illinoian tills having an eastern (Lake Michigan Lobe) source area. Therefore, the Illinoian tills contain more illite that the Pre-Illinoian tills which entered Iowa from the northwest. The eastern source area for Illinoian tills is also evident in the abundance of Pennsylvanian rock fragments present in the sand-size through pebble and cobble fractions of the Illinoian tills (Hallberg et al., 1980). Additionally, Illinoian tills contain two to four times as much dolomite as calcite, and Pre-Illinoian tills contain anywhere from more calcite than dolomite to less than twice as much dolomite (Lineback, 1979).

One of the most distinguishing characteristics of the Kellerville Member is the high dolomite content in the matrix carbonates (Kemmis and Hallberg, 1980). The Kellerville has a much lower calcite to dolomite (C/D) ratio (less than 0.40) than the Pre-Illinoian deposits of the Wolf Creek and Alburnett formations (95% of which all have C/D ratios greater than 0.40). The Kellerville also exhibits a particularly high total sedimentary grain content in the very coarse sand fraction and an abundance of coal and black shale fragments in the sand fraction. The abundance of Pennsylvanian lithologies in the pebble fraction is also an important characteristic and often distinguishes the Kellerville in the field (Hallberg et al., 1980).

Alburnett Formation

The Alburnett Formation is composed of multiple “undifferentiated” till units, a variety of fluvial deposits, and associated minor paleosols. Throughout eastern Iowa, these deposits fill and bury deep bedrock channels. The properties for the type areas are described in detail by Hallberg (Hallberg et al., 1980). The Alburnett Formation is defined by its stratigraphic position and distinctive clay mineralogy. The clay mineralogy of the Alburnett Formation in southeast Iowa is very similar to that in the type areas, and the particle size is also similar. In comparison with the Wolf Creek Formation, the Alburnett tills have significantly lower percentages of expandable clay minerals and higher kaolinite plus chlorite. The Alburnett Formation contains 44% expandables, 24% illite, and 32% kaolinite plus chlorite (Hallberg et al., 1980). The Alburnett Formation in southeast Iowa contains 18.7% clay, 36.8% silt, and 44.4% sand (Hallberg et al., 1980). The Alburnett Formation tills have not been widely recognized in southeast Iowa, but a limited number of deep core holes have been drilled in the region. Both core holes drilled to bedrock (Yarmouth Core and Mediapolis-1) encountered the Alburnett Formation. Only one Alburnett Till was recognized in each hole, with a maximum thickness of 19 feet (5.8 m). The till was underlain directly by Mississippian bedrock with some minor inclusions of sand and gravel toward the base of the unit. The Alburnett was overlain directly by the Aurora Till Member of the Wolf Creek Formation. In outcrop, the Alburnett tills have only been identified in exposures near the bluffs of the Mississippi River in Des Moines and Lee counties. In these areas, the lower contact is not exposed and the top of the unit is overlain directly by the Wolf Creek Formation. (Hallberg et al., 1980).

Wolf Creek Formation

The Wolf Creek Formation members (Winthrop, Aurora, and Hickory Hills) consist of basal tills and intertill stratified sediments. These members are separated by soil stratigraphic units. The type areas for the Wolf Creek Formation and associated members are in east-central Iowa, but the units in southeast Iowa show the same general characteristics. The Aurora and Hickory Hills Till members are widespread throughout southeast Iowa, but the Winthrop Till Member has only been identified at a few localities.

The upper boundary of the Wolf Creek Formation is marked by the unconformable contact with Illinoian age deposits of the Glasford Formation. The Wolf Creek is underlain by either the Alburnett Formation or Paleozoic bedrock. Where Illinoian age deposits are present, the Yarmouth Paleosol is formed in the Wolf Creek Formation. Beyond the reaches of the Illinoian deposits, the Glasford Formation Yarmouth-Sangamon soil is developed in the Wolf Creek Formation deposits. The individual till members of the Wolf Creek Formation may be directly overlain by each other or be separated by undifferentiated sediments, glaciofluvial deposits, or paleosols.

Clay mineral composition is used to distinguish the Wolf Creek Formation from the Alburnett Formation. The Wolf Creek Formation averages 50-60% expandable clays (slightly lower in the southeast portion of the state), 16-19% illite, and 22-24% kaolinite plus chlorite (Hallberg et al., 1980). The three members are differentiated based on particle size and matrix carbonate data. The texture is typically loam, with the Winthrop Member ranging to light clay loam. The

Hickory Hills Member has relatively more sand, and the Aurora and Winthrop members are relatively silty. The Aurora has a higher clay content than the Winthrop. Average clay, silt, and sand percentages are listed below (Hallberg et al., 1980):

Member	Clay %	Silt %	Sand %
Hickory Hills (n=87)	21.6±1.7	32.9±2.4	45.5±2.3
Aurora (n=41)	20.5±2.2	41.6±3.4	38.0±3.5
Winthrop (n=21)	27.6±1.9	42.0±2.3	30.4±1.9

Texturally the Aurora Till Member is very similar to the Kellerville Till Member, but mineralogically (clay mineraology and matrix carbonate) they are very different. Also, 2% of the Aurora Till Member sand-fraction samples show traces of coal in the southeastern portion of Iowa. Although this is a characteristic of Illinoian tills, the Kellerville Till Member shows traces of coal and Pennsylvanian lithologies in much greater abundance than that of the Aurora Till Member (Hallberg et al., 1980). Throughout southeast Iowa the Hickory Hills Till Member is typically very uniform in properties, both vertically and laterally, and may contain some block inclusions of substrate materials in its lowermost portions (Hallberg et al., 1980).

Glasford Formation- Kellerville Till Member

The Kellerville Till Member of the Glasford Formation is the oldest of the Illinoian age tills and is the only one present in Iowa. The Kellerville is separated into two till facies (a subglacial or basal till facies and a superglacial facies) based on stratigraphic position, sedimentological properties, and the consistency-density-consolidation properties (Hallberg et al., 1980; Wickham, 1980; Lineback, 1979). The two facies are similar in mineralogy (both clays and sand-fraction lithology), but not physical characteristics.

The subglacial till facies has a firm, dense uniform till matrix with texture ranging from silt loam to a light clay loam (on average a loam till relatively high in silt). In contrast, the superglacial facies may be composed of till, diamicton (reworked till such as superglacial debris flows), sorted fluvial and lacustrine sediments, and peat beds. Deposits may be interbedded or occur as a contorted melange of sediments. The superglacial till is highly variable in density and texture, and contains a wide variety of sediments. The texture values tend to cluster if samples are from the same general area. The overall range of matrix texture includes sandy loam, loam, silt loam, silty clay loam, clay loam, and clay. The stratified deposits within the superglacial till facies vary in texture from sand, sand and gravel, to very heavy clay. The superglacial facies also contains some peats and organic silts (Hallberg and Baker, 1980).

Although the superglacial and subglacial facies vary widely in texture, they are very similar in mineralogy, especially the clay abundance and the sand-fraction lithology. The clays typically contain 46% expandables, 34% illite, and 20% kaolinite plus chlorite. Additional clay mineral characteristics (including a high illite to kaolinite plus chlorite ratio, moderate amounts of expandables, and the frequent occurrence of identifiable chlorite peaks) are used to distinguish the Kellerville from other units in the area. Clay mineral data from the Kellerville Member are more variable (wider range) than for the Pre-Illinoian tills (Hallberg et al., 1980; Lineback, 1979). Pebble lithologies commonly include coal fragments and other Pennsylvanian clasts.

NELSON QUARRY

Site Description

The Quaternary materials at Nelson Quarry offer one of the best exposures of Illinoian age deposits in southeast Iowa. A large part of the stratigraphic work conducted in the region was from core studies, and exposures of this thickness are uncommon. The general Quaternary stratigraphy of the quarry consists (base to top) of the Kellerville Till Member of the Glasford Formation (subglacial and superglacial facies present), the Sangamon Geosol, and a Wisconsin Loess cap. The subglacial facies of the Kellerville till member is a uniform dense till. The superglacial facies is highly variable throughout the quarry ranging from sandy loam to silty clay loam, but the matrix is predominantly loam to silt loam. Throughout the 13.5 m (44 ft) section the physical properties of the deposits change and thicknesses are variable. A stratigraphic section was described in the northeast portion of the quarry.

Other interesting features present are a second till unit along the southeast quarry wall and a swale fill of unknown origin below the Illinoian age till. The till along the southeast wall is truncated from the main quarry wall and either represents a Pre-Illinoian till or a basal Illinoian till. Additional samples were collected from the isolated till deposit along the south wall to try to determine the relationship with the deposits in the northeast portion of the quarry.

Stratigraphic Section Description

The main section description was taken from the northeastern portion of the quarry (see Figure 2). This section consisted of the Kellerville Till Member, Sangamon Geosol, and Wisconsin loess deposits. The Kellerville Till Member was deposited on top of a swale fill. The contact was abrupt, and the nature of the relationship could not be determined. It must also be noted that the physical characteristics of the Kellerville deposits (especially grain-size) had a wide range throughout the quarry. Several areas appeared to have slump or slope deposits. Coal and Pennsylvanian lithologies were abundant throughout the till and the clast size increased to the base.

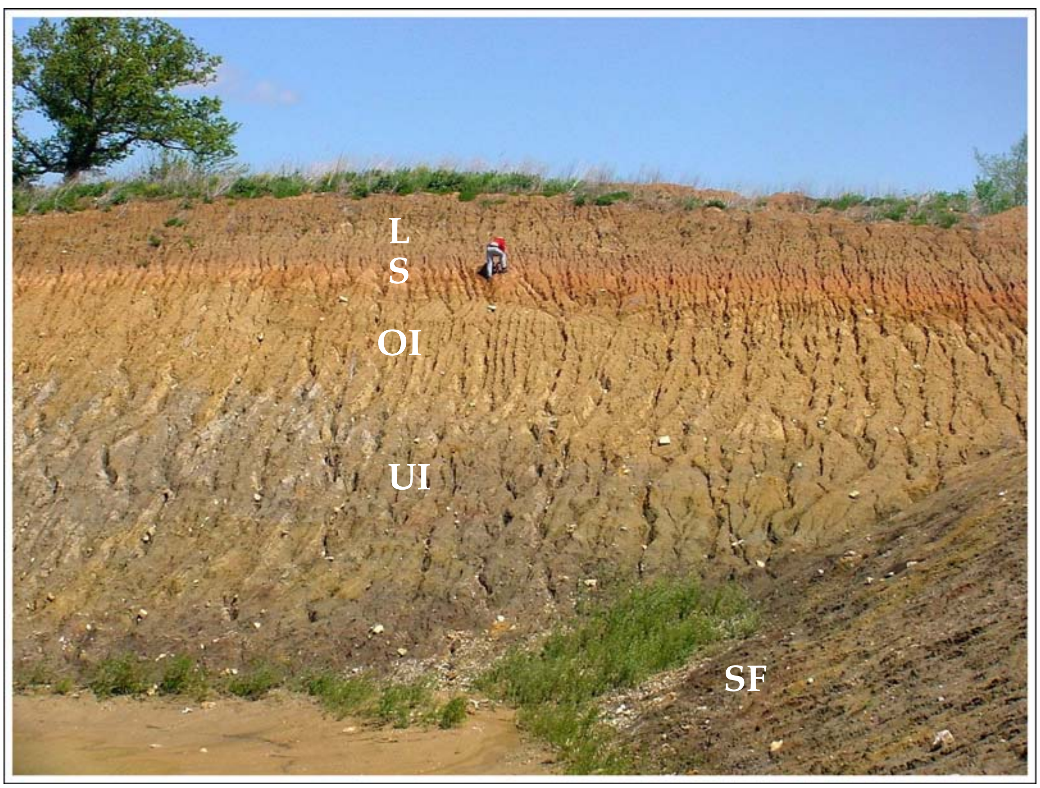

Figure 2. Stratigraphic section described in the northeast portion of the Nelson Quarry. A basal swale fill (SF) is overlain by an Illinoian till, unoxidized (UI) and oxidized (OI). The Sangamon Geosol (S) is formed in the Illinoian till. The section is capped by an oxidized loess (L).

The stratigraphic section from the northeastern portion of the quarry is presented below (*All colors moist):

WISCONSINAN LOESS
0.0-1.5m (0.0-4.9 ft)
OL- Yellowish brown (10YR 5/4) silt (loess), weak fine subangular blocky, friable, few fine to medium yellowish brown (10YR 5/8) Fe accumulations, few clay coatings

SANGAMON GEOSOL
1.5-3.4m (4.9-11.2 ft)
Strong brown (7.5YR 4/6) with dark yellowish brown (10YR 4/6) areas and few yellowish brown (10YR 5/6) areas at base grading upward to yellowish brown (10YR 5/6) with common strong brown (7.5YR 4/6) areas and few dark yellowish brown (10YR 3/6) areas, loam to sandy loam, weak fine to medium subangular blocky, friable to loose, leached

ILLINOIAN TILL
3.4-4.8m (11.2-15.7 ft)
OU TILL- Yellowish brown (10YR 5/4) silt loam, massive, firm, strongly effervescent, common black (10YR 2/1) Mn accumulations, few dark yellowish brown (10YR 4/6) Fe accumulations

4.8-7.0m (15.7-23.0 ft)
UJU/RJU TILL- UU- Dark gray (5Y 4/1) silt loam, massive, firm, strongly effervescent, common joints with oxidized/reduced colors yellowish brown (10YR 5/6) and light olive brown (2.5Y5/6); RU- Olive brown (2.5Y 4/3) and dark olive brown (5Y 3/2) silt loam, weak medium subangular block to platy, strongly effervescent; heavily jointed areas- dark yellowish brown (10YR 4/6), light olive brown (2.5Y 5/4), and dark gray (5Y 4/1) colors along joint surfaces in oxidized and depleted zones

7.0-9.6m (23.0-31.5 ft)
UJU/RU TILL- UU- Very dark grayish brown (2.5Y 3/2) loam, massive, firm, dark yellowish brown (10YR 4/6) Fe coatings along joints, few fine twigs, strongly effervescent, prominent joints several cm in diameter; common reduced areas- olive brown (2.5Y 4/3 to 2.5Y 4/4) massive, firm, strongly effervescent

9.6-11.9m (31.5-39.0 ft)
UU TILL- Dark olive brown (5Y 3/2) silt loam, uniform, massive, moderately effervescent, few fine wood fragments, common olive brown (2.5YR 4/3) reduced areas

SWALE FILL
11.9-12.2m (39.0-40.0 ft)
LAMINATED DEPOSITS- Olive brown (2.5Y 4/4) and gray (2.5Y 5/1) silt, thinly laminated, strongly effervescent

12.2-13.5m (40.0-44.0 ft)
MASSIVE DEPOSITS- Gradational color change- Dark gray (2.5Y 4/1) and olive brown (2.5Y 4/4) at base to very dark gray (5Y 3/1) to dark grayish brown (10YR 3/2) at top; gradually fining upward from loam to silt; friable; structural range from moderate fine to medium subangular blocky upward to weak to moderate fine subangular blocky; moderately effervescent; few fine dark yellowish brown (10YR 4/4) Fe accumulations; few dark gray (10YR 4/1) areas and common dark yellowish brown (10YR 3/4) areas near the top of the unit

Swale Fill

The swale fill is present in the northeast portion of the quarry. This fill is greater than 1.5m (5 ft) thick and is truncated along the middle of the east wall by a bedrock high. The base could not be identified, but the increasing grain-size and the addition of pebbles near the base indicates that bedrock is likely not far below the base of the exposed section. Organic materials and possible aquatic snails were identified in these deposits, but the origin and relationship with the overlying till is unclear. Its formation may have been related to variations in bedrock topography or ice-marginal position. The swale fill contains organic rich areas as well as laminated deposits at irregular spacings, but usually occur near the upper surface. Vertical variations are noted in both the texture and the nature of the deposit.

Pre-Illinoian(?) Till

Near the middle of the east wall of the quarry, a change in bedrock elevation truncates the basal portion of the Kellerville Till (see Figure 3). An unoxidized and unleached till is present in the southeastern corner of the quarry. Based on the physical properties of the till (homogeneity, grain-size data, lack of coarser pebbles and clasts, and lack of Pennsylvanian lithologies), it is believed this till may be Pre-Illinoian in age. However, the clay mineralogy data was not sufficient to make this determination. If it is not a Pre-Illinoian till, it is a basal Illinoian till.

The till is a dark olive brown (5Y 3/2) clay loam to loam with pebbles up to 3cm diameter. It is uniform, massive, very firm and stongly effervescent. It did not seem to have as many coarse pebbles and cobbles as the other till. Stratigraphically, 4-5m (13-16 ft) of the unoxidized unleached till was overlain by the Sangamon Geosol and Wisconsinan loess as seen elsewhere along the quarry face. Seems (up to 15cm thick) of fine to medium uniform brown to tan sand are common throughout the till.

Figure 3. Pre-Illinoian (?) till (between white arrows) truncated by bedrock surface in the southeast portion of the Nelson Quarry.

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BEDROCK GEOLOGY IN THE BURLINGTON AREA, SOUTHEAST IOWA

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INTRODUCTION

The Mississippian System (now Subsystem) was historically proposed for the succession of strata exposed in the Mississippi River Valley between Burlington, Iowa, and southern Illinois. Therefore, the bedrock exposures in and around the City of Burlington take on special significance as they comprise part of the historic "body stratotype" on which the concept of the Mississippian System was defined and based. Until recently, the Mississippian has been primarily a North American chronostratigraphic label roughly synonymous with the Lower Carboniferous of the Old World. As recently approved by the Subcommission on Carboniferous Stratigraphy (in 1999) and ratified by the International Union of Geological Sciences and the International Commission on Stratigraphy (in 2000), the Carboniferous System has been officially subdivided into lower and upper subsystems, the Mississippian and Pennsylvanian, respectively. As such, the Mississippian now has meaning and application as a major subdivision of geologic time not only in North America, but as a globally defined subsystem. The bedrock strata at Burlington provide a significant historic reference for the Mississippian. This Tri-State field conference can be viewed as geologic pilgrimage to the classic Mississippian succession in its type area.

The term "Mississippi Group" was introduced by Winchell (1869, p. 79) who wrote, "I propose the use of this term [Mississippi Group] as a geographical designation for the Carboniferous Limestones of the United States which are so largely developed in the valley of the Mississippi River." Williams (1891, p. 135) first used the Mississippian as a time-stratigraphic term by designating it a series, and Chamberlain and Salisbury (1906, p. 496) subsequently elevated it to the status of a geologic period. North American stratigraphers have considered the Mississippian to be a system since 1915 when the USGS recognized it as such (Lane and Brenckle, 2001).

Bedrock strata are well exposed in the Burlington area at places along the Mississippi River bluffs, its tributary valleys, and in numerous abandoned and operating quarries. The exposed stratigraphic succession begins in the lower reaches of the valley walls in an interval of Upper Devonian shale and siltstone. Proceeding upward, the Devonian strata are capped by a picturesque and fascinating succession of Mississippian strata including limestone, dolomite, chert, and siltstone lithologies. The thick limestone and dolomite interval that caps the Mississippi River bluffs at Burlington comprises the type area of the Burlington Formation, world-renowned for its exceptional fossil faunas, especially of crinoids. The reader is referred to a more comprehensive summary of Mississippian rocks in southeast Iowa by Witzke et al. (1990) for further information. Field guides to the Burlington area by Glenister et al. (1987) and Witzke and Tassier-Surine (2001) provide additional information. The succeeding overview in this guidebook of the bedrock stratigraphy seen in the Burlington area (including Des Moines and Lee counties, Iowa) provides field trip participants with the general background to the bedrock geology.

UPPER DEVONIAN SHALE AND SILTSTONE STRATA

English River Siltstone and "Maple Mill" Shale

A succession of shale and siltstone strata begins at the level of the Mississippi River at Burlington and rises up the lower valley walls a short distance (locally varying between about 20 to 45 feet [6-14 m] in thickness). The upper siltstone-dominated portion of this interval (about 20-25 ft [6-7.5 m] thick at

Burlington) has been generally referred to the English River Siltstone (beginning with Laudon, 1931), a formation whose name derives from exposures along the English River in Washington County, Iowa ("English river gritstone" of Bain, 1896). This siltstone was earlier called the "Chonopectus sandstone," named after a distinctive and abundant brachiopod (*Chonopectus fischeri*) found in these strata at Burlington. This siltstone interval is conformable and gradational with blue-gray shale strata below, and these shales have been traditionally assigned to the Maple Mill Shale across southeast Iowa (Laudon, 1931). The Maple Mill Shale also derives its name from exposures along the English River in Washington County (Bain, 1896).

The lower shale and siltstone interval exposed at Burlington was once included as the lower part of the "Kinderhook" beds or group (beginning in 1861 with the James Hall era and continuing well into the 20th century, e.g., Laudon, 1931). The term derived from nearby Kinderhook, Pike County, Illinois, and the entire succession of "Kinderhook beds" (capped by the Burlington Limestone) was considered to be of Carboniferous (Mississippian) age. However, it was subsequently discovered that the lower shale-siltstone package at Burlington is, in fact, an Upper Devonian interval, not part of the Carboniferous at all. As such, these strata were removed from the "Kinderhook group," and the name Kinderhook was applied only to the overlying succession, above the English River Siltstone and below the Burlington Limestone. The Kinderhookian is now a chronostratigraphic label for the basal Series of the Mississippian System.

Most geologists and paleontologists working in the area now recognize the English River Siltstone as an Upper Devonian rock unit (not Kinderhookian). This age assignment is based on the recovery of Devonian conodonts from the English River Siltstone at Burlington (Cascade Station) and in the type area along the English River in Washington County (Collinson, 1961; Scott and Collinson, 1961; Straka, 1968). In addition, clymeniid ammonoids (*Cyrtoclymenia strigata, Imitoceras opimum*) have also been collected in the Burlington area that clearly indicate an Upper Devonian (upper Famennian) age (House, 1962; Glenister et al., 1987). The English River Siltstone elsewhere in southeast Iowa has also yielded Upper Devonian conodont faunas which Pavlicek (1986) assigned to the *Polygnathus delicatulus* Zone of upper Famennian age.

Nevertheless, Carter (1988, p. 11) reached the following conclusion about correlation of the rich brachiopod fauna from the English River Siltstone at Burlington: "the Glen Park of Missouri, . . . the Horton Creek . . . and lower Hannibal . . . of Illinois, and the English River Sandstone of southeastern Iowa are the same age and can be confidently correlated." Because the Glen Park, Horton Creek, and lower Hannibal are all Mississippian (lower Kinderhookian) units, Carter therefore included the English River Siltstone at Burlington within the Mississippian. The conodont and ammonoid biostratigraphy, however, clearly indicates that his correlations are in error. There is great similarity between the brachiopod faunas of the late Famennian (English River) and early Kinderhookian (Glen Park, Horton Creek, lower Hannibal) in the area, and many brachiopod species apparently ranged across the Devonian-Mississippian boundary unchanged.

The English River Siltstone at Burlington contains a rich and varied fauna, and their fossils are seen as well-preserved internal and external molds. Fossils are most abundant in the upper beds of the siltstone unit, and the lower beds are variably fossiliferous (commonly burrowed but with sparse shelly fauna). The shelly fauna at Burlington is highly diverse, typically dominated by bivalves (clams) and brachiopods. About 25 species of brachiopods (Weller, 1900; Carter, 1988) and 32 species of bivalves (Weller, 1900) are recognized. In addition, gastropods (21 species), cephalopods, scaphopods, conularids, bryozoans, and crinoid debris are also noted.

It is generally accurate to correlate the lower shale-siltstone succession at Burlington with the Maple Mill-English River succession along the English River in Washington County, but some stratigraphic observations are pertinent. First, the English River Siltstone at its type locality, as originally defined by Bain (1896), included an upper siltstone interval that is now known to be of Mississippian age (and which correlates with the Prospect Hill Siltstone at Burlington; see Straka, 1968). Only the lower two-thirds of the siltstone interval at the English River type locality actually correlates with the Devonian siltstone at Burlington, and this interval now represents the re-defined type section of the English River Siltstone

(Straka, 1968). Second, the English River Siltstone and Maple Mill Shale have traditionally been considered to be separate formations. Nevertheless, a distinctive and consistent separation of these two units is not always possible across southeast Iowa, and the contact is commonly gradational. An interval of interbedded shale and siltstone is common in many sections. In addition, regional subsurface stratigraphic investigations of the Upper Devonian shale-siltstone interval across southeast Iowa reveal that the English River Siltstone regionally shares lateral lithofacies relationships with the upper Maple Mill Shale. That is, the English River Siltstone is locally replaced by shale-dominated facies that are included in the Maple Mill Shale. As such, the lithostratigraphic separation of English River and Maple Mill strata does not mark a regionally correlatable stratigraphic datum, but merely reflects a complex siltstone-shale facies transition. Because of these facies relationships, the English River logically could be included as an upper member within a larger and thicker shale-dominated formation.

The Maple Mill Shale was originally defined and historically used as a lithostratigraphic label to include the entire succession of thick Upper Devonian shale found above the Cedar Valley Limestone in southeast Iowa. This broadly defined shale interval can and should be further subdivided into formation-level or member-level lithostratigraphic packages, as recommended by Dorheim, Koch, and Parker (1969), who grouped the Upper Devonian shale and siltstone succession into the "Yellow Spring Group" (type locality in Des Moines County). They restricted usage of the "Maple Mill"to the upper half of this succession below the capping English River Siltstone. Two additional Upper Devonian shale formations have been previously applied to the lower half of this succession in southeast Iowa, namely the Sweetland Creek Shale (or Lime Creek Shale) and Sheffield Shale. The Sweetland Creek Shale is an Upper Devonian shale interval named for a cut-bank exposure in southeast Iowa (Muscatine County), and this unit has been shown to have stratigraphic utility as a formation across the Illinois Basin (Cluff et al., 1981). As an Iowa-defined term, the Sweetland Creek Shale should be recognized as a distinct lithostratigraphic unit in southeast Iowa. By contrast, the Sheffield Shale, a term introduced from north-central Iowa, has been used inappropriately in southeast Iowa (primarily on well logs), and further use of this term is discouraged in southeast Iowa. As used by Dorheim et al. (1969), the so-called "Sheffield" of southeast Iowa is not a precise lithostratigraphic or chronostratigraphic equivalent of the type Sheffield of north-central Iowa (see conodont correlations of Pavlicek, 1986, and Metzger, 1989). The term "Yellow Spring Group" seems largely synonymous with the "New Albany Shale Group" as presently used in Illinois (e.g., Cluff et al., 1981).

Maple Mill-English River Succession in Southeast Iowa

The Upper Devonian shale-dominated succession in southeast Iowa (Lee and Des Moines counties) ranges between about 200 and 250 feet (60-76 m) in thickness. Only the upper portion of this succession is exposed at the surface in southeast Iowa, but complete sections of this shale interval are available in subsurface exploration cores in Lee County (Fig. 1). Fortunately, conodont biostratigraphy has been examined from one of these cores (H-29; Pavlicek, 1987). Above the condensed Sweetland Creek Shale (an upper Frasnian interval equivalent to the Lime Creek Fm of northern Iowa), the main mass of the overlying Famennian shale interval can be subdivided into two major cyclic units. These units each comprise a transgressive-regressive stratigraphic sequence, and the two units correlate with T-R cycles IIe-1 and IIe-2 (see Woodruff, 1990, and Johnson et al., 1985, for summary of Devonian T-R cycles). Each of these Famennian cycles displays two general facies: 1) laminated brown organic shales (mostly unburrowed) in the lower part, and 2) green-gray silty shales with burrow fabrics in the upper part. Silt laminae and thin siltstone beds are scattered within the green-gray shale intervals, and sparse fossils (in addition to the burrow mottles) include lingulid and chonetid brachiopods and branchiopod crustaceans (*Estheria*). An interbedded transition between these two general shale facies is also displayed in each cycle (see Fig. 1).

These two major shale units are each interpreted to represent major transgressive-regressive (deepening-shallowing) cycles of deposition within a stratified seaway (Witzke, 1987). The brown organic-rich laminated shale facies were deposited under dysoxic to anoxic bottom waters. The bottom environments were apparently hostile to most benthic organisms, as suggested by the paucity of burrowing and absence of shelly faunas in these facies. These facies are interpreted to have been deposited in the deepest environments of the cycle. By contrast, the green-gray silty shales were deposited in shallower environments in which the bottom conditions were slightly better oxygenated, probably deposited in dysoxic and low-oxygen environments within and above the pycnocline. Oxygenation, albeit in low concentrations, enabled burrowing organisms to thrive at times, and shelly benthic faunas (low-diversity lingulid and chonetid associations) were present during episodes of highest bottom oxygenation. Episodic influx of silt laminae and thin siltstone lenses into the green-gray facies may represent winnowing and transportation by distal storm currents.

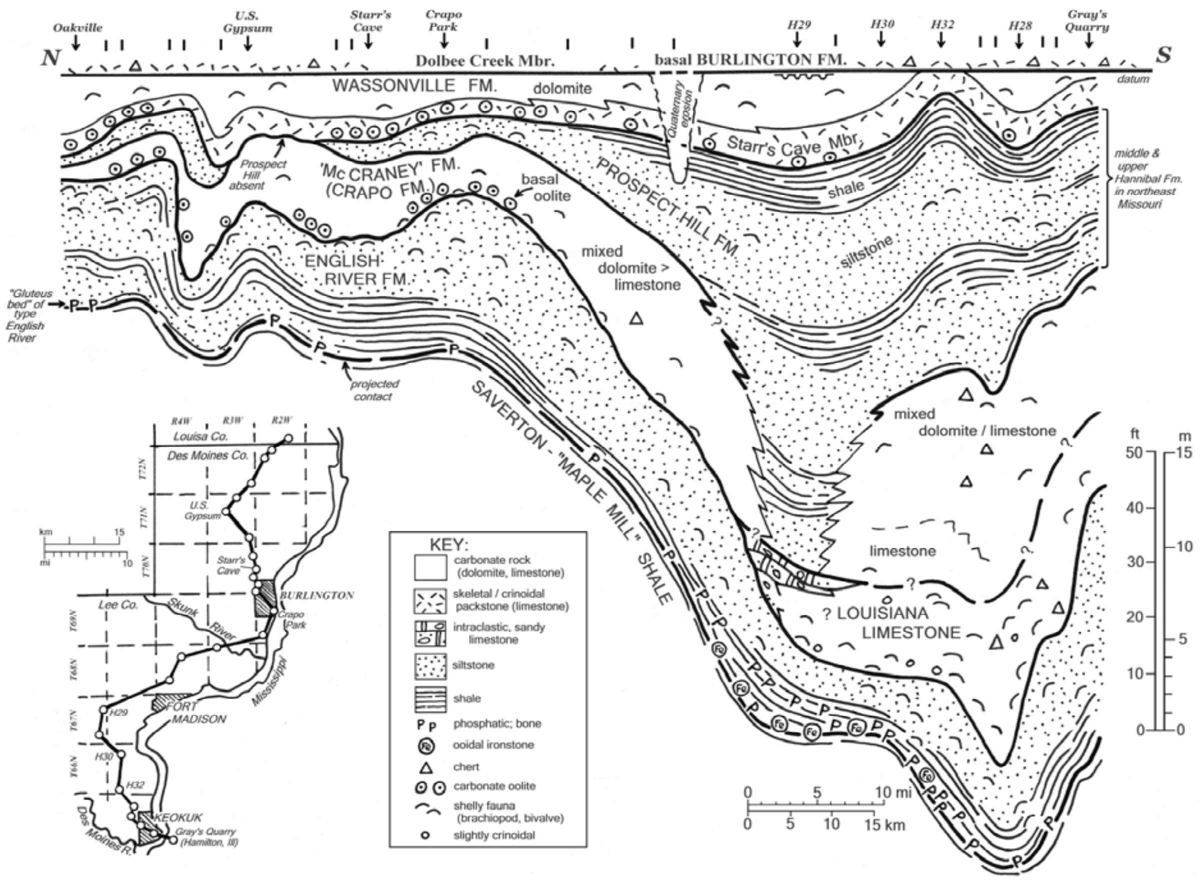

Figure 1. Schematic north-south cross section of uppermost Devonian (English River Fm.) and Kinderhookian (Lower Mississippian) strata across Des Moines and Lee counties, Iowa. Datum is base of Burlington Formation. Uncertain stratigraphic relationships are queried. The Prospect Hill Formation is thin to locally absent in Des Moines County, but thickens dramatically southward into Lee County. The base of a widespread phosphatic zone (and local ooidal ironstone) is proposed to mark the base of a re-defined English River Formation.

In western Illinois, the upper green-gray silty shale interval in the Famennian shale succession is assigned to the Saverton Shale, but the contact tween this shale and the underlying organic-rich shale (included in the upper Grassy Creek Shale) represents a gradational facies transition. As such, the base of the upper green-gray shale interval (Saverton) does not represent a stratigraphic datum, and the Saverton-Grassy Creek formational boundary is transitional (Cluff et al., 1981, p.40). A traceable lithostratigraphic datum can be drawn at the contact between the lower and upper parts of the Grassy Creek Shale (see Fig. 1), which also corresponds to a sequence boundary in the region. Upper Grassy Creek shales overstep older Devonian units to directly overlie Middle Devonian strata (Cedar Valley Group) in parts of northeast Missouri (Woodruff, 1990). These relationships further underscore the lithostratigraphic separation of lower and upper Grassy Creek strata in the region.

The Grassy Creek Shale (named after exposures in northeast Missouri) of Missouri and western Illinois is dominated by organic-rich shale facies, but green-gray shales occur within the brown organic shale interval in the region of southeast Iowa (Fig. 1). Burrowed gray to green-gray shales are also known in the middle part of the Grassy Creek interval in west-central Illinois (Cluff et al., 1981, p. 61) and northeast Missouri (Woodruff, 1990). Northwestward in Iowa, equivalent strata of the lower Grassy Creek become dominated by green-gray shale facies (including the type Sheffield Shale of northern Iowa).

The highest part of the Famennian succession in southeast Iowa comprises a third depositional cycle (T-R cycle IIf of Johnson et al., 1985), and it is this succession that is exposed in the Burlington area (Figs. 1, 2). This interval is notably thinner than the underlying Famennian units and includes, in ascending order, 1) a lower phosphatic lag or ooidal ironstone/phosphorite, inter-burrowed with gray shale; 2) a middle silty gray to green-gray shale, burrowed with scattered brachiopods (lingulids, chonetids); and 3) an upper siltstone and argillaceous siltstone, commonly burrowed and with scatttered to abundant shelly fauna (especially upwards). The upper siltstone interval comprises the English River Siltstone, whereas the lower two units are generally assigned to the upper Maple Mill or upper Saverton Shale. Of note, the lower shaly units of this cycle locally interfinger with or are laterally replaced by siltstone beds of the English River Siltstone, and in the type English River area of Washington County, the basal phosphatic lag is bounded by siltstone strata.

The lower bed or interval of this upper cycle is characterized by phosphatic enrichment and ooidal ironstones (Figs. 1, 2), interpreted to be the condensed bed of the sequence. Phosphatic clasts (apatite grains and pebbles), phosphorite "flaxseed" (ooids), and bone phosphate are common to abundant, and a distinctive fish tritor is locally abundant in the phosphatic bed

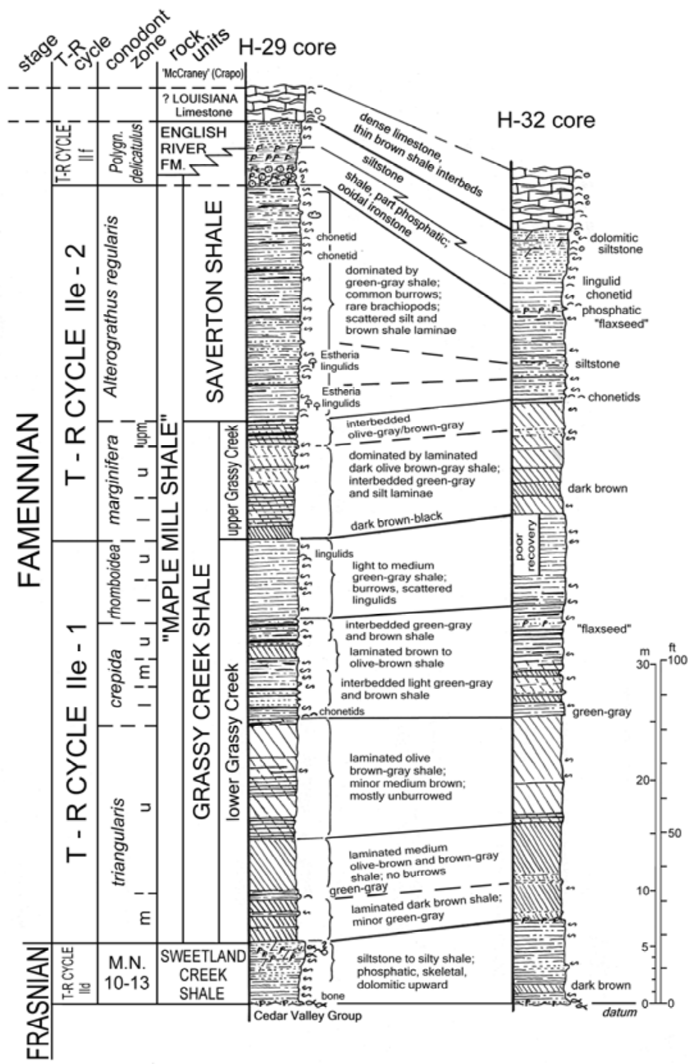

Figure 2. Representative graphic stratigraphic sections of Upper Devonian shale strata from two cores in southeast Iowa (Lee County) and their proposed correlations. See Figure 1 for location of core sections. Conodont zonation derived from H-29 core is adapted from Pavlicek (1986) with modifications by Woodruff (1990). Symbols largely follow the keys in figures 1 and 7. S-shaped squiggles indicate burrow mottling. Cross-ruling denotes intervals of organic shales (mostly brown to olive gray), unburrowed to slightly burrowed; organic-rich (brown shale) horizons are shown by dark horizontal lines. Increased density of cross-ruling qualitatively reflects increased preservation of organic material (darker colors). Ooids below English River siltstones in H-29 core are entirely ooidal ironstones. Transgressive-regressive (T-R) cycles after Johnson et al. (1985) and Woodruff (1990).

in the type English River area ("*Gluteus* bed" of Davis and Semken, 1975). Ooidal ironstones (comprised of ferric oxides and/or ferrous clay minerals with phosphatic to pyritic inter-laminae) are locally well developed in southeast Iowa (e.g., H-29, Figs. 1, 2), and thin ooidal ironstone facies are widespread at this stratigraphic position across much of Iowa, trending into eastern Nebraska and Kansas (basal Boice Shale). A zone of phosphatic nodules ("Falling Run Bed") near the top of the New Albany Group is widely traceable across much of Illinois and Indiana (Cluff et al., 1981, p. 30), and may correlate with this horizon in Iowa. The phosphatic or phosphatic/ironstone bed in the upper Maple Mill-English River interval is an important stratigraphic marker across the region, and could be used to delineate a lithostratigraphic boundary in the shale succession.

Some Lithostratigraphic Recommendations Concerning the Upper Devonian Shale Succession of Southeast Iowa

The Upper Devonian shale interval exposed at Burlington and elsewhere in southeast Iowa (including Washington County) correlates with the shale succession in nearby areas of western Illinois and northeast Missouri. It is appropriate that previously named and correlable lithostratigraphic units in nearby Missouri and Illinois should be considered for possible use in southeast Iowa (in part to promote regional lithostratigraphic harmony). The upper Maple Mill-English River interval is lithostratigraphically indistinguishable from the Saverton Shale, a shale and siltstone unit whose name derives from the town of Saverton in northeast Missouri (Ralls County). The Saverton Shale is also used as a stratigraphic term across western Illinois. The full Upper Devonian shale succession in western Illinois and adjacent Missouri includes, in ascending order, the Sweetland Creek, Grassy Creek, and Saverton shales (Cluff et al., 1981). These same lithostratigraphic subdivisions can be recognized in southeast Iowa, and it may be appropriate to use them in Iowa to promote regional lithostratigraphic harmony with surrounding states. In northern Iowa, the Famennian succession includes a lower Sheffield Shale (dominantly green-gray shale), a middle fossiliferous carbonate (Aplington Fm), and an upper unnamed green-gray shale with a capping siltstone (probably equivalent to the Saverton-English River interval, but commonly termed the "Maple Mill" Shale). In the absence of any Aplington carbonate in southeast Iowa, it is difficult to apply these stratigraphic subdivisions across southeast Iowa.

It is tentatively recommended that the three T-R cycles in the Famennian succession of southeast Iowa be labeled, respectively: 1) lower Grassy Creek Shale; 2) upper Grassy Creek-Saverton Shale; and 3) English River Formation. The Maple Mill Shale could conceivably be retained as a local term for the Saverton Shale equivalent in southeast Iowa, but there seems little to be gained by retaining this term. We propose here that the English River be redefined to include all strata of T-R cycle IIf (Johnson et al., 1985), conveniently marked at its base by a widespread phosphatic or ooidal ironstone bed. This usage would remove the basal shaly strata of this cycle from the upper Saverton Shale and place these beds in the English River interval. Formerly, the term English River was applied only to the siltstone-dominated portion of the depositional cycle (IIf), but the interbedded aspect of siltstone and shale strata and the gradational nature of the shale-to-siltstone transition made the formational boundary a potentially difficult one to consistently define. The newly-proposed lower boundary for the English River Formation, instead, is a widely traceable stratigraphic datum. Further discussion and comments are welcome.

KINDERHOOKIAN (LOWER MISSISSIPPIAN) STRATIGRAPHY IN SOUTHEAST IOWA

The lower series of the Mississippian Subsystem in North America is termed the Kinderhookian, initially named by Meek and Worthen (1861, p. 288) for exposures near the village of Kinderhook, Pike County, western Illinois. The main reference section is generally recognized near Kinderhook, Illinois, but Keyes (1941) considered outcrops at Burlington, Iowa, to be the type section (Lane and Brenckle, 2001, p. 84). Moore (1928) described Kinderhookian sections in the Mississippi Valley area, and he has an interesting subheading for his stratigraphic discussion entitled, "The Type Sections of the Kinderhook" (emphasis on sections, plural, not just one section). He apparently consisered two different localities to comprise the Kinderhookian type section: 1) Kinderhook, Illinois, and 2) Burlington, Iowa. Although modern stratigraphic definitions should be constrained at a single type locality, it is of historical note that the section at Burlington, Iowa, was considered of considerable importance in the formulation and definition of the Kinderhookian Series of the Lower Mississippian.

Strata of the Grassy Creek, Saverton, and Maple Mill shales in Illinois, Missouri, and Iowa were included within the Kinderhookian by early workers, but subsequent studies revealed that these historic "lower Kinderhook" strata are actually of Devonian age. The Kinderhookian was thereby restricted to include only the post-Devonian and pre-Burlington interval in the Mississippi Valley (Collinson, 1961). At Burlington, the Kinderhookian succession includes, in ascending order: 1) the "McCraney" Formation (limestone, dolomite); 2) the Prospect Hill Siltstone; and 3) the Wassonville Formation (lower Starrs Cave limestone, upper dolomite)(see Fig. 2).

The "McCraney," Prospect Hill, and lower Wassonville (Starrs Cave) interval was lumped together by Laudon (1931) to comprise the "North Hill formation." North Hill is the historic name for the prominent bluff that occupies the northern part of the City of Burlington, immediately north of the downtown area (and centered near the North Hill School), and the "North Hill" type section is seen lower in the bluff facing the Mississippi. These strata were subsequently elevated to group status (Workman and Gillette, 1956). Witzke et al. (1990, p. 8) pointed out the lithologic gradation and faunal continuity of Starrs Cave with overlying Wassonville strata, and they suggested that the "North Hill Group has little or no stratigraphic utility in southeast Iowa or elsewhere. . . Its continued use artificially separates the Starrs Cave from the overlying Wassonville, and creates an impression of stratigraphic ordering that obfuscates facies relationships."

"McCraney" (Crapo) Formation

The McCraney Formation derives its name from McCraney Creek a short distance north of Kinderhook, Illinois, and the unit is part of the type Kinderhookian succession. This sparsely fossiliferous limestone and dolomite unit was originally defined by Moore (1928) as the "McKerney," but the term has been replaced subsequently by the current spelling of the creek's name, McCraney. From its inception, Moore (1928) correlated the "McKerney" at Kinderhook, Illinois, with the carbonate interval above the English River Siltstone at Burlington, and this correlation has been promulgated by subsequent workers (e.g., Witzke et al., 1990; Witzke and Bunker, 2001). However, we have recently questioned the presumed stratigraphic equivalency of the McCraney interval at Kinderhook with the basal Mississippian carbonate unit at Burlington. This is because the type McCraney section at Kinderhook lies at the *top* of the Kinderhookian succession, *above* Kinderhookian siltstones of the Hannibal Formation (Collinson et al., 1979, p. 26), whereas the so-called "McCraney" at Burlington lies at the *base* of the Kinderhookian succession, *below* Kinderhookian siltstones of the Prospect Hill Formation. While biostratigraphic evidence is presently insufficient to demonstrate whether or not the carbonate units at Kinderhook and Burlington are correlative, the significant difference in the relative stratigraphic positions of these carbonate units at Burlington and Kinderhook should raise a cautionary flag. Because we believe stratigraphic equivalency of the so-called "McCraney" at Burlington and the type McCraney at Kinderhook has not been established, we feel it prudent to query this correlation by informally terming the interval at Burlington as the "McCraney" (in quotes). Should the correlation of these two units prove to be incorrect, we provisionally propose to rename the basal Kinderhookian unit at Burlington the Crapo Formation (type locality in Crapo Park, Burlington; Stop 1). Rocks of "McCraney-like lithology" also occur in parts of northeast Missouri in the upper part of the Kinderhookian succession, where this interval has also been informally termed the "McCraney" (Thompson, 1986, p. 63). These strata occupy a similar stratigraphic position to the type McCraney in Illinois, but correlation is uncertain.

The "McCraney" (Crapo) interval in southeast Iowa ranges in thickness from 0 to 65 ft (20 m), but commonly is about 10 to 15 ft (3-4 m) thick. Person (1976, p. 21) succinctly summarized the general lithologic character of the "McCraney" in the Burlington area, as being "composed of alternating beds of sparsely fossiliferous, sublithographic limestone, and dark brown, coarser-grained unfossiliferous dolomite. A thin chonetid brachiopod-rich and oolitic layer is present at the base." The alternation of light and dark lithologies imparts a “strikingly banded appearance” on outcrop that is of diagenetic origin (Glenister et al., 1987), and the wavy- to nodular-bedded aspect of these strata is distinctive. Some of the limestone beds display calcite and dolomite-filled fractures, probably originating as syneresis cracks in the nodular bedforms. Fossils are generally sparse, except in the basal layer, although the known fauna includes brachiopods, bivalves, gastropods, ostacodes, corals, echinoderm fragments, foraminifers, and conodonts. Weller (1900), Moore (1928), and VanTuyl (1923) recorded a moderately diverse assemblage of brachiopods from this interval at Burlington that included *Paryphorhynchus striatocostatum*, *Rhynchopora pustulosa*, *Rugosochonetes gregarius*, *Allorhynchus heteropsis*, and a number of additional forms (totaling 15 to 20 species). Many, but not all, of these taxa are found in the Hannibal Formation of northeast Missouri.

The exact stratigraphic relationships of the “McCraney” (Crapo) interval in southeast Iowa have not been fully resolved. It disconformably overlies Devonian strata of the English River Formation in the Burlington area, and a basal oolite is commonly recognized (Fig. 2). A short distance southward into Lee Co., Iowa, however, the basal relationships are complicated by the presence of an interval of variably fossiliferous wavy-bedded lime mudstone (interbedded with thin brown shaly partings) above the English River and below the more typical sparsely fossiliferous mixed limestone-dolomite interval of the “McCraney” (Figs. 1, 2). The basal oolite is absent in this area. In one core (H-29; Fig. 1), the lower fossiliferous limestone is capped by a unit of intraclastic to sandy lime mudstone (‘sublithographic’), unlike any lithology seen northward or southward. Although biostratigraphic analysis of this lower fossiliferous interval has not yet been undertaken, it is lithologically indistinguishable from strata of the Louisiana Limestone exposed in northeast Missouri. In addition, chemostratigraphic evidence does not support a Mississippian correlation, as the interval in Iowa captures a major carbon-isotope excursion (M. Saltzman, 2001, personal communication) not known from Kinderhookian sections elsewhere in North America. This interval is tentatively labeled “?Louisiana Limestone” in Lee Co., Iowa (Figs. 1, 2).

Historical confusion between the Louisiana Limestone (named after exposures near the town of Louisiana in northeast Missouri) and strata referred to the McCraney Formation has been understandable as both units have similar lithologies and occupy similar stratigraphic positions. However, the lower part of the Louisiana Limestone contains a Devonian conodont fauna (Scott and Collinson, 1961; assigned to the highest Famennian by Klapper et al., 1971), whereas the type McCraney and Iowa “McCraney” contain Mississippian (Kinderhookian) conodont faunas (Scott and Collinson, 1961; Straka, 1968). This lower limestone interval in Lee County, Iowa, may represent the northernmost occurrence of the Devonian Louisiana Limestone, but further study is needed. This limestone is paraconformably overlain by a mixed dolomite/limestone interval that apparently correlates with the “McCraney” (Crapo) interval at Burlington.

The McCraney Limestone in its type area shares lateral and subjacent facies relationships with the Hannibal Formation (see Collinson et al., 1979; Moore, 1928). The Hannibal is a widespread Kinderhookian shale and siltstone unit that comprises the bulk of the Kinderhookian succession in eastern Missouri and western Illinois. The “McCraney” (Crapo) interval of southeast Iowa also apparently shares lateral facies relationships with shale and siltstone strata of the lower Prospect Hill Formation (which is a northern equivalent of the middle to upper Hannibal Fm in Iowa; see also Fig. 1, queried boundaries).

Locally in northern Missouri and western Illinois, the base of the Kinderhookian succession below the Hannibal Formation is marked by a Kinderhookian carbonate interval known as the Horton Creek Formation (which includes wavy-bedded silty limestone, oolitic to skeletal limestone, dolomite, and siltstone/shale). The Horton Creek and the Iowa “McCraney” occupy the same relative stratigraphic position within the Kinderhookian succession (above Famennian strata, at the base of the Kinderhookian, and below Kinderhookian siltstone/shale). It is, therefore, tempting to suggest that these units may share some sort of lithofacies relationship within the Kinderhookian succession. Carter (1988, p. 27), in discussing the common Iowa “McCraney” brachiopod *Rugosochonetes gregarius* (which also occurs in the Horton Creek), suggested that “it is possible that ‘Bed 3’ [i.e., lower ‘McCraney’] is a northern extension of the [Horton Creek] . . . in any case the stratigraphic position and lithologic similarity of the two widely separated units invites comparison of several faunal elements.” Such a suggestion certainly has compelling aspects. The only potential drawback is that conodont faunas from the lower Horton Creek (*Siphonodella sulcata*) have been considered to be older than faunas from the lower "McCraney" (*S. duplicata, S. cooperi*). If true, the base of these two carbonate units may not correlate. and a northward onlap of Kinderhookian strata in the Mississippi Valley may be indicated. However, the stratigraphic ranges of *S. sulcata*, *S. duplicata*, and *S. cooperi* are known to overlap (Sandberg et al., 1978), and biostratigraphic separation between these two units is not necessarily demonstrated. The Horton Creek and "McCraney" are interpreted to form a potentially diachronous and widespread basal Kinderhookian carbonate lithofacies package extending from west-central Illinois into southeast Iowa. The underlying Louisiana-Glen Park carbonates of the upper Famennian share many lithologic and faunal similarities with the Kinderhookian Horton Creek-"McCraney" interval, and similar depositional settings are inferred for both.

Prospect Hill Formation

A distinctive siltstone interval above the "McCraney" at Burlington was named the Prospect Hill Siltstone by Moore (1928), who regarded this unit as a member of the Hannibal Formation. As noted previously, the Hannibal Formation is a siltstone and shale interval named after Hannibal in northeast Missouri. Workman and Gillette (1956) separated the Prospect Hill from the Hannibal and elevated it to formational status. The name Prospect Hill derives from the major bluff that occupies the southeastern part of the City of Burlington, immediately south of the downtown area and north of Crapo Park. The exposures of the Prospect Hill Formation seen on this field trip represent typical sections in the historic type area of the formation. The formation extends northwestwards across much of Iowa, where it generally forms the basal unit of the Kinderhookian ("McCraney" absent across most of Iowa).

The Prospect Hill Formation is dominated by siltstone in the Burlington area, slightly argillaceous with scattered shaly partings. The siltstones locally display horizontal laminations and low-angle cross stratification to hummocky bedforms. Vertical to horizontal burrow fabrics are locally prominent. Fossil molds are variably common to absent in individual sections, but some beds locally contain abundant fossil molds. The shelly fauna is very similar to that seen in the older English River siltstones, and the Prospect Hill fauna is generally dominated by bivalves and brachiopods (especially chonetids). Gastropods, cephalopods, scaphopods, bryozoans, and crinoid debris are also noted.

As displayed in the Burlington area, the Prospect Hill Formation is a relatively thin interval only about 4 to 8 feet (1.2-2.4 m) thick. The formation overlies a slightly eroded surface on the McCraney, locally displaying up to 16 inches (40 cm) of relief (as seen at Starrs Cave Preserve). The top of the Prospect Hill Formation also locally shows some minor erosional relief (to 2 inches) in the Burlington area, and it is probable that the formation is bounded above and below by unconformity surfaces. The formation locally displays concentrations of fish bone (bone bed) at its base (as at the Mediapolis Quarry, Des Moines County), and its thickness varies dramatically across southeast Iowa (locally reaching thicknesses to 90 feet; 27 m). Where the formation is thick, it generally includes significant shale and shaly siltstone facies.

The Prospect Hill Formation displays interesting but poorly-understood stratigraphic relationships across southeast Iowa (Fig. 1). A relatively thin Prospect Hill interval (< 8 ft thick; 2.4 m) is noted across Des Moines County, but the formation is completely absent at some localities (U.S. Gypsum core, Fig. 1; see also Stony Hollow, Stop 3) where the Starrs Cave directly overlies the "McCraney." By contrast, southward into Lee County, Iowa, the Prospect Hill dratically thickens and contains two or more laterally persistent shale units (Fig. 1). Where the formation is thickest, the lower siltstone/shale beds overlie the ?Louisiana Limestone and are interpreted to share lateral stratigraphic relationships with the "McCraney" carbonates. Southward stratigraphic relationships with the Hannibal Formation in Missouri are not yet delineated. However, similar siltstone-shale lithofacies in the Hannibal and Prospect Hill formations support the idea that the Propect Hill is a northern equivalent of some part of the Hannibal, primarily the middle and upper Hannibal. The northward thinning of the Prospect Hill Formation in southeast Iowa (as seen in the Burlington area) reflects, in part, erosional beveling of upper strata prior to the deposition of the Starrs Cave Member. The complete absence of the Prospect Hill Fm across portions of eastern and central Des Moines County (U.S. Gypsum, Fig. 1; Stony Hollow, Stop 3) underscores the significance of pre-Starrs Cave erosion. Such erosional beveling may have been structurally influenced, with maximum truncation across the crests of local anticlines (see Fig. 1). Northward thinning also may have been accentuated by northward onlap of upper Prospect Hill siltstone facies above the "McCraney," but further study is needed.

Wassonville Formation

An interval of carbonate rock, including fossiliferous to oolitic limestone and dolomite, cherty in part, overlies the Prospect Hill Siltstone at Burlington. The basal part of this interval was named the Starrs Cave Formation by Workman and Gillette (1956); the type locality is located at Starrs Cave Preserve along Flint Creek a short distance north of Burlington (Stop 2). The Starrs Cave is a relatively thin limestone unit that is characteristically a fossiliferous oolitic packstone to grainstone, although the interval is a sparsely oolitic to non-oolitic skeletal packstone to grainstone at some localities in southeast Iowa. This interval is generally only 1.5 to 5 feet (0.5-1.5 m) in thickness in the Burlington area, but it thickens westward (to 15 ft; 4.5 m) across southeast Iowa (Witzke et al., 1990). Limestones of the Starrs Cave are locally absent at some localities (especially in Washington County) where Wassonville dolomites directly overlie the Prospect Hill Siltstone. However, dolomitized oolitic strata at the base of the Wassonville in that area indicate that Starrs Cave equivalents are actually present (Straka, 1968).

The contact between the skeletal to oolitic Starrs Cave limestone and overlying dolomite strata is gradational and interbedded. The contained benthic faunas (especially the brachiopods) are identical in the Starrs Cave limestones and the overlying dolomite beds. The upper contact of the Starrs Cave is arbitrarily selected at the base of the lowest dolomite in the succession. As suggested by Witzke et al. (1990, p. 11), the gradational character of Starrs Cave and overlyling Wassonville dolomite strata indicates that the two units should be naturally grouped (and the contact not used to mark the top of a formation or the bounding top of the "North Hill Group"). They wrote: "Although the thin Starrs Cave interval has been accorded formational status and separated from overlying Wassonville strata in most previous reports, it may be desirable at some point to re-assign the Starrs Cave as a member of the Wassonville Formation." Since no serious objection has been expressed over this suggestion, Witzke and Bunker (2001, p. 15) formally proposed that the Starrs Cave be considered the basal member of the Wassonville Formation, and not a separate formation by itself.

The Wassonville Formation is a dolomite-dominated unit, but interbedded limestone and dolomitic limestone lithologies are present (especially in the lower part), and the basal portion is limestone (the Starrs Cave Member) at most localities. Wassonville strata locally display silicification fabrics and nodular chert bands, but these are irregular in their distribution. The dolomite beds commonly display obscure or faint irregular laminations. Thin interbedded fossiliferous limestones are seen as stringers or starved bedforms, commonly with abundant brachiopods (especially chonetids). The dolomite-dominated portion of the Wassonville Formation above the basal Starrs Cave Member represents the upper member of the formation. These upper strata have not been formally named as yet, and they are here informally termed the "upper member." A number of representative sections of this interval can be found in southeast Iowa, and good candidates for the type section of the "upper member" include Crapo Park, Mediapolis Quarry and West Chester Quarry (see Witzke et al., 1990, p. 13).

The Wassonville Formation contains a diverse and abundant fossils fauna (see Weller, 1900; VanTuyl, 1923; Moore, 1928; Laudon, 1931; Witzke et al., 1990, p. 14), especially in the lower skeletal packstone beds and lenses. These contain diverse assemblages of brachiopods and other fossils (crinoids, bryozoans, corals, trilobites, etc.) that closely resemble certain faunas from the Missouri Chouteau Fm and the Maynes Creek Fm of central Iowa. Upper dolomitic strata are less fossiliferous, but lenses and thin skeletal-rich beds contain a variety of fossils, most commonly chonetid brachiopods (*Rugosochonetes multicostus*).

THE SUB-BURLINGTON DISCONFORMITY

The upper contact of the Wassonville Formation with the overlying Burlington Formation in southeast Iowa is sharp, although the nature of this contact is a difficult one to understand in a regional context. The contact marks the boundary between the Kinderhookian and Osagean series in the area. Minor relief (to 4 inches) on this contact is locally seen in the Burlington area, although at many localities it appears more-or-less planar with no obvious erosional relief or evidence of subearial exposure. At some localities the contact is marked by a prominent hardground surface (probably of submarine origin) or thin interval of multiple stacked hardground surfaces. Even though this contact does not appear to be deeply eroded, regional truncation of Wassonville strata is evident on regional scale. Progressive eastward and southward thinning of the Wassonville is seen across southeast Iowa which appears to bevel the upper member (see Witzke et al., 1990, p. 13). The formation thins from 60 to 35 feet (18-10 m) eastward across Keokuk and Washington counties. In the Burlington area, the Wassonville thins from 19 feet (6 m) at the Mediapolis Quarry (north of Burlington) to only 7.4 feet (2.3 m) at Crapo Park. Southward across Lee Co., Iowa, and into adjoining areas of west-central Illinois, the Wassonville Formation shows further thinning. In western Illinois, the entire Wassonville is erosionially truncated locally, and the Burlington Formation directly overlies the Prospect Hill Siltstone or lower strata in some areas (Workman and Gillette, 1956).

Although not a focus for this field trip, it is of particular note that farther northwestward, strata of the upper Maynes Creek and basal Gilmore City formations in central and northern Iowa (s. 3, 4) have no equivalent strata in southeast Iowa. These strata thin southeastward and vanish along the sub-Burlington disconformity surface in southeast Iowa. The magnitude of the sub-Burlington disconformity surface in southeast Iowa is considerable, and strata spanning parts of the upper Kinderhookian and lower Osagean are not present at Burlington (Figs. 3, 4). Locally in western Illinois, lower Kinderhookian strata are also truncated beneath this surface (Workman and Gillette, 1956).

What is the origin of this southeastward beveling of the Wassonville Formation and other Kinderhookian strata beneath the Burlington Formation? Is it simply subaerial erosional truncation along a major unconformity surface? If this is the case, the direction of beveling seems anomalous. The general shoreward direction during Mississippian deposition was to the northwest in Iowa (toward the Transcontinental Arch). This is clearly reflected by increasing depositional shallowing to the northwest (especially oolite shoals and restricted back-shoal mudstone facies), along with the increasing presence of peritidal and mudflat facies (fenestral mudstones, mudcracked exposure surfaces, stromatolites) in that direction. By contrast, no unequivocal evidence of subaerial exposure has been identified in the Kinderhookian-Osagean succession of southeast Iowa. Therefore, shoreward areas during the Mississippian lie to the northwest, not the southeast. But how can this be? Shouldn't the erosional beveling of Kinderhookian strata expand in a shoreward direction, not in an offshore direction? The southeastward beveling and truncation of Wassonville and other Kinderhookian units seems strangely perplexing.

The southeastward expanding hiatus that separates Kinderhookian strata from the overlying Burlington Formation in Iowa may conceivably be explained by one of two possible explanations. First, some structural upwarping across the shallow shelf may have temporarily disrupted the general deepening-and-shallowing trends across the seaway and reversed the direction of erosional beveling. Although we cannot categorically dismiss this suggestion, the complete absence of peritidal deposition in southeast Iowa (and the common presence of such facies in northern Iowa) seriously undermines this idea. Secondly and alternatively, the southeastward erosional beveling of Kinderhookian strata may have resulted from lower rates of sediment accumulation and increased erosional beveling in an offshore direction. This suggestion initially seems counter-intuitive. Although many details of regional Mississippian sedimentation need to be worked out, Witzke and Bunker (1996) proposed that the sub-Burlington discontinuity may actually be a broad submarine surface marked by widespread sediment starvation in offshore areas (of the "middle shelf"). The beveling of sub-Burlington strata may, therefore, represent submarine erosional planation (perhaps related to recurring storm current activity that

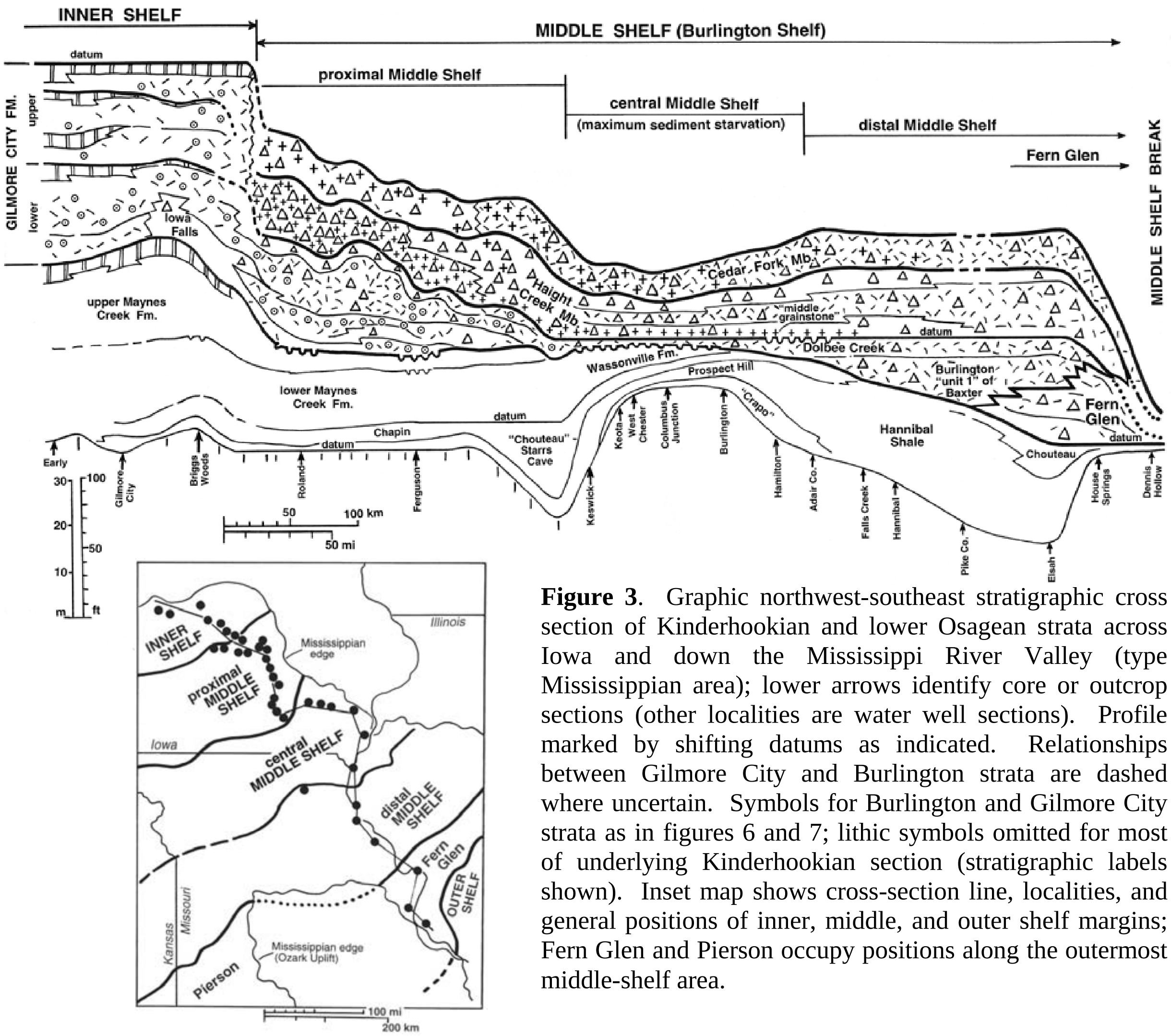

Figure 3. Graphic northwest-southeast stratigraphic cross section of Kinderhookian and lower Osagean strata across Iowa and down the Mississippi River Valley (type Mississippian area); lower arrows identify core or outcrop sections (other localities are water well sections). Profile marked by shifting datums as indicated. Relationships between Gilmore City and Burlington strata are dashed where uncertain. Symbols for Burlington and Gilmore City strata as in figures 6 and 7; lithic symbols omitted for most of underlying Kinderhookian section (stratigraphic labels shown). Inset map shows cross-section line, localities, and general positions of inner, middle, and outer shelf margins; Fern Glen and Pierson occupy positions along the outermost middle-shelf area.

episodically eroded and transported material from broad areas of the "middle shelf"). As we examine the Wassonville-Burlington contact on this field trip, it would be well to ponder the regional relationships and ramifications of this surface. A broader understanding of this disconformity surface may have far-reaching effects on our comprehension of the nature of the stratigraphic record in cratonic areas.

THE BURLINGTON FORMATION (LOWER OSAGEAN, MISSISSIPPIAN)

Burlington Formation in the Burlington Area

Owen (1852) described the "encrinital group of Burlington," and the term "Burlington limestone" was introduced by Hall (1857) and Hall and Whitney (1858) for the succession of crinoidal limestones exposed in the Mississippi River bluffs at Burlington, Iowa. These exposures at Burlington have long been famous for their rich paleontologic resources, especially the fantastic crinoid faunas (Wachsmuth and Springer, 1897; see contribution by Gahn in this guidebook) and to a lesser extent the brachiopods (Weller, 1914). We will examine some of the classic exposures in the type area of the formation for this field trip. The Burlington is presently used as a formational term over a broad area of the Midcontinent, from Iowa to Arkansas, and from Illinois to Kansas. The Burlington Formation is beautifully exposed not only in the Burlington area, but prominent exposures are also well displayed for long stretches along the Mississippi River bluffs farther downstream in Illinois and Missouri. Because of its historic significance, as well as its distinctive fossiliferous lithologies, the Burlington Formation can be considered the quintessential and characteristic rock unit in the Mississippian type area. The Burlington Formation forms the lower part of the Augusta Group (named after the town of Augusta west of Burlington), a term resurrected by Witzke et al. (1990). The Burlington Formation comprises the lower portion the Osagean Series in the Mississippi Valley area.

Although crinoidal ("encrinital") limestones (packstones and grainstones) are an important and distinctive part of the Burlington Formation across its extent, thick intervals of dolomite subdivide the succession into several distinct stratigraphic units. Harris and Parker (1964) subdivided the Burlington Formation in southeast Iowa into three members, in ascending order: the Dolbee Creek Member (dominated by crinoidal limestones), the Haight Creek Member (dominated by cherty dolomite strata), and the Cedar Fork Member (dominated by cherty crinoidal limestone). These members are best characterized in the Burlington, Iowa, area (Fig. 5) and the type localities for all three members are designated in Des Moines County, Iowa. The Burlington Formation averages about 65 feet in thickness in the Burlington area (full thickness where capped by the overlying Keokuk Formation). However, the Mississippi River bluff edge in the City of Burlington generally occupies a position no higher than the Cedar Fork Member, and the uppermost part of the formation is erosionally missing. The Burlington Formation varies between about 55 and 80 feet (16-24 m) thick across southeast Iowa, and it is thinnest at Columbus Junction (Fig. 3).

The **Dolbee Creek Member** varies between about 6 and 13 feet in thickness in the Burlington area, where it is characterized by a stacked succession of crinoidal packstones and grainstones. The crinoidal

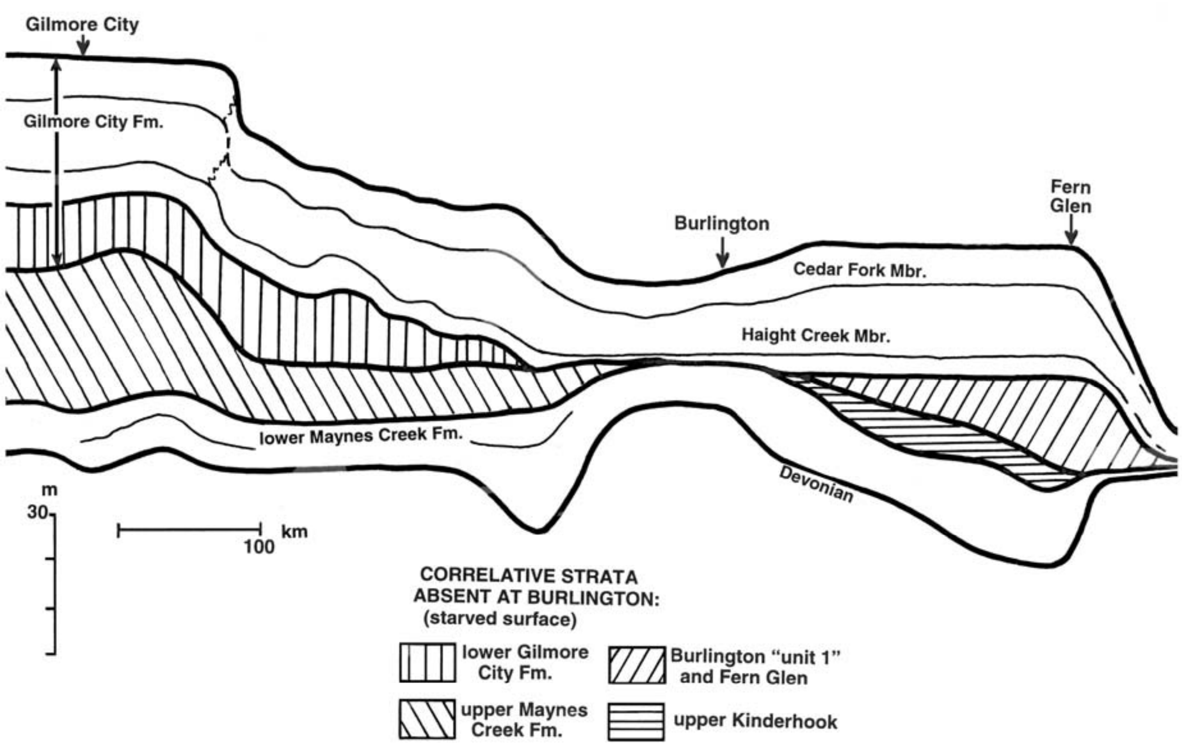

Figure 4. Graphic cross-section of Kinderhookian and lower Osagean strata, Iowa and Mississippi Valley (this is the same profile as Figure 3). Upper Kinderhookian and lower Osagean units that are absent beneath the Burlington Formation at Burlington are highlighted, including the upper Maynes Creek and lower Gilmore City formations of northern Iowa and uppermost Kinderhookian and basal Burlington-Fern Glen strata in Illinois-Missouri. This diagram emphasizes the considerable magnitude of the sub-Burlington disconformity at Burlington and across the central middle-shelf area, a region of maximum sediment starvation. Upper arrows mark the locations of the type sections of the Gilmore City, Burlington, and Fern Glen formations.

limestone beds may be graded (coarsest crinoid grains at the base), and many individual beds display lenses, stringers, and starved megaripple bedforms of fine to coarse crinoid debris. Such crinoidal beds commonly amalgamate into thicker intervals of crinoidal grainstone. The crinoidal beds are interbedded in a complex manner with less fossiliferous mudstone and wackestone lithologies, usually seen as dolomite and dolomitic limestone interbeds (Frig. 5). Some of the dolomite beds are cherty, and large nodules and laterally-extensive chert beds locally occur. The dolomite interbeds are replaced by crinoidal limestones over short lateral distances, and dolomite beds are entirely absent in some sections (as at the Mediapolis Quarry). Unlike the underlying Wassonville Formation, the Dolbee Creek Member generally thins westward into Washington County but thickens southward in the Mississippi Valley. Near St. Louis, Missouri, the lower Burlington crinoidal limestones are further underlain by an additional basal Osagean formation (Fern Glen Fm.) not seen at Burlington (Fig. 3).

The Dolbee Creek Member is overlain by the dolomite-dominated **Haight Creek Member** (30-33 feet thick in the Burlington area). These dolomite strata are cherty to very cherty in part, but in the Burlington area these strata are commonly poorly exposed along the bluff slopes. The dolomites are generally sparingly fossiliferous, although molds of crinoid debris, brachiopods, and other fossils are seen. Many of the dolomites display faint laminations, possibly relict hummocky stratification. Prominent large chert nodules and bedded cherts (up to 30 cm thick) are common in the interval, and these whitish cherts were widely used by aboriginal peoples for thousands

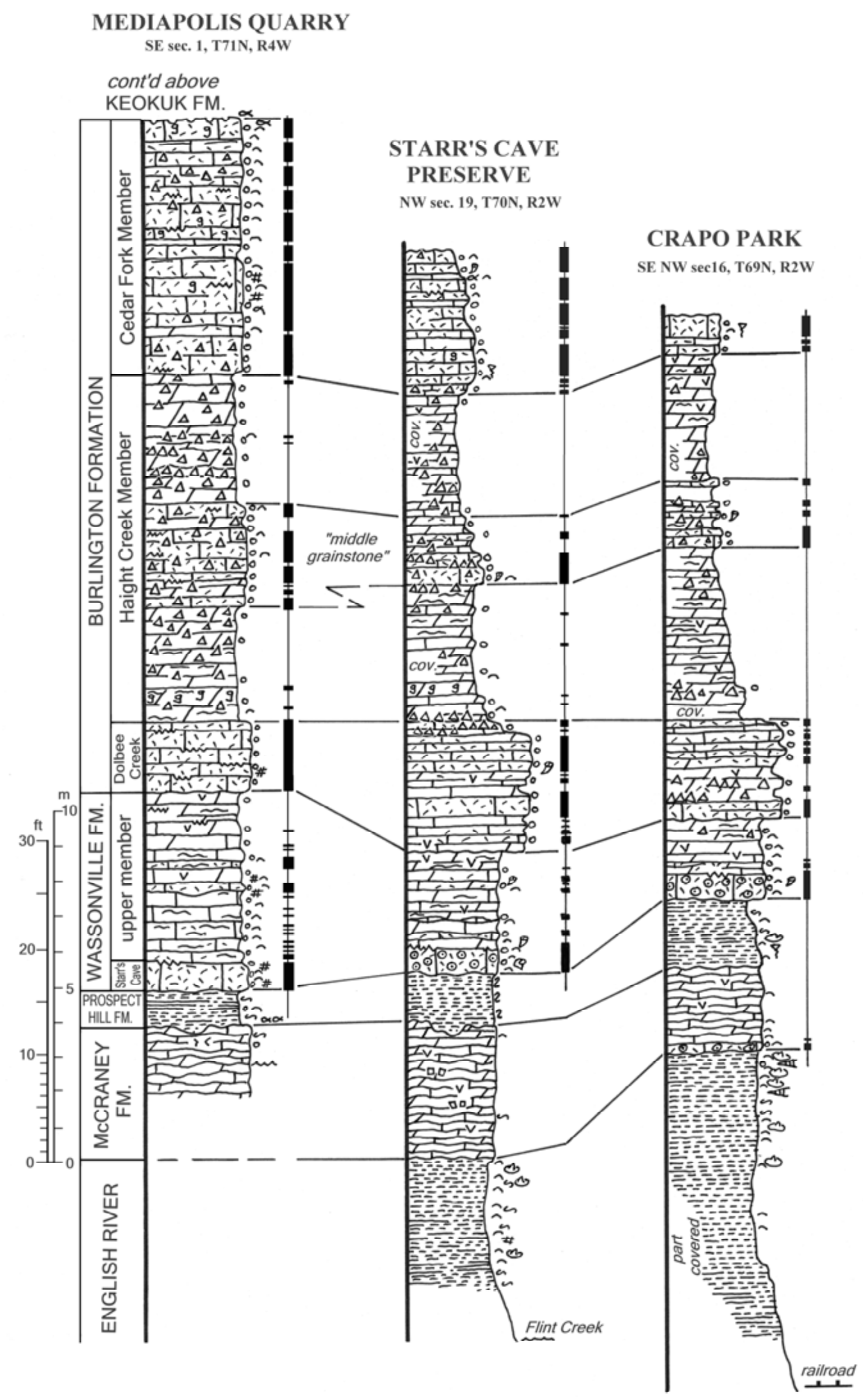

Figure 5. Correlation of Burlington Formation and sub-Burlington strata at three localities in Des Moines Co., Iowa (including Stops 1 and 2). Datum is base of Haight Creek Member. Black bars denote positions of packstone-grainstone intervals. Symbols as in figures 1, 6, 7; g = glauconitic, o = crinoidal, # = bryozoans.

of years because of their exceptional quality for flint knapping. A distinctive glauconitic unit (a greenish-colored bed) occurs at or near the base of the Haight Creek Member throughout its extent in Iowa. This glauconitic bed forms a major stratigraphic marker within the Burlington Formation (as first recognized by Harris and Parker, 1964). A shaly unit is locally recognized above the glauconitic bed in parts of southeast Iowa and adjacent Illinois. An interval of more resistant limestone strata (crinoidal packstones and grainstones similar to those seen in the Dolbee Creek and Cedar Fork members) is found in the middle part of the Haight Creek Member in the Burlington area and throughout most of southeast Iowa (Fig. 5). This crinoidal limestone interval has been termed the "middle grainstone" unit by Witzke et al. (1990).

The **Cedar Fork Member** comprises the upper part of the Burlington Formation, and these strata are dominated by crinoidal limestones (packstones and grainstones) similar in many respects to those seen in the Dolbee Creek and "middle grainstone" intervals. However, chert nodules are generally more common in the Cedar Fork Member than seen in the Dolbee Creek, and some of the crinoidal limestones of the Cedar Fork are glauconitic to varying degrees (the glauconite is seen as small green pellets <1 mm in size). In a few beds, the glauconite can be so abundant that the limestone displays a prominent green color when freshly broken. Hardground surfaces are locally seen beneath one or more glauconitic beds in the member. Minor dolomite and dolomitic limestone interbeds are locally present within the member, but these are laterally discontinuous at the scale of individual quarries. Although brachiopods occur in varying abundance within all crinoidal limestone units of the Burlington Formation, some of the Cedar Fork limestones display prominent large brachiopods (commonly silicified), especially the very large spiriferid known as *Spirifer grimesi*. Any of the crinoidal limestones of the Burlington Formation can potentially produce articulated crinoid cups, and crinoid cups are locally abundant in some beds of the Cedar Fork Member. Concentrations of fish bone (especially bradyodontid shark teeth) can be found on some bedding surfaces within the Cedar Fork Member, and bones are particularly abundant and prominent at the top of the member (the widespread Burlington-Keokuk [B-K] bone bed). Pioneering paleontologists Wachsmuth and Springer (1878) referred to the B-K bone bed as "one of the best stratigraphic landmarks that we know in this formation, and it is found over a wide area in localities over a hundred miles apart and always in the same position."

The Burlington Formation has provided an abundance of fossils in the Burlington area and throughout its outcrop, and many distinguished paleontologists have studied these fossils. Fossils of stalked echinoderms, especially their disarticulated plates and columnals, are the most significant and abundant fossils in the formation, and remarkably diverse assemblages of crinoids and blastoids (known from their articulated cups) are identified (see Gahn, this field guide). In addition to the crinoids and blastoids, brachiopod assemblages are also diverse and noteworthy (Weller, 1914), but brachiopod abundances pale in comparison to the crinoid debris. Additional fossils identified in the Burlington Formation include bryozoans, corals, gastropods, bivalves, trilobites, conularids, sponges, foraminifera, conodonts, and fish (bone and teeth) (Van Tuyl, 1923; Laudon, 1929; Witzke et al., 1990). Fossils are abundant in the skeletal packstones and grainstones, but the intervening dolomitized mudstones (like those of the Haight Creek Mbr) are typically only sparsely fossiliferous. The mudstones most commonly contain small molds of crinoid debris and/or sponge spicules, but other fossils are sometimes noted.

Regional Relationships and Deposition of the Burlington Formation

The Burlington Formation is thinnest in southeast Iowa, and the formation thickens southward (along the Mississippi Valley in Missouri and Illinois) and northwestward into central Iowa (see cross-section, Fig. 3). The magnitude of the sub-Burlington disconformity is also greatest in southeast Iowa (Fig. 4). The Burlington Formation accumulated during the early Osagean across a vast subtidal epicontinental shelf (commonly termed the "Burlington shelf") that stretched from Illinois and Iowa into central Kansas and Oklahoma (Lane, 1978; Witzke et al., 1990, p. 55). Iowa occupied a geographic position in the southern tropics at that time. Correlative limestone strata of the Lake Valley Formation cojoin with Burlington strata across the Texas Panhandle and extend into New Mexico ("Lake Valley shelf") and northward into western Arizona ("Redwall shelf"). This elongate belt from Illinois to Arizona contains similar carbonate lithofacies (especially crinoidal packstone-grainstones) and fossil faunas (crinoids, brachiopods) across its vast extent, and peritidal/mudflat carbonates are characteristically absent. This facies association characterizes the "middle shelf" facies of the Mississippian interior sea of North America (Witzke and Bunker, 1996). Middle-shelf facies terminate southward at a shelf break (Fig. 3), and correlative deeper-water sedimentation across the "outer shelf" (Tennessee, southern Arkansas-Oklahoma, central Texas) was marked by a belt of thin condensed phosphatic shales and carbonates that bordered the submerged continental margin (where strata correlative with the Burlington Fm are only a few inches to a few feet thick). By contrast, a more shoreward belt ("inner shelf") of carbonate strata paralleled the ancient shoreline which bordered the Transcontinental Arch (from northern Iowa across Kansas and Colorado into central Arizona). Correlative lower Osagean strata across this "inner shelf" include a variety of shallow-water marine and restricted-marine facies, oolite shoals, and peritidal/mudflat facies. The Gilmore City Formation of northern Iowa displays characteristic inner-shelf facies for the lower Osagean.

Constraining northwestward stratigraphic relationships of the Burlington Formation remains one of the most vexing problems in Iowa stratigraphy. Witzke and Bunker (2001, p. 79) "tentatively suggested that the Gilmore City and Burlington formations are partially correlative units representing inner-shelf and middle-shelf facies tracts, respectively, during the early Osagean" (bounding relationships queried on . 5). Lower Gilmore City oolitic and skeletal strata extend southeastward into proximal or transitional areas of the middle shelf (Figs. 3, 6), and these interbed with Burlington-style coarse crinoidal grainstones and cherty dolomite. Therefore, part of the lower Gilmore City interval shares facies relationships with the Dolbee Creek Member. However, the basal Gilmore City is older than any strata seen at Burlington (sub-Burlington disconformity), so it is likely that only the upper part of this interval shares relationships with the Burlington. These lower Gilmore City strata across the proximal middle shelf are overlain by middle and upper Burlington units of the Haight Creek and Cedar Fork members. The Haight Creek Member in this area contains characteristic cherty dolomite lithologies, but the interval is significantly more glauconitic than seen southeastward; much of the member is green and glauconitic, resembling the lower Haight Creek glauconitic unit of southeast Iowa. The Cedar Fork Member is dominated by crinoidal packstone fabrics (partly cherty and glauconitic), although generally the member in central Iowa is pervasively dolomitized.

The proximal facies of the Haight Creek and Cedar Fork members abruptly terminate at the inner-shelf margin of the upper Gilmore City Formation (Figs. 3, 6), and intertonguing of Burlington and Gilmore City strata is not clearly evident at this position. Nevertheless, coarse crinoidal lithologies, some resembling those seen in the Cedar Fork, are locally evident in the upper half of the upper Gilmore City suggesting possible lateral relationships. Upper Gilmore City peritidal facies are overlain by dolomite, shale, and limestone strata tentatively assigned to the Keokuk Formation (Witzke, 2002). If this correlation is correct, the same relative position of the Gilmore City and Burlington formations beneath the Keokuk Fm further supports their stratigraphic equivalency.

Across the middle-shelf area of Iowa, lower Gilmore City strata and the Dolbee Creek Member of the Burlington overlie the same disconformity surface above the Maynes Creek and Wassonville formations (Fig. 3). Locally this surface is recognized to be a submarine hardground surface or amalgamated stack of multiple hardgrounds (Witzke et al., 1990), and the Dolbee Creek Member is greatly thinned above this surface in southeast Iowa. Southward from southeast Iowa, the Burlington thickens along the Mississippi Valley in Illinois and Missouri (central middle shelf to distal middle shelf; Fig. 3). Of particular note, the lower Burlington expands considerably in thickness in that direction, incorporating strata not seen at Burlington (Fig. 3; "unit 1" of Baxter and Haines, 1990). The entire Burlington Formation is replaced southward by a greatly thinned Fern Glen Formation as the middle-shelf break is approached (Fig. 3).

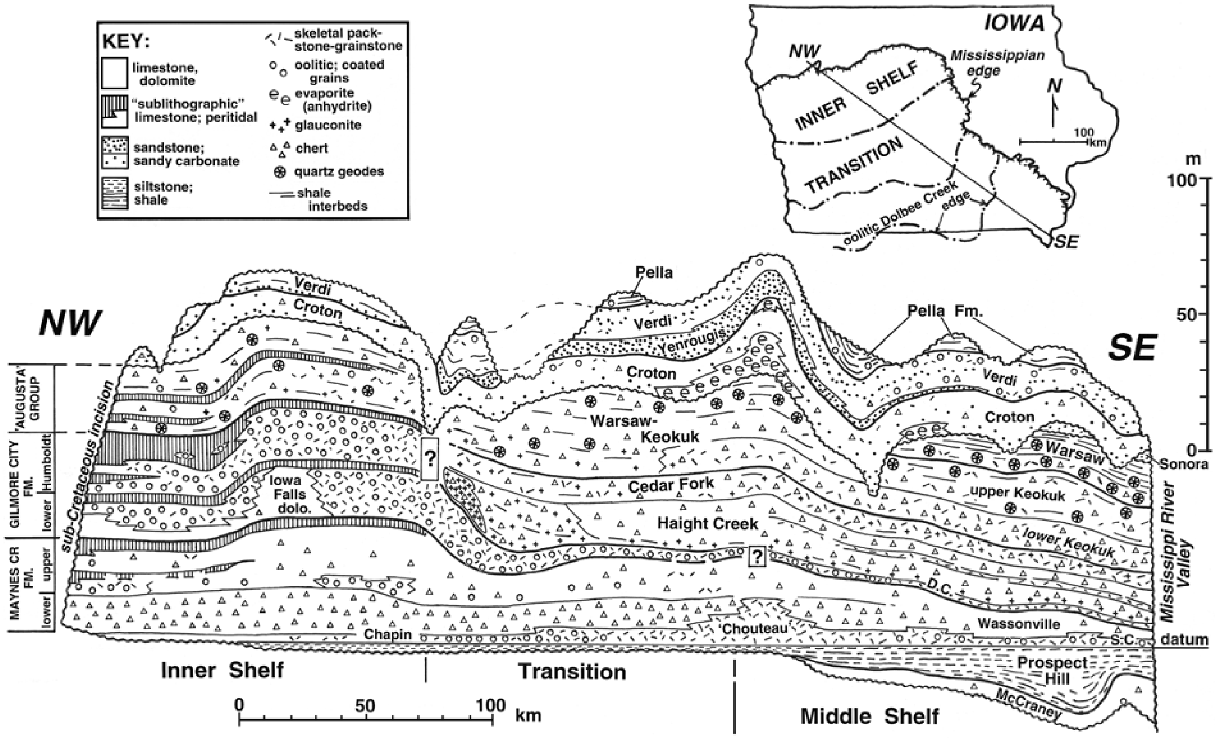

Figure 6. Northwest-southeast stratigraphic cross section of Mississippian strata in Iowa, spanning inner-shelf, transitional middle-shelf, and middle-shelf facies tracts. The diagram is based largely on well penetrations, but it includes numerous outcrop sections in southeast Iowa. Uncertain stratigraphic relations between Gilmore City and Burlington formations are queried. Abbreviations: S.C. = Starrs Cave Member; D.C. = Dolbee Creek Member. The magnitude of the sub-Croton unconformity surface is evident, and the general distribution of evaporites within the Croton Member is shown. Glauconite-rich facies become prominent northwestward within the Haight Creek Member, but these abruptly terminate at the upper Gilmore City margin.

Regardless of the actual correlations between inner- and middle-shelf areas, it is clear that the Burlington Formation is restricted entirely to the middle-shelf area. Witzke and Bunker (2001, p. 73) noted that "the Burlington and Keokuk formations are considered to be good examples of middle-shelf deposits, characterized by slow average rates of subtidal sediment accumulation and influenced by varying degrees of storm current activity dependent on water depths." The complete absence of peritidal and mudflat/sabkha facies or exposure surfaces within the Burlington strongly supports the idea that deposition of the Burlington Formation was entirely subtidal (submarine) through the early Osagean. During Burlington-Keokuk deposition, sediment accumulation was relatively slow and never filled up accommodation space for prolonged periods of time (hundreds of thousands to millions of years). Graded to amalgamated bedforms of crinoidal packstone and grainstone are interpreted to be the result of winnowing and transportation by episodic storm-current activity across the middle shelf. These are sometimes preserved as thin megaripple and lensatic bedforms which provide evidence of bottom current activity during single storm events. However, interbedding grainstones with dolomitized or silicified mudstones indicates that bottom conditions were relatively quiet and incapable of winnowing muds between storm events. The Burlington is not a shoal-water deposit, and the abundant grainstones in the formation are not the result of shallowing into wave-washed shoals as some previous interpretations have suggested.

The succession of Burlington and Keokuk strata is punctuated by several prominent condensed units and/or starved surfaces, regionally characterized by sculpted hardgrounds and/or phosphatic enrichment (bone beds). The bone beds (especially the one at the top of the Burlington) are interpreted to be submarine discontinuities, coinciding with a significant slowdown in sediment accumulation (Harris, 1982, p. 39). The resistant bones and teeth of fish (especially the remains of bradyodont, cladodont, and ctenacanth sharks) accumulated as a sedimentary lag at times when carbonate sediment accumulation was slow to absent. Some bedding surfaces and bone beds contain large thick-shelled brachiopods (especially *Spirifer grimesi*) that show evidence of corrosion and boring, possibly indicating long exposure on the seafloor and episodic carbonate dissolution.

Glauconitic enrichment is of particular note in the lower Haight Creek green dolomite bed as well as in some limestone beds of the Cedar Fork. Glauconite is a term that encompasses various green-colored iron-rich clay minerals, and sedimentary geologists commonly use its occurrence as an indicator of very slow sedimentation. Witzke and Bunker (1996, p. 322) interpreted the lower Haight Creek glauconitic bed to mark a condensed transgressive deposit formed during an interval of seaway deepening. Glauconitic and phosphatic (bone bed) enrichment in some beds of the Cedar Fork Member underscores the slow and condensed aspect of much of Burlington deposition. Some of the Cedar Fork bone beds apparently have regional stratigraphic continuity. Extreme glauconitic enrichment ("super-glaucony") is noted in the Haight Creek Member of central Iowa in proximal areas of the middle shelf below the inner-shelf margin (Figs. 3, 6), where much of the dolomite succession is notably green in color (laminated in part). For reasons not yet understood, this area was a locus for glauconite deposition.

Witzke and Bunker (1996) subdivided the Burlington Formation into two large-scale transgressive-regressive (deepening-shallowing) cycles of deposition. They interpreted the carbonate mudstones (now dolomitized or silicified) to represent the deepest-water deposits, generally "deposited in quiet subtidal environments below storm wave base" (ibid., p. 322). Crinoidal packstone-grainstone beds were interpreted to record relative "depositional shallowing above storm wave base" (ibid.). The Dolbee Creek Member was assigned to transgressive-regressive (T-R) cycle 4 of the Mississippian succession, although the transgressive portion of this cycle apparently is represented by a starved surface of nondeposition and submarine planation across southeast Iowa. In more shoreward directions of the inner shelf, transgressive deposits of T-R cycle 4 are probably represented by lower facies of the lower Gilmore City Formation (Fig. 3).

Witzke and Bunker (1996) assigned the Haight Creek-Cedar Fork interval of the Burlington to Mississippian T-R cycle 5, and subdivided into two "subcycles" at the position of the "middle grainstone" in the Haight Creek Member. The lower Haight Creek green marker (glauconitic) is considered to be the condensed transgressive bed of this cycle. Recent investigations of Burlington and Gilmore City strata across central and northern Iowa suggest that these subdivisions within cycle 5 may be overly simplistic. The remarkable lateral continuity of the Cedar Fork Member across its regional extent (Fig. 3), the apparent absence of interfingering or gradational facies between upper Haight Creek and lower Cedar Fork strata, and the local identification of a submarine hardground surface at the base of the Cedar Fork may suggest that the Cedar Fork comprises a separate T-R cycle. Middle and upper Gilmore City strata likely are stratigraphic equivalents of the Haight Creek-Cedar Fork interval (Witzke, 2002), and these inner-shelf strata also comprise two separate T-R cycles, each marked by subtidal skeletal limestones at their base and capped by a shallowing succession of oncolitic and peritidal/mudflat facies.

The modern shelf of south Australia recently described by James et al (2001) may be an appropriate modern analog for Burlington-Keokuk deposition. They documented a sedimentary regime influenced by strong currents and storm activity in which skeletal packstone and grainstone deposition (dominantly bryozoans and mollusks) occurs across a broad region of 50 to 200 m (160-650 ft) depth which they termed the "middle shelf." Spiculitic carbonate mudstones and wackestones dominate at depths of 200 to 500 m (650-1600 ft) on the south Australian shelf (James et al., 2001), and these facies may be analogous to the cherty and spiculitic carbonate mudstones (mostly dolomitized) of the Burlington-Keokuk interval.

THE KEOKUK FORMATION (UPPER OSAGEAN, MISSISSIPPIAN)

The Keokuk limestone was named by Owen (1852) for bluff and creek valley exposures of cherty carbonate rocks at Keokuk, Lee County, Iowa. Hall (1857; Hall and Whitney, 1858) defined the Keokuk Formation to include a lower cherty limestone interval, a middle crinoidal limestone unit, and upper geode-bearing strata. Keyes (1895) designated these lower beds the “Montrose cherts,” named for exposures at Montrose, Lee County, Iowa, but he included these strata in an expanded Burlington Formation. Van Tuyl (1923) returned the cherty “Montrose” interval to the Keokuk Formation, and assigned the geode-bearing strata to the overlying Warsaw Formation. Subsequent usage in Iowa and Illinois has followed Van Tuyl’s definition of the Keokuk (Harris and Parker, 1964). Recent attempts to designate a specific type locality for the “Montrose Member” revealed a rather unfortunate discovery. Keyes (1895) regarded the quarries and exposures at Montrose to characterize this interval, but restudy of the exposures and well penetrations at Montrose indicated that the lower Keokuk interval lies entirely below the level of the Mississippi River at Montrose (22-55 ft below river level). Only the uppermost part of the Keokuk Formation and the lower part of the Warsaw Formation are actually exposed at Montrose. This discovery seemingly necessitates the abandonment of the term “Montrose” as it has come

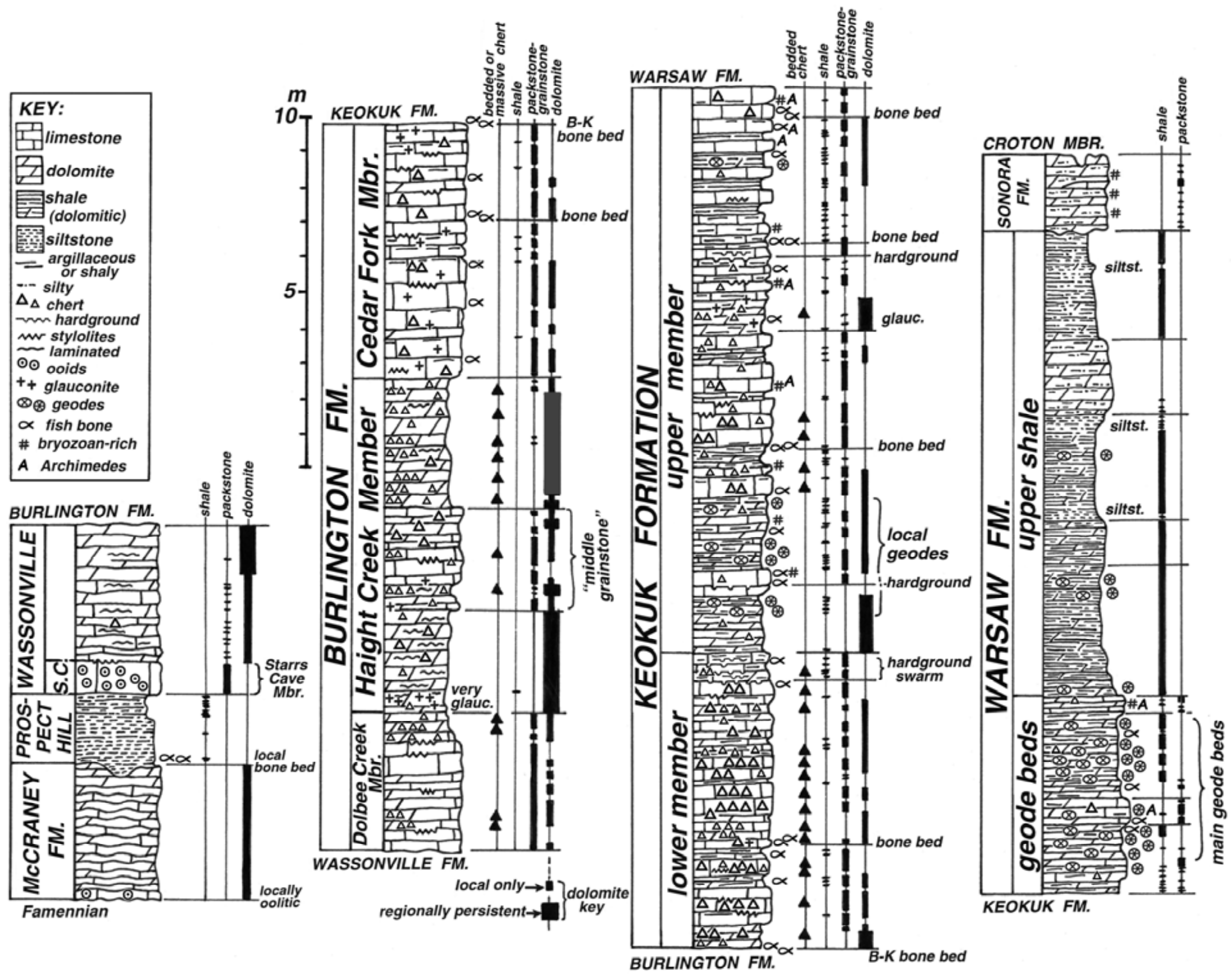

Figure 7. Composite stratigraphic section of sub-“St. Louis” (sub-Croton) stratigraphic units in southeast Iowa, primarily Des Moines, Lee, and Van Buren counties, the type region for the classic Burlington-Keokuk succession. The stratigraphic distribution of massive or bedded cherts, shale, packstone-grainstone units (especially crinoidal), and dolomite facies are indicated. Abbreviations: S.C. = Starrs Cave member; B-K = Burlington-Keokuk contact.

to be used, and new stratigraphic nomenclature for the lower and upper members of the Keokuk Formation in Iowa and adjoining areas of Illinois and Missouri should be considered. For now, the Keokuk Formation is informally subdivided into unnamed upper and lower members (Fig. 7).

The Keokuk Formation occupies the middle part of the Augusta Group, and it is characterized by an interbedded succession of skeletal limestone, dolomite, nodular to bedded chert, and thin shale in southeast Iowa. Where overlain by the Warsaw Formation, the Keokuk averages about 65 to 75 feet (20-23 m) thick in southeast Iowa (Fig. 7). The lower member ("Montrose chert") is characterized by a 23 to 30 foot (7-9 m) thick interval of interbedded fossiliferous packstone-grainstone (limestone), sparsely fossiliferous dolomite, prominent large chert nodules and chert beds, and minor green-gray shale. The limestones, dolomites, and cherts closely resemble those seen in the Burlington Formation, although the Keokuk grainstones are commonly more brown-colored than those of the Burlington. Massive nodular to bedded cherts are prominent in the lower member, and these are generally pale buff to gray smooth cherts replacing carbonate mudstone fabrics (with some silicified packstone). Keokuk strata are erosionally beveled beneath Quaternary units over much of Des Moines County, Iowa, and only exposures of the lower member are generally accessible in the greater Burlington area (Fig. 8). Upper Keokuk and Warsaw strata are best seen south of Burlington in Lee County (the type Keokuk area), but exposures are also present in southwestern Des Moines County.

Representative lower Keokuk sections (within a 12-miles radius of Burlington) are shown in Figure 8. The lower Keokuk can be subdivided into three large-scale lithologic units: 1) a lower interval with common fossiliferous packstones and bedded cherts, 2) an upper unit dominated by argillaceous dolomite, and 3) highest beds with common packstones and shale interbeds. The lower interval is very fossiliferous, with an abundance of crinoid debris (scattered articulated cups), bryozoan grains (including common fenestellids), and brachiopods (especially spiriferids and productids) with scattered trilobites (notably more common than in the Burlington), solitary corals, rostroconchs, gastropods, bivalves, conodonts, and fish debris (see faunal lists by Van Tuyl, 1923). The upper unit is sparingly fossiliferous, but small crinoid debris and sponge spicule molds are noteworthy. Thin discontinuous fossiliferous packstones interbed with the argillaceous dolomites. A comparison of the Keokuk sections at the Nelson and South Augusta quarries (Fig. 8) reveals remarkable similarities, not only in the gross lithologic succession, but also in the positions of individual packstone-grainstone beds. The apparent correlatability of individual beds supports the idea of widespread depositional uniformity acrosss regions of the middle shelf, and the widespread thin packstones may represent individual or amalgamated storm deposits. As for Burlington deposition, units with enrichment in bone and/or glauconite in the Keokuk succession likely mark episodes of relative sediment condensation.

The upper member of the Keokuk Formation in southeast Iowa is similar in gross respect to the lower member. However, the upper member displays considerably fewer large nodular to bedded cherts, it contains more shale interbeds (especially upward), and some dolomite units within the member (locally with quartz geodes) are regionally traceable (Fig. 7). Interbedded crinoidal packstone-grainstone units in the upper member resemble those of the lower member, but some horizons in the upper part contain vast numbers of very large brachiopods, some up to 4 inches (10 cm) in diameter (especially *Orthotetes keokuk*, "*Spirifer*" *keokuk*). Bryozoans are common in many of the Keokuk grainstones, and the distinctive screw-shaped *Archimedes* first appears in the upper half of the upper member in southeast Iowa. Several regionally persistent bone bed horizons and hardground surfaces are identified in the upper member, and some strata are variably glauconitic (Fig. 7). The contact between the Keokuk and overlying Warsaw formations is generally drawn at the top of an interbedded succession of fossiliferous packstones with thin shales and beneath an argillaceous dolomite unit with quartz geodes. This contact apparently is conformable, but a prominent bone bed horizon a few feet (<1 m) below the contact likely marks a thin condensed interval.

WARSAW FORMATION (UPPER OSAGEAN, MISSISSIPPIAN)

Warsaw Stratigraphy

The Warsaw Formation comprises the upper part of the Osagean Series in southeast Iowa, but the formation is not exposed at Burlington or over most of Des Moines County. The Warsaw Formation was originally named by Hall (1857; Hall and Whitney, 1858) for a succession of interbedded shale and limestone found above the "geode beds" at Warsaw, Illinois, about 3 miles (5 km) down the Mississippi Valley from Keokuk, Iowa. Van Tuyl (1923) expanded the Warsaw Formation to include the "geode beds," an interval of argillaceous dolomite and shale with prominent quartz geodes that was previously included in the Keokuk Formation. Using this definition, the Warsaw Formation, thereby, consists of two general members: 1) the lower "geode beds" (and associated shale strata)[Lower Warsaw], and 2) a fossiliferous limestone and shale interval of the upper Warsaw (about 25 to 30 ft thick; 7.5-9 m)[Upper Warsaw]. The boundary between lower and upper Warsaw was used to define the contact between the Osagean and Meramecian Series (Kammer et al., 1990). Characteristic fossiliferous limestones (with *Archimedes*) seen in the upper Warsaw at Warsaw, Illinois, have not been recognized in Iowa, and it is likely that the entire Warsaw succession in southeast Iowa represents only the lower member. The thickest sections of the Warsaw Formation in Iowa (about 60 feet; 18 m) are thinner than the type Warsaw section, where the entire formation is about 80 feet (24 m) thick.

The lower member of Warsaw Formation shows a similar succession of strata in Lee County, Iowa, and at Warsaw, Illinois: 1) a lower geode-bearing argillaceous dolomite interval, and 2) an upper shale-dominated interval (with scattered geodes). The lower 14 to 20 feet (4-6 m) is dominated by argillaceous to shaly dolomite containing scattered to abundant quartz geodes. This interval contains the largest and best geodes, and it is this interval that apparently comprises the "geode bed" of Hall (Hall and Whitney, 1858, p. 100-101, wrote that the "geode bed" does not generally exceed 20 to 25 feet in thickness). The geode-bearing argillaceous dolomite beds are only sparingly fossiliferous, but scattered skeletal debris molds and burrows are noted. This geode-bearing interval also contains discontinuous fossiliferous limestone beds, commonly with lensatic or megarippled bedforms. These limestone beds are primarily skeletal packstones containing crinoid debris, brachiopods, and bryozoans (including *Archimedes*). Thin shales and shaly dolomite interstratifies in the interval. Chert nodules are generally rare. In many respects, this lower argillaceous dolomite interval duplicates lithologies seen in the Keokuk Formation.

The upper portion of the lower member of the Warsaw Formation is dominated by shale, and this interval is informally termed the "upper shale." The gray shales are silty and dolomitic to varying degrees, and the "upper shale" is less resistant than underlying units and commonly weathers to a recessive slope. This "upper shale" interval in southeast Iowa reaches thicknesses to 42 feet (13 m) in Lee County, and it is about 30 feet (9 m) thick at Warsaw, Illinois (where it is overlain by the fossiliferous Upper Warsaw beds). Fossils are characteristically very sparse in the "upper shale" interval, although fossiliferous lenses are noted (Snyder, 1991). Thin siltstone beds occur within the shale interval, and argillaceous and silty dolomite strata are common in the middle and upper parts of the interval. The upper shale interval contains scattered quartz/chalcedony geodes, especially in the lower half, but these geodes are not particularly well preserved (commonly collapsed or incompletely formed).

The "upper shale" interval of the Warsaw Formation is overlain by strata of the Sonora or "St. Louis" formations in southeast Iowa. The Warsaw Formation is locally overlain by the Sonora Formation in a small area of southeast Iowa and western Illinois, and this contact is known to display minor erosional relief at some localities. Across most of Iowa, however, the Warsaw Formation is erosionally beveled and incised by a major erosional unconformity that separates Warsaw-Keokuk strata from the overlying "St. Louis" Formation. This surface displays up to130 feet (40 m) of erosional relief across southeast Iowa, and it is locally incised as low as the lower Keokuk Formation. This major erosional episode marks a time of complete seaway withdrawal from Iowa and long-term subaerial exposure and erosion. This erosional episode was largely coincident with deposition of shallow-water facies of the Salem Formation southward in Illinois and Missouri.

Origin of Geodes

Quartz geodes are common in the lower Warsaw Formation of southeast Iowa and adjoining areas of Illinois and Missouri. These distinctive geodes are so renowned among rock and mineral collectors that the Iowa General Assembly declared the geode as the official “state rock” in 1967. The geodes typically show an outer rind of chalcedony with quartz crystals lining (or filling) the interior. A variety of secondary minerals locally occur in the geode interiors (most commonly botryoidal linings of chalcedony). Most Warsaw geodes range between about 1 and 6 inches (3-15 cm) in diameter, but significantly larger specimens are known from the area. Although the lower Warsaw Formation is the source of most geodes in southeast Iowa, there are also geode-bearing argillaceous dolomite strata in the upper Keokuk Formation that also contain similar quartz geodes (Fig. 7; Van Tuyl, 1923; Witzke and Bunker, 2000). Geodes and quartz-crystal-lined voids are also locally developed in dolomite strata of the Burlington Formation (as seen in Haight Creek strata at Grays Quarry, Hamilton, Illinois). Therefore, quartz geodes can potentially occur at multiple stratigraphic positions within the Augusta Group. Geodes are typically hosted in argillaceous dolomites beds and less commonly in dolomitic shales, but they are not known from limestone beds.

Quartz geodes clearly are not primary sedimentary features, but compactional deformation of enclosing strata suggests that the nodular forms are likely of fairly early diagenetic origin. The origin of geodes has vexed geologists for some time, and several different ideas have been put forward. However, there is general agreement concerning several points: 1) geode precursors were nodules or concretions of a non-quartz mineral (either calcite or anhydrite); 2) the interiors of these nodules were dissolved, leaving a hollow space; and 3) the minerals now seen inside geodes were transported in groundwater solutions and then precipitated as replacements of the geode walls and as crystalline growths within their interior cavities.

The best case can be made for the precursor nodules being composed of anhydrite (a calcium sulfate evaporite mineral) rather than calcite based on two significant observations. First, virtually all Warsaw geodes contain traces of anhydrite preserved within the outer chalcedony shells. These anhydrite inclusions may be microscopic, but the ubiquitous occurrence of anhydrite in the geodes lends support to the idea that the nodules were originally composed of anhydrite. Second, Keokuk and lower Warsaw strata in the subsurface of south-central Iowa (as seen in bedrock cores) still preserve intervals with anhydrite nodules. Of particular note, partial replacement of these anhydrite nodules by chalcedony is locally seen (i.e., these are incipient geodes). In addition, if the nodules were originally composed of calcite, their subsequent dissolution is difficult to understand in light of the abundance of undissolved calcite (as limestone beds) in the Keokuk-Warsaw formations. The lumpy cauliflower-like form of many geodes closely resembles the lumpy forms seen in many anhydrite nodules.

If the precursor nodules were composed of anhydrite, however, an additional question must be asked: where did the anhydrite come from? This question is a more difficult one to answer, primarily because anhydrite is an evaporite mineral more characteristically associated with precipitation in hypersaline environments. However, geode-bearing strata are known to contain fossils generally associated with environments of normal-marine salinity (not hypersaline), especially crinoids and fenestellid bryozoans. Therefore, the anhydrite nodules cannot be regarded as primary sedimentary features, but must be a post-depositional early-diagenetic growth within uncompacted sediments. The precipitation of these anhydrite nodules likely occurred as hypersaline waters regionally percolated through the sediments of the Augusta Group (see Loope et al., 1996, for a similar model of geode formation in Indiana). As noted above, an episode of seaway restriction and withdrawal across Iowa occurred following Warsaw deposition, resulting in deep erosional incision of Warsaw strata. The final stages of seaway withdrawal are not preserved in the Iowa Warsaw succession, and the expected nearshore and restricted-marine facies associated with such seaway shallowing are now eroded across the region. It is not unreasonable to hypothesize that hypersaline environments would migrate across the region during seaway withdrawal. When shallow seas once again returned to the area, precipitation of anhydrite was widespread across southern Iowa (“St. Louis” Fm; see Fig. 6).

The following sequence of events is proposed for the formation of quartz geodes in the Keokuk and Warsaw formations. 1) Anhydrite nodules grew in uncompacted sediments (precipitated from hypersaline groundwaters). 2) Silica was mobilized in groundwater solutions. 3) The anhydrite nodules were partially replaced by chalcedony (espcially the outer shells). 4) Anhydrite is further dissolved by less saline groundwater influx leaving hollow geode interiors (this dissolution probably accompanied early chalcedony replacement, and steps 2, 3, and 4 are coeval in part). Some thin-shelled geodes probably collapse under compaction at this stage. 5) Continuing influx of silica-bearing solutions produces further precipitation of quartz crystals and chalcedony. 6) Infilling of geodes by other minerals (especially calcite) occurred during later stages.

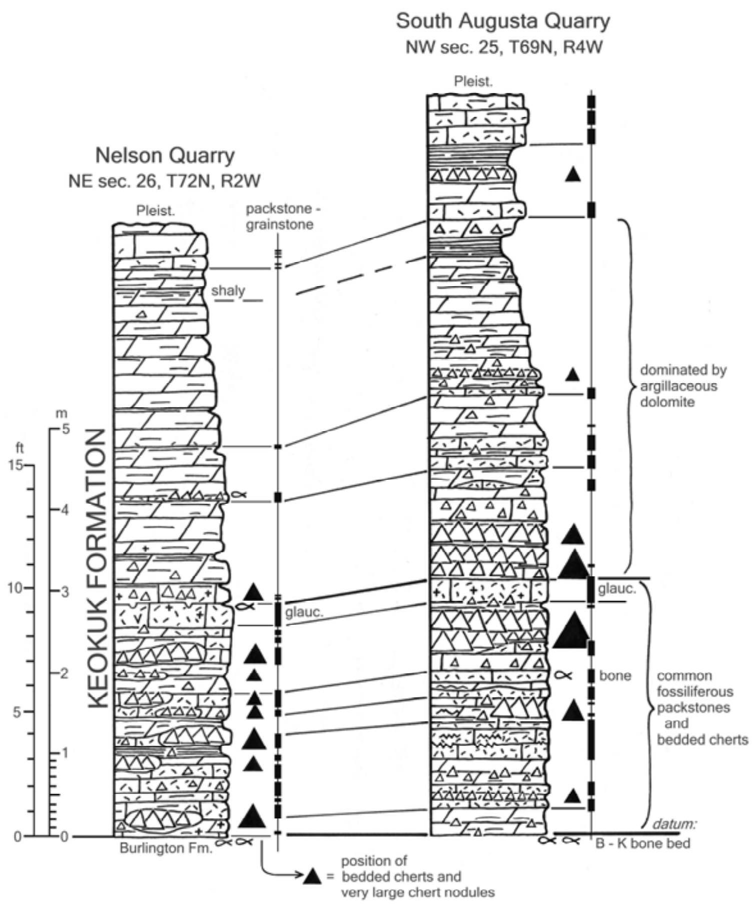

Figure 8. Representative graphic sections of lower Keokuk strata within a 12-mile radius of Burlington, southeast Iowa. Symbols as in figures 1, 6, and 7. Positions of bedded cherts are indicated. Correlation of individual packstone-grainstone beds shows remarkable lateral continuity.

The development of a reasonable hypothesis for the origin of geodes includes two critical issues: 1) identifying sources for the quartz (silica), and 2) identifying mechanisms for moving tremendous volumes of silica-bearing groundwater through the sediments (large volumes are inferred because of the low solubility of quartz in solution). Quartz geodes form a volumetrically minor part of the total volume of silica preserved in the Augusta Group when compared to the prominent masses of nodular and bedded chert (microquartz) which also represent silica replacements (of carbonate sediments). Large volumes of silica are needed to account for the replacement-precipitation of chert, grain silicification, and geode growth, but where did all this silica come from? Volcanic ash has been considered a potential source by some geologists, but the complete absence of any recognizable volcanic material in these strata makes this source seem unlikely. Instead, a biogenic (biologically-precipitated) source seems more reasonable, derived from the siliceous skeletons of certain plants and animals. Siliceous sponge spicules are common small fossils in the geode-bearing succession, and these spicules are commonly seen as hollow molds in dolomite (spicules now dissolved). The dissolution of siliceous sponges and other siliceous fossils in the sediments potentially could yield large volumes of silica to the migrating groundwater solutions. Biogenic silica (opaline silica) is much more soluble than ordinary quartz, making it easier to account for the mobilization and precipitation of so much silica in these strata. Biogenic (opaline) silica is unstable over geologic time, and it mineralogically inverts to ordinary quartz. The migration of large volumes of silica-bearing groundwater solutions is necessary, but given the large amount of time represented by the post-Warsaw unconformity, this may not be as large a problem as it initially may seem. Coastal aquifers likely moved large volumes of water through these strata over a period of time that spanned many hundreds of thousands of years.

POST-WARSAW MISSISSIPPIAN STRATA IN SOUTHEAST IOWA

Sonora Formation

Keyes (1895) named the Sonora Formation for an interval of dolomite and dolomitic sandstone in the Sonora Quarry north of Nauvoo, Illinois (south across the Mississippi River from Fort Madison, Iowa). The formation is seen in parts of southeast Iowa and west-central Illinois, and it has a relatively limited geographic distribution. It reaches maximum thicknesses to about 25 to 35 feet (8-11 m). The Sonora is discontinuous in southeast Iowa, and it is commonly absent (where "St. Louis" directly overlies Keokuk or Warsaw). At the Geode Quarry in southwest Des Moines Co., Iowa, the Sonora occurs as a large lenticular ("reef"-like) body of cross-bedded sandy dolomite strata with a diameter not much larger than the quarry itself.

The Sonora Formation apparently is disconformable with underlying Warsaw strata in southeast Iowa. The contact in this area locally displays minor erosional relief (as at Orba-Johnson terminal road, Sunday field stop; and at type Warsaw section) but is roughly planar at many localities. Regionally, the Sonora overlies progressively older units northwestward. At the type Warsaw section, the Sonora overlies the strata near the top of the Upper Warsaw; at Orba-Johnson (Sunday field stop) north of Keokuk, the Sonora overlies strata near the top of the Lower Warsaw; and in western Des Moines County it overlies strata near the middle of the Lower Warsaw (Van Tuyl, 1923, p. 227). These stratigraphic relationships underscore the unconformable nature of the Warsaw-Sonora contact in Iowa. However, southward in Illinois the Sonora and uppermost Warsaw become laterally gradational (Collinson, 1964). Where capped by younger Mississippian strata in Iowa, the Sonora is unconformably overlain by the "St. Louis" Formation, and the contact is generally irregular. In western Illinois, the Sonora grades southward into carbonate facies of the Salem Formation, and the "contact between the Sonora and Salem is conformable" (Snyder, 1991, p. 14).

The Sonora Formation is dominated by variably-sandy porous dolomite beds containing abundant molds (and dolomitized replacements) of fenestellid and polyporid bryozoans. The bryozoan-rich dolomites contain scattered to abundant quartz sand and/or silt, and thin sandy argillaceous partings separate some beds. The strata commonly show low-angle cross-stratification to planar bedding of fossiliferous and sandy laminae. Higher-angled crossbeds are also noted. Bryozoans are overwhelmingly dominant, but crinoid and blastoid debris, brachiopods (a moderately diverse fauna), gastropods, solitary corals, and trilobites are also present (Van Tuyl, 1923, p. 217). The sandy bryozoan-rich dolomite beds locally contain discontinuous thin lenses to beds of limestone (including crinoidal packstones). Shale and sandy shale beds are locally prominent in Iowa. Thin sandstone interbeds are seen, and thicker sections of crossbedded dolomitic sandstone are locally noteworthy (generally more common northwestward in southeast Iowa).

The Sonora Formation in southeast Iowa probably likely represents a shoreward facies equivalent of some part of the Salem Formation in western Illinois. The Sonora marks the farthest the Salem seaway encroached to the northwest into the continental interior. Quartz sand and silt probably originated as fluvial influx of quartz detritus derived from the eroding landscape of the continental interior. The abundance of comminuted bryozoan debris and crossbedding suggests relatively high-energy wave-washed marine environments, possibly in shoal settings. Back-shoal areas may have been sites of finer grained sedimentation. The Sonora is the oldest Meramecian unit over most of southeast Iowa.

"St. Louis" Formation in Iowa

Carbonate, evaporite, and sandstone facies of the so-called "St. Louis" Formation were deposited across much of Iowa above the major unconformity surface on eroded strata of the Augusta Group and Sonora Formation. These strata have traditionally been assigned to the St. Louis Formation in Iowa (e.g., Van Tuyl, 1923) even though the contained rocks seem lithologically distinct from facies in the type St. Louis area. In addition, there has been a paucity of biostratigraphic control to verify correlation between Iowa and Missouri. In particular, a significant portion of the Iowa "St. Louis" interval is composed of siliciclastic deposits (especially sandstone) that are strikingly dissimilar to the carbonate-dominated sections in the type St. Louis area. Preliminary biostratigraphic studies reported by Woodson (*in* McKay et al., 1987) first suggested that the upper part of the so-called "St. Louis" interval in Iowa was, in fact, entirely younger than the type St. Louis succession. A more detailed description of the "St. Louis" interval across southeast Iowa is given by Witzke et al. (1990).

The so-called "St. Louis" interval in Iowa has been subdivided into four stratigraphic units in southeast Iowa (McKay et al., 1987; Witzke et al., 1990), in ascending order: Croton Member, Yenruogis Member, Verdi Member, and Waugh Member. Formational terminology has not been formally adopted for Iowa, but new formational names are encouraged to replace the awkward and inappropriate label of "St. Louis." Only the Croton and Yenruogis members in Iowa probably correlate with the St. Louis succession in Missouri. The lower Croton Member (named by Van Tuyl, 1923) includes evaporites (gypsum-anhydrite) in southern Iowa (Fig. 6), and extensive collapse breccias are seen in the carbonate-dominated Croton interval across much of Iowa outcrop belt. The scale of brecciation and sizes of the contained breccia clasts show tremendous variation at outcrop scale. The Croton is locally seen in an unbrecciated "undisturbed phase" (Van Tuyl, 1923), where it is composed of bedded dolomitic limestone and sandy dolomite with minor interbedded sandstone. The upper Croton beds are locally fossiliferous, and the St. Louis guide fossil *Acrocyathus floriformis* ("*Lithostrotion*" corals) is identified. Calcareous microfossils recovered from this interval include the foraminiferal genus *Eoendothyranopsis* and the problematic alga *Koninckopora* (Witzke et al., 1990, p. 24), both not known to range above the St. Louis Formation in the Illinois Basin. The Croton Member is overlain by a southeastward prograding sandstone body named the Yenruogis Sandstone (ibid.), but this sandstone body does extend into extreme southeastern Iowa. Collectively, the Croton and Yenrougis members constitute a single depositional cycle.

The upper part of the so-called "St. Louis" interval in Iowa includes a lower marine limestone and sandstone interval (the Verdi Member) and an upper nonmarine interval of limestone, shale, and sandstone (Waugh Member). The Waugh Member is only developed in the western part of southeast Iowa, and it does not extend to the Mississippi River. The Verdi Member contains a moderately diverse marine invertebrate fauna (Witzke et al., 1990, p. 34). Woodson (*in* McKay et al., 1987) documented a variety of calcareous microfossils from the Verdi of southeast Iowa which generally support a correlation of these strata with the Ste. Genevieve Formation of the Illinois Basin rather than with the St. Louis Formation. Verdi conodonts include *Hindeodus cristulus* and *Cavusgnathus unicornis* (McKay et al., 1987). The apparent absence of several characterisitic St. Louis conodont taxa, especially species of *Apatognathus* and *Spathognathodus scitulus*, is of note, further supporting a Ste. Genevieve correlation for the Verdi.

The Waugh Member includes fluvial, lacustrine, brackish, and terrestrial facies that have yielded a diverse and remarkable vertebrate fauna (McKay et al., 1987; Bolt et al., 1988; Witzke et al., 1990). Abundant tetrapod fossils (protoanthracosaurs [*Whatcheeria*], colosteids [*Greererpeton*]) were recovered within a collapse feature in Keokuk Co., Iowa (Bolt et al., 1988; Lombard and Bolt, 1995), and a variety of fish taxa occur in the lacustrine facies (including xenacanths, petalodonts, acanthodians, palaeoniscoids, rhipidistians, rhizodonts, dipnoans). The combined Verdi-Waugh interval records a transgressive-regressive cycle of deposition that has been interpreted to correlate with the lower Ste. Genevieve cycle of the Illinois Basin (Witzke et al., 1990). Thin coal and *Stigmaria* rooting are locally seen at the top of the Waugh Member.

The Croton Member correlates with the true St. Louis Formation of Missouri and, therefore, is a Meramecian interval. The upper "St. Louis" Verdi beds probably correlate with the lower Ste. Genevieve Formation of Illinois and Missouri. The Ste. Genevieve Formation is now included in the lowermost part of the Chesterian Series (Lane and Brenckle, 2001), but some earlier workers considered the Ste. Genevieve to be an upper Meramecian interval.

Pella Formation

The Pella considered to be correlative with the Ste. Genevieve Formation of the Illinois Basin, and some stratigraphers have synonymized the Pella and Ste. Genevieve. The formation records a major marine transgression, but the upper part of the Pella has been erosionally truncated beneath the Pennsylvanian across Iowa. The Pella is retained as a distinct lithostratigraphic unit in Iowa characterized by a thin basal limestone unit (locally oolitic) and a thicker upper fossiliferous calcareous shale ("marl") interval. The abundant macrofauna and microfauna of the Pella is consistent with a Ste. Genevieve correlation (McKay et al., 1987).

The Verdi-Waugh interval and the Pella Formation both probably correlate with the Ste. Genevieve Formation, but each represents a separate transgressive-regressive depositional cycle. Of note, the Ste. Genevieve in the Illinois Basin area also contains two separate depositional cycles (Swann, 1963) each marked by a lower marine limestone (marine transgression, Fredonia and Karnak-Joppa members) and an upper sandstone (regression, Spar Mountain Member and Aux Vases Sandstone).

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CRINOID AND BLASTOID BIOZONATION AND BIODIVERSITY IN THE EARLY MISSISSIPPIAN (OSAGEAN) BURLINGTON LIMESTONE

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INTRODUCTION

During the first geological survey of Iowa, Wisconsin, and Minnesota (1848-49), Owen (1850) named the crinoidal limestones along the bluffs of the Mississippi River the "Encrinital Group of Burlington," for the limestones exposed in the vicinity of Burlington Iowa, and the "Reddish-brown Encrinital Group of Hannibal," for similar grainstones exposed in the vicinity of Hannibal, Missouri. Owen (1850) was also the first to describe fossil crinoids and blastoids from these strata; however, he believed that the crinoidal limestones of Hannibal were stratigraphically younger than those exposed in Burlington. The same strata were referred to as the "Encrinital limestone" in the first geological survey of Missouri (Swallow, 1855; Shumard, 1855); however, Swallow (1855) recognized that Owen's (1850) "Encrinital Groups" of Burlington and Hannibal were actually part of the same geologic formation. Hall (1857) concurred with Swallow and formally named the "Encrinital limestone" the "Burlington limestone," for the well-exposed encrinites of Burlington, Iowa.

Since the 1850's, the Burlington Limestone has received much attention from stratigraphers, economic geologists, and paleontologists, with the latter paying particular attention to the extremely high concentration of crinoidal material. Furthermore, many researchers noted that the Burlington was not uniform in its lithologic or biotic composition and began subdividing the formation based on these differences (White, 1860, 1870; Niles and Wachsmuth, 1866). The purpose of this paper is to present an historical account of attempts to divide the Burlington Limestone lithologically and paleontologically and discuss confusion that has arisen around the position of these boundaries. Moreover, I hope to provide a framework within which, any Burlington researcher will be able to easily recognize the primary faunal associations. Because crinoids and blastoids are the most abundant fossils in the Burlington, this work follows the lead of White (1860), Niles and Wachsmuth (1866), Rowley (1908), Laudon, (1937, 1973) and others in focusing on the distribution of these pelmatozoan, or stalked echinoderms as biostratigraphical tools. Understanding faunal assemblages and the stratigraphic ranges of each species is essential for documenting multi-scale spatiotemporal paleoecological and evolutionary patterns. That these distributions are clearly understood for the pelmatozoans of the Burlington Limestone is particularly crucial as the Burlington Limestone represents the most diverse concentration of crinoids in the geologic record. Therefore, I also provide a culled listing of currently recognized crinoids and blastoids from this formation, which includes their occurrence and relative abundance. Moreover, I propose herein, recommendations for reporting the stratigraphic occurrence of paleontological samples from the Burlington Limestone and suggest directions of future study.

HISTORICAL DIVISION OF THE BURLINGTON LIMESTONE

White (1860) was one of the first geologists to recognize that the Burlington Limestone could be naturally divided based on lithological and paleontological criteria. He described three divisions of the Burlington, including: 1) a basal crinoidal limestone, 2) alternating layers of limestone, mudstone, and chert and 3) and an upper crinoidal limestone. White (1860, 1870) referred to the lower two portions as the "lower division," and called uppermost portion the "upper division" of the Burlington Limestone. White (1870, p. 203) also suggested that "…the accession of silicious material to the waters of that epoch resulted in or at least was followed by the extermination of all the species of crinoids then existing…," (capitalization, spelling, and grammar in this and proceeding quotations exactly follow the original text) suggesting that the interbedded chert and dolomitic mudstone of the "lower division" formed a significant paleontological boundary between the two crinoidal limestones. Niles and Wachsmuth (1866) proposed to divide the Burlington Limestone into two distinct geological formations based on these paleontological differences, naming White's "lower division" the "Lower Burlington limestone," and the "upper division" the "Upper Burlington limestone." However, White (1870; who was at the time, the state geologist of Iowa) rejected the formal division of the Burlington Limestone into two separate formations based on his observations that the distinction between the two divisions could only be recognized locally. Nevertheless, Niles and Wachsmuth (1866), as well as many subsequent workers, reported the occurrence of Burlington Limestone species as occurring in the informal "lower" or "upper" Burlington Limestone.

The practice of dividing the Burlington Limestone into lower and upper divisions and reporting species in relation to their boundaries is still a common practice (see Gahn and Kammer, 2002); but it does not adequately (or accurately) reflect the natural divisions within this formation. There has been considerable confusion concerning the placement of the lower-upper Burlington boundary by various authors resulting in its inconsistent application. This confusion is centered in the lithological variability of White's (1870) second division of the Burlington Limestone, which is roughly equivalent to the Haight Creek Member (Harris and Parker, 1964) (Fig. 1). The Haight Creek Member is typically characterized in southeast Iowa by having abundant layers of interbedded chert and dolomitic mudstone. However, the Haight Creek Member also contains layers of crinoidal packstone and grainstone that can vary in abundance from being sparse to the dominant lithotype. The Haight Creek Member often contains a thick encrinite near its middle and top that is very similar to the crinoidal limestones of the underlying Dolbee Creek and overlying Cedar Fork Members (Harris and Parker, 1964). This "middle grainstone" was noted by Van Tuyl (1922, p. 121) and further discussed by Witzke et al. (1990, p. 16). The Haight Creek Member carries a unique pelmatozoan assemblage that is more similar to the fauna of the Cedar Fork Member than that of the Dolbee Creek Member.

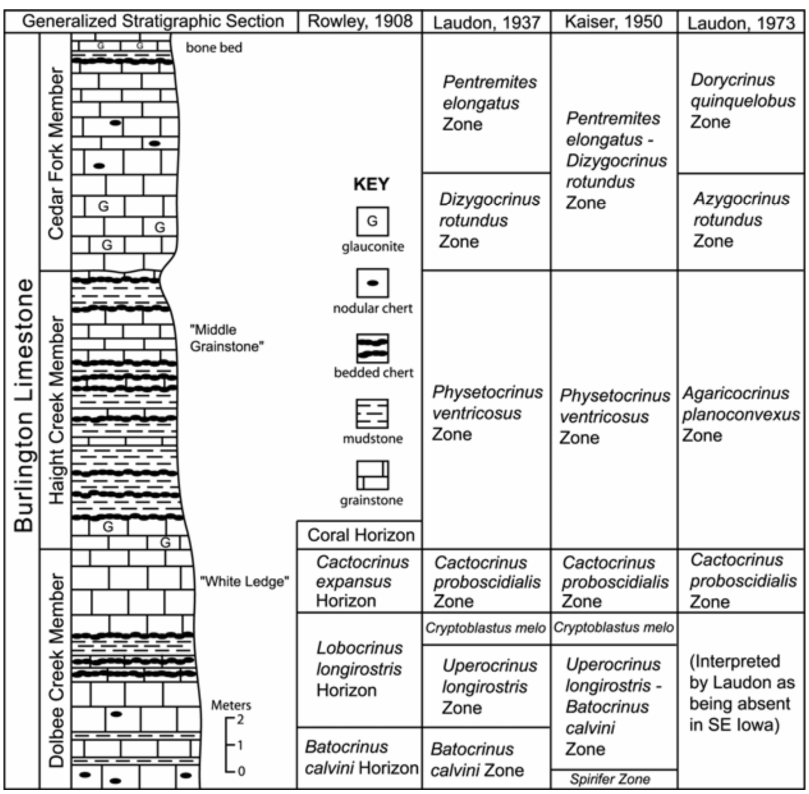

Figure 1. Historical biozonation of the Burlington Limestone. Faunal zones of Rowley (1908), Laudon (1937, 1973), and Kaiser (1950) are plotted against a generalized section of the Burlington Limestone. The zonation proposed by Rowley (1908) was based on the Burlington section at Louisiana, Missouri. The zonation proposed by Laudon (1973) is a composite group of biozones created from Burlington sections in Hannibal, Missouri and southeast Iowa. The zonation proposed by Kaiser (1950) was based on several Burlington sections in southwest Missouri. The second zonation proposed by Laudon (1973) was restricted to southeast Iowa.

The currently accepted interpretation of the lower and upper Burlington places their boundary at the base of the Cedar Fork Member, with the entirety of the Dolbee Creek and Haight Creek members being confined to the lower Burlington (Van

Tuyl, 1922; Laudon, 1973; Witzke et al., 1990). Nevertheless, the fauna from the Haight Creek grainstones were most commonly assigned to the upper Burlington. This unfortunate circumstance results in a paleontologically defined lower-upper Burlington boundary that conflicts with the recognized lithological lower-upper Burlington boundary.

However, it is quite possible that White (1860, 1870) and Niles and Wachsmuth (1866) originally placed the lower-upper Burlington boundary at the base of the "middle grainstone." White (1870) suggested that the lower and upper Burlington divisions are approximately equivalent in thickness, which would be consistent with a lower-upper Burlington boundary at the base of the "middle grainstone." Niles and Wachsmuth (1866, p. 4) recognized the alternating layers of chert and mudstone of the Haight Creek member as being part of the lower Burlington, and delineated the lower-upper Burlington boundary by "the uppermost stratum of chert, which attains any considerable extent and thickness." This is a particularly enigmatic boundary definition as thick, persistent chert beds can be present locally in the vicinity of Burlington, Iowa at the base of the middle grainstone of the Haight Creek Member and at the base of the Cedar Fork Member (and even extending into the lower beds of the latter). Wachsmuth and Springer (1897) refer to many of the typical representatives of the Haight Creek fauna as occurring in the "lower part of the Upper Burlington limestone," confirming the practice of assigning these beds to the upper Burlington on paleontological grounds. More recent studies have also variably placed the lower-upper Burlington boundary. For example, Van Tuyl (1922, p. 121, horizon 2) assigned strata equivalent to the Haight Creek Member to the lower Burlington, including the "middle grainstone" in Burlington, Iowa. However, in Augusta, Iowa, he assigned the "middle grainstone" to the upper Burlington, possibly because he was able to collect what he interpreted as an "upper Burlington" fauna from these strata (Van Tuyl, 1922, p. 132, horizon 3). Furthermore, Moore (1928, p. 171) assigned strata equivalent to the Haight Creek Member almost wholly to the upper Burlington. Understanding the placement of the boundary between the lower and upper Burlington Limestone by those who originally defined it is difficult enough, but the inconsistent use of these divisions on local and regional scales makes the distinction between the lower and upper Burlington essentially meaningless; and therefore, I recommend a cessation of the formal use of these divisions.

Rowley (1908) made the earliest attempt to further subdivide the Burlington Limestone into discrete biozones (Fig.1). He separated the lower Burlington strata of Louisiana, Missouri into four divisions including, in ascending order: 1) the *Batocrinus calvini*, 2) *Lobocrinus longirostris*, 3) *Cactocrinus expansus*, and 4) Coral Horizons. Laudon (1934) expanded Rowley's (1908) work by establishing seven "life zones" in the Burlington. He retained Rowley's division of the Dolbee Creek Member, but separated out the upper five feet of the *Lobocrinus longirostris* Horizon (which he named the *Uperocrinus longirostris* Zone) in Hannibal, Missouri, and called this interval the *Cryptoblastus melo* Zone. Moreover, Laudon renamed Rowley's *Cactocrinus expansus* Horizon as the *Cactocrinus proboscidialis* Zone (probably because *C. expansus* is not a formally defined species; Rowley may have actually been referring to *C. exerptus* (Hall) or *C. extensus* Wachsmuth and Springer, but this is uncertain), and renamed Rowley's "Coral horizon" the *Physetocrinus ventricosus* Zone, which he extended to include the full extent of Haight Creek-equivalent strata. Laudon also established two "life zones" in the Cedar Fork Member, including the *Dizygocrinus rotundus* Zone and the overlying *Pentremites elongatus* Zone.

The "zones" recognized by Rowley (1908) and Laudon (1934) were based exclusively on exposures of the Burlington Limestone in southeastern Iowa and northeastern Missouri. However, Kaiser (1950) applied the same zonation scheme to the Burlington Limestone of southwestern Missouri, albeit with a few changes. He recognized an additional horizon at the base of the Burlington Limestone that he referred to as the "Spirifer zone." He was also unable to distinguish between the *Batocrinus calvini* and *Uperocrinus longirostris* Zones or the *Dizygocrinus rotundus* and *Pentremites elongatus* Zones, and thus combined them.

The most recent echinoderm zonation scheme for the Burlington Limestone was proposed by Laudon (1973) for exposures in southeast Iowa, exclusively. The lowest zone that he recognized in the Burlington Limestone of southeast Iowa was the *Cactocrinus proboscidialis* Zone; believing that the underlying zones were not deposited in the area (see Laudon, 1937). Oddly, Laudon changed the name of

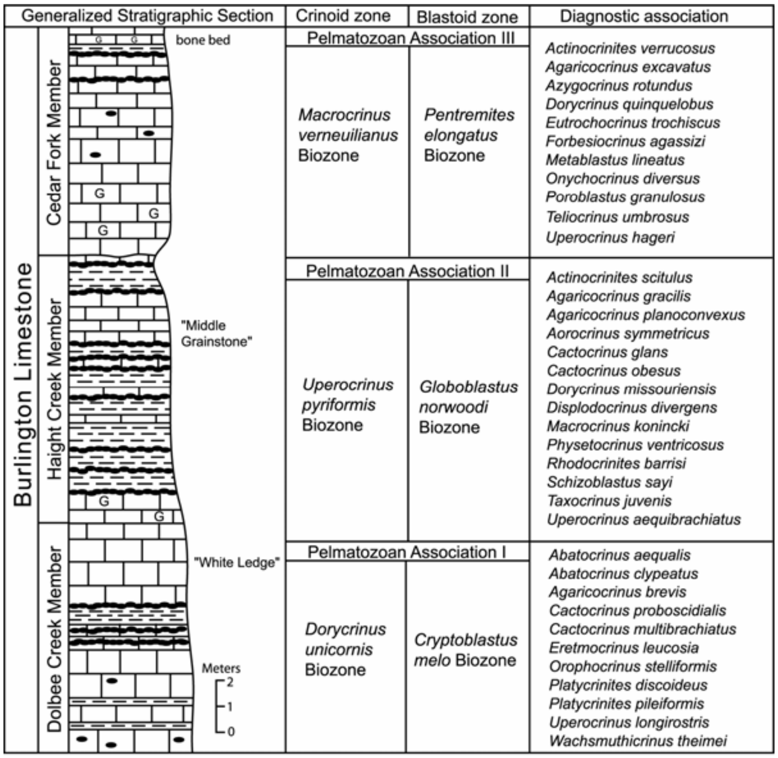

Figure 2. Pelmatozoan Associations of the Burlington Limestone. This figure summarizes the pelmatozoan associations described herein, including their relative stratigraphic positions, corresponding crinoid and blastoid biozones, and species characteristic of each association. Please refer to the text for further discussion of these associations. The key for the stratigraphic section is presented in Figure 1.

the overlying *Physetocrinus ventricosus* Zone to the *Agaricocrinus planoconvexus* Zone, even though they are exactly equivalent. He also changed the *Dizygocrinus rotundus* Zone to the *Azygocrinus rotundus* Zone (after Lane's 1963 amendment of *Dizygocrinus*), and the *Pentremites elongatus* Zone to the *Dorycrinus quinquelobus* Zone (without a clear reason for doing so).

A REVISED BIOZONATION

The zonation schemes discussed above were based on lithological as well as paleontological characteristics. For example, Rowley's (1908) *Cactocrinus expansus* Horizon was originally named for the "White Ledge" of northcentral Missouri; a name given by local quarry men to an economically valuable, massive crinoidal grainstone (Laudon, 1937). In addition, Laudon originally named his *Dizygocrinus rotundus* Zone for glauconitic grainstones found at the base of the Cedar Fork Member as well as for the local abundance of *Azygocrinus rotundus* (Yandell and Shumard) in southeast Iowa. He also suggested that the fauna of this zone is not represented in Hannibal, Missouri, and Kaiser (1950) did not recognize this zone in southwestern Missouri. However, the fauna of the *Dizygocrinus rotundus* Zone is present throughout Missouri, but it cannot be characterized by the glauconitic grainstone or the unusually high abundance of *A. rotundus* that is present in southeast Iowa. The latter point illustrates another problem with the currently proposed biozonation schemes; several of the biozones are characterized by locally abundant or restricted species. Although *A. rotundus* occurs abundantly in southeast Iowa, it is rare throughout Missouri. Additionally, Rowley's (1908) *Batocrinus calvini* Horizon is defined by the presence of *Abatocrinus calvini* (Rowley) an uncommon and locally restricted species.

Because many of the biozones were originally defined by lithologic differences and locally abundant or restricted species, it is difficult to use them outside of the limited geographic regions in which they were described. Herein, I propose a biozonation scheme that is applicable over the full geographic distribution of the Burlington Limestone, and one that is based entirely on paleontological data. The biozonation scheme presented below is based on a decade of personal field experience in the Burlington Limestone and the examination of museum collections housed in the Springer Room of the United States National Museum of Natural History, Harvard's Museum of Comparative Zoology, and the University of Iowa. Special attention was also given to stratigraphic collections from various Burlington Limestone localities made by D. B. Macurda and D. L. Meyer in the late 1960's and early 1970's that are housed at the University of Michigan. The biozonation scheme proposed herein for the Burlington Limestone is divided into three parts that are generally equivalent to the positions of the Dolbee Creek, Haight Creek, and Cedar Fork members. Although the proposed biozones roughly track the three Burlington members, it is important to note that the faunal assemblages described below are not defined by these members and can be traced even where lithological distinction of the members is not possible. Crinoid and blastoid biozones were established for each paleontological association. The names of these biozones were carefully chosen to represent species that: 1) are common representatives of the association over the entire geographic extent of the Burlington Limestone, 2) reach their acme, or maximum abundance within the confines of the biozone, 3) are easily recognizable, but not easily confused with other species by non-specialists and, 4) are reasonably stable taxonomically. If the taxa used in previous biozonation schemes met these criteria, then I honored the names used by prior authors; otherwise, I explain the designation of a new name. Figures and general descriptions of the species chosen to represent the biozones in this study are found in "Index Fossils of North America," by Shimer and Shrock (1944).

The naming of these biozones is secondary in importance to their faunal compositions; understanding the make up of each pelmatozoan association is critical for addressing evolutionary and paleoecological questions pertinent to the crinoids and blastoids of the Burlington Limestone. I have listed a few of the diagnostic species from each pelmatozoan association (2), but it would be more valuable to know the approximate stratigraphic ranges of each pelmatozoan species in the formation. Thus, I attempted to create a complete table of the crinoids and blastoids present in the Burlington Limestone and their known distributions in relation to the associations recognized herein (Table 1). This table is discussed further under the subsequent section on crinoid and blastoid diversity.

Burlington Pelmatozoan Association I: The stratigraphically oldest association recognized in this study is referred to as the Burlington Pelmatozoan Association I (BPAI), and includes the *Dorycrinus unicornis* and *Cryptoblastus melo* Biozones. The biozones of the *D. unicornis*-*C. melo* Association generally encompass the zones discussed in this paper that are equivalent to the Dolbee Creek Member of the Burlington Limestone (Rowley, 1908; Laudon, 1937, 1973; Kaiser, 1950). *Dorycrinus unicornis* (Owen and Shumard) was chosen as the key index crinoid for this zone rather than one of the previously used "zone species" because it possesses a diagnostic morphology that makes it difficult to confuse with any other species. Several species of *Abatocrinus* and *Cactocrinus* are also common in this zone, but they are currently in need of taxonomic revision, and the many species of these genera that are presently in the BPAI can be difficult to distinguish by non-specialists. *Cryptoblastus melo* (Owen and Shumard) was retained as the key index fossil of the blastoid biozone following Laudon (1937) and Kaiser (1950). *Cryptoblastus melo* is an excellent name for this blastoid biozone as the species is probably the most abundant echinoderm occurring therein. *Dorycrinus unicornis* and *C. melo* are also restricted to the BPAI, as well are the majority of crinoids and blastoids that occur in this association. The *D. unicornis*-*C. melo* Association also includes many rare genera that were carried over from underlying Kinderhookian strata, including species of *Belemnocrinus*, *Gilmocrinus*, *Holcocrinus*, *Nactocrinus*, *Megistocrinus* and *Paracosmetocrinus*.

Burlington Pelmatozoan Association II: The second association recognized in the study is referred to as the Burlington Pelmatozoan Association II (BPAII), and includes the *Uperocrinus pyriformis* and *Globoblastus norwoodi* Biozones. The biozones of the *U. pyriformis*-*G. norwoodi* Association include Rowley's (1908) Coral Horizon, the *Physetocrinus ventricosus* Zones of Laudon (1937) and Kaiser (1950), and the *Agaricocrinus planoconvexus* Zone of Laudon (1973). This association is roughly equivalent to the strata deposited in the Haight Creek Member of the Burlington Limestone. However, the transition from the BPAI to the BPAII may occur within the "White Ledge" or Laudon's (1937) *Cactocrinus proboscidialis* Zone of north central Missouri; strata considered to be equivalent to the Dolbee Creek Member of the Burlington Limestone. I observed a particularly fossil-rich exposure of the "White Ledge" in Hannibal, Missouri that contained the typical BPAI fauna throughout most of its thickness. However, the upper portion of this bed contained an abundance of BPAII forms such as *Agaricocrinus planoconvexus* (Hall) and *Physetocrinus ventricosus* (Hall) and a conspicuous absence of

BPAI forms such as *D. unicornis*, *C. melo*, and diagnostic species of *Abatocrinus* and *Cusacrinus*. This may suggest that the transition from the *D. unicornis*-*C. melo* Association to the *U. pyriformis*-*G. norwoodi* Association began within a single facies.

Uperocrinus pyriformis was chosen as the key index crinoid for this biozone even though *P. ventricosus* and *A. planoconvexus* can be equally abundant. *Physetocrinus ventricosus* was not retained as the namesake for this biozone because the stellate plates of this species may lead a non-specialist to confuse it with one of the many actinocrinitids that occur in the underlying BPAI. *Agaricocrinus planoconvexus* was not chosen because it is only abundant in the lower beds of the BPAII. Furthermore, there are several morphologically similar species of *Agaricocrinus* described from the Burlington Limestone and the taxonomy of this group is in need of revision before the Burlington *Agaricocrinus* species can be fully utilized as index fossils. *Globoblastus norwoodi* was chosen as the key index blastoid for this interval because it is the only abundantly occurring blastoid in this zone. Both *U. pyriformis* and *G. norwoodi* first occur in this zone and range into the uppermost strata of the Burlington Limestone; however, they reach their greatest abundance in the BPAII. Many other important crinoid and blastoid species reported as occurring in the upper Burlington first occur in this association including: *Actinocrinites scitulus* Miller and Gurley, *Cactocrinus glans* (Hall), *Macrocrinus koninki* (Shumard), *Strotocrinus glyptus* (Hall), and *Schizoblastus sayi* (Shumard). Interestingly, species that have been reported as occurring only in the lower Burlington Limestone also occur commonly in this association including *Cactocrinus obesus* (Keyes) and *Displodocrinus divergens* (Hall), again illustrating confusion surrounding the position of the lower-upper Burlington contact.

Burlington Pelmatozoan Association III: The stratigraphically youngest association recognized in this study is referred to as the Burlington Pelmatozoan Association III (BPAIII), and includes the *Macrocrinus verneuilianus* and *Pentremites elongatus* Biozones. The biozones of the *M. verneuilianus*-*P. elongatus* Association incorporate all of the zones discussed in this study that are referable to the Cedar Fork Member of the Burlington Limestone (Rowley, 1908; Laudon, 1937, 1973; Kaiser, 1952). *Macrocrinus verneuilianus* (Shumard) was chosen as the key index crinoid of this biozone rather than *Azygocrinus rotundus* (Yandell and Shumard) because (as discussed above) *A. rotundus* is only a dominant element of PBAIII in southeast Iowa and is rare elsewhere. Nevertheless, *A. rotundus* is an extremely useful index crinoid for this zone in southeast Iowa, occurring in densities as high as 100 individuals per m^2. *Macrocrinus verneuilianus* occurs frequently with *A. rotundus*, but is a more useful index fossil, as it is a common member of this association throughout the entire geographic extent of the Burlington Limestone. Laudon (1937, 1973) referred to the upper portion of the Cedar Fork Member as the *Dorycrinus quinquelobus* and the *Pentremites elongatus* Zones. *Dorycrinus quinquelobus* (Hall) is a diagnostic crinoid of the BPAIII; however, it is relatively uncommon. *Pentremites elongatus* (Shumard) was retained as the index fossil for the blastoid biozone as it is one of the most abundant blastoids in this association. The only other blastoid that that reaches equally high abundance (and frequently more so) is *Poroblastus granulosus* (Meek and Worthen). However, *P. granulosus* is only abundant in southwestern Missouri and rare elsewhere. The crinoids *Teliocrinus umbrosus* (Hall), *Uperocrinus nashvillae subtractus* (White), *Eutrochocrinus trochiscus* (Meek and Worthen), and the blastoid *Arcuoblastus shumardi* (Meek and Worthen) do not occur in the underlying biozones.

The fact that the biozones described above roughly mirror the Dolbee Creek, Haight Creek and Cedar Fork members of the Burlington Limestone is likely a function of fluctuations in sea level. Witzke et al. (1990) and Witzke and Bunker (1996) divided the Mississippian strata of Iowa into 10 third-order transgressive-regressive cycles of approximately one to three million years duration. The Dolbee Creek Member comprises Cycle 4 (or the Dolbee Creek Cycle) and the Haight Creek and Cedar Fork members are included in Cycles 5A and 5B, respectively (or the Haight Creek Cycle). The grainstones of the Dolbee Creek Cycle contain crinoids of the *Dorycrinus unicornis*-*Cryptoblastus melo* Association. The crinoids of this association are very distinct from either of the overlying associations; very few species carry over into the BPAII or the BPAIII. Conversely, the grainstones of the Haight Creek Cycle contain many of the same species. Many commonly occurring crinoids and blastoids of the *Uperocrinus* *pyriformis-Globocrinus norwoodi* Association extend into the *Macrocrinus verneuilianus-Pentremites elongatus* Association, but the latter contains several species that are constrained therein.

CRINOID AND BLASTOID BIODIVERSITY

Anyone who has collected crinoids and blastoids from the Burlington Limestone should be able to relate to the sentiments expressed by Rowley (1891, p. 71) who mentioned that "…the Burlington Limestone is, perhaps, the most interesting to the intelligent collector, not that its fossil treasures are more perfectly preserved or more abundant than individuals in the Keokuk or Chester divisions, but from the diversity of its Crinoidal remains and the great number of species of Echinoderms. The collector is always happening on something new, and his artistic eye is in constant rapture over the beautiful and ever changing sculpture of the calyx plates of the Actinocrinoids and the granular ornamentation of the Blastoids." Indeed, the Burlington Limestone contains the most speciose assemblage of crinoids and blastoids in the geologic record. Over 600 species of crinoids and blastoids have been described from the Burlington Limestone. Nevertheless, only about 400 species of crinoids and 30 species of blastoids are currently recognized as valid, and many of these are synonymous.

There are many reasons for redundant species descriptions in the Burlington Limestone. Early paleontologists did not work under a clear species concept and new species were introduced based on such minor characteristical deviations as differences in arm number, ornamentation, or interray plating. Moreover, many species were considered formation- or locality-specific. This led several workers into the pitfall of circular reasoning, which is in part why Niles and Wachsmuth (1866, p.4-5) proclaimed that, "We have examined the species of Crinoids and noticed their stratigraphical distribution with care, and have found no evidence of any species occurring in both the Lower and Upper Burlington limestones." Taphonomic process also resulted in the naming of redundant species because compressed or otherwise distorted material appears different from perfect specimens. In addition, many species were described from incomplete material, such as the basal circlets of *Platycrinites*. Several of the earliest species descriptions of Burlington crinoids are incomplete or enigmatic. This, coupled with the fact that many of these species were never figured, resulted in the redundant description of several crinoids and blastoids. Wachsmuth and Springer (1897, p. 19) certainly expressed the sentiments of many paleontologists when they wrote, "These descriptions, in many cases, were so indefinite that the identification of the species was almost impossible, and this created considerable annoyance and labor to later writers." Another matter of concern is that many "new species" were described from personal collections and are now lost. Similarly, several holotypes were destroyed. For example, Wachsmuth and Springer (1897, p. 5) reported that "McChesney's types were all destroyed in the great Chicago fire." They also believed that all of Owen and Shumard's types were destroyed in a fire in Burlington or Keokuk, Iowa; however, Springer (1920, p. 7) discovered that these were "rescued from a rubbish barrel at the old David Dale Owen headquarters in New Harmony, Indiana," and are now reposited (at the Field Museum of Natural History) in Chicago. Unfortunately, ego also played a role in the current state of Burlington crinoid and blastoid taxonomy. Wachsmuth and Springer (1897, p. 652-653) clearly addressed this issue when they said that, "The earlier authors…may readily be excused for describing their species from such material as they had. But at the present day the only excuse for this class of work that can be found is the desire of the authors to see their names appended to the greatest possible number of species… All we have in many cases is the assurance of the author that the species is so unlike any other that a comparison is unnecessary. We have found in practice that a declaration of this kind is a badge of suspicion, and is one of the most common indications of a synonym."

Many crinoid clades in the Burlington are in dire need of taxonomic "housekeeping". Wachsmuth and Springer (1897) presented an excellent summary of the camerate crinoids, and Springer (1920) treated most of the flexible crinoids. Nevertheless, the cladids of the Burlington Limestone are poorly understood. Kirk (1938, 1940, 1941, 1943b, 1945, 1947) described many new genera and species of cladids from the Burlington Limestone, but left many unresolved problems. Recently, Gahn and Kammer (2002) and Kammer and Gahn (2003) have revised the non-pinnulate cladids from the Burlington Limestone, and Kammer (in prep) is currently revising the pinnulate cladids.

Although several of the described Burlington species are synonymous, new species continue to be found and described from old museum holdings and new field collections. I have attempted to compile a table comprising the currently recognized crinoids and blastoids from the Burlington Limestone. This list was compiled from Webster's (1973, 1977, 1986, 1988, 1993) "Bibliography and index of Paleozoic crinoids," and includes the originally reported (lower and upper Burlington) range and relative frequency of each species according to their distribution in the Burlington pelmatozoan assemblages. This table should be considered as a working draft, but aims to be a useful summary of Burlington pelmatozoan taxonomy and distribution. I have cursorily examined most of the original species descriptions and culled about 100 species from the original list of nearly 430. I eliminated species that are probable synonyms; however, it is very likely that additional synonyms remain in the list. Furthermore, I may have been overzealous in my efforts, and there may be species that may need to be reinstated. Many of the species that were culled include those (approximately 50) described by Miller and Gurley from 1893-1897. Kirk (1943a, p. 264) explained that Miller and Gurley "described every specimen they could get their hands on—good bad or indifferent. As was well known to their contemporaries, the main purpose was to forestall the work of Wachsmuth and Springer." I have examined all of the species descriptions and figures published by Miller and Gurley from this period, and the great majority is assignable to previously described and common representatives of the Burlington fauna. In fact, it is unlikely that any of their Burlington crinoid species are valid. As such, I have even-handedly, but tentatively eliminated every species described by Miller and Gurley from the table presented herein.

Despite the problems associated with the taxonomy of Burlington pelmatozoans, the Burlington Limestone constitutes an incredibly diverse and evolutionarily important fauna. This is well illustrated by the fact that the majority of the underlying Kinderhookian faunas are represented by fewer than 50 species of crinoids and a handful of blastoids (Laudon, 1933; Laudon and Beane, 1937; Peck and Keyte, 1938). The Burlington Limestone, with approximately 300 crinoids and 25 blastoids represents a six-fold increase in diversity over a few million years. Monobathrid camerate crinoid families such as the Actinocrinitidae, Batocrinidae, and Platycrinitidae underwent incredible morphological diversification on the Burlington Shelf (Lane, 1978). Moreover, flexibles, which are typically represented by only a few species in any given formation, were represented by 8 genera and nearly 20 species in the Burlington Limestone. The same can be said for the blastoids that are represented by 15 genera and approximately 25 species. The Burlington cladid fauna is transitional between the Kinderhookian and Late Osagean faunas and shares genera with each. The evolutionary importance of the cladid faunas will be better understood following publication of Kammer's (in prep) current research on the group. The cladid genera *Barycrinus* and *Cyathocrinites* underwent considerable diversification on the Burlington Shelf, and phylogenetic studies of these genera by Gahn and Kammer (2002) and Kammer and Gahn (2003) suggest that many of these species originated from single, abundant, geographically widespread, and geologically long-ranging species such as *Barycrinus rhombiferus* (Owen and Shumard) and *Cyathocrinites iowensis* (Owen and Shumard). Gahn (in prep) will demonstrate that the majority of these species arose through anagenesis (wherein a known ancestor evolves directly into a new species) and budding (wherein a known ancestor persists and gives rise to a new species), rather than through cladogenesis (wherein an ancestor "bifurcates" or gives rise to two new species).

The high species diversity of crinoids and blastoids in the Burlington Limestone is partially attributable to a fierce collecting effort from the 1850's to the present and extensive exposures throughout southeast Iowa, west-central Illinois, Missouri, and northwestern Arkansas. Nevertheless, the regional encrinites that define the Burlington Limestone promoted incredible pelmatozoan diversity (Ausich 1997, 1999). The carbonate grains deposited in the Burlington Limestone were generated almost entirely by the disarticulation of fossil crinoids and blastoids. These echinoderms were essentially living on a mobile substrate generated by their forbearers. Evidence that these sediments were unconsolidated is provided by the crinoids themselves; well-articulated crinoid crowns are often found buried by coarse crinoidal grainstone. Furthermore, the graded and low-angle cross-stratified crinoidal limestones are indicative of storm-generated sedimentary processes that transported the mobile disarticulated remains of these echinoderms. Many Burlington grainstones represent amalgamated storm beds, and as a result, only the most taphonomically resilient components of the pelmatozoan fauna are typically preserved. The calyxes of monobathrid camerate crinoids and blastoid thecae are among the most taphonomically robust skeletal constructions, and their abundance relative to other stalked echinoderm groups is likely inflated by taphonomic processes.

Crinoid and blastoid abundance and diversity was probably enhanced through the positive taphonomic feedback generated by the unconsolidated echinoderm bioclasts. Brachiopods, mollusks, bryozoans, corals, and other sessile marine invertebrates were present in the Burlington Limestone, but their diversity and abundance pale in comparison to the pelmatozoan echinoderms. The mobile sediments produced by the crinoids and blastoids likely inhibited the successful proliferation of many other fixed invertebrates, while at the same time producing a suitable substrate for their own attachment. Crinoids and blastoids used a variety of attachment strategies to adapt to the mobile substrates of the Burlington Limestone. Some species had extremely robust holdfasts with extensive radicular cirri, while others had a distally tapering holdfast that sat freely on the substrate. Many of the diplobathrid camerates had prehensile distal stalks that could be used to wrap around the stalks of other crinoids. Furthermore, small encrusting holdfasts have been observed on large distal columnals of *Platycrinites* and the taphonomically resistant calyx rim of *Strotocrinus glyptus*. Crinoids and blastoids also served as hosts to other invertebrates. For example, *Tremichnus* borings are commonly found on the plates of the calyx and stalk of these echinoderms. These are particularly abundant in genera with very large and/or stellate plates such as those found in the Actinocrinitidae, Dichocrinidae, and Platycrinitidae. Moreover, platyceratid gastropods, or diagnostic scars and boreholes from the same, have been found on several genera of Burlington pelmatozoans. Other echinoderms have even been observed using crinoids as hosts, such as ophiuroids on the genus *Actinocrinites* (personal collection of Karl Stuekerjuergen).

Resource partitioning was another factor in the generation of pelmatozoan diversity on the Burlington Shelf, and may explain the much higher diversity attained by crinoids than blastoids. Fossil crinoids partitioned food resources through modifications of their feeding filtration fan and by differences in stalk length (Ausich, 1980). The food gathering morphology of crinoids is incredibly diverse, whereas that of the blastoids is more generalized. It is possible that the monomorphic feeding construction of the blastoids prevented them from diversifying to the same extent as the crinoids. Nevertheless, a greater proportion of blastoid species (than crinoid species) reached high levels of abundance. In fact, blastoids are numerically superior to crinoids at most localities in the Burlington Limestone, but their typically smaller size makes them less conspicuous.

Parasitism and predation were also likely factors in the morphological diversification of crinoids during the deposition of the Burlington Limestone through processes such as evolutionary escalation (Vermeij, 1987). Platyceratid gastropods are often found positioned over the anal opening of fossil crinoids, and have typically been interpreted as commensals that fed on crinoid excrement (Bowsher, 1955). However, a few studies have proffered evidence suggesting that at least some platyceratids were parasitic (Rollins and Brezinski, 1988; Baumiller and Gahn, 2002a; Gahn and Baumiller, 2002c). If these gastropods were detrimental to their hosts, then natural selection would favor those crinoids with parasite-resistant morphological features. Gahn and Baumiller (2001) demonstrated that crinoids with long anal tubes were less frequently infested by platyceratid gastropods than crinoids that lacked them. They also demonstrated that anal tubes evolved several times within the *Compsocrinia* from parasitized, tubeless ancestors. These studies suggest that parasitism by platyceratid gastropods may have influenced the morphological diversification of fossil crinoids. Crinoids that are known to be infested by platyceratid gastropods in the Burlington Limestone include species of *Actinocrinites*, *Aryballocrinus*, *Cusacrinus*, *Dorycrinus*, *Eucladocrinus*, *Gilbertsocrinus*, *Physetocrinus*, *Platycrinites*, and *Strotocrinus*.

Predation may have provided another extrinsic evolutionary influence on crinoids and the teeth of durophagous or shell-crushing sharks are common in the Burlington Limestone, especially in the uppermost strata of the formation. Although predator-prey interactions are difficult to document in the fossil record, shark coprolites containing abundant remains of fossil crinoids have been reported (Laudon, 1957). More common are damaged and regenerated arms, spines, and calyxes of fossil crinoids. Damage and regeneration in Recent crinoids has often been attributed to predation; indeed, Meyer et al. (1984) observed crinoid arms dangling from the mouth of a saddled coralfish over the Great Barrier Reef, Australia. Similarly, regeneration patterns in fossil crinoids appear to be best explained by predation (Meyer and Ausich, 1983). Laudon (1957) suggested that the abundance of crinoid stalk material and paucity of skeletal material representing the crowns of fossil crinoids in the Burlington Limestone indicated that shell-crushing sharks utilized crinoids as an important food source and essentially grazed over vast "crinoidal gardens." Signor and Brett (1984) demonstrated a coincident diversification in Paleozoic durophagous predators and an increase in the spinosity and plate thickness of fossil crinoids. They argued that predation on crinoids in the middle Paleozoic may have been sufficient to drive morphological change and evolutionary innovations in crinoids that would facilitate predator avoidance. Gahn and Baumiller (2002a, 2002b) have recently provided evidence suggesting that regeneration frequencies may have been higher in the Paleozoic than previously recognized; reporting regeneration frequencies as high as 25% for Mississippian crinoids. This provides further support to claims that predation was a significant factor in the evolutionary development of Paleozoic crinoids. Many genera and species of crinoids in the Burlington Limestone have robust spines on the tegmen (e.g. *Dorycrinus* and *Displodocrinus*), dorsal cup (e.g. *Gilberstocrinus* and *Wachsmuthicrinus*), anal tube (e.g. *Uperocrinus*), or anal sac (e.g. *Coeliocrinus* and *Pelecocrinus*) that may represent independently derived, anti-predatory characters. The development of broad medial calyx rims (e.g. *Eutrochocrinus* and *Strotocrinus*), dorso-ventrally flattened calyxes (e.g. *Agaricocrinus* and *Plemnocrinus*), defectively pinnulate and paddle-shaped distal arm brachials (e.g. *Cusacrinus* and *Eretmocrinus*), and either very large or very small body size may have also assisted in predator-avoidance or damage reduction. Broad medial calyx rims and paddle-shaped distal arms may have helped the crinoids with these traits avoid predation by making them appear larger than they actually were, similar to the predator-avoidance strategy of Australia's frilled lizard. Defectively pinnulate arms, or those that lack pinnules distally, evolved independently at least twice in camerate crinoids. The non-pinnulate arms of these crinoids may have permitted the loss of a substantial portion of the arm without a great loss in feeding efficacy. Dorso-ventral flattening and reduction of the visceral mass of the calyx may have reduced the probability of lethal predatory attacks by increasing the probability of the arms being damaged rather than regions that are more vital. Although entertaining anti-predatory hypotheses for these structures is engaging, they prove difficult, if not impossible to test. Coincidentally, many of the most spinose and seemingly best-defended crinoid genera are present in the *Macrocrinus verneuilianus-Pentremites elongatus* Association, which also contains the greatest concentration and diversity of shell-crushing shark remains. Many of these well-defended camerates abruptly declined at the end of the Osagean and became entirely extinct by the Meramecian. Waters and Maples (1991) suggested that the diminished dominance of this clade was caused by predator-mediated community reorganization.

CONCLUSIONS

The Burlington Limestone is renowned for incredible crinoid and blastoid diversity. However, not all of these species lived contemporaneously. At least three faunal associations can be distinguished in the Burlington Limestone and appear to coincide with significant fluctuations in sea level. The crinoids and blastoids of the *Dorycrinus unicornis-Cryptoblastus melo* Association are mostly restricted to the crinoidal grainstones of the Dolbee Creek Cycle, and the overlying pelmatozoan associations are restricted to the Haight Creek Cycle. The stark difference in faunal composition between the Dolbee Creek and Haight Creek Cycles and the similarity shared by the *Uperocrinus pyriformis-Globocrinus norwoodi* and *Macrocrinus verneuilianus-Pentremites elongatus* Associations is consistent with a sea-level fluctuation of greater magnitude occurring above the Dolbee Creek Member than in the "middle grainstone" of the Haight Creek Member. Whether the species of these associations were able to track the encrinites (and continue evolving in "greener pastures") during intervals of sea-level change, tolerate or adapt to the flooding of the carbonate shelf, or went extinct is unclear and requires further study. However, it is apparent that the associations recognized herein can be traced over the expanse of the Burlington shelf and perhaps beyond. The crinoid and blastoid faunas of the Lake Valley (New Mexico) and Redwall (Arizona) Formations are strikingly similar to those of the *Dorycrinus unicornis-Cryptoblastus melo* Association of the Burlington Shelf (Brower, 1970; Macurda, 1970), and similarities between the crinoid and blastoid fauna of the Nada Member of the Borden Formation (Kentucky) and the *Uperocrinus pyriformis-Globoblastus norwoodi* Association are incredible (Lane and DuBar, 1983). The observation that many of the species that characterize these pelmatozoan associations extend well beyond the Burlington Shelf and occur in a myriad of facies suggests that the associations are not confined to a single environment. Therefore, it seems likely that many of the common and widely distributed species of Burlington crinoids should persist though facies changes in the Burlington Limestone, unless these changes represented rapid and drastic changes in sea level. The integrity of the echinoderm associations recognized herein over hundreds, if not thousands of miles suggests the presence of extensive epicontinental seas that were relatively free of physical and oceanographic barriers.

A more detailed analysis of species-level taxonomy and spatiotemporal distribution of Burlington crinoids and blastoids and those of coeval formations may yield insight into spatiotemporal morphological variation and endemism. Such information would be beneficial to the understanding of ancestor-descendant relationships and evolutionary processes acting upon the crinoids and blastoids during this pivotal interval of diversification. This is not an unrealistic task considering the abundance of exposures and echinoderms in the Burlington Limestone. However, if such a goal is to be met, then amateur and professional paleontologists alike must develop a clear understanding of the stratigraphic and taxonomic complexities (or simplicities, if you prefer) of the Burlington Limestone and keep this information with the specimens they collect. I certainly hope that this paper will be a helpful step in such an endeavor.

The three pelmatozoan assemblages defined in this study should be easy to recognize in the field over the entire extent of the geologic distribution of the Burlington Limestone, but they should not be used exclusively. The zones of Rowley (1908), Laudon (1937, 1973) and Kaiser (1950) can be recognized and be very useful at local scales. The best data of course, would be exact positional measurements of specimens from a diagnostic stratigraphic marker bed. However, I am aware that many Burlington fossils are collected as float and can only be traced back to a more generalized biozone. Regardless of what zonation scheme is chosen, I strongly encourage discontinuing the use of the "lower" and "upper" Burlington in reference to anything other than historical discussions of the Burlington Limestone and in reporting the stratigraphic occurrence of specimens from old collections.

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Table 1. Pelmatozoan echinoderms of the Burlington Limestone. A culled listing of currently recognized crinoids and blastoids of the Burlington Limestone, including the author of each species and reported lower vs. upper Burlington occurrences (lB = lower Burlington, uB = upper Burlington, Bu = Burlington undifferentiated). Refer to Webster (1973, 1977, 1986, 1988, 1993) for the citations listed in the table. The distribution and relative frequency of all species is also given for each Burlington pelmatozoan association (except for those that are unknown). Please refer to the text and Figure 2 for an explanation of these associations. This table should be used cautiously as it likely requires substantial revisions. However, it should be useful as a general guide to the crinoids and blastoids of the Burlington Limestone. The frequencies are indicated and defined as follows: (a) abundant – species that are extremely numerous at some localities, but only common at others; (c) common – species that are represented at almost every outcrop visited; (u) uncommon – species that are found only after considerable collecting effort; (r) rare – species that are only represented by very few species in all available collections.

Pelmatozoan Echinoderms of the Burlington Limestone

#	CRINIODS	Author	Division	ASSOCIATION		
	Monobathrids			**I**	**II**	**III**
1	*Aacocrinus arrosus*	(Miller, 1892)	Bu	r		
2	*Abatocrinus aequalis*	(Hall, 1858)	lB	c		
3	*A. calvini*	(Rowley, 1890)	lB	u		
4	*A. clypeatus*	(Hall, 1859)	lB	c		
5	*A. curiosus*	(Rowley, 1908)	lB	r		
6	*A. laura*	(Hall, 1861)	uB		r	u
7	*A. lepidus*	(Hall, 1859)	lB	u		
8	*A. pistillus*	(Meek and Worthen, 1865)	uB		r	u
9	*A. rotadentatus*	(Rowley and Hare, 1891)	lB	r		
10	*A. tuberculatus*	(Wachsmuth and Springer, 1897)	lB	r		
11	*A. turbinatus*	(Hall, 1858)	lB	c		
12	*Actinocrinites eximimus*	(Kirk, 1943)	uB			u
13	*A. multiradiatus*	(Shumard, 1857)	uB			c
14	*A. probolos*	Ausich and Kammer, 1991	uB			r
15	*A. scitulus*	Meek and Worthen, 1860	uB		c	r
16	*A. verrucosus*	(Hall, 1858)	uB			c
17	*Agaricocrinus bellatrema*	Hall, 1861	uB			r
18	*A. bellatrema major*	Wachsmuth and Springer, 1897	uB			r
19	*A. brevis*	(Hall, 1858)	lB	c		
20	*A. bullatus*	(Hall, 1858)	uB		c	c
21	*A. convexus*	(Hall, 1859)	uB		u	
22	*A. excavatus*	(Hall, 1861)	uB			u
23	*A. gracilis*	Meek and Worthen, 1861	uB		u	
24	*A. inflatus*	Hall, 1861	uB		u	r
25	*A. louisianensis*	Rowley, 1900	lB	r		
26	*A. nodosus*	Meek and Worthen, 1869	uB			c
27	*A. planoconvexus*	Hall, 1861	lB	r	c	
28	*A. pyramidatus*	(Hall, 1858)	lB	r		

29	*A. stellatus*	(Hall, 1858)	lB		u	c
30	*Ancalocrinus spinobrachiatus*	(Hall, 1859)	lB	u	u	
31	*Aorocrinus canaliculatus*	(Meek and Worthen, 1869)	lB	r		
32	*A. subaculeatus*	(Hall, 1858)	lB	r		
33	*A. symmetricus*	(Hall, 1858)	Bu		c	c
34	*A. wachsmuthi*	Rowley, 1901	lB	r		
35	*Aryballocrinus tenuidiscus*	(Hall, 1861)	lB	r		
36	*A. whitei*	(Hall, 1861)	lB-uB	u	u	u
37	*Auliskocrinus crassitestus*	(White, 1862)	uB			r
38	*Azygocrinus andrewsianus*	(McChesney, 1860)	uB			u
39	*A. dodecadactylus*	(Meek and Worthen, 1861)	uB			u
40	*A. rotundus*	(Yandell and Shumard, 1855)	uB			a
41	*Cactocrinus clarus*	(Hall, 1861)	lB	u		
42	*C. extensus*	Wachsmuth and Springer, 1897	lB	r		
43	*C. glans*	(Hall, 1859)	uB		c	u
44	*C. multibrachiatus*	(Hall, 1858)	lB	c		
45	*C. obesus*	(Keyes, 1894)	lB		c	
46	*C. opusculus*	(Hall, 1859)	lB	u		
47	*C. proboscidialis*	(Hall, 1858)	lB	c		
48	*C. reticulatus*	(Hall, 1861)	lB	u		
49	*C. sexarmatus*	(Hall, 1859)	lB	r		
50	*C. thalia*	(Hall, 1861)	IB	u		
51	*Camptocrinus praenuntius*	Springer, 1926	uB		u	r
52	*Coelocrinus concavus*	(Meek and Worthen, 1861)	uB			r
53	*Cusacrinus asperrimus*	(Meek and Worthen, 1869)	IB	r		
54	*C. chloris*	(Hall, 1861)	lB	r		
55	*C. coelatus*	(Hall, 1858)	lB	u		
56	*C. denticulatus*	(Hall, 1863)	IB	r		
57	*C. ectypus*	(Meek and Worthen, 1869)	IB	r		
58	*C. gracilis*	(Wachsmuth and Springer, 1897)	IB	r		
59	*C. limabrachiatus*	(Hall, 1861)	IB	r		
60	*C. longus*	(Meek and Worthen, 1869)	lB	r		
61	*C. penicillus*	(Meek and Worthen, 1869)	IB	r		
62	*C. spinotentaculus*	(Hall, 1859)	lB	r		
63	*C. tenuisculptus*	(McChesney, 1860)	lB	u		
64	*C. thetis*	(Hall, 1861)	lB	r		
65	*C. tuberculosus*	(Wachsmuth and Springer, 1897)	uB		r	
66	*Cytidocrinus sculptus*	(Hall, 1858)	lB	u	u	u
67	*Dichocrinus conus*	Meek and Worthen, 1860	lB-uB	u	u	u
68	*D. gracilis*	Broadhead, 1981	uB			r
69	*D. lachrymosus*	Hall, 1859	uB			r
70	*D. laevis*	Hall, 1859	lB	r		
71	*D. pocillum*	Hall, 1861	uB			u
72	*Displodocrinus divergens*	(Hall, 1859)	lB	u	u	
73	*Dorycrinus cornigerus*	(Hall, 1858)	uB		c	c
74	*D. missouriensis*	(Shumard, 1855)	uB		u	r
75	*D. pentagonus*	Rowley, 1900	uB		r	
76	*D. quinquelobus*	(Hall, 1859)	uB			u

77	*D. roemeri*	Meek and Worthen, 1860	uB			r
78	*D. subturbinatus*	(Meek and Worthen, 1860)	lB	r		
79	*D. unicornis*	(Owen and Shumard, 1850)	lB	a		
80	*D. unispinus*	(Hall, 1861)	lB	r		
81	*Eretmocinus brevis*	Rowley, 1902	uB		r	
82	*E. calyculoides*	(Hall, 1860)	uB		u	u
83	*E. calyculoides nodosus*	Wachsmuth and Springer, 1897	uB			r
84	*E. clio*	(Hall, 1861)	lB	r		
85	*E. cloelia*	(Hall, 1861)	uB	r		
86	*E. corbulis*	(Hall, 1861)	lB	u		
87	*E. coronatus*	(Hall, 1859)	lB	r		
88	*E. depressus*	Keyes, 1894	uB			u
89	*E. expansus*	Keyes, 1894	lB	r		
90	*E. leucosia*	(Hall, 1861)	lB	u		
91	*E. matutus*	(Hall, 1861)	uB		u	
92	*E. minor*	Wachsmuth and Springer, 1897	uB		r	
93	*E. neglectus*	(Meek and Worthen, 1868)	lB	u		
94	*E. rugosus*	Wachsmuth and Springer, 1897	lB	r		
95	*Eucladocrinus pleurovimenus*	(White, 1862)	uB			u
96	*E. praenuntius*	(Wachsmuth and Springer, 1878)	uB			u
97	*Eutrochocrinus christyi*	(Shumard, 1855)	uB		u	c
98	*E. lovei*	(Wachsmuth and Springer, 1881	uB		r	r
99	*E. trochiscus*	(Meek and Worthen, 1868)	uB			u
100	*Macrocrinus gemmiformis*	(Hall, 1859)	lB	r		
101	*M. konincki*	(Shumard, 1855)	uB		c	u
102	*M. verneuilianus*	(Shumard, 1855)	uB		u	a
103	*Megistocrinus evansii*	(Owen and Shumard, 1850)	lB	u	r	r
104	*M. evansii crassus*	White, 1862	lB	r		
105	*Nunnacrinus locellus*	(Hall, 1861)	lB	u		
106	*N. puteatus*	(Rowley and Hare, 1891)	lB	r		
107	*Paradichocrinus liratus*	(Hall, 1861)	uB			u
108	*Physetocrinus asper*	(Meek and Worthen, 1869)	uB		r	r
109	*P. dilatatus*	(Meek and Worthen, 1869)	uB		r	r
110	*P. ornatus*	(Hall, 1858)	lB	u		
111	*P. ventricosus*	(Hall, 1858)	lB-uB	r	a	u
112	*Platycrinites americanus*	(Owen and Shumard, 1852)	lB	c	u	
113	*P. aqualis*	(Hall, 1861)	uB			u
114	*P. asper*	(Meek and Worthen, 1861)	uB			r
115	*P. burlingtonensis*	(Owen and Shumard, 1850)	lB	c		
116	*P. brevinodus*	(Hall, 1861)	lB-uB	u	u	r
117	*P. corbuliformis*	(Rowley and Hare, 1891)	lB	r		
118	*P. davisi*	(Wachsmuth and Springer, 1897)	lB	r		
119	*P. discoideus*	(Owen and Shumard, 1850)	IB	c		
120	*P. excavatus*	(Hall, 1861)	uB	u	u	
121	*P. geometricus*	(Wachsmuth and Springer, 1897)	uB		r	r
122	*P. glyptus*	(Hall, 1861)	uB		r	u
123	*P. nodostriatus*	(Wachsmuth and Springer, 1897)	lB-uB	r	u	
124	*P. ornogranulus*	(McChesney, 1860)	lB	c		

125	*P. pocilliformis*	Hall (1858)	lB	c	u	
126	*P. parvinodus*	(Hall, 1861)	lB	r		
127	*P. planus*	(Owen and Shumard, 1850)	lB	c	u	
128	*P. regalis*	(Hall, 1861)	lB	r		
129	*P. saffordi*	(Hall, 1858)	lB			u
130	*P. scobina*	(Meek and Worthen, 1861)	lB	u		
131	*P. sculptus*	(Hall, 1858)	lB	u	u	
132	*P. spinifer*	(Wachsmuth and Springer, 1897)	lB	r		
133	*P. spinifer elongatus*	(Wachsmuth and Springer, 1897)	lB	r		
134	*P. subspinulosus*	(Hall, 1859)	uB		u	
135	*P. trunculatus*	(Hall, 1858)	lB	u		
136	*P. verrucosus*	(White, 1865)	lB	u		
137	*P. wortheni*	(Hall, 1858)	lB	r		
138	*P. yandelli*	(Owen and Shumard, 1850)	lB	r		
139	*P. yandelli perasper*	(Meek and Worthen, 1865)	lB	r		
140	*Plemnocrinus beebei*	Kirk, 1946	uB		u	r
141	*P. homalus*	Kirk, 1946	lB			r
142	*P. occidentalis*	(Miller, 1891)	Bu			r
143	*P. subspinosus*	(Hall, 1858)	lB-uB	u	u	u
144	*P. tuberosus*	(Hall, 1858)	uB			r
145	*P. eminulus*	(Hall, 1861)	lB	r		
146	*Pleurocrinus halli*	(Shumard, 1866)	uB			r
147	*P. incomptus*	(White, 1863)	uB		u	r
148	*P. pileiformis*	(Hall, 1858)	lB	u		
149	*P. quinquenodus*	(White, 1862)	uB		r	r
150	*Springeracrocrinus praecursor*	(Springer, 1926)	uB		r	r
151	*Steganocrinus burlingtonensis*	Brower, 1965	uB			r
152	*S. concinnus*	(Shumard, 1855)	uB		r	r
153	*S. elongatus*	Kirk, 1943	uB		c	u
154	*S. multistriatus*	Brower, 1965	Bu			r
155	*S. pentagonus*	(Hall, 1858)	lB	c	u	r
156	*S. planus*	Brower, 1965	uB			r
157	*S. robustus*	Brower, 1965	uB		u	
158	*S. validus*	(Meek and Worthen, 1860)	uB			r
159	*Strimplecrinus ovatus*	(Owen and Shumard, 1850)	lB	r		
160	*S. pendens*	(Wachsmuth and Springer, 1897)	uB		r	r
161	*S. pisum*	(Meek and Worthen, 1869)	lB	u		
162	*S. plicatus*	(Hall, 1861)	uB			r
163	*S. striatus*	(Owen and Shumard, 1850)	lB-uB		u	c
164	*Strotocrinus glyptus*	(Hall, 1860)	uB		u	u
165	*Teliocrinus adolescens*	Wachsmuth and Springer, 1897	lB-uB	r	r	r
166	*T. liratus*	(Hall, 1859)	uB			r
167	*T. umbrosus*	(Hall, 1858)	uB			c
168	*Uperocrinus aequibrachiatus*	(McChesney, 1860)	uB		c	u
169	*U. aequibrachiatus astericus*	(Meek and Worthen, 1860)	uB		c	u
170	*U. hageri*	(McChesney, 1860)	uB		r	c
171	*U. inflatus*	(Rowley and Hare, 1891)	lB	c		
172	*U. longirostris*	(Hall, 1858)	lB	c		

173	*U. nashvillae subtractus*	(White, 1862)	uB			u
174	*U. pyriformis*	(Shumard, 1855)	uB		a	c
	Diplobathrids					
1	*Cribanocrinus wachsmuthi*	(Hall, 1861)	lB	r		
2	*C. whitei*	(Hall, 1861)	lB	r		
3	*C. wortheni*	(Hall, 1858)	IB	r		
4	*Gilbertsocrinus fiscellus*	(Meek and Worthen, 1860)	lB	r		
5	*G. obovatus*	Meek and Worthen, 1869	uB			r
6	*G. tuberculosus*	(Hall, 1859)	uB		r	r
7	*G. typus*	(Hall, 1859)	lB-uB	r	u	u
8	*Rhodocrinites barrisi*	(Hall, 1861)	uB		u	
9	*R. barrisi striatus*	Wachsmuth and Springer, 1897	ub		r	
10	*R. truncatus*	(Wachsmuth and Springer, 1897)	uB		r	
	Disparids					
1	*Catillocrinus wachsmuthi*	(Meek and Worthen, 1866)	uB			r
2	*Halysiocrinus dactylus*	(Hall, 1860)	lB-uB	u	u	u
3	*Synbathocrinus dentatus*	Owen and Shumard, 1852	uB		c	c
4	*S. papillatus*	Hall, 1861	Bu			
5	*S. wachsmuthi*	Meek and Worthen, 1869	uB		u	u
6	*S. wortheni*	Hall, 1858	uB			c
	Cladids					
1	*Abrotocrinus cf. A. unicus*	(Hall, 1861)	uB			r
2	*Acylocrinus striatus*	(Meek and Worthen, 1869)	lB	r		
3	*A. tortuosus*	(Hall, 1861)	Bu			
4	*A. tumidus*	Kirk, 1947	lB	r		
5	*Aphelecrinus delicatus*	(Meek and Worthen, 1869)	lB-uB	r	r	r
6	*A. meeki*	(Kirk, 1941)	lB	R		
7	*Ascetocrinus rusticellus*	(White, 1863)	uB			r
8	*A. scoparius*	(Hall, 1861)	lB	r		
9	*A. whitei*	(Hall, 1861)	lB	r		
10	*Atelestocrinus delicatus*	Wachsmuth and Springer, 1886	lB	r		
11	*A. robustus*	Wachsmuth and Springer, 1885	lB	r		
12	*Barycrinus crassibrachiatus*	(Hall, 1860)	uB		u	u
13	*B. magister*	(Hall, 1858)	uB			r
14	*B. rhombiferus*	(Owen and Shumard, 1852)	lB-uB	r	u	c
15	*B. sampsoni*	(Miller and Gurley, 1896)	lB	r		
16	*B. scitulus*	(Meek and Worthen, 1860)	lB	r		
17	*B. spurius*	(Hall, 1858)	lB-uB	r	u	u
18	*Belemnocrinus pourtalesi*	Wachsmuth and Springer, 1877	lB	r		
19	*B. typus*	White, 1862	lB	r	r	
20	*Blothrocrinus cultidactylus*	(Hall, 1859)	lB-uB	r	r	
21	*B. swallovi*	(Meek and Worthen, 1860)	uB			r
22	*Bursacrinus confirmatus*	White, 1862	lB	r		
23	*B. wachsmuthi*	Meek and Worthen, 1861	uB			r
24	*Cercidocrinus bursaeformis*	(White, 1862)	lB	r		

25	*Coeliocrinus dilatatus*	(Hall, 1861)	lB	r		
26	*C. subspinosus*	White, 1863	uB			r
27	*C. ventricosus*	(Hall, 1861)	lB-uB	r	u	u
28	*Corythocrinus tenuis*	Kirk, 1946	uB			r
29	*Costalocrinus cornutus*	(Owen and Shumard, 1850)	lB-uB	r	u	u
30	*Cyathocrinites barrisi*	(Hall, 1861)	lB	r		
31	*C. barydactylus*	(Wachsmuth and Springer, 1878)	uB			r
32	*C. deroseari*	Kammer and Gahn, 2003	lB	r		
33	*C. gilesi*	(Wachsmuth and Springer, 1878)	uB			r
34	*C. iowensis*	(Owen and Shumard, 1850)	lB-uB	c	c	c
35	*C. kelloggi*	(White, 1862)	lB-uB	r	r	r
36	*C. lamellosus*	(White, 1863)	uB			r
37	*C. rigidus*	(White, 1865)	lB	r		
38	*C. sampsoni*	(Miller, 1891)	lB	r		
39	*"Cyathocrinites" formosus*	(Rowley, 1905)	lB	r		
40	*Cydrocrinus robbi*	(Roy, 1929)	Bu			
41	*Decadocrinus scalaris*	(Meek and Worthen, 1869)	uB		r	r
42	*Eratocrinus elegans*	(Hall, 1858)	lB-uB	u	u	c
43	*E. ramosus*	(Hall, 1858)	uB			r
44	*Gilmocrinus cf. G. oneali*	Laudon and Beane, 1937	lB	r		
45	*Goniocrinus incipiens*	(Hall, 1861)	lB	r		
46	*Graphiocrinus simplex*	(Hall, 1858)	uB			r
47	*G. spinobrachiatus*	Hall, 1861	uB			r
48	*G. subimpressus*	(Meek and Worthen, 1861)	lB	r		
49	*G. whitei*	(Meek and Worthen, 1869)	uB			r
50	*Histocrinus juvenis*	(Meek and Worthen, 1869)	lB	r		
51	*Holcocrinus spinobrachiatus*	(Hall, 1861)	lB	r		r
52	*H. wachsmuthi*	(Meek and Worthen, 1861)	lB	r		
53	*Hypselocrinus calyculus*	(Hall, 1858)	Bu			
54	*H. fusiformis*	(Hall, 1861)	Bu			
55	*H. macrodactylus*	(Meek and Worthen, 1869)	lB-uB	c	c	c
56	*H. tethys*	(Meek and Worthen, 1869)	lB-uB	r	r	r
57	*Lanecrinus halli*	(Hall, 1861)	uB			u
58	*Linocrinus asper*	(Meek and Worthen, 1869)	lB	u		
59	*L. penicillus*	(Meek and Worthen, 1869)	lB-uB	u	c	c
60	*L. perangulatus*	(White, 1862)	uB			c
61	*L. scobina*	(Meek and Worthen, 1869)	uB		u	u
62	*Nactocrinus antiquus*	(Meek and Worthen, 1869)	lB	r		r
63	*N. nitidus*	Kirk, 1947	lB	r		
64	*Pachylocrinus carinatus*	(Hall, 1861)	uB			r
65	*P. clio*	(Meek and Worthen, 1869)	uB			r
66	*P. cuneatus*	(Quenstedt, 1876)	Bu			
67	*P. dichotomus*	(Hall, 1858)	uB			r
68	*P. liliiformis*	(Meek and Worthen, 1869)	uB			r
69	*P. ramulosus*	(Hall, 1861)	uB			r
70	*Paracosmetocrinus cf. P. strakai*	Strimple, 1967	lB	r		
71	*Parisocrinus labyrinthicus*	(Miller, 1891)	lB	r		
72	*P. tenuibrachiatus*	(Meek and Worthen, 1861)	lB-uB	u	u	u

73	*Pelecocrinus aqualis*	(Hall, 1859)	lB	u	r	
74	*P. insignis*	Kirk, 1941	uB			r
75	*Pellecrinus sp.*	Kammer and Gahn, 2003	lB	r		
76	*Poteriocrinites notabilis*	Meek and Worthen, 1869	lB	r		
77	*P. obuncus*	(White, 1862)	lB	r		
78	*P. waltersi*	(Rowley and Hare, 1891)	lB	r		
79	*Ramulocrinus rudis*	(Meek and Worthen, 1873)	uB		r	r
80	*Scytalocrinus cf. S. dodecadactylus*	(Meek and Worthen, 1860)	uB			r
81	*Springericrinus doris*	(Hall, 1861)	uB	r	u	u
82	*S. macropleurus*	(Hall, 1861)	lB	c		
83	*Tropiocrinus carinatus*	Kirk, 1947	uB		r	r
84	*Whiteocrinus florifer*	(Wachsmuth and Springer, 1877)	uB		r	
85	*Zygotocrinus enormis*	(Meek and Worthen, 1861)	lB	r		

Flexibles

1	*Forbesiocrinus agassizi*	Hall 1858	uB			r
2	*F. burlingtonensis*	Springer, 1920	uB			r
3	*Mespilocrinus chapmani*	Springer, 1920	uB		r	r
4	*M. konincki*	Hall, 1859	lB-uB	r	r	
5	*M. thiemei*	Springer, 1920	lB	r		
6	*Methichthyocrinus burlingtonensis*	(Hall, 1858)	lB	r		
7	*Nipterocrinus arboreus*	Worthen in Meek and Worthen, 1873	lB	r		
8	*N. wachsmuthi*	Meek and Worthen, 1868	uB		r	r
9	*Onychocrinus asteriaeformis*	(Hall, 1861)	uB			u
10	*O. diversus*	Worthen 1866	uB			u
11	*Parichthyocrinus nobilis*	(Wachsmuth and Springer, 1879	uB			r
12	*Taxocrinus juvenis*	(Hall, 1861)	lB	u	u	
13	*T. ornatus*	Springer, 1920	lB	r		
14	*T. ramulosus*	(Hall, 1859)	uB			r
15	*Wachsmuthicrinus bernhardinae*	Springer, 1920	lB	r		
16	*W. iowensis*	Springer, 1920	uB		r	r
17	*W. spinifer*	(Hall, 1861)	lB	r		
18	*W. thiemei*	(Hall, 1861)	lB	u		

BLASTOIDS

Fissiculates

1	*Hadroblastus whitei*	(Hall, 1861)	uB			r
2	*Orophocrinus catactus*	(Rowley, 1908)	lB	r		
3	*O. gracilus*	(Meek and Worthen, 1870)	lB	r		
4	*O. stelliformis*	(Owen and Shumard, 1865)	lB	c		
5	*Phaenoschisma gracillimum*	(Rowley and Hare, 1891)	lB	r		
6	*P. laeviculum*	(Rowley, 1900)	lB-uB	r	r	r

Granatocrinids

1	*Arcuoblastus shumardi*	(Meek & Worthen, 1895)	uB			r
2	*Auloblastus clinei*	Beaver, 1961	uB			r
3	*Carpenteroblastus magnibasus*	(Rowley, 1895)	uB			r
4	*C. pentalobus*	(Rowley, 1901)	Bu			r
5	*Cryptoblastus melo*	(Owen & Shumard, 1850)	lB	a		
6	*C. pisum*	(Meek & Worthen, 1869)	lB-uB	r	r	
7	*Decemoblastus melonoides*	(Meek and Worthen, 1869)	uB			r
8	*Dentiblastus sirius*	(White, 1862)	uB		r	r
9	*Lophoblastus inopinatus*	(Rowley and Hare, 1891)	lB	u		
10	*L. tenuistriatus*	(Hambach, 1903)	Bu	r		
11	*Poroblastus granulosus*	(Meek & Worthen, 1865)	uB			a
12	*Schizoblastus aplatus*	(Rowley and Hare, 1891)	lB	r		
13	*S. marginulus*	(Rowley, 1901)	uB			r
14	*S. moorei*	(Cline, 1936)	uB			r
15	*S. sayi*	(Shumard, 1855)	uB		u	c

Pentremitids

1	*Globoblastus norwoodi*	(Owen and Shumard, 1850)	uB		a	c
2	*Pentremites elongatus*	(Shumard, 1858)	uB		r	a
3	*P. kirki*	(Hambach, 1903)	lB	r		

Troosticrinids

1	*Metablastus lineatus*	(Shumard, 1858)	lB-uB	r	r	u