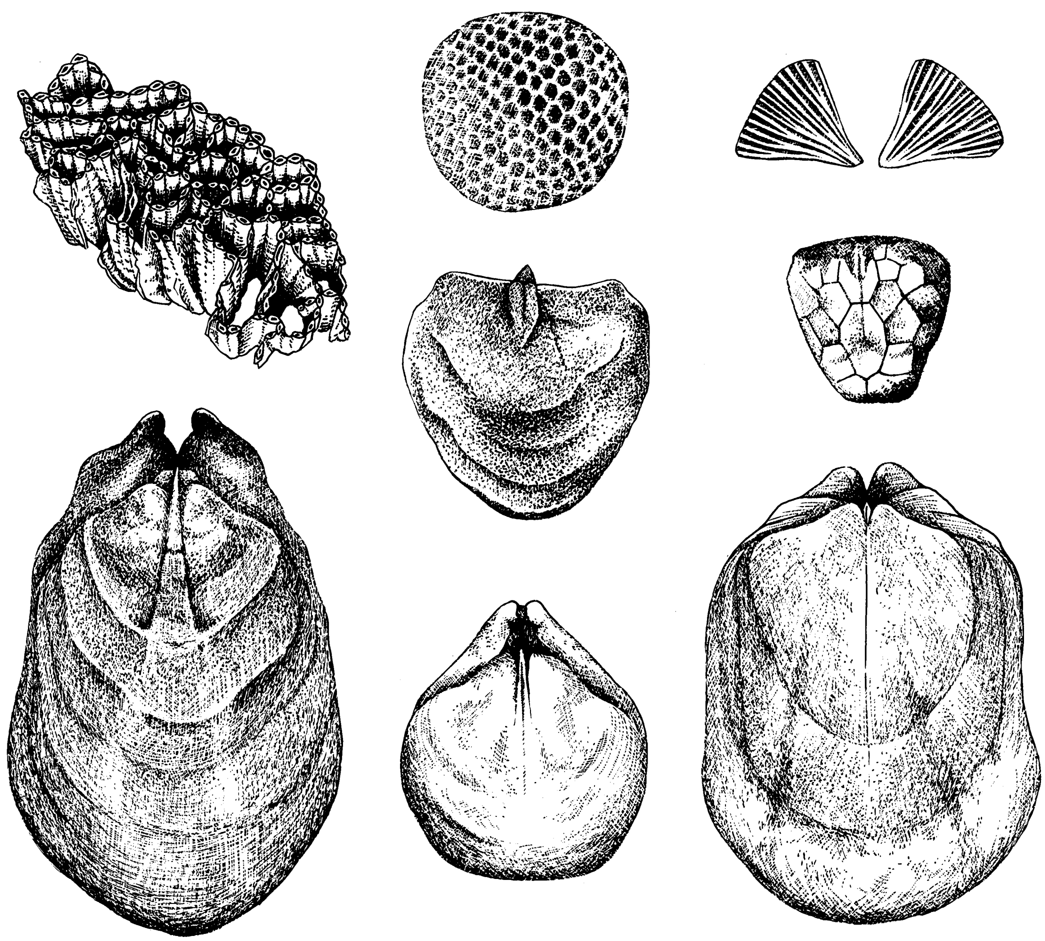
SILURIAN STRATIGRAPHY AND CARBONATE MOUND FACIES OF EASTERN IOWA

Field Trip Guidebook to Silurian Exposures in Jones and Linn Counties

GUIDEBOOK SERIES NO. 11



Iowa Department of Natural Resources
Larry J. Wilson, Director
April 1992

Cover illustration

top row:	*Halysites labyrinthicus,* *Cyclocrinites dactioloides,* *Petalocrinus mirabilis* arm-fans
middle right:	*Stricklandia laevis,* *Eucalyptocrinites ornatus*
bottom row:	*Pentamerus oblongus,* *Harpidium (Isovella) maquoketa,* *Pentameroides subrectus*

Cover drawings by Kay Irelan

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with additional contributions

Michael J. Bounk:
The development of Indian Bluff Cave

Greg A. Ludvigson, Brian J. Witzke, and Luis A. Gonzalez:
Observations on the diagenesis and stable isotopic compositions of Silurian carbonates in Iowa

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Iowa Department of Natural Resources
Larry J. Wilson, Director

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PART I

SILURIAN STRATIGRAPHY, DEPOSITION AND DIAGENSIS

SILURIAN STRATIGRAPHY AND CARBONATE MOUND FACIES OF EASTERN IOWA

by

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INTRODUCTION

Silurian rocks are extensively exposed in a broad belt across eastern Iowa (Fig. 1). The interval of Silurian carbonate rocks, now largely dolomitized, forms one of the most resistant bedrock sequences in Iowa, and a prominent erosional escarpment marks the northeastern edge of Silurian rocks in Iowa. The Mississippi River deflects southeastward roughly coincident with this escarpment, and continues in that direction as Silurian rocks thicken southward across east-central Iowa. Southeastward flowing tributaries of the Mississippi in east-central and northeast Iowa display steep bedrock valley walls along segments of their drainages that are incised into the resistant Silurian dolomites. Both active karst and paleokarst features are visible in these valleys. We will visit exposures along three river drainages in eastern Iowa during this field trip, namely the Maquoketa, Wapsipinicon, and Cedar rivers.

Aside from their aesthetic quality in natural exposures along various river drainages, Silurian rocks are economically important to many counties in eastern Iowa. Vast aggregate resources are quarried in many areas of east-central and northeast Iowa, and some Iowa Silurian lithologies may represent the highest quality concrete aggregates recognized in the state (W. Dubberke, 1989, pers. comm.). The bulk of sawed dimension stone produced in the state is also quarried from Silurian strata, primarily the laminated "Anamosa stone" from the Stone City area (STOP 4). Silurian rocks host the widespread Silurian aquifer, which forms the primary bedrock aquifer across much of eastern Iowa. The Silurian aquifer is extensively utilized for rural and municipal water needs, and an understanding of the stratigraphy, structure, and rock properties of the Silurian interval in Iowa provides important constraints for defining hydrogeologic parameters within this important aquifer.

The Silurian sequence of eastern Iowa is dominated by fossiliferous dolomites (dolostones) and cherty dolomites. These rocks may appear to the casual observer to represent a hopelessly monotonous interval ill-suited for lithostratigraphic subdivision. However, the efforts of various geologists over the past 150 years have combined to establish an objective Silurian lithostratigraphy that, although historically frought with misconceptions and miscorrelations, now provides a workable framework for subsequent study of the Silurian System in Iowa. Identification of gross rock fabrics is essential for recognition of lithostratigraphic units in the Silurian sequence, which include: argillaceous content, presence or absence of chert, dolomite crystal size, occurrence of skeletal molds or dolomitized skeletal grains, depositional fabrics (mudstone, wackestone, packstone), porosity, and presence of laminations. Recognition of certain key fossils greatly aids in evaluating stratigraphic assignments, but is not essential for establishment of the actual lithostratigraphy (in other words, the stratigraphic sequence is not a paleontologic zonation masquerading as rock-stratigraphy).

HISTORICAL BACKGROUND

A summary of the historical development of Silurian stratigraphic nomenclature in eastern Iowa and adjacent northwestern Illinois is provided in chart format (Figs. 2, 3). David Dale Owen first ventured into the Upper Mississippi Valley in 1839, including eastern Iowa, and he published reports on the geology of the region between 1840 and 1852 (see Johnson, 1977a). Owen was the first to recognize rocks and fossils of Silurian age in Iowa, only a few years after Murchison defined the Silurian System in 1835 from exposures in the Welsh Borderland. Owen correlated the "Coralline and Pentamerus beds" of the "Magne-

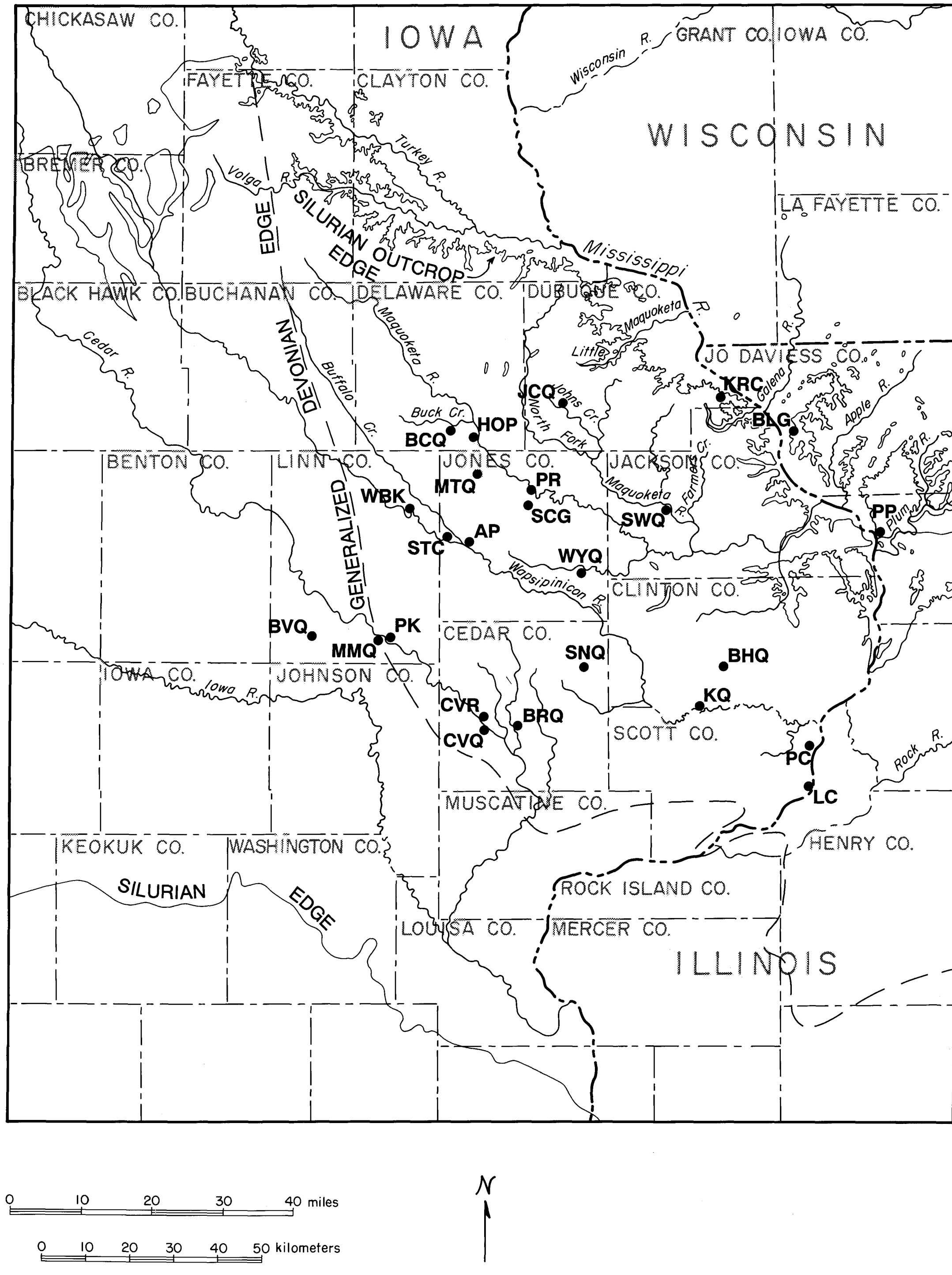

Figure 1. General extent of Silurian strata in eastern Iowa and adjacent portions of northwest Illinois and southwest Wisconsin. Silurian outcrop and subcrop edges delineated; position of Devonian edge is schematic only. Locations of sites noted in text are marked by letter designations.

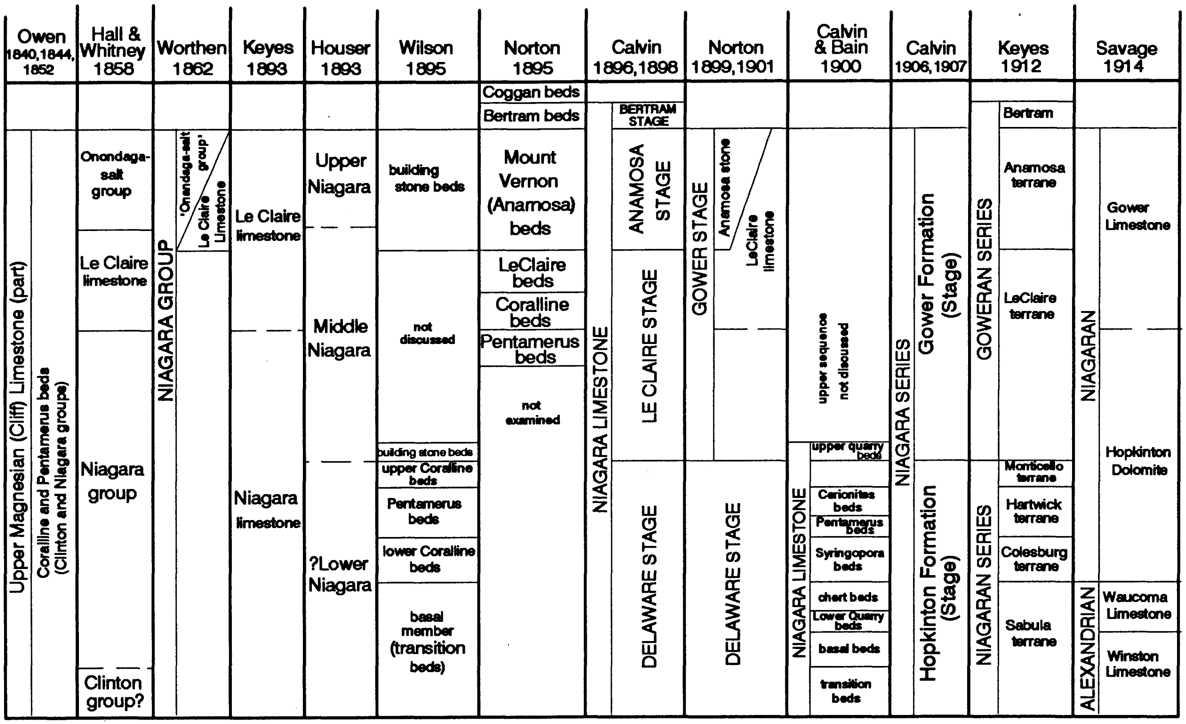

Figure 2. Development of Silurian stratigraphic nomenclature in eastern Iowa, 1840-1914.

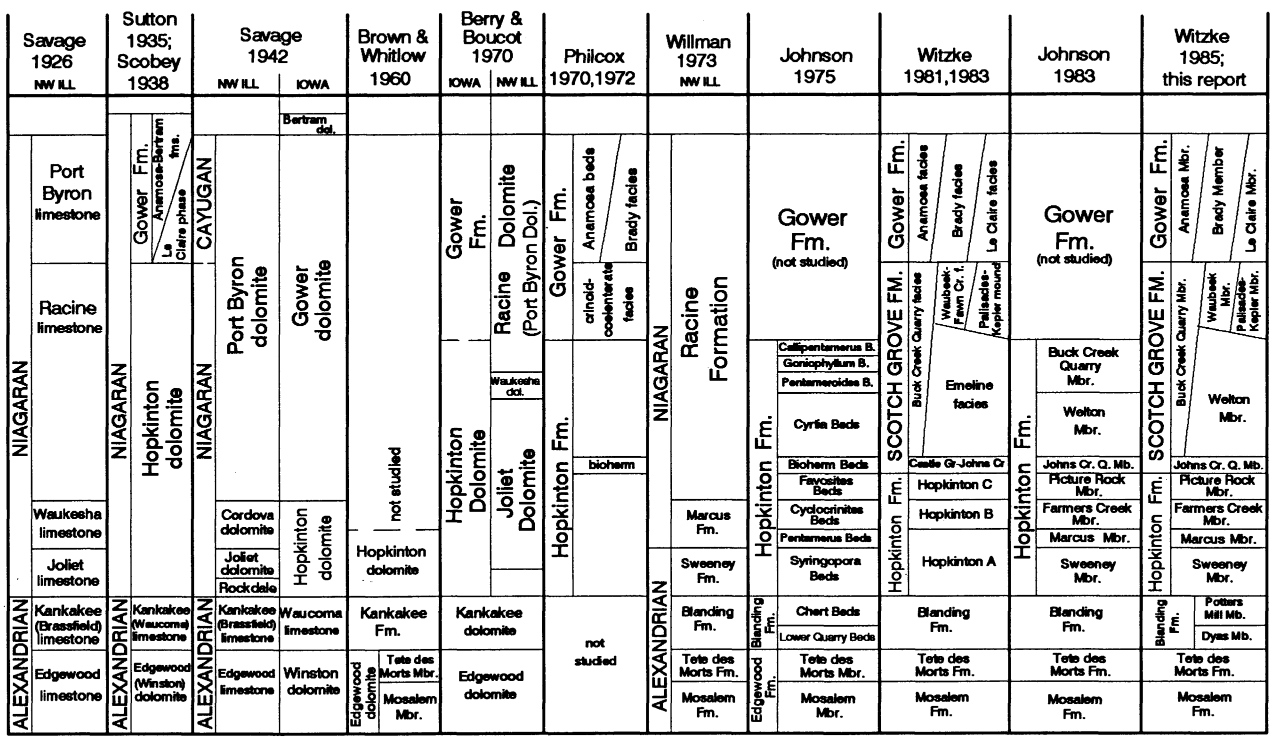

Figure 3. Development of Silurian stratigraphic nomenclature in eastern Iowa and northwestern Illinois, 1926-1992.

sian Cliff Limestone" in Iowa with the Clinton and Niagara groups of New York (Owen, 1852, p. 624). Subsequent studies by Hall and Whitney (1858) assigned portions of the Iowa Silurian sequence to the Clinton and Niagara groups, and they proposed a new stratigraphic name, the "LeClaire limestone," for exposures above the "Niagara" in eastern Iowa. They assigned an upper laminated dolomite interval to the "Onondaga-salt group," a name derived from a Devonian rock unit in New York. Worthen (1862) recognized the so-called "Onondaga" as a lateral facies of the Le Claire limestone, which he included as part of an expanded Niagara group.

Significant advancements in understanding the vertical sequence of Silurian strata in Iowa occurred during the 1890s, particularly the geologists at the newly-formed Iowa Geological Survey under the direction of Samuel Calvin. The general sequence in the lower part of the Silurian was first outlined by Wilson (1895), and subsequent work by the Survey geologists expanded on Wilson's stratigraphic subdivisions (Calvin, 1896, 1898; Norton, 1895, 1899, 1901; Calvin and Bain, 1900). New formal and informal stratigraphic units were proposed in their reports that included the Hopkinton Formation (first termed the "Delaware stage") and an overlying Gower Formation ("stage"). Although these studies laid out a relatively complete stratigraphic breakdown of Iowa"s Silurian sequence, certain errors in correlation and fossil identification gave rise to some critical stratigraphic misconceptions for portions of the sequence. In particular, Norton (1895) and Calvin and Bain (1900) "confused the occurrence of *Pentamerus oblongus* from the Pentamerus Beds with *Pentameroides subrectus* from younger strata . . . The mistake caused Calvin to exclude some additional beds from his classification" (Johnson, 1977b, p. 28-29). The confusion between these species also played a role in the placement of a Hopkinton-Gower contact in some sections at a position much higher in the stratigraphic sequence than where the top of the Hopkinton ("Delaware") was originally defined to be (see Witzke, 1981a). Calvin's and Norton's reports also varied considerably on the relationships between the LeClaire and Anamosa beds, which, in turn, influenced the definition of the Gower Formation and the placement of its basal contact. Although the "Gower Stage" was defined (Norton, 1899) to include the laminated "Anamosa stone" and the "LeClaire limestone," Norton (1899, 1901) recognized these units as lateral facies equivalents whereas Calvin (1896, 1899) stacked them vertically.

Savage's (1914, 1926, 1942) studies in Iowa and adjacent northwest Illinois resulted in an influential stratigraphic classification that was adopted by the Iowa and Illinois geological surveys during the 1920s through 1970s. Savage (1914) initially erected two new stratigraphic units at the base of the Silurian sequence, the Winston and Waucoma limestones, thereby restricting the base of Calvin's Hopkinton Formation to a higher stratigraphic position. Savage (1926) subsequently dropped the terms "Winston" and "Waucoma" and largely substituted a stratigraphic nomenclature derived from other areas. He correlated the old "Winston" unit with the Edgewood limestone of Missouri, and the "Waucoma" through mid Hopkinton interval was assigned to and correlated with, in ascending order, the Kankakee, Joliet, and Waukesha formations of northeastern Illinois. Savage (1942) subsequently introduced additional terminology for various Silurian units in northwest Illinois, including the Cordova dolomite for *Pentamerus*-bearing strata.

Brown and Whitlow (1960) subdivided Savage's "Edgewood" into two distinctive lithic units, the Mosalem and Tete des Morts members. Studies by Philcox (1970a,b; 1972) recognized several important stratigraphic relations in the eastern Iowa Silurian sequence: 1) a late Llandoverian biohermal facies was identified in the "Hopkinton"; 2) the laminated "Anamosa beds" were demonstrated to stratigraphically overlie biohermal (reefal) "crinoid-coelenterate facies" of the lower "Gower" Formation; 3) the "Anamosa beds" were shown to be lateral facies equivalents of the brachiopod-rich biohermal rocks (newly named "Brady facies"), not crinoidal biohermal rocks as previously thought. Philcox's studies paved the way for subsequent re-evaluation of Gower Formation stratigraphy, and clarified several misconceptions about relationships between biohermal (reefal) and inter-biohermal (inter-reefal) strata in the Iowa Silurian.

Studies during the 1970s and 1980s recognized significant problems with Savage's (1926, 1942) stratigraphic nomenclature, and a revised and updated stratigraphic classification emerged to take its place. Willman (1973) demonstrated that gross lithostratigraphic miscorrelations between northwestern Illinois and other areas (northeast Illinois, eastern Wisconsin, eastern Missouri) led to an inappropriate and inaccurate stratigraphic nomenclature in northwestern Illinois under Savage's scheme. Willman (1973) rejected Savage's supposed correlation and extension of "Edgewood," "Kankakee," "Joliet," and "Waukesha" units into the Upper Mississippi Valley, and new stratigraphic names were proposed to replace them. However, Willman did not reconcile or adopt previously named and correlative units across the Mississippi River in eastern Iowa, particularly the Hopkinton, Gower, and LeClaire.

In retrospect, it seems remarkable that *Savage had miscorrelated all Silurian units in the area*, but that nevertheless seems to be the case: 1) Savage's "Edgewood" does not lithically resemble nor does it chronostratigraphically correlate with the Edgewood sequence in Missouri and southern Illinois (Witzke, 1981a); 2) Savage's "Kankakee" is a different rock-stratigraphic unit than the type Kankakee interval in northeast Illinois (Willman, 1973); 3) Savage's "Joliet" is lithically unlike the type Joliet sequence in northeast Illinois, but closely resembles the type Kankakee (Willman, 1973); and 4) Savage's *Pentamerus*-bearing "Waukesha" is significantly different lithically from the type Waukesha interval in Wisconsin, and is also older than at the type section (Willman, 1973; Witzke, 1981a). Savage's (1942) local reassignment of the "*Pentamerus* beds" to the "Cordova dolomite" also proved to be unsatisfactory, as "Savage was under the misapprehension that the beds in the quarry at Cordova, which are Racine [i.e., LeClaire Member], belong to this unit" (Willman, 1973, p. 38).

Willman (1973) proposed new names to replace Savage's inappropriate nomenclature in northwestern Illinois, substituting the Blanding, Sweeney, and Marcus formations for the "Kankakee," "Joliet," and "Waukesha," respectively. Willman (1973) also elevated Brown and Whitlow's (1960) members of the "Edgewood" to formational rank. Willman's (1973) much-needed restructuring of the Silurian stratigraphy in the Upper Mississippi Valley proved to be a significant advance. However, one correlation remained unchecked, namely the extension of the "Racine Formation" in the area, which Willman (1973) extended downward to the top of the *Pentamerus*-bearing Marcus Formation. Further study of these lower "Racine" strata would demonstrate noteworthy lithic, paleontologic, and chronostratigraphic differences with strata of the type Racine sequence in eastern Wisconsin (Johnson, 1975, 1983; Witzke,1981a).

Johnson (1975, 1977b) described a sequence of informal stratigraphic units in the Lower Silurian of eastern Iowa that largely paralleled those of Calvin and Bain (1900), but he expanded the sequence upward to include additional units not recognized by earlier workers (Fig. 3). Units in the Hopkinton sequence were named after distinctive fossils, and each was characterized by lithic characters and recurrent associations of benthic paleocommunities (Johnson, 1975). Johnson (1983) later proposed formal lithostratigraphic names for these units (Fig. 3), largely adapting Willman's (1973) Silurian units in northwestern Illinois as members within the lower Hopkinton interval, and a series of new member names were erected for strata above the Marcus (Fig. 3). Johnson's (1983) stratigraphic subdivisions have proven to have significant utility throughout much of eastern Iowa, and all of his member names are retained for use in this guidebook.

Witzke (1981a, 1981b, 1983) used a stratigraphic classification that largely paralleled that of Johnson (1975, 1977b). However, Witzke's initial informal lithostratigraphic framework (Fig. 3) incorporated more restricted concepts of the Hopkinton and Gower formations than were used in many previous studies. The complex facies relationships within and between intervals previously assigned to the upper "Hopkinton," Gower, Anamosa, Brady, and LeClaire underscored the need for stratigraphic restructuring, and Witzke (1985) formally introduced a new formation, the Scotch Grove, in order to retain previously-defined stratigraphic units in a consistent manner. Calvin (1896, 1906, 1907) originally marked the top of the Hopkinton Formation below the "upper quarry beds" (Johns Creek Quarry Member of Johhson, 1983), and Witzke (1985) retained the Hopkinton below this prominent lithic boundary. The Gower Formation was retained only for that interval that incorporates the type Gower, type Anamosa, and type LeClaire sections. Some intervals that were assumed to correlate with these sections, particularly strata termed "LeClaire" by many earlier workers, were shown by Witzke (1981a, 1985) to lie in stratigraphic position below Gower and Anamosa strata.

The introduction of the Scotch Grove Formation (Witzke, 1985) enabled the interval above the type Hopkinton and below the type Gower sections to be consistently labelled without misapplying original concepts of the Hopkinton and Gower. In addition, no regionally consistent lithic boundary was found within the Scotch Grove interval that could have served to erect the previously undefined and elusive Hopkinton-Gower contact. Instead, the interval was found to be a complex of biohermal, inter-biohermal, and cherty facies (Witzke, 1981a). Each facies in the Scotch Grove was given formal member status (Witzke, 1985), incorporating Johnson's (1983) members of the upper "Hopkinton" as facies within the lower half of the new formation. Additional members were introduced for upper Scotch Grove facies, the Palisades-Kepler and Waubeek members (Fig. 3). The Gower Formation was constrained to the interval that includes the laminated Anamosa Member and its laterally equivalent biohermal (carbonate mound) facies of the LeClaire and Brady members (Witzke, 1985). Witzke (1985, p. 36) noted: "Most previous workers included the Palisades-Kepler Member . . . within the 'LeClaire facies' and incorrectly inferred lateral equivalency of these mounds with laminated Anamosa rocks. This procedure has promulgated serious stratigraphic errors and led to an over-simplification of the complex facies relationships in the Scotch Grove and Gower formations."

The stratigraphic nomenclature used in this report essentially expands on the general framework erected by Calvin and Bain (1900), Willman (1973), Johnson (1975, 1977b, 1983), Witzke (1981a, 1985), and others. A general stratigraphic column for eastern Iowa, as presently recognized at the Geological Survey Bureau, is given in Figure 4. Interpretations of Iowa's Silurian stratigraphy and depositional history will undoubtedly be modified or changed as additional research is undertaken. It is hoped that such studies will refine and correct our understanding of the Silurian rocks in Iowa, enabling Iowa's exceptional Silurian sequence to become a standard of comparison for the cratonic interior of North America.

STRUCTURAL SETTING

Silurian rocks are buried beneath Middle Devonian strata along the western edge of the Silurian outcrop belt in eastern Iowa. Silurian rocks extend westward beneath Devonian strata in the Iowa subsurface, constricting and thinning in central Iowa, then expanding and thickening towards the southwest corner of the state (Fig. 5). Iowa Silurian strata are contiguous with Silurian rocks in adjacent areas of eastern Nebraska, northwestern Missouri, and northwestern Illinois, and scattered erosional outliers extend into southwestern Wisconsin.

The thickest interval of Silurian rocks in Iowa is found in the outcrop belt of eastern Iowa (Jones County) adjacent to the Plum River Fault Zone (480 ft; 146 m; Bunker et al., 1985, p. 81). In general, Silurian stratigraphic units thicken eastward across east-central Iowa (Fig. 6), but units are erosionally bevelled northward in the outcrop belt and by incision of the Mississippi drainage. Silurian strata thin beneath the Devonian cover to the south, west, and northwest of the outcrop belt (Fig. 5). The occurrence of the thickest Silurian sequences in Iowa within the outcrop belt is especially noteworthy, and suggests that eastern Iowa was the location of a pre-Middle Devonian basin. Silurian rocks that were deposited within this

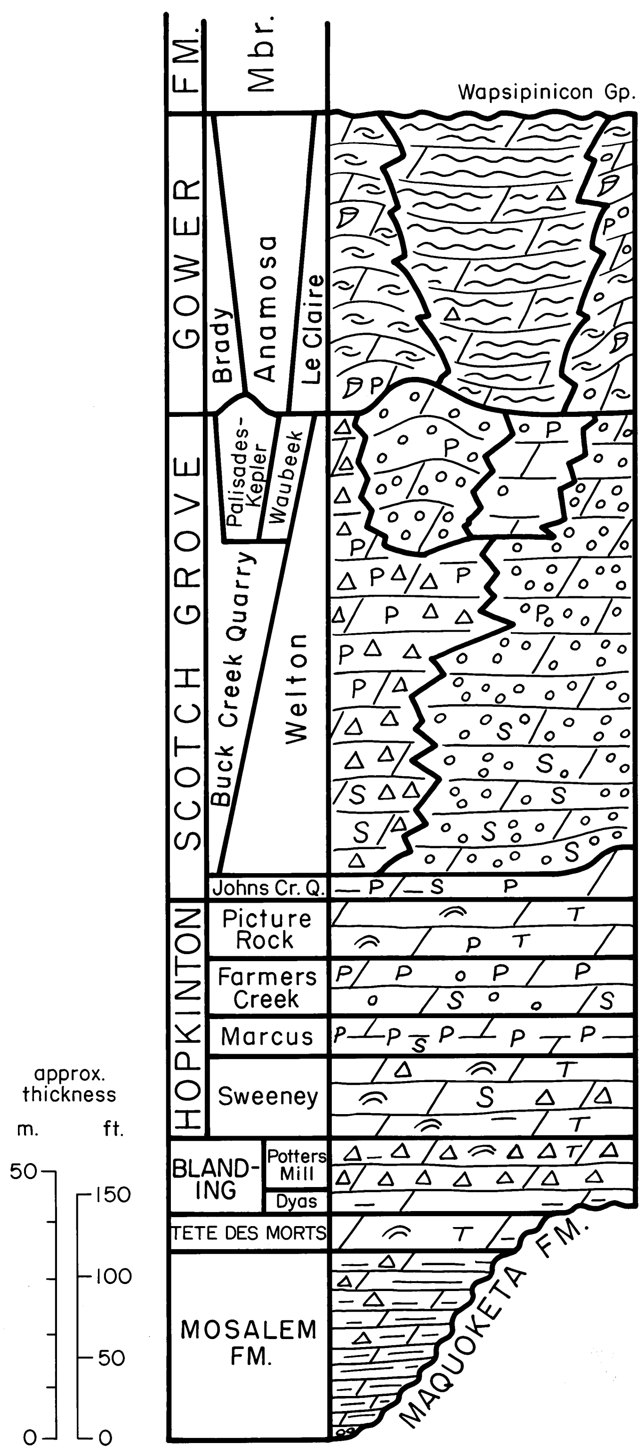

Figure 4. General Silurian stratigraphic column for eastern Iowa. See page 86 for lithologic symbols.

paleo-basin, termed the East-Central Iowa Basin, were uplifted and erosionally bevelled to the northeast prior to regional Pennsylvanian sedimentation (Bunker et al., 1985; Lee, 1946). Stratigraphic and depositional relationships of Silurian strata within the East-Central Iowa Basin indicate that the locus of maximum sediment accumulation shifted southward in eastern Iowa during the Silurian and that downwarping or faulting along the Plum River Fault Zone was coincident in part with Silurian sedimentation (Bunker et al., 1985).

The southwestward thickening of Silurian strata in southwestern Iowa also reflects the presence of a pre-Middle Devonian basin in that area, the North Kansas Basin (Lee, 1946). Maximum Silurian thicknesses in the North Kansas Basin (to about 500 ft; 150 m) are identified in southeastern Nebraska, closely comparable to maximum thicknesses in the East-Central Iowa Basin. The close parallels in thickness and extent of Silurian strata in both basins are noteworthy, and the East-Central Iowa Basin and North Kansas Basin can be regarded as "sister" basins physically joined by a structural sag across central Iowa. Pennsylvanian movements along the Nemaha Uplift (southeast Nebraska, northeast Kansas) were responsible for erosional truncation of Silurian rocks in the central part of the North Kansas Basin. Therefore, Silurian stata in the central areas of both the East-Central Iowa and North Kansas basins were uplifted and eroded prior to their burial by Pennsylvanian strata.

Silurian rocks in the central and southern craton of North America are preserved primarily in pre-Middle Devonian structural basins (Fig. 7). With few exceptions, the intervening arches and uplifts have been erosionally stripped of their Silurian cover prior to the Middle Devonian. Silurian strata of the North Kansas and East-Central Iowa basins are bounded by a series of structural highs (Fig. 7): 1) the broad Transcontinental Arch to the north and northwest, 2) the Chautauqua Arch and Ozark Uplift to the south, 3) the Central Kansas Uplift to the west, and 4) the Wisconsin Arch to the east. Silurian strata in the East-Central Iowa Basin are physically connected across Illinois with the Michigan Basin the east and the proto-Illinois Basin to the southeast (Fig. 6).

Silurian carbonate sedimentation apparently was much more extensive across the North American craton than the modern distribution of Silurian rocks may seem to indicate. Silurian rocks in the central craton are dominated by remarkably pure carbonate sediments, suggesting that most siliciclastic source terranes in the craton became inundated during transgression of an extensive Silurian epeiric sea, thereby minimizing detrital input (Johnson, 1987; Witzke, 1990). Even the broad Transcontinental Arch (Fig. 7) was inundated by Silurian seas, as evidenced by the fortuitous preservation of "clean" carbonate rocks in isolated kimberlite diatremes in the Laramie Range of Wyoming-Colorado (Chronic et al., 1969). Therefore, it seems reasonable to suggest that Silurian sedimentation was contiguous between the Iowa-Nebraska area and the Williston Basin of the Dakotas (Fig. 7) across the Transcontinental Arch during times of maximum marine transgression. Likewise, Silurian seaways also probably expanded across the Chautauqua Arch at times, joining the North Kansas Basin with the Oklahoma Basin and contiguous shelf environments along the Ouachita continental margin (Fig. 7). The preservation of Silurian limestones in the Decaturville impact structure of central Missouri further indicates that Silurian seas covered portions of the central Ozark Uplift (Offield and Pohn, 1979). The prolonged period of erosion, probably in excess of 25 million years, that separated Silurian and Middle Devonian sedimentation in the Iowa area was responsible for extensive truncation and removal of Silurian strata from structurally higher regions.

A series of Silurian basins and arches are well displayed by the sub-Middle Devonian distribution and thickness patterns of Silurian rocks across the North American craton (Fig. 7). However, subsidence patterns were modified in the Iowa area during the Middle to Late Devonian, and the former loci of maximum Silurian sediment accumulation, namely the East-Central Iowa and North Kansas basins, lost their character as discrete basins. Instead, the area of maximum sediment accumulation shifted during the Middle Devonian to the structural saddle between the two Silurian basins in central Iowa, marking the development of the Iowa Basin (Bunker et al., 1988; Witzke et al., 1988).

Silurian and Devonian structural patterns were significantly modified by structural re-organization in the central United States during the Late Mississippian and Pennsylvanian, probably in response to closure and collision along the Ouachita continental margin to the south. In the Iowa area, these changes were manifested by two structural shifts that were largely responsible for the development of the present-day structural configuration of Silurian rocks: 1) uplift of the Nemaha Uplift and coincident asymmmetrical basinal subsidence in the newly-formed Forest City Basin of southwest Iowa and adjacent states, and 2) uplift along the trend of the Wisconsin Arch. These changes resulted in the general southwestward sloping structural surface

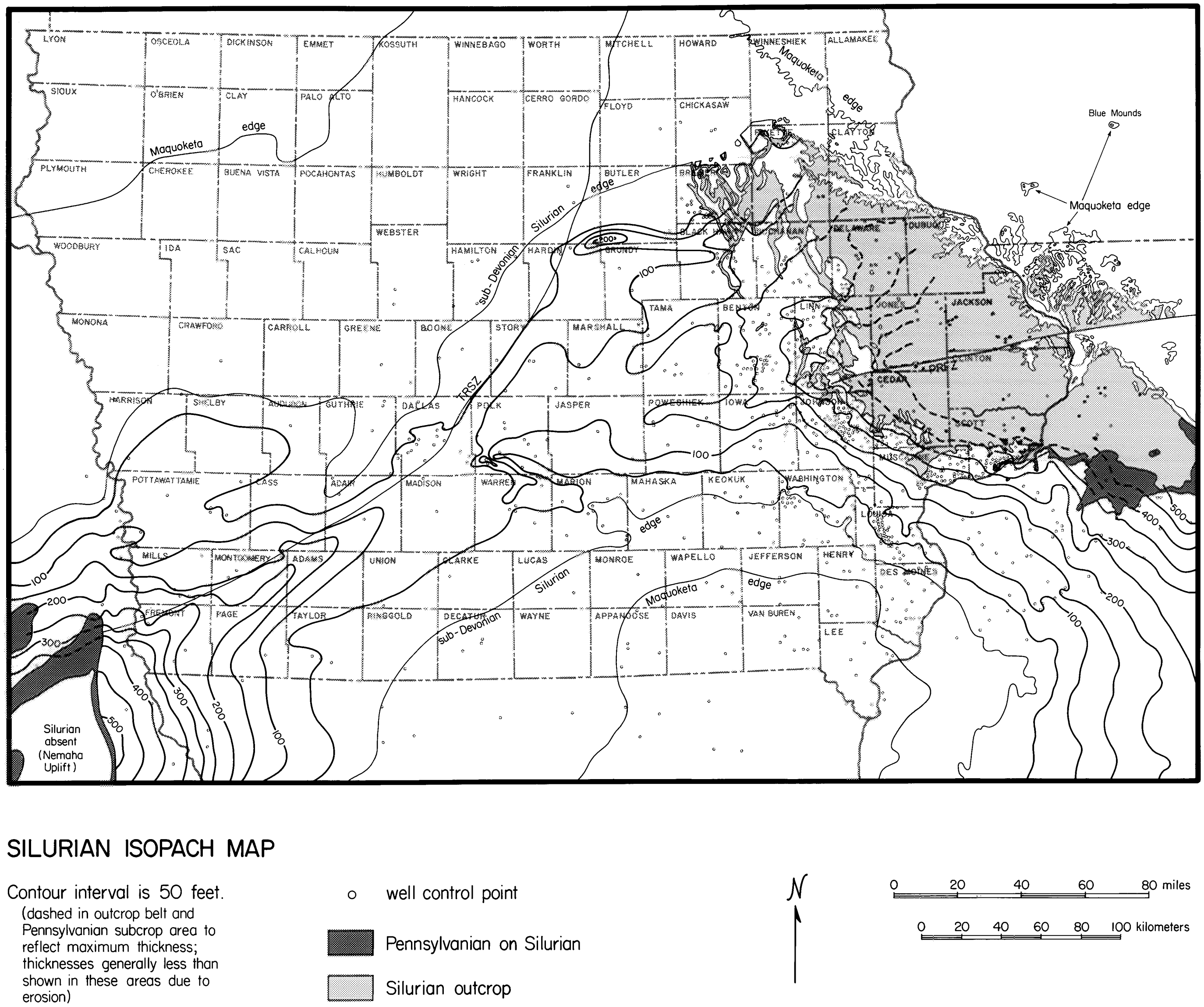

Figure 5. Isopach map of Silurian strata in Iowa and adjacent areas. Edges of Maquoketa Formation (Upper Ordovician) and Silurian strata noted. Isopachs dashed in outcrop area to reflect maximum thicknesses only. PRFZ (Plum River Fault Zone), TRSZ (Thurman-Redfield Structural Zone).

of Silurian rocks that stretches across Iowa from southeast to southwest. Because of this general slope, Silurian rocks regionally achieve their highest elevations in northeast Iowa (>1200 ft; >365 m) and southwest Wisconsin (at West Blue Mound, 1716 ft; 523 m) and their lowest elevations (to at least -1619 ft; -494 m) in southwest Iowa.

Although post-Pennsylvanian structural movements are documented in the Iowa area (Bunker et al., 1988), the general structural configuration of Silurian rocks in Iowa was largely emplaced during the late Paleozoic. Pennsylvanian and post-Pennsylvanian structural movements are evident particularly along the trend of the Thurman-Redfield Structural Zone in central to southwestern Iowa, where Paleozoic strata are deformed by faulting, flexuring, and doming. Silurian strata locally

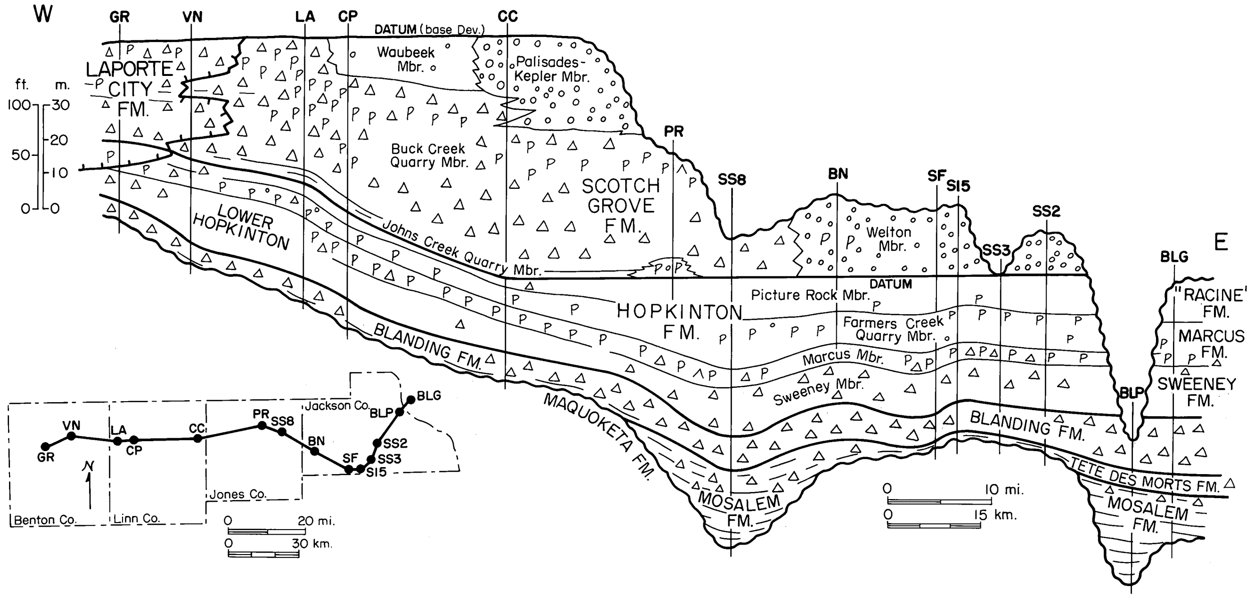

Figure 6. Stratigraphic cross-section of Silurian dolomite units, east-central Iowa. LaPorte City Formation is a western limestone facies. Control points include both core and outcrop sections. Datum shifts from base of Devonian in the west to base of Scotch Grove Formation in the eastern outcrop. See page 86 for lithologic symbols.

display up to 1000 ft (300 m) of structural relief around domes associated with this zone in southwest Iowa. Broad-scale epeirogenic upwarping or isostatic adjustments during the late Mesozoic and Cenozoic slightly modified the regional southwestward structural slope.

GENERAL SILURIAN LITHOLOGIES AND DIAGENESIS

Siliciclastic Content

Silurian rocks exposed in eastern Iowa are overwhelmingly dominated by dolomite (dolostone) strata, with secondary components of chert and limestone. Argillaceous and detrital quartz content is extremely low through most of the Silurian sequence, although argillaceous to shaley strata characterize the basal unit (Mosalem Fm.) in portions of the Silurian outcrop belt. These basal clay and quartz sediments were derived from the underlying Ordovician Maquoketa shales, but the input of detrital sediments was greatly reduced following the regional burial of the Maquoketa shale surface. Portions of the Blanding, Hopkinton, and lower Scotch Grove formations display minor argillaceous partings, especially along stylolitic surfaces, although total clay content is extremely low in these strata. Presumably, siliciclastic source terranes lay at considerable distance from the eastern Iowa area during most of the Lower and mid Silurian, probably in the Taconic uplands of the eastern U.S. and regions along the Transcontinental Arch and Canadian Shield.

Dolomite Fabrics

All dolomite lithologies in the Iowa Silurian are interpreted to be of secondary diagenetic origin (Witzke, 1981a). Although diagenetically compromised, original depositonal fabrics are preserved to varying degrees within most of these dolomites. Fortuitously, many Silurian dolomite lithologies seen in eastern Iowa can be directly compared with stratigraphically equivalent and geographically contiguous limestone facies preserved in Iowa. These Silurian limestone units include the Waucoma Formation of northeast Iowa (equivalent to the Tete des Morts, Blanding, and lower Hopkinton formations in the dolomite facies) and the LaPorte City Formation of northeast and central Iowa (equivalent to the upper Hopkinton and lower to mid Scotch Grove formations; Witzke, 1981a).

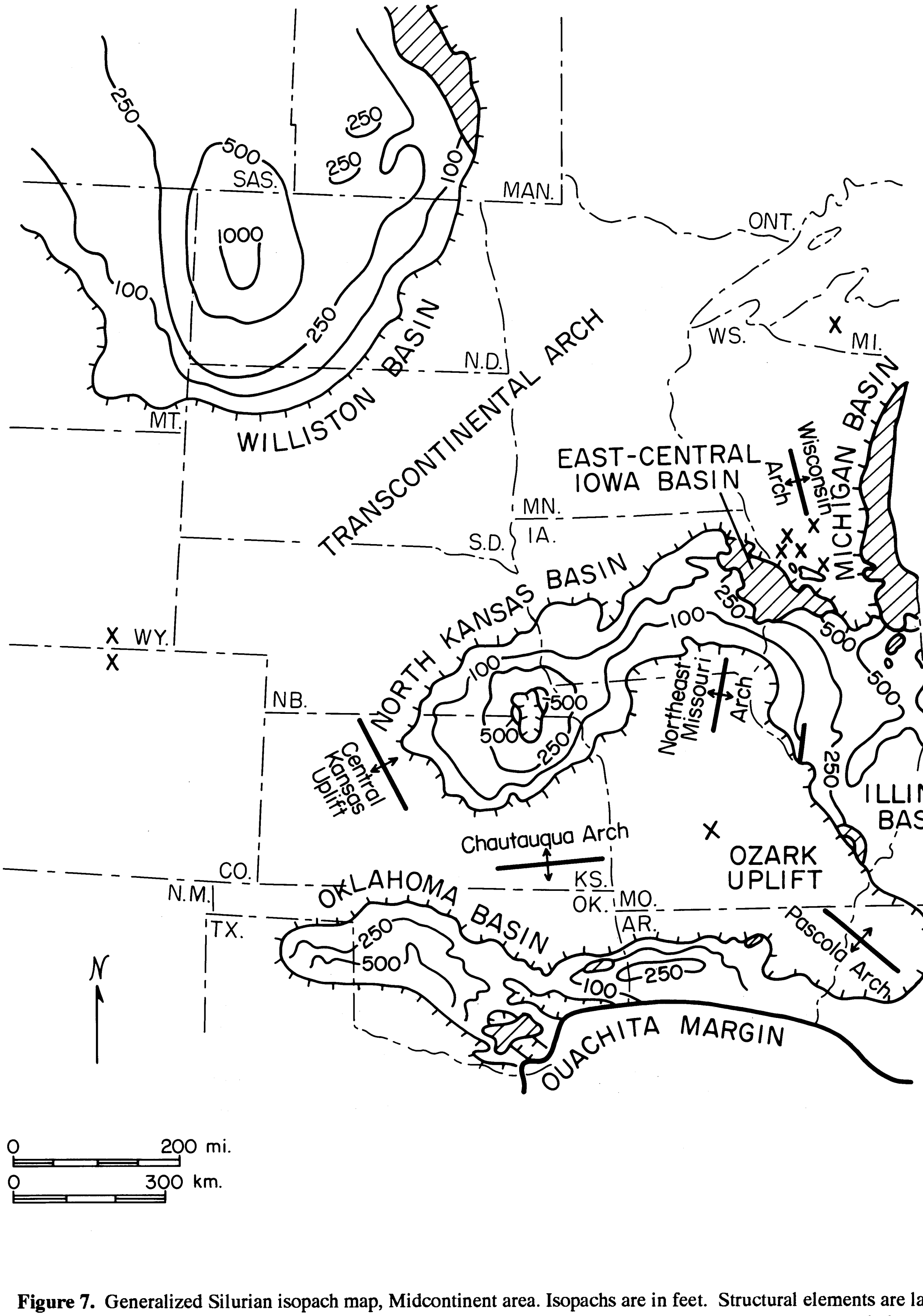

Figure 7. Generalized Silurian isopach map, Midcontinent area. Isopachs are in feet. Structural elements are labeled. Location of Silurian outliers marked by X. Silurian edge is hachured. Silurian outcrop areas are cross-ruled.

As such, most dolomite fabrics can be unambiguously interpreted in terms of original carbonate depositional fabrics, and the dolomites are categorized accordingly using a modified Dunham-type classification (Dunham, 1962). The bulk of Silurian dolomite strata in eastern Iowa display the following carbonate fabrics: 1) skeletal mudstone, 2) skeletal wackestone, 3) skeletal packstone, and 4) nonskeletal laminated mudstone. In addition, minor grainstone, bafflestone, pelletal, stromatolitic, and isopachous cement fabrics are recognized in the dolomitized rocks.

Dolomite crystal sizes, as seen in thin section, contrast in different stratigraphic intervals and facies, with the finest crystal sizes (2-50 microns) noted in the mudstone and wackestone units. Coarser dolomite crystals (commonly 50-750 microns) are seen in skeletal packstone intervals (replacing skeletal grains in part) and lining some voids. The coarsest dolomite crystals in the Iowa Silurian have replaced echinoderm grains in packstone to grainstone fabrics, with some crinoid ossicles and columnals replaced by unit dolomite crystals as large as 2 cm. Most matrix dolomite is composed of anhedral to subhedral crystals in porous hypidiotopic to dense xenotopic crystal fabrics, and euhedral crystal margins are commonly seen along void margins. "Ghost" fabrics are commonly preserved in the matrix dolomite, best seen under plane polarized light, which display fine skeletal grains, pellets, or cement fabrics in various units. Extremely finely crystalline dense xenotopic dolomites that do not display ghost fabrics and that contain only scattered skeletal molds (or lack molds) are interpreted to be dolomitized carbonate mud.

The bulk of skeletal grains in skeletal mudstone to wackestone fabrics have been solutionally removed, but the locations of the former grains are clearly recognizable as moldic porosity. These molds commonly preserve skeletal morphologies with remarkable detail, faithfully preserving skeletal structures both externally and internally (steinkerns). As such, the moldic preservation of many benthic invertebrate skeletons, especially brachiopods and crinoids, provides an adequate basis for taxonomic assessment. Skeletal-moldic porosities vary in different facies, from less than 1% in some sparse mudstones to greater than 30% in some wackestones (Witzke, 1981a). Most skeletal grains in packstone and grainstone fabrics, however, are not displayed as moldic porosity. It would be logically impossible to preserve packstones-grainstones with moldic grains, as these grain-supported fabrics would be reduced to a rock with no internal support. Instead, skeletal packstones and grainstones are typically preserved in the Iowa Silurian rocks by dolomite replacement of the skeletal grains. In particular, the unit-calcite crystals of echinoderm grains are commonly replaced by large unit-dolomite crystals. Minor skeletal moldic porosity is commonly noted co-occurring with skeletal grain replacement in many dolomitized packstones.

In addition to skeletal molds, vugular porosity is well developed in some Silurian stratigraphic units, primarily in dolomitized mudstones and wackestones. These vugs typically range in diameter from about 1 to 10 cm (0.4-4 inches). Many vugs formed prior to the completion of dolomite and silica precipitation in the Silurian sequence, as coarse dolomite and chalcedony or megaquartz void linings are commonly seen. Later-stage calcite spar fills vugs at some localities, but open vugs are characteristic at most localities, both in exposure and in the subsurface. Interconnected vugular networks and solutional channels are common in particular stratigraphic units, especially the Farmers Creek Member, and these probably formed in response to post-Silurian to recent groundwater movement. The vugular Farmers Creek Member forms the primary producing interval in the Silurian aquifer across much of eastern Iowa.

Relict carbonate cement fabrics are preserved in the dolomitized rocks in several stratigraphic units with carbonate mound ("reef") facies, and these fabrics are observable both in thin section and hand sample. Relict fibrous cements lining pore spaces are commonly seen in thin sections from carbonate mound facies, and isopachous cement rims are seen surrounding skeletal grains in some packstone/grainstone intervals (Witzke, 1981a). These relict cements resemble fibrous calcite or aragonite cements characteristic of active marine phreatic environments, and they are interpreted to provide evidence of submarine cementation in the mounds. In hand specimen, the relict fibrous cements often form distinctive botryoidal coatings around skeletal grains and pores. Minor syntaxial dolomite overgrowths around crinoid grains have also been noted in some packstone fabrics. These may conceivably represent replacements of pre-existing calcite syntaxial overgrowths or aggrading dolomite crystal growth into matrix areas.

Silica Fabrics

Silicification is a noteworthy feature of several stratigraphic units in the Iowa Silurian sequence, particularly in the Blanding and lower Hopkinton formations and the Buck Creek Quarry Member of the Scotch

Grove Formation. Minor silicification occurs in some other units, although grain and matrix silicification is conspicuously absent in a number of stratigraphic units (especially the Welton, Palisades-Kepler, Brady, and LeClaire members). Nodular cherts provide the most conspicuous evidence of silicification in the Iowa Silurian. The cherts occur in two general forms: 1) smooth chert (hard with glassy fracture), and 2) chalky chert (less resistant with "chalky" feel). Many nodules consist of smooth chert in the middle, rimmed by an outer layer of chalky chert, although many nodules are composed entirely of chalky chert. The smooth cherts are characteristically composed of microquartz (1-5 microns), with small spherical aggregates (10-20 microns) of microquartz preserved in some nodules which may be relict opal-CT lepispheres. By contrast, the chalky cherts contain up to 60% (generally 10-40%) dolomite rhombs within a microquartz matrix. Pyrite crystals are scattered to abundant in some cherts, generally concentrated within the outer edges of the nodules.

Skeletal grain silicification is common in some units, which occurs both within chert nodules and as isolated grains within the dolomite matrix. Skeletal grains are variably replaced by microquartz, fibrous chalcedony, or chalcedony spherules with radial fibrous microstructure. Echinoderm grains are typically replaced by microquartz, rarely displaying an optically-continuous mosaic. Brachiopod grains are most commonly replaced by fibrous chalcedony, which parallels original shell microstructure in some cases. Corals and stromatoporoids are commonly silicified in some units, especially the Hopkinton Formation and portions of the Buck Creek Quarry Member.

Megaquartz commonly lines vugs and pores within the dolomites in some cherty as well as non-cherty intervals. Some vugs, as in the Waubeek Member, are locally lined with terminated quartz crystals over 5 mm in length. In thin section, megaquartz fills and lines scattered fossil molds and pores within the dolomite groundmass. Additionally, some vugs and pores in the dolomites are partially or completely filled with lumpy to botryoidal chalcedony, and a few are filled by rare banded agate nodules. Some chert nodules include voids and fractures that are lined with megaquartz. Therefore, most of the megaquartz/chalcedony pore and vug linings apparently post-date chert nodule formation as well as the development of vugs and skeletal molds in the dolomites.

Diagenesis

Petrographic evidence clearly indicates that the Silurian sediments were originally deposited as carbonate mud and skeletal grains, and that dolomitization is entirely of post-depositional origin. Non-dolomitized limestone units in northeast Iowa reflect original compositions, and provide evidence that regional dolomitization was not pervasive. Likewise, chert nodule growth and grain silicification is an early diagenetic replacement of original carbonate sediments. A paragenetic sequence of diagenetic processes was documented for the eastern Iowa Silurian sequence by Witzke (1981a), and three general stages can be summarized. The first stage is interpreted to have occurred penecontemporaneous with marine deposition, and includes 1) marine phreatic cementation (especially in carbonate mound facies) and early fracturing of the lithified carbonates, 2) development of micrite envelopes around grains in some settings, and 3) deposition of early pyrite framboids within the sediment.

These early processes were succeeded by a middle stage of diagenesis wherein several important diagenetic events, especially silicification and dolomitization, were completed. The middle diagenetic stage is interpreted to have occurred primarily, if not entirely, within the so-called "mixing zone" (Witzke, 1981a), a diagenetic environment characterized by the mixing of marine pore fluids with freshwater and marked by salinities and water chemistry intermediate between marine and meteoric waters. Dolomitization and silicification of calcium carbonate sediments is known to be thermodynamically possible in such a mixing zone. The mixing zone will, of course, occupy an intermediate position between freshwater recharge areas (land) and the adjacent seaway, and mixing zone environments will prograde seaward during marine regression, possibly providing a mechanism for regional dolomitization of Silurian sediments in the continental interior (see accompanying article by Ludvigson et al.). Kinetic factors are also important, as dolomitization is presumably not a rapid process. Gradual and stepwise offlap of the seas during the Late Silurian would provide a mechanism for sediments to occupy prolonged and recurrent episodes within mixing zone environments.

A general ordering of middle-stage diagenetic events can be determined petrographically. Chert nodule growth was initiated in the skeletal mudstones and wackestones, and was succeeded by dolomitization of the carbonate mud. The silica needed for chertification was probably derived from biogenic silica within the sediment (especially sponge spicules), which was mobilized and then precipitated (probably originally as metastable opal-CT) around specific nucleation sites, primarily in mud-rich precursor sediments. Carbonate sediments are compactionally deformed around some chert nodules, suggesting that chertification accompanied, in part, compaction. In addition, chertification overlapped in part with dolomitization, as smooth chert cores without dolomite are rimmed by chalky cherts intimately intergrown with dolomite rhombs (and locally with concentrations of pyrite crystals). The completion of dolomitization post-dated chert nodule growth.

The exquisite preservation of some silicified fossils, when contrasted with dolomitized rocks, further suggests that grain silicification preceded the completion of dolomitization. The development of skeletal molds and some vugs apparently accompanied dolomitization of mudstone and wackestone lithologies, whereas grain replacement characterized packstone and grainstone facies. Later-stage dolomite and silica precipitation includes chalcedony, megaquartz, and coarse dolomite void and mold linings, clearly indicating that the development of skeletal molds and some other voids preceded the final stages of dolomite and silica precipitation. Pressure solution (stylolite surfaces and stylo-swarms) apparently post-dated chert nodule growth and is interpreted to have accompanied, at least in part, the later stages of dolomitization.

The final diagenetic stages post-date dolomitization and silicification, and include several important, but volumetrically insignificant, processes. These primarily include further solutional enlargement of vugs and fractures and calcite spar void filling. Such processes probably occurred in freshwater phreatic environments that repeatedly incurred through the Silurian interval during latest Silurian through recent times. In addition, minor calcitization of dolomite and sulfide void linings are of late diagenetic origin.

Porosity Development

The formation of vugular, moldic, and intercrystalline porosities in the Silurian dolomites is of special interest, as such porosities are a characteristic feature of the eastern Iowa Silurian sequence. As noted, skeletal-moldic porosities, as well as at least some of the vugular porosity (especially those with later-stage dolomite or silica void linings), are interpreted to have formed contemporaneous with dolomitization. Dolomitization is considered to be a solution-precipitation phenomenon whereby the original calcium carbonate precursor is replaced by dolomite. Dolomite has a molar volume about 13% smaller than that of calcite, and an ion-for-ion replacement of calcite should result in a 13% volume decrease, causing a corresponding porosity increase. However, the complete replacement of skeletal grains by dolomite in the Iowa Silurian shows no evidence of any porosity increase, and many dense xenotopic dolomites (after micritic precursors) have porosities far less than 13%. I agree with Smith's (1967, p. 89) observations that there is a general "lack of evidence for a volume decrease resulting from dolomitization" of skeletal grains in the Iowa Silurian, and dolomitization "must be a volume-for- volume and not an ion-for-ion process."

Because a molar volume of dolomite is about 13% smaller than that of the calcite precursor, additional carbonate must have been added to the rock in order to accomplish a volume-for-volume replacement. An obvious local source of extra carbonate would be through dissolution of skeletal grains, and many skeletal-moldic dolomites in the Iowa Silurian may have been produced in such a manner. Vugs, small pores, and intercrystalline porosity may have formed through dissolution of the matrix micrite during dolomitization. Such dissolution provided a local source of carbonate necessary for completion of dolomitization in the remaining sediment. Dolomitization was generally initiated in the carbonate mud matrix (as seen in partially dolomitized limestone facies; Witzke, 1981a), and excess carbonate was initially added to the system by dissolution of nearby skeletal grains. If not enough excess carbonate was available from the skeletal grain fraction to complete the dolomitization process, dissolution of additional undolomitized matrix carbonate would probably serve as the local source.

Why, then, are dolomite-replaced skeletal grains observed in some Iowa Silurian rocks? Although excess dissolution of carbonate is necessary for dolomitization, how much is necessary? Since the molar volume of dolomite is about 13% smaller than that of calcite, a 13% volume decrease in the total system would be expected. It seems reasonable to suggest that no more than a 13% increase in porosity is needed in the entire system in order to complete the process. Therefore, within some intervals, primarily those with an abundance of skeletal grains, not all grains need to be dissolved during pervasive dolomitization. This conclusion fits with the observation that dolomite-replaced skeletal grains are most frequently observed in packstone fabrics (packstones generally have much more than 13% skeletal grains by volume). Many skeletal-replaced packstone and wackestone fabrics in the Iowa Silurian also include scattered to abundant skeletal molds, and, therefore, grain replacement and grain dissolution can clearly take place in the same rock.

Some intervals in the Silurian sequence of Iowa include porosities in excess of 13%, and porosities as high as 39% have been observed in some facies (Witzke, 1981a). Other intervals have porosities far less than 13%, with some dense mudstone fabrics approaching 0% porosity. It is suggested that the excess removal of carbonate from some intervals probably helped to achieve total mass balance in the system as dolomitization went to completion. This implies that dolomitizing fluids circulated through significant intervals, transporting ions between stratigraphic units. In other words, carbonate removed from some intervals was utilized to complete the dolomitization process in other intervals. For example, carbonate removed from very porous and skeletal-moldic facies in the upper Scotch Grove Formation (average 20% porosity) may have been transported and utilized in adjacent more sparsely skeletal-moldic mudstone and wackestone facies (1-10% porosities typical) during dolomitization.

FAUNAL ASSOCIATIONS

The Silurian sequence of eastern Iowa is richly endowed with paleontologic resources, and the succession of faunas provides significant insights into depositional environments, relative sea-level changes, and biostratigraphic resolution. Johnson (1975, 1977b) first interpreted a series of benthic paleocommunities in the Llandoverian and lower Wenlockian sequence in Iowa that partly paralleled those in the Welsh Borderland described by Ziegler (1965). Johnson's initial studies were influential in establishing the relative bathymetric position of various benthic associations, and, thereby, in inferring relative changes in water depth from the succession of faunas. Witzke (1981a, 1983) built upon Johnson's lead and recognized additional benthic faunal associations higher in the stratigraphic sequence (upper Scotch Grove and Gower formations). These studies were integrated by Witzke and Johnson (1992) into a series of Boucot-style benthic communities, primarily named after characteristic brachiopod taxa or coralline associations. Echinoderms are extremely important sediment contributors in many Silurian units, and diverse echinoderm faunas are recognized in several stratigraphic intervals (Witzke, 1976). A series of echinoderm-dominated benthic communities were delineated for the Iowa Silurian by Frest et al. (1992).

Portions of the contained brachiopod faunas, especially the pentamerids and stricklandiids, have been described and illustrated in the literature (Johnson, 1979), but much of the fauna has never been monographed. The Iowa Silurian coral and stromatoporoid faunas have not been fully documented, and listed taxa are tentatively identified to generic and family level only. Some Iowa Silurian trilobite faunas have received preliminary study (Mikulic, 1979), and portions of the echinoderm faunas have been systematically described (Witzke, 1976; Witzke and Strimple, 1981). The remainder of the fauna has received little or no systematic study, and much paleontologic study remains to be accomplished.

A series of generalized faunal associations, some of which recur at more than one stratigraphic position, are briefly outlined in this section. These associations provide a means of contrasting and comparing faunal content in various stratigraphic units, which, by inference, will provide a basis for recognizing similarities or differences in depositional environments. Brachiopod faunas are particularly useful for such comparisons, but coralline and echinoderm-rich faunas are significant in some units. The Iowa Silurian benthic associations were grouped into two general environmental categories (Witzke, 1983): 1) open marine (stable, normal marine salinities) and 2) restricted marine (varying degrees of hypersalinity). Open-marine associations typically contain diverse benthic assemblages with common to abundant echinoderm debris. By contrast, restricted-marine associations are generally characterized by less diverse faunas that lack an echinodermal component.

Coralline Faunas

Scattered tabulate corals occur through much of the Iowa Silurian sequence, but several Llandoverian units contain a distinctive and recurring association of tabulate corals and stromatoporoids (the Tabulate Coral-Lamellar Stromatoporoid Community of Witzke and Johnson, 1992). The conspicuous fauna of tabulate corals (especially tabular *Favosites, Halysites*) co-occurs with distinctive flattened disc-shaped stromatoporoids (*Ecclimadictyon*, possibly others). The corals and stromatoporoids, which typically range between about 5 and 50 cm in diameter, are commonly partially silicified. Less common rugose corals (solitary forms, *Arachnophyllum*) are often present. Fine crinoidal debris consistently occurs with this coralline association, commonly forming packstones, but the coralline component is the most conspicuous macrofauna visible on outcrop. Rare brachiopods (especially *Flabellitesia*), trilobites, bryozoans, and molluscs also co-occur in the association.

An additional coral-rich benthic association occurs in the middle Scotch Grove Formation, termed the Tabulate-Rugose Coral Community by Witzke and Johnson (1992). Tabulate (especially *Favosites*), solitary rugose (*Cystiphyllum*, *Ptychophyllum*, *Goniophyllum*, etc.), and colonial rugose (*Arachnophyllum*, "*Diphyphyllum*", etc.) corals are abundant, but stromatoporoids (*Clathrodictyon*, *Stromatopora*) are proportionately less common. Colonial corals are represented by massive, hemispherical, fasciculate, and tabular morphologies, which contrasts with the dominant tabular morphologies present in the Tabulate Coral-Lamellar Stromatoporoid Community; the varied morphologies probably represent in part a hydrodynamic response to less agitated environments (ibid.). Individual colonies of "*Diphyphyllum*" are known to reach dimensions to 15 m in the vicinity of STOP 1 (Thomas, 1917). This distinct coralline association co-occurs with scattered brachiopods, especially *Pentameroides*, and additional fauna includes inadunate crinoids (especially *Petalocrinus*), molluscs, trilobites, and bryozoans. This community is not geographically widespread in eastern Iowa, unlike the Tabulate Coral-Lamellar Stromatoporoid Community, and probably forms localized coralline bank or biostromal facies.

Additional coralline faunas are recognized in various carbonate mound and inter-mound facies, both in open-marine and restricted-marine associations. These are noted in subsequent paragraphs.

Pentamerid Associations

Pentamerid brachiopods are conspicuous macrofaunal elements in a number of stratigraphic intervals in the eastern Iowa Silurian sequence. These form distinctive open-marine faunal associations, and the pentamerid shells variably occur in a preservational spectrum ranging from disarticulated valves to whole shells in life position. Many of the pentamerid occurrences apparently represent storm accumulations. Many pentamerid-bearing units intergrade (both stratigraphically and paleontologically) with coral-stromatoporoid associations on one hand, and with stricklandiid faunas on the other, suggesting an intermediate position for the pentamerid-dominated associations. Several pentamerid genera are recognized at different stratigraphic positions in the Iowa Silurian.

The oldest pentamerid (virginianid) association occurs near the base of the Waucoma Formation (correlative with lower Blanding strata), where an abundance of virgianids (aff. *Platymerella*) is noted at a single locality. *Pentamerus oblongus* is abundantly represented at two general stratigraphic positions in the Hopkinton Formation, commonly forming virtual monospecific shell accumulations. The Marcus Member contains dense *Pentamerus* accumulations over a broad geographic area that covers an area larger than the entire Silurian outcrop belt of eastern Iowa and northwest Illinois. *Pentamerus* locally ranges upward into the basal Farmers Creek Member, where it co-occurs with *Stricklandia*. *Pentamerus* accumulations also occur in the lower Picture Rock area, but these are not as geographically widespread.

A distinctive globular-shaped pentamerid, *Harpidium* (*Isovella*) *maquoketa*, is abundant in parts the Farmers Creek Member, particularly in the upper part of the member in the outcrop belt. *Harpidium* shell accumulations resemble those seen for the *Pentamerus*-bearing units, although *Harpidium* faunal associations are typically more diverse and include a variety of additional brachiopod taxa as well as significant molluscan and echinodermal components and calcareous algae (*Cyclocrinites*) (Witzke and Johnson, 1992).

Pentameroides subrectus, a phyletic descendent of *Pentamerus*, is abundantly represented at two general stratigraphic positions in the lower and middle Scotch Grove Formation in the Johns Creek Quarry and Buck Creek Quarry members. Accumulations of *Pentameroides* shells are commonly disarticulated valves in tempestite layers, but specimens in life position are also found. *Pentameroides* most commonly occurs in high dominance, moderate to low diversity faunal associations. Dense accumulations of *Pentameroides* are not geographically or stratigraphically continuous, and are apparently restricted to discontinuous bank-like developments. A variation on this community includes accumulations of *Pentameroides* (*Callipentamerus*) *corrugatus* in the mid Scotch Grove, commonly associated with *Costistricklandia*. *Callipentamerus* co-occurs with *P. subrectus* at some localities, but also locally forms monospecific accumulations in a basinward direction. Although the presence of corrugated ornament and shorter outer plates generally separates *Callipentamerus* from *P. subrectus*, a degree of apparent gradation between the two forms is noted where they co-occur. *Callipentamerus* is tentatively interpreted as a slightly deeper water ecophenotypic variant of *Pentameroides subrectus*. Further biometric study is needed.

Additional pentamerid associations occur in carbonate mound and inter-mound strata in the upper Scotch Grove and Gower formations. A large ribbed pentamerid, *Rhipidium*, is noted in horizontal strata in the upper Waubeek Member, and dense accumulations locally occur. In addition, *Rhipidium* occurs, sometimes abundantly, within small carbonate mounds of the Palisades-Kepler Member (Cedar Co.). An additional ribbed pentamerid, *Harpidium* (*Lissocoelina*), locally forms scattered to dense accumulations in the central to flanking beds of carbonate mounds of the Palisades-Kepler and LeClaire members in eastern Iowa and northwestern Illinois. Finally, clusters of an another large pentamerid, *Harpidium* (*Isovella*) sp. (aff. *H. luckeyensis*), locally occur in the central core of Brady Member carbonate mounds, associated with trimerellids and other brachiopods.

Stricklandiid and Related Associations

Johnson (1975, 1979) recognized recurring benthic faunal associations in the Iowa Llandoverian characterized by the presence of distinctive stricklandiid brachiopods. Based on biometric comparison with typical European specimens, Johnson assigned the Iowa stricklandiids to *Stricklandia* (in the Hopkinton) and *Costistricklandia* (in the Scotch Grove). I concur with his assignments, although it should be noted that these Iowa stricklandiids were given independent generic names by Boucot and Ehlers (1963), *Microcardinalia* and *Plicostricklandia*, respectively. Based on analogy with the Welsh communities (Ziegler, 1965), Johnson interpreted the Iowa stricklandiid associations to represent the deepest-water paleocommunities in the Silurian sequence.

The oldest occurrences of stricklandiids in the Silurian sequence are from the lower Blanding Formation. These include *Stricklandia* sp. (aff. *S. lens intermedia*) from the subsurface of southeast Iowa (identifications courtesy M. Johnson, 1990, pers. comm.), and an undetermined stricklandiid (presumably the same taxon) from the outcrop belt of northwestern Illinois (Witzke and Johnson, 1992; Frest et al., 1992). The lower Blanding faunal associations differ from those in the upper part of the formation, and include sparse brachiopods (especially resserellids), bryozoans, small corals, and trilobites in a mudstone to wackestone matrix. These lower Blanding faunas probably are biofacies equivalents of the scattered stricklandiid occurrences elsewhere in the area.

Abundant stricklandiids occur slightly higher in the sequence (mid to upper Sweeney Member), where scattered to dense accumulations of *Stricklandia lens progressa* occur with tabulate corals and other fossils in high-dominance associations. This stricklandiid species locally co-occurs with *Pentamerus* in the upper Sweeney and lower Marcus members. *Stricklandia laevis* is common in the lower to mid Farmers Creek Member, occurring in moderate to high diversity faunal associations. Unlike older Iowa stricklandiid associations, *S. laevis* co-occurs with a diverse brachiopod assemblage, in addition to noteworthy molluscan, echinodermal, and coralline components. The distinctive calcareous alga, *Cyclocrinites*, is also a common associate. *Pentamerus* locally co-occurs in the basal part, and *Harpidium* assumes dominance upward in the Farmers Creek sequence (the *Harpidium-Stricklandia* Community of Witzke and Johnson, 1992).

Ribbed stricklandiids are common in the lower to mid Scotch Grove Formation, primarily *Costistricklandia castellana*. *Costistricklandia* first occurs in the lower Scotch Grove, Johns Creek Quarry Member, associated with *Pentameroides*. Slightly higher in the lower Welton Member (and locally in the lower Buck Creek Quarry Member), *C. castellana* occurs in exceptionally diverse faunal associations without *Pentameroides*, the *Costistricklandia-Eospirifer* Community of Witzke and Johnson (1992). These strata contain the most diverse benthic fauna known from the entire Silurian sequence of Iowa. In addition to the name-bearers, typical brachiopods include *Atrypa*, *Camerella*, *Ferganella*, *Cyrtia*, *Dalejina*, and others. Bryozoans, especially fenestellids and branching cryptostomes, are extremely abundant in some beds, and small tabulate corals and solitary rugosans (especially *Porpites*) are common. A relatively diverse trilobite (Mikulic, 1979) and molluscan fauna is noted, and a truly exceptional echinoderm fauna containing in excess of 30 genera is identified (Witzke and Strimple, 1981; Frest et al., 1992).

Costistricklandia castellana co-occurs with *Pentameroides* and *Callipentamerus* in the mid Scotch Grove, but stricklandiid abundance is subordinate to the pentamerids and locally abundant rugose-tabulate corals. These stricklandiid occurrences are the youngest recognized in Iowa, dominantly *C. castellana* but also locally including the more finely ribbed *C. multilirata* in the same interval.

Some faunally-related benthic associations occur in the lower to mid Scotch Grove Formation that do not contain stricklandiids, but share many taxa in common. These are briefly considered here, although the absence of stricklandiids may suggest slightly shallower deposition or other minor environmental differences. Witzke and Johnson (1992) recognized the *Ferganella* Community in lower to mid Scotch Grove (Welton) strata, named after a distinctive rhynchonellid brachiopod. This community shares many brachiopod taxa in common with the *Costistricklandia-Eospirifer* Community (including *Ferganella, Atrypa, Eospirifer, Cyrtia, Camerella, Dalejina*, etc.), and further contains virtually identical coralline, bryozoan, and echinoderm faunas. An additional association, the Bryozoan-*Dicoelosia* Community (ibid.) in the lower Scotch Grove (lower Welton and lower Buck Creek Quarry members), is characterized by an abundance of bryozoans (fenestellids and branching cryptostomes) with scattered brachiopods (*Atrypa, Dicoelosia*, others) and other fossils. This bryozoan-rich association shares lateral facies relationships with *Costistricklandia*-bearing lower Welton strata.

Open-Marine Carbonate Mound Associations

Carbonate mounds ("reefs") in the lower and upper Scotch Grove Formation contain distinctive benthic faunal associations, and similar faunas also occur in the LeClaire Member of the Gower Formation. These faunas are interpreted to be stenohaline (open-marine) and commonly include abundant echinoderm debris. Faunal associations are grouped for convenience into two general categories based on their relative positions within the mounded features: 1) central mound (so-called "reef core" of some workers) and 2) mound flanks (commonly with graded beds and debris flows). Although much has been made of "framework" cores within various Silurian "ecologic reefs" of the Midwest and Great Lakes area (e.g., Shaver, 1989, 1991), no evidence has yet come to light in the Silurian carbonate mounds of Iowa that unequivocally identifies a framework core at any known locality. Instead, the central mound facies in the Scotch Grove Formation are primarily composed of discrete beds of carbonate mud and fine to coarse skeletal debris (especially crinoidal) with varying amounts of relict marine cements.

Tabulate corals and stromatoporoids occur in varying abundance within the central mounds at some localities, but these are often patchy in their distribution and locally absent. In general, corals in the central mounds of the Palisades-Kepler and LeClaire members typically do not exceed 20 to 30 cm in diameter. Favositids are generally dominant, but tabular to hemispherical stromatoporoids are locally noted. Corals most commonly are overturned or at varying attitudes, but locally are preserved in growth position within the carbonate matrix (Philcox, 1971). Corals are locally more abundant in the central mounds of the Johns Creek Quarry Member (e.g., STOP 2), where individual colonies in growth position are known to reach dimensions to 2 m high (more commonly 10-40 cm) (Philcox, 1970a). The largest corals include fasciculate tabulates ("*Syringopora*") and rugosans ("*Diphyphyllum*"), which may have acted as local sediment baffles, but no boundstone framework pervades the central mound areas. Carbonate mud and crinoidal debris are the dominant components in the central mound areas, although brachiopods and molluscs are locally noted. Scattered "pockets" within the central mounds contain accumulations of trilobites (especially *Bumastus*) or nautiloids.

The mound flanks contain faunas similar to those of the central mounds, although the skeletal material is typically segregated in graded debris flows and wedge beds. Mud is generally less abundant than in the central mound areas. Crinoid stems and plates are generally dominant among the skeletal debris, commonly displaying packstone (some grainstone) fabrics with scattered relict marine cements. A variety of crinoid cups have been identified, although *Eucalyptocrinites* is typically dominant. Fenestellid bryozoans are common in some beds. Additional fauna is present in lesser abundance, including brachiopods, gastropods, bivalves, nautiloids, corals, calcareous algae, and trilobites. Brachiopod faunas in both central mounds and mound flanks are similar, and include pentamerids (see earlier discussion), rhynchonellids (*Stegerhynchus, Rhynchotreta, Ferganella*), atrypids, spiriferids (*Hedeina*), orthids (*Dalejina*, others), strophomenids (*Leptaena, Protomegastrophia*), and trimerellids (*Dinobolus, Trimerella*).

Open-Marine Inter-Mound Associations

Horizontally-bedded strata in intermound positions within the lower and upper Scotch Grove Formation contain faunas distinctive from those in the coeval carbonate mounds. Inter-mound strata of the Johns Creek Quarry Member are dominated by mud-rich lithologies generally with sparse faunas. However, dense accumulations of *Pentameroides* are locally noteworthy. Laterally discontinuous skeletal-rich (crinoidal to coralline) biostromal units (horizontally bedded) display some faunal similarities with the central mound faunas, particularly in the local occurrence of large (to 90 cm) corals (*Syringopora*, "*Diphyphyllum*", *Halysites*), overturned in part (e.g., STOP 1). These biostromal intervals differ from the central mounds, however, not only in bedding relations, but also in the occurrence of faunal components not noted in the mounds (especially *Pentameroides*, *Petalocrinus*).

Inter-mound strata in the upper Scotch Grove Formation contain open-marine level-bottom faunal associations that differ considerably from faunas in the mound facies. Inter-mound facies at most localities are dominated by sparse skeletal debris, primarily fine crinoidal material (not like the coarse crinoidal material in the mounds) with scattered brachiopods and other fauna. Skeletal debris increases in abundance adjacent to the mounds. Several faunal associations have been recognized in inter-mound strata of the Waubeek and upper Buck Creek Quarry members (Witzke and Johnson, 1992). The *Hedeina*-Gypidulid Community (ibid.), or variations of this community, characterize Waubeek faunas at most localities; these are moderate to high diversity low-dominance associations. Brachiopods commonly include *Hedeina*, *Atrypa*, *Gypidula*, *Resserella*, and *Dalejina*, but many additional taxa are present. Small tabulates and solitary rugosans are generally present, and bryozoan, molluscan, and trilobite (calymenids, encrinurids, proetids) components are commonly associated. Fine crinoid debris occurs in the skeletal mudstones, probably dominated by inadunate crinoids. However, crinoid diversity generally increases adjacent to the mounds, and camerate crinoid taxa, allied in part with mound faunas, become noteworthy (*Dimerocrinites* locally dominant). Inter-mound accumulations of large pentamerids (*Rhipidium* Community), locally occur in a faunal association containing elements shared with the *Hedeina*-Gypidulid Community.

An inter-mound faunal association (*Porpites* Community of ibid.) is noted in cherty upper Scotch Grove strata, characterized by an abundance of the small "button" coral *Porpites* (or *Palaeocyclus*). This distinctive coral occurs throughout much of the Scotch Grove sequence, but nowhere in the relative abundance seen in this community. Associated fauna resembles that seen in other upper Scotch Grove inter-mound units, but diversity is reduced.

Restricted-Marine Associations

Distinctive faunal associations, generally of high-dominance and low to moderate diversity, occur in units of the Gower Formation (especially Brady Member) that interbed down-dip with barren laminated carbonates of the Anamosa Member. These Gower units lack several stenohaline faunal groups that consistently occur in underlying open-marine associations, particularly the echinoderms, bryozoans, trilobites, and a number of characteristic brachiopods (Witzke, 1983). The absence of many stenohaline elements is interpreted to be a response to varying salinity stresses during Gower deposition (Witzke, 1987). These Gower faunas are categorized as restricted-marine, reflecting benthic salinity stresses that are interpreted to have ranged from near normal marine into hypersaline conditions (at or near the precipitation of gypsum). No evidence for significant freshwater input has been identified, and salinity stresses probably did not range into brackish conditions.

Skeletal-rich carbonate mound facies of the Brady Member contain distinctive faunal associations unlike any seen in underlying strata. These skeletal lithologies are known to interbed down-dip from the mounds into flat-lying strata of the Anamosa Member. Similar faunas locally occur in thin units within Anamosa strata with no demonstrable relationship to nearby mounds, and these may represent distal accumulations or short-lived events that permitted episodic benthic habitation in the otherwise hostile Anamosa environments. Shelly faunas of the Brady and Anamosa members are generally dominated by brachiopod-rich accumulations, characterized by an abundance of rhynchonellids and/or protathyrids (the Rhynchonellid-Protathyrid Community of Witzke and Johnson, 1992). The dominant rhynchonellid cannot be assigned with confidence to any genus, and is likely an undescribed taxon. Protathyrid brachiopods have previously been assigned to *Protathyris* and *Hyattidina*, but some (or most) can probably be included in *Greenfieldia*. Protathyrids display delicate spiralia in the hollow shell interiors at some localities. Brachiopod diversity is low to moderate, and associated taxa locally include *Atrypa*, *Meristina* (locally dominant), *Spirinella*, and *Coolinia*. Brachiopods numerically overwhelm this association, and literally millions of specimens can be seen at some localities. However, additional fauna is locally noteworthy, which includes small tabulate corals, *Pycnostylus*, gastropods, bivalves, and ostracodes.

Laterally contiguous and locally gradational with the Rhynchonellid-Protathyrid Community in the Brady mounds is a coral-rich variant, the *Pycnostylus*-Tabulate Coral Community (ibid.). This community is characterized by dense clusters or solitary individuals of the rugosan *Pycnostylus*, commonly associated with small tabulate corals (favositids, *Halysites*). Although most typical of the Brady Member, this community also locally occurs within the LeClaire Member. Scattered *Pycnostylus* are noted within crinoidal open-marine facies of the LeClaire Member at some localities, indicating that associations of this coral occur in both restricted and open-marine settings. The LeClaire Member contains both restricted and open-marine faunal associations, and commonly displays crinoidal mound biofacies closely similar to those seen in the Palisades-Kepler Member. However, non-crinoidal units within both mounded and non-mounded facies of the LeClaire Member contain faunas similar to those seen in the Brady; a variant of the Rhynchonellid-Protathyrid Community locally contains additional taxa, including a unique atrypid (?*Nalivkinia*).

Inter-mound strata in the Gower Formation are primarily characterized by unfossiliferous laminated strata of the Anamosa Member, although, as noted, interbeds containing faunas of the Rhynchonellid-Protathyrid Community occur. The great bulk of these inter-mound strata display no evidence of shelly or burrowing benthos, and the bottom environments must have been hostile to most invertebrates, probably due to hypersaline conditions. Nevertheless, enigmatic fossils locally known as "rods" (cylindrical objects about 1 cm by 2 mm in size) occur as scattered to packed accumulations along some laminae within the laminated dolomite sequence, which constitute the "Rod" Community of Witzke and Johnson (1992). Although of unknown biologic affinities, Henry (1972) suggested that the "rods" were soft, gelatinous dwelling tubes of a worm-like organism. The best preserved "rods" display a hollow central chamber, lending credence to their interpretation as dwelling tubes. Clearly, the "Rod" Community was adapted to harsh environmental conditions unsuitable for other organisms. "Rod"-bearing strata are observed to interbed with skeletal units around Brady mound margins with the Rhynchonellid-Protathyrid Community.

SILURIAN STRATIGRAPHY OF EASTERN IOWA

Stratigraphic units in the Silurian dolomite sequence of eastern Iowa are described in this section. Silurian limestone units (Waucoma and LaPorte City formations) will not be examined on this field trip, and additional information on their character and distribution is available elsewhere (Witzke, 1981a; Bowman, 1985). Basal Silurian dolomite formations (Mosalem, Tete des Morts, Blanding) also will not be visited on the trip, but descriptions are included here for reference.

Mosalem Formation

Brown and Whitlow (1960) originally defined the Mosalem as a member of the Edgewood Formation, and Willman (1973) elevated it to formational status. It is named for exposures in Mosalem Township, Dubuque Co., Iowa (Loc. KRC, Fig. 1). The Mosalem Formation is restricted to paleotopographic lows, probably paleovalleys, developed on an eroded Maquoketa (Upper Ordovician) shale surface. The Mosalem Formation ranges in thickness between about 0 and 100 feet (0-30 m) in the outcrop belt of eastern Iowa and northwestern Illinois, and significant variations in thickness are displayed over short distances. It is absent where underlying Maquoketa strata are thickest, and the Mosalem and Maquoketa share generally complementary thickness relations.

The Mosalem Formation is a wavy-bedded, argillaceous dolomite with chert nodules and shale partings. It is argillaceous throughout, but argillaceous content generally decreases upward in the sequence. Argillaceous carbonate mudstone fabrics dominate, but scattered to common skeletal debris (silicified or dolomitized) is present in the middle to upper parts of the formation. Sedimentary structures include horizontal laminations to low-angle ripple laminations. Wavy bedding is characteristic, and small-scale hummocky stratification and nodular bedding is locally noted in the upper part of the formation. Chert nodules are present in varying quantities at different sections, and nodular chert bands are commonly present in the middle to upper parts of thick Mosalem sections. A basal conglomeratic zone is locally recognized, containing clasts of reworked material from underlying Maquoketa strata. Where the Mosalem is thin, a stromatolitic bed may be locally present at the base (Johnson, 1977).

Except for an abundance of *Chondrites* and other burrows, the Mosalem is only sparsely fossiliferous in most beds. However, two general benthic associations are recognized (Witzke and Johnson, 1992): 1) a lower Mosalem *Lingula*-Orbiculoid Community, and 2) an upper Mosalem *Dalmanella- Eospirigerina* Community. Carbonaceous debris in the lower interval includes graptoloid and dendroid graptolites, and a soft-bodied fauna is locally preserved

(*Lecthaylus-Sphenothallus* Community of Frest et al., 1992). Recovered graptolites from the lower Mosalem suggest an Early Llandoverian (Rhuddanian) age (Ross, 1964), and the formation is constrained below mid-Llandoverian (early Aeronian) strata of the Blanding Formation.

Tete des Morts Formation

Brown and Whitlow (1960) originally defined the Tete des Morts as a member of the Edgewood Formation, and Willman (1973) elevated it to formational status. Its type locality (King roadcut; Loc. KRC, Fig. 1; ibid.) lies in the drainage of the Tete des Morts River in Dubuque Co., Iowa. The formation is recognized through much of the Silurian outcrop belt of eastern Iowa and adjacent northwest Illinois and southwest Wisconsin, but it loses its identity westward and southward in the subsurface. It ranges between 5 and 25 feet (1.5-7.5 m) in thickness in the outcrop belt, but significant thickness variations are observed over relatively short distances (Willman, 1973). The Tete des Morts conformably overlies the Mosalem at most localities, but it unconformably overlies the Maquoketa Formation (Ordovician) at some localities where the Mosalem is absent.

The Tete des Morts was originally described as a cliff-forming, massive-bedded dolomite, in part glauconitic (Brown and Whitlow, 1960). The dolomites are generally very fine to medium crystalline, with some medium to coarsely crystalline lithologies displaying crinoidal packstone fabrics in part. The massive dolomites weather to produce rough uneven to pitted vertical surfaces. The formation is characteristically thick to massive bedded, and is commonly exposed in bold faces above the recessive slopes of the underlying Mosalem and Maquoketa formations.

The Tete des Morts contains a fauna similar to that seen in parts of the overlying Blanding and Hopkinton formations. Lamellar stromatoporoids and corals (*Favosites*, solitary rugosans) are common, with scattered brachiopods noted. The stratigraphic position above the Mosalem and below the Blanding suggests a mid-Llandoverian (late Rhuddannian and/or early Aeronian) age.

Blanding Formation

Willman (1973) named an interval of cherty dolomite the Blanding Formation after the village of Blanding, Jo Daviess Co., Illinois (type section, Loc. BLG, Fig. 1). This interval was incorrectly labelled the "Kankakee" Formation in many previous studies. The Blanding Formation is recognized across most of east-central and northeast Iowa and northwest Illinois, with outliers noted in southwest Wisconsin. The Blanding Formation varies regionally in thickness, being thickest in northeast Iowa and adjacent Illinois (35 to 60 ft; 10 to 18 m). It is 50 to 55 feet (15-17 m) thick in the type area. The formation thins southward and southwestward from the outcrop belt into the subsurface across east-central Iowa, generally ranging between 20 and 35 feet (6-10 m) in thickness across much of its subsurface extent. The Blanding is less than 20 feet (6 m) thick in southwestern areas of the eastern Iowa subsurface. The Blanding is characterized by its very cherty character, but the interval loses this distinction in areas of northeast and southeast Iowa.

A high chert content distinguishes the Blanding Formation in both subsurface and outcrop sections. Chert commonly forms about 20-30% by volume of the formation. Chert is recognized both in nodular bands and discrete chert beds. The formation displays decreasing chert content westward from the northeast Iowa outcrop, and becomes progressively more difficult to distinguish westward in the northern subsurface. Dolomites of the Blanding Formation vary between dense mudstones, slightly argillaceous, and skeletal-moldic and skeletal- replaced wackestones to packstones (especially upward), very fine to medium crystalline. The basal unit of the formation in the outcrop belt, generally 5 to 19 feet (1.5-5.8 m) thick, is non-cherty and well-bedded, characterized by dense, very finely crystalline lithologies, slightly argillaceous in part. This unit is the "Lower quarry beds" of Calvin and Bain (1900), which contrasts notably from overlying cherty strata. The lower non-cherty and upper cherty units are informally termed the Dyas and Potters Mill members, respectively, on Figure 4. These members will be formally proposed in a forthcoming detailed report on the Silurian stratigraphy of Iowa (Witzke, in preparation). The type section for both members will be designated in the roadcut west of Bellevue (see Ludvigson and Bunker, 1988, p. 240).

The Blanding Formation conformably overlies the Tete des Morts Formation throughout most of the outcrop belt, but the Blanding overlies the Mosalem Formation southward in east-central Iowa and Carroll Co., Illinois (where the upper Mosalem is a facies of the Tete des Morts). The Blanding oversteps the Mosalem-Tete des Morts edge southwestward in the eastern Iowa subsurface to directly and unconformably overlie upper Maquoketa (Ordovician) strata.

Lower Blanding strata are sparsely fossiliferous, with occurrences of resserellid brachiopods, bryozoans, and calymenid trilobites noted; stricklandiids (*Stricklandia* sp. aff. *S. lens intermedia*) are locally present. The remainder of the Blanding is generally more fossiliferous, with corals the most noteworthy (especially *Favosites* and solitary rugosans). Lamellar stromatoporoids are common in the upper parts of the formation. Crinoid debris molds are scattered to common, and sponge spicule molds are observed in some beds. Occurrences of *Cryptothyrella* and *Stricklandia* in the formation suggest a mid-Llandoverian (early to mid Aeronian) age for the Blanding. Its position above the Mosalem (Rhuddanian) and below the Sweeney Member (with mid or late Aeronian *S. lens progressa*) further supports the age assignment.

Hopkinton Formation

Originally termed the "Delaware Stage" (Calvin, 1896), a preoccupied stratigraphic nomen, the interval was renamed the Hopkinton Formation (Calvin, 1906) after exposures along the Maquoketa River near the town of Hopkinton in Delaware Co., Iowa (type section, Loc. HOP, Fig. 1). As presently defined, the Hopkinton Formation is comprised of four members. The Hopkinton is recognized as a formation throughout most the Silurian outcrop of eastern Iowa, and the formation trends westward in the subsurface across central Iowa to join with Hopkinton strata in the North Kansas Basin area. Although the Hopkinton has not been introduced as a stratigraphic term by Illinois geologists in their state (e.g., Willman, 1973), it is proposed that the Hopkinton Formation has stratigraphic precedence and utility in northwest Illinois. No stratigraphic distinctions are apparent between the Hopkinton sequence in Iowa and correlative strata in adjacent northwest Illinois, and unification of stratigraphic nomenclature seems desirable. Hopkinton strata in northwest Illinois have previously been referred to the Sweeney, Marcus, and lower "Racine" formations (ibid.).

The Hopkinton Formation in the Silurian outcrop belt of eastern Iowa and northwestern Illinois, where capped by the Scotch Grove Formation, ranges between 100 and 160 feet (30-49 m) in thickness and commonly averages about 130 feet (40 m). Constituent members of the Hopkinton are described in subsequent sections, which serve to characterize the lithic and faunal sequence within the formation. In general, the Hopkinton is dominated by dolomite, skeletal-moldic in part, with very fine to coarse crystalline fabrics. Nodular chert forms a secondary component within the sequence, and chert is generally most common in the lower half of the formation within the outcrop belt. Argillaceous content is remarkably low in most units, although minor argillaceous partings and stylolitic surfaces occur, primarily in the lower half of the interval. The Hopkinton Formation conformably overlies the Blanding Formation throughout the Silurian outcrop belt. Where covered by younger Silurian dolomite units, the Hopkinton is overlain by the Scotch Grove Formation with apparent conformity. This contact may become locally disconformable to the southwest in some subsurface sections. The Hopkinton is entirely of Llandoverian age, mid to late Aeronian at the base and mid-Telychian (C_5) at the top.

Sweeney Member

The Sweeney was introduced by Willman (1973) as a formational term in northwest Illinois, named for the Sweeney Islands in Carroll County adjacent to the type locality in Mississippi Palisades State Park (Loc. PP, Fig. 1). This stratigraphic interval was first described in Iowa as the "lower Coralline beds" by Wilson (1895), and subsequently labelled the "Syringopora beds" by Calvin and Bain (1900). Johnson (1983) proposed that the Sweeney be recognized as the basal member of the Hopkinton Formation in Iowa. The Sweeney crops out over a wide area of eastern Iowa and northwest Illinois, with outliers in southwest Wisconsin. The member is also recognizable in the subsurface across much of Iowa, trending into the North Kansas Basin. The Sweeney Member ranges between 33 and 60 feet (10-18 m) in thickness in the outcrop belt. The member thins southwestward in the subsurface, where thicknesses between 19 and 45 feet (6-14 m) are observed. The Sweeney is conformable and locally gradational with both overlying and underlying units.

The Sweeney Member is characterized by thin- to thick-bedded dolomite, ranging between extremely fine (micritic) and medium to coarse crystalline. The dolomites display skeletal-moldic wackestone and dolomite-replaced skeletal wackestone to packstone fabrics; dolomitized crinoidal packstones, generally medium to coarse crystalline, commonly occur in thin stringers or lenses. Bands with common to abundant large brachiopod molds (stricklandiids, pentamerids) are commonly observed in the middle to upper parts of the member. The member ranges from dense to vuggy. Chert content is variable at different localities, being most common in a zone of finely crystalline dolomites near the middle of the member. Skeletal grain silicification is common in the member, with silicified tabulate corals and stromatoporoids being most conspicuous.

The characteristic Sweeney megafauna is the Tabulate Coral-Lamellar Stromatoporoid Community of Witzke and Johnson (1992). The cherty unit in the middle Sweeney is locally non-coralline, and scattered to common fenestellid bryozoans are observed. Isolated to packed accumulations of stricklandiid brachiopods (*Stricklandia lens progressa*) first appear within or slightly above the middle cherty unit, and stringers of scattered to abundant stricklandiids and pentamerids (*Pentamerus oblongus*) are locally developed in the upper Sweeney. The stricklandiid occurrences in the mid to upper Sweeney suggest a late Aeronian (C_{1-3}) age.

Marcus Member

Willman (1973) introduced the Marcus as a formational term in northwest Illinois, named after the village of Marcus near the type locality in Mississippi Palisades State Park (Loc. PP, Fig. 1). The original description of the type Marcus section by Willman (1973) included characteristic accumulations of *Pentamerus* only in the lower part of the unit, overlain by vesicular massive dolomite in the upper Marcus. However, these distinctive upper vesicular beds were previously assigned to the "*Cerionites* beds" in Iowa (Calvin and Bain, 1900), and the characteristic alga *Cyclocrinites* (formerly *Cerionites*) occurs within the upper part of the type Marcus sequence. However, Willman (1973, p. 30) correlated the "*Cerionites* beds" of Iowa with the lower "Racine" in northwest Illinois, and apparently was unaware that his upper Marcus beds are exactly equivalent to the "*Cerionites* beds." Johnson (1983) assigned the distinctive upper vesicular beds with *Cyclocrinites* to a new stratigraphic unit, the Farmers Creek Member, thereby restricting the definition of the Marcus. Johnson's usage of the Marcus is followed here, restricting the unit to the interval containing prominent accumulations of large brachiopods (*Pentamerus*).

The Marcus Member is recognized across much of the Silurian outcrop belt of eastern Iowa and northwest Illinois, and *Pentamerus*-rich Marcus strata are known to trend into the subsurface in eastern Iowa. Occurrences of *Pentamerus* are not sufficient in themselves to identify Marcus strata, as the Marcus is intended to be a lithostratigraphic unit, not a paleontologic unit, and *Pentamerus* is known to occur in all four members of the Hopkinton. The Marcus is recognized by distinct lithologies that are characterized by abundant accumulations of large brachiopods, usually moldic, below the distinctive Farmers Creek Member. In the absence of such brachiopod-rich lithologies, the Marcus cannot be recognized as a discrete lithic unit. In areas of the Iowa subsurface, the Marcus is not consistently distinguishable from the Sweeney Member, where the Hopkinton inteval below the Farmers Creek is informally grouped as "lower Hopkinton."

Marcus lithologies are characterized by thin- to medium-bedded dolomite, primarily very fine to fine crystalline, but including medium to coarse crystalline crinoidal packstones. "The dense concentrations of brachiopods form a distinctive lithology that contrasts markedly with underlying Sweeney strata" (Witzke, 1985, p. 24). *Pentamerus* shells occur in varied associations: 1) shells in life position, 2) coquinoid accumulations of disarticulated valves, and 3) scattered shells, both articulated and disarticulated. Some horizons display truncation of shells beneath bored hardground surfaces (Johnson, 1977b). A fauna of tabulate corals and lamellar stromatoporoids indistinguishable from that of the Sweeney interbeds with pentamerid-rich intervals in the Marcus. The Marcus Member is conformable and gradational with underlying Sweeney strata, and the two members are, in part, lateral facies equivalents.

Farmers Creek Member

Johnson (1983) introduced the term "Farmers Creek" as a member within the Hopkinton Formation, named after Farmers Creek Township in Jackson Co., Iowa (type section, Loc. SWQ, Fig. 1). These strata were termed the "*Cerionites* beds" by Calvin and Bain (1900) and the "*Cyclocrinites* beds" by Johnson (1975), named after the distinctive green alga. The entire interval of the Farmers Creek was apparently included within the original definition of the Marcus Formation in northwest Illinois (Willman, 1973). However, Johnson's (1983) classification is adopted here because significant lithologic and paleontologic differences clearly separate Farmers Creek strata from underlying *Pentamerus*-rich Marcus strata, and a lithostratigraphic distinction is warranted. The Farmers Creek Member is recognized across much of the Silurian outcrop belt of eastern Iowa and northwestern Illinois. The member occurs as a distinct lithostratigraphic and hydrostratigraphic unit through much of the subsurface of eastern Iowa, regionally forming the dominant water-yielding interval within the Silurian aquifer. Where capped by younger Silurian strata, the Farmers Creek Member varies from 28 to 48 feet (8.5-14.6 m) thick in the outcrop belt of eastern Iowa and northwestern Illinois. It thins southwestward in the subsurface of eastern Iowa, varying between about 6 to 25 feet (2-8 m) in thickness. The Farmers Creek Member sharply but conformably overlies the Marcus Member in the outcrop belt, and a prominent lithologic change marks the contact.

Johnson (1983, p. 14) described the Farmers Creek Member as a "... massive, very finely crystalline to micritic dolomite" with scattered to abundant pentamerid brachiopod molds in the upper two-thirds. The member commonly weathers into steep to overhanging cliff faces on outcrop, forming a massive sequence of dense to vuggy dolomite largely uninterrupted by prominent bedding planes. Small crinoid debris molds are scattered to abundant throughout the member, forming the dominant skeletal fabric in the lower half of the member at many exposures. Skeletal wackestone fabrics dominate, but coquinoid brachiopod and crinoidal accumulations occur in the middle to upper parts of the member in the outcrop belt. Prominent vuggy and vesicular zones are of particular note at most localities. Porous vugular networks commonly coalesce to form solutional conduits and caves within the member, and irregular vuggy surfaces commonly pock-mark the steep exposure surfaces. The member is generally free of chert, but sparse chert nodules locally occur.

The Farmers Creek Member contains the most diverse faunas known within the Hopkinton Formation, although fossils may be difficult to distinguish in the lower half of the member in many outcrop sections. Within the outcrop belt, the Farmers Creek Member displays a faunal succession assigned, in ascending order, to three basic faunal associations (Witzke and Johnson, 1992): 1) *Stricklandia laevis* Community, 2) *Harpidium maquoketa-Stricklandia laevis* Community, and 3) *Harpidium maquoketa* Community. Westward in the subsurface, the first noted community is generally absent, and pentamerid (*Harpidium*) associations become dominant in the lower half of the member (Witzke, 1983). In addition to the characteristic brachiopod faunas, the member has yielded diverse assemblages of echinoderms and molluscs which are unique to the member. Articulated echinoderm cups are identified through much of the member, and diverse assemblages dominated by camerate crinoids are characteristic (Witzke, 1976; Witzke and Strimple, 1981; Frest et al., 1992). Additional fauna includes bryozoans, solitary rugose and tabulate corals, and an interesting trilobite fauna (Mikulic, 1979). The occurrence of *Stricklandia laevis* suggests an early Telychian (C_4) age for the member.

Picture Rock Member

Johnson (1983) proposed the Picture Rock as a member of the Hopkinton Formation for the interval he formerly termed the "*Favosites* beds" (Johnson, 1975). It derives its name from the county park at the type locality in Jones County (STOP 1; Loc. PR, Fig. 1). The member is exposed at numerous localities in eastern Iowa, and is also recognized in northwestern Illinois. It trends westward in the subsurface of eastern Iowa. The Picture Rock Member is known to range in thickness in the outcrop belt between 17 and 50 feet (5-15 m). It thins westward in the subsurface, generally 4 to 16 feet (1-5 m), but it is locally absent or unrecognizable in the Iowa City area. The Picture Rock Member sharply but conformably overlies the Farmers Creek Member in the outcrop belt. In some subsurface sections, the contact between the two members appears gradational, and locally the two members are not clearly distinguishable. Where capped by younger Silurian dolomite units, the Picture Rock is sharply overlain by basal units of the Scotch Grove Formation. This upper contact is apparently conformable through the outcrop belt, but erosional bevelling beneath the Scotch Grove is suggested in some subsurface sections.

Johnson (1983, p. 16) described the Picture Rock Member as a "very thick bedded or sometimes massive, medium crystalline, highly vuggy dolomite." The massive character of the member is disrupted by irregular and discontinuous bedding breaks, locally weathering into thin- to medium-bedded intervals. The dolomites are generaly fine to medium crystalline, but thin coarse crystalline horizons are present (preserving crinoidal packstone fabrics). Much of the dolomite displays small skeletal debris molds, and wackestone to mixed wackestone-packstone fabrics are typical. The member is non-cherty throughout much of the outcrop belt, but silicified corals and stromatoporoids are common. The Picture Rock commonly forms bold cliff faces in natural exposures.

The Picture Rock Member contains a fauna closely similar to that seen in the Sweeney Member, with a conspicuous macrofauna of lamellar stromatoporoids and tabulate corals. Basal strata locally include horizons rich in large pentamerid brachiopods (*Pentamerus oblongus*). Stratigraphic position indicates a mid Telychian ($C_{4\text{-}5}$) age for the member. Conodonts recovered from the upper Picture Rock (Witzke, 1981a) include *Llandoverygnathus celloni, Aulacognathus* *bullatus*, *Distomodus staurognathoides*, and *Ozarkodina polinclinata*, consistent with assignment to the *celloni* Zone (C_5).

Scotch Grove Formation

Witzke (1985, p. 26) formally proposed the Scotch Grove as a formational term for the "dolomite and cherty dolomite interval above the Picture Rock Member and below the base of the laminated and mounded dolomites of the Gower Formation." The Scotch Grove was previously introduced as a informal stratigraphic term by Witzke (1981a, 1983). It derives its name from the type locality in Scotch Grove Township, Jones Co., Iowa (Loc. SCG, Fig. 1). No single locality exposes the entire formation, and only the lower portion of the formation (40 ft; 12 m) is preserved in the type section. Constituent members, described subsequently, serve to characterize the lithic sequence within the formation. The Scotch Grove Formation is exposed over broad areas of eastern Iowa and northwestern Illinois, and it trends westward in the subsurface across much of eastern Iowa. Additional subsurface control indicates that the formation extends westward into central Iowa and the North Kansas Basin. Southeastward from the outcrop belt, the formation is probably replaced by units assigned to the Joliet and lower Racine formations in the subsurface of central Illinois.

Where capped by the Gower Formation in the subsurface of eastern Iowa, the Scotch Grove is known to vary from 94 to 170 feet (29-52 m) in thickness. The Scotch Grove Formation is erosionally truncated beneath Devonian strata westward, although cored sequences as thick as 209 to 240 feet (64-73 m) are documented in these areas (Witzke, 1981a) providing constraints on maximum thicknesses. "The maximum thickness of the Scotch Grove is not precisely known, although considering the vertical dimensions of carbonate mounds in the upper Scotch Grove . . . the Scotch Grove Formation reasonably can be inferred to reach thicknesses to 300 feet (90 m)," within grabens along the Plum River Fault Zone (Witzke, 1985, p. 28). The Scotch Grove Formation sharply but conformably overlies the Hopkinton Formation throughout the outcrop belt, although possible disconformable relations may occur in Johnson County. The Scotch Grove is overlain by the Silurian Gower Formation in part of the outcrop belt, but Scotch Grove strata are truncated westward in the subsurface beneath Devonian units.

The Scotch Grove Formation is a relatively pure dolomite and cherty dolomite sequence. Distinctive dolomite lithofacies characterize the formation: 1) flat-lying, dense, cherty to very cherty, sparsely fossiliferous dolomite; 2) flat-lying, dense non-cherty, sparsely fossiliferous dolomite; 3) flat-lying, porous, fossiliferous dolomite (common skeletal molds, some packstones); 4) mounded ("reefal") crinoidal dolomite (wackestone, packstone, some grainstone); and 5) mounded, dense, sparsely fossiliferous dolomite (mudstone to wackestone). These general lithofacies are assigned member status within the formation, as discussed below. Fossiliferous units in the formation display small molds (0.5-2 mm) of echinoderm debris in varying abundance, and crinoid dorsal cups are common in some beds (especially camerate crinoids). Distinctive arm-fans of the inadunate crinoid *Petalocrinus mirabilis* are restricted to the lower and middle parts of the formation. Bryozoan debris is particularly common in the lower part of the formation, and small solitary corals are scattered through much of the formation. Larger colonial corals are sparse to absent through most of the formation, but occur in some units. Diverse faunas are recognized in the Scotch Grove, and associations of various brachiopod taxa are especially instructive (see earlier discussion).

The lowest Iowa occurrences of *Pentameroides subrectus* and *Costistricklandia castellana* in basal Scotch Grove strata indicate a mid or late Telychian (late Llandoverian; C_{5-6}) age (Johnson, 1979). Conodont faunas from the lower Scotch Grove are consistent with assignment to the *celloni* and *amorphognathoides* Zones, supporting the brachiopod correlations (Witzke, 1981a). The middle part of the Scotch Grove (with the highest *Pentameroides* and *Costistricklandia*) contains the highest occurrences of *amorphognathoides* Zone conodonts, suggesting an early Wenlockian age for these strata. The recovery of the graptolite, *Monograptus priodon*, in the mid Scotch Grove further supports an early Wenlockian age (ibid.). Upper Scotch Grove strata have been more difficult to constrain biostratigraphically, but the presence of *Rhipidium* and a few noteworthy conodonts (*Ozarkodina sagitta rhenana*, *Kockelella* cf. *walliseri*) supports a late early to mid Wenlockian age.

Johns Creek Quarry Member

Johnson (1983) proposed the Johns Creek Quarry Member for strata above the Picture Rock Member and below the Welton Member, which he had earlier termed the "Bioherm Beds" (Johnson, 1975). The member encompasses strata previously referred to the "Upper Quarry beds" of Calvin and Bain (1900). Philcox (1970a) was the first to identify carbonate mound facies (bioherms) at this statigraphic position in Iowa. Witzke (1985) included the Johns Creek Quarry Member as the basal unit of the Scotch Grove Formation. The member derives its name from the type quarry adjacent to Johns Creek in Dubuque Co., Iowa (Johnson, 1983; Loc. JCQ, Fig. 1). The member is known from exposures in east-central Iowa, and it is identified in the subsurface westward in eastern Iowa. It varies significantly in thickness in eastern Iowa, in part a reflection of facies variations within the member. The member is thickest where developed as carbonate mound facies, ranging from about 16 to 50 feet (5-15 m) in thickness. Where characterized by horizontally-bedded inter-mound facies, the member ranges from 1.5 to 16 feet (0.5-5.5 m) in thickness. It is locally absent in areas of Jackson County, where the Welton Member directly overlies the Hopkinton, and in some subsurface sections, where Welton or Buck Creek Quarry strata overlie the Hopkinton.

Four general lithologies are recognized in the Johns Creek Quarry Member: 1) flat-lying, dense, well-bedded, sparsely fossiliferous dolomite; 2) flat-lying dense to porous, fossiliferous dolomite with large brachiopods; 3) mounds of dense, coral-bearing dolomite flanked by crinoid-rich strata; and 4) dense, sparsely fossiliferous dolomite with argillaceous partings or streaks (primarily in the western subsurface). Dolomite lithologies are typically very finely crystalline, but more coarsely crystalline fabrics occur locally in the mound facies. Flat-lying inter-mound facies are typically evenly bedded and well suited for quarrying of building stone (the "Upper Quarry beds" of Calvin and Bain, 1900). Carbonate mound facies and coralline biostromes are known from the outcrop belt. The mounds range to about 40 or 50 feet (12-15 m) in thickness and vary between about 50 to 500 feet (15-150 m) in lateral dimensions. The central portion, or "core," of these mounds is dominated by "micritic" dolomite with scattered to common large colonial corals and echinoderm debris (Philcox, 1970a; Johnson, 1983). The central mounds are, in turn, flanked by dipping beds (0-30°) of differing lithology, dominantly crinoidal dolomite (packstones) in graded wedge-shaped debris flows. The flanking and capping beds primarily post-date deposition of the central mounds, and display, in part, a facies transition into strata of the Welton Member. Relict isopachous and botryoidal submarine cement fabrics are observed in the central mound and flanking facies (STOP 2). Horizontal to gently dipping basal Scotch Grove strata locally are porous and skeletal-moldic with scattered large colonies (to 1 m) of *Halysites* and rugosans, possibly a biostromal variant of the mound facies.

Pentameroides subrectus is scattered to abundant locally in intermound facies of the member, associated with *Costistricklandia castellana* and other brachiopods at some localities. Small echinoderm debris is scattered to common, and distinctive arm-fans of *Petalocrinus* are locally noteworthy. Carbonate mound and bank facies contain a fauna distinct from that of the inter-mound units. Colonial rugosans and tabulates are prominent in some central mound facies, reaching dimensions to 6.5 feet (2 m); the largest corals (>40 cm) are tentatively assigned to "*Diphyphyllum*," *Entelophyllum*, *Halysites*, and *Syringopora*. Crinoidal lithogies which flank and bury the mound facies contain a fauna which includes abundant crinoids (especially *Eucalyptocrinites*, *Manticrinus*, *Dimerocrinites*) with brachiopods (rhynchonellids, *Atrypa*, *Eospirifer*, *Costistricklandia*, etc.), fenestellids, and trilobites (Philcox, 1970a; Frest et al., 1991; Mikulic, 1981). These flanking faunas are dissimilar to those in central mound and inter-mound strata, but are closely allied with lower Welton paleocommunities. A late Telychian age (late C_5 or early C_6) is indicated.

Welton Member

The Welton Member was formally proposed by Johnson (1983) for an interval of fossiliferous dolomite which he had previously termed the "*Cyrtia* Beds" (Johnson, 1975). Although Johnson (1983) assigned the Welton to the upper Hopkinton Formation, Witzke (1985) recognized the Welton as a member of the Scotch Grove Formation. Some previous workers erroneously assigned strata now included in the Welton Member to the LeClaire facies of the Gower Formation (e.g., Norton, 1899, 1901; Rowser, 1929, 1932). The member is named for quarry exposures south of the town of Welton, Clinton Co., Iowa (Loc. BHQ, Fig. 1).

The Welton forms the dominant lithology of the Scotch Grove Formation across much of the Silurian outcrop belt of east-central Iowa and adjacent northwestern Illinois. The member is also known from subsurface sections to the west, trending into the North Kansas Basin area. Maximum thicknesses of the Welton Member in the outcrop belt are unknown due to erosional truncation, although exposed intervals up to 75 feet (23 m) thick are noted. The lower 41 feet (12.5 m) of the member is exposed at the type locality. The member is known to reach thicknesses of 130 to 160 feet (40-50 m) in Muscatine and Scott counties. Thicknesses of 200 to 250 feet (60-75 m) are inferred within grabens along the Plum River Fault Zone in

Jackson and Carroll counties, where virtually the entire Scotch Grove Formation is represented by Welton strata. Westward in the subsurface, Welton strata interbed with other units in the lower to middle Scotch Grove, where the member is not known to exceed 65 feet (20 m) in thickness.

The Welton displays complex stratigraphic relations with other members in the Scotch Grove Formation, although the seemingly complex facies architecture shows regionally significant geographic patterns. In general, the Welton forms the dominant lithologic unit of the formation in the eastern area (Jackson, Muscatine, Clinton, Scott, Carroll counties). In that area the Welton generally overlies a thin Johns Creek Quarry Member and is conformably and gradationally overlain by strata of the Waubeek or Palisades-Kepler members. However, Welton strata extend to, or nearly to, the top of the Scotch Grove Formation in some sections. Welton strata interbed westward in a complex manner with cherty facies of the Buck Creek Quarry Member. This region of interbedding marks a broad facies transition between Welton-dominated sequences to the east and cherty Buck Creek Quarry-dominated sections to the west. Welton lithologies are sufficiently distinct and geographically widespread to warrant member status.

The Welton Member was originally described as a "tan, poorly bedded, very finely crystalline dolomite" containing diverse assemblages of brachiopods and other fossils (Johnson, 1983, p. 17). Witzke (1985, p. 31) characterized the member as a "horizontally-bedded, porous, abundantly fossiliferous (typically very crinoidal), non-cherty dolomite." Porosity is primarily related to skeletal molds, with vugular porosity significant in some beds. Welton strata are notably more fossiliferous than adjacent Silurian units, although some interbeds within the member are sparsely skeletal. Moldic wackestone fabrics are characteristic, but some dolomite-replaced wackestone to packstone fabrics occur. The member appears thick bedded to massive in most exposures.

The Welton contains an abundant and diverse benthic fauna. Small molds of crinoidal debris are generally the dominant fossils, and articulated crinoid cups are common in some beds, typically dominated by camerate genera (especially *Eucalyptocrinites*, *Dimerocrinites*, *Siphonocrinus*, and *Macrostylocrinus*). Flat-branching cryptostomes (cystodictyonids) and fenestellid taxa are the dominant bryozoans in the Welton. The contained brachiopod faunas are particularly significant, and in excess of 35 genera are noted in lower Welton strata (especially *Atrypa*, *Eospirifer*, *Costistricklandia*, *Cyrtia*, and *Ferganella*). Large pentamerids are generally absent in the Welton, but *Callipentamerus* locally occurs in the mid Scotch Grove. Conodonts recovered from Welton exposures are assigned to the *celloni* and *amorphognathoides* Zones of late Telychian to early Wenlockian age. As the Welton locally spans the entire Scotch Grove Formation, it locally includes strata as young as mid Wenlockian.

Buck Creek Quarry Member

Johnson (1983) formally proposed the Buck Creek Quarry as a member of the Hopkinton Formation, named after the quarry on the east edge of the village of Buck Creek, Delaware Co., Iowa (Loc. BCQ, Fig. 1). Johnson (1975) earlier termed this interval the "*Pentameroides* Beds." The Buck Creek Quarry Member is best developed in the western portion of the eastern Iowa Silurian outcrop belt, and it trends westward into the subsurface beneath the Devonian cover all the way into the North Kansas Basin area. The member is generally absent in the eastern Silurian outcrop, but thin cherty intervals included within the member are locally recognized in that area. Where capped by younger Silurian strata, the member reaches maximum thicknesses ranging between 120 and 170 feet (37-52 m). It is thinner where interbedded with Welton or Waubeek strata, commonly 3 to 90 feet (1-27 m). The Buck Creek Quarry Member is locally a lateral facies equivalent of *all* members defined in the Scotch Grove Formation (Witzke, 1981a, 1985). The member can potentially occupy any stratigraphic position within the formation, but is most commonly recognized in the middle to upper Scotch Grove. The Buck Creek Quarry Member is bounded conformably by units of the Johns Creek Quarry, Welton, Waubeek, and Palisades-Kepler members. The Buck Creek Quarry Member in Jones and Linn counties locally occupies a position in the upper Scotch Grove, where it is overlain by strata of the Gower Formation.

The Buck Creek Quarry Member was defined as a interval of "finely crystalline dolomite with numerous white chert horizons" (Johnson, 1983, p. 17). The member is dominantly a dense, sparsely fossil-moldic, very cherty dolomite (skeletal mudstone to wackestone) that locally displays abundant molds (packstones) of large pentamerid brachiopods (*Pentameroides subrectus*). The cherts are dominantly chalky (dolomitic) and form nodular bands in the sequence. Silicified crinoidal and coralline packstone to grainstone fabrics are rarely associated with the pentamerid-rich units. Dolomite crystals are extremely fine to microcrystalline, and the rocks are relatively dense (particularly when contrasted with the Welton or Palisades-Kepler members). Vugular porosity is generally not well developed.

Fossils occur in varying abundance in the Buck Creek Quarry Member. Faunas generally resemble those seen in the Welton Member, although *Pentameroides*-rich biofacies are noteworthy in the Buck Creek Quarry but absent in the Welton. In addition, camerate crinoids are rare in the member, but are common to abundant in the Welton. Accumulations of large pentamerid brachiopods in the Buck Creek Quarry Member are most commonly recognized in middle Scotch Grove strata, but *Pentameroides* is known to range into the lower parts of the formation at some localities. The sporadic distribution of *Pentameroides*-rich units in the member suggests that this large brachiopod formed discontinuous shell-bank facies. Abundant silicified corals locally occur within the member in the mid Scotch Grove (see earlier discussion), and laterally extensive colonies are known in the vicinity of STOP 1 to reach diameters to 50 feet (15 m). Buck Creek Quarry strata in the upper Scotch Grove locally contain abundant *Porpites* (see earlier discussion). Stratigraphic relations show that Buck Creek Quarry strata locally span the entire interval of the Scotch Grove Formation, and the age of the member is thereby constrained from late Telychian to mid Wenlockian.

Waubeek Member

Witzke (1981a, 1983) first introduced the "Waubeek Facies" as a stratigraphic label, and the Waubeek was subsequently designated as a formal member in the upper Scotch Grove Formation (Witzke, 1985). The name derives from the type locality northeast of the town of Waubeek, Linn Co., Iowa (ibid.; Loc. WBK, Fig. 1). Strata assigned to the Waubeek Member are recognized over much of western and southern edge of the Silurian outcrop belt of eastern Iowa, and the member trends westward in the subsurface beneath the Gower Formation or Devonian units. It is recognized in the North Kansas Basin area. Where capped by the Gower Formation, the Waubeek commonly averages about 40 feet (12 m) in thickness through most of its extent, locally reaching thicknesses to 56 feet (17 m). Thinner Waubeek intervals locally occur at the top of the Scotch Grove in the eastern sections. The Waubeek Member is not geographically continuous, sharing lateral facies relations with other members in the upper Scotch Grove (Buck Creek Quarry, Welton, Palisades-Kepler members). The Waubeek forms the dominant inter-mound facies of the coeval Palisades-Kepler mounds. Waubeek strata occupy a position above the highest occurrences of *Pentameroides* in the Scotch Grove sequence, and the member conformably overlies the Welton or Buck Creek Quarry Member regionally. Where capped by younger Silurian strata, the Waubeek is generally overlain by the Anamosa Member of the Gower Formation; the contact is conformable to gradational at most localities, but locally disconformable relations are noted.

Waubeek strata were defined to include "dense to vuggy, sparsely fossil-moldic dolomites," in part with quartz-lined vugs (Witzke, 1981b, p. 12). Although sparsely fossiliferous through most, some beds are locally slightly more fossiliferous and resemble portions of the Welton Member. The dolomites are finely crystalline to microcrystalline. The Waubeek Member lacks chert. Horizontal bedding in generally medium to thick. Although only sparingly fossiliferous, the Waubeek has yielded a variety of fossils. Small crinoid debris molds are scattered through much of the member, but are absent locally in the upper part. Brachiopods are uncommon, although a relatively diverse assemblage is present in the member (*Hedeina*-Gypidulid Community of Witzke and Johnson, 1992). Accumulations of large pentamerids (*Rhipidium*) are locally noted. Additional fauna occurs (see earlier discussion). When compared with the total Waubeek fauna, trilobites (especially encrinurids and calymenids) are proportionately more common in the Waubeek than in most Iowa Silurian units. Uppermost Waubeek faunas locally differ significantly from underlying faunas, characterized by an assemblage of rhynchonellid and protathyrid brachiopods and a general absence of echinoderm debris and trilobites (Rhynchonellid-Protathyrid Community). This rhynchonellid-rich association is allied with faunas in the overlying Gower Formation. Conodonts from upper Waubeek strata include *Ozarkodina sagitta rhenana* and *Kockelella* cf. *K. walliseri* suggesting an early mid-Wenlockian age (*ranuliformis* and *amsdeni* Zones; Witzke, 1981a). Stratigraphic position and faunal constraints suggest an early to mid Wenlockian age for the Waubeek Member.

Palisades-Kepler Member

Complex crinoidal mound and skeletal bank facies are identified in the upper Scotch Grove. Erroneously assigned to the "LeClaire facies" of the Gower Formation by many previous workers, Witzke (1981a, 1983) recognized that these strata pre-date the type LeClaire sequence and proposed an informal name for them, the "Palisades-Kepler Mound facies." The Palisades-Kepler Member was later proposed as a formal member in the upper Scotch Grove Formation (Witzke, 1985), named after the type locality at Palisades-Kepler State Park, Linn Co., Iowa (STOP 5; Loc. PK, Fig. 1). The member shares lateral facies relationships with inter-mound strata of the Waubeek and Buck Creek Quarry members. An additional unit, the Fawn Creek Member, was proposed for flat-lying skeletal-rich strata in the same stratigraphic position (Witzke, 1985). However, it is proposed herein to suppress the Fawn Creek as a formal unit and include both carbonate mound and contiguous flat-lying vuggy and porous skeletal-moldic facies within a slightly expanded Palisades-Kepler Member to simplifly nomenclature. The type Fawn Creek locality, at Wapsipinicon State Park (Jones Co.), is known to laterally adjoin with mounded facies (Witzke, 1981a), indicating the intimate relationship between Palisades-Kepler mounds and similar but flat-lying Fawn Creek strata.

The Palisades-Kepler Member is exposed at numerous localities in eastern Iowa, commonly seen in river cuts and quarries within topographic klints (bedrock knobs). The member is characterized by mounded carbonate buildups ("reefs"), including isolated single mounds or coalesced complexes of mounds and contiguous flat-lying units. The type locality of the member is within a coalesced complex of six or more carbonate mounds that extends laterally for about 1.5 miles (2.4 km) (Witzke, 1987). Single isolated mounds in the member range from about 200 to 3000 feet (60-900 m) in lateral extent, and vertical dimensions vary between about 30 to 200 feet (9-60 m), the range of thickness for the member (Witzke, 1985). Dips within the mounds vary between 0 and 50° (more commonly 0-30°), and significant portions of many mounded complexes are commonly horizontally bedded or nearly so (0-5° dips). The member is known to interfinger with more sparsely skeletal inter-mound strata of the Waubeek and Buck Creek Quarry members. It conformably overlies Welton or Buck Creek Quarry strata above the middle part of the Scotch Grove Formation (above strata with *Pentameroides*).

The dominant lithologies of the Palisades-Kepler Member display skeletal (primarily crinoidal) wackestone to packstone fabrics. Skeletal material is variably seen as molds or as dolomitized grains, and large unit dolomite crystals are commonly seen as volume-for-volume replacements of original crinoidal calcite. Wedge-shaped debris flows in the flanking units are dominantly dolomitized crinoidal packstones with minor grainstone. Extremely fine to fine crystalline dolomite matrix (with varying amounts of skeletal material or molds) volumetrically dominates the mounds, interpreted as replaced carbonate mud. Relict isopachous submarine cements are commonly observed as void fillings in the more grain-rich lithologies. Corals and stromatoporoids are scattered to locally common in the mounds, but no organic framework or boundstone has been identified with confidence at any of the numerous exposures of the member in all of eastern Iowa. Portions of some central mounds are highly fractured to brecciated, interpreted as evidence of extensional fracturing of rigid mounds contemporaneous with deposition. Some mounds display "pockets" filled with trilobite or nautiloid-rich sediments, suggesting the development and filling of large voids or fractures during mound deposition. Horizontally-bedded strata in the member, previously termed the "Fawn Creek facies" (Witzke, 1981a), are dominantly skeletal-moldic crinoidal wackestones with numerous vugs, similar in some respects to Welton strata.

Post-Silurian erosion has bevelled the crests of most of the Palisades-Kepler mounds, although contact relations with overlying Gower strata are displayed at several localities around the mound flanks. The contact between these units is marked by a lithologic change from crinoidal wackestones and packstones below and brachiopod-rich packstones and laminated dolomites above. It is typically abrupt, but gradation over a few meters of thickness is locally seen. The contact between the Palisades-Kepler and Brady (Gower) members may be locally disconformable, and Philcox (1972, p. 704) noted a "possible erosion surface" at one locality, suggesting that the tops of the mounds may have been subaerially exposed prior to basal Gower deposition. Gradation between the two units in more flanking positions indicates, however, that only the crests of the mounds may have been locally emergent.

Faunas of the Palisades-Kepler Member have been discussed in an earlier section (see also Witzke, 1983). The stratigraphic position of the member suggests an early to mid Wenlockian age, and the faunas, although not clearly diagnostic, tend to support this age assignment.

Gower Formation

The Gower Formation is named after the old Bealer's Quarry (presently in Cedar Valley County Park) in Gower Township, Cedar Co., Iowa (Loc. CVQ, Fig. 1), where 108 feet (33 m) of laminated dolomites were exposed during quarry operations (Norton, 1901). Quarrying adjacent to the type locality displays lateral relationships between laminated strata and carbonate mound facies. As defined by Norton, the Gower Formation includes not only the distinctive horizontally-bedded laminated dolomites (known as "Anamosa stone"), but also laterally equivalent mud-rich to skeletal-rich carbonate mounds (originally labelled the "LeClaire Limestone"). Unfortunately, historic usage expanded the "LeClaire" to include strata that underlie laminated Gower rocks (see Introduction). These strata are now included in the Palisades-Kepler Member of the Scotch Grove Formation (Witzke, 1985), and the LeClaire is restricted to include strata that share lateral facies relationships with laminated Gower units, as in the type LeClaire area. Three general carbonate lithofacies are presently recognized in the Gower Formation, each given member rank. Subsequent discussion of each member will serve to characterize lithologies, faunas, and stratigraphic relations within the Gower Formation.

The Gower Formation is exposed along the southern and southwestern margins of the Silurian outcrop belt in eastern Iowa, and its subsurface extent is limited. Gower outliers are also noted within grabens along the Plum River Fault Zone of east-central Iowa (Bunker et al., 1985). Although not formally recognized across the Mississippi River in northwest Illinois, both laminated and mounded units indistinguishable from those seen in Iowa occur there, and it seems reasonable to assign these to the Anamosa and LeClaire members of the Gower Formation (presently included as part of the "Racine Formation"; Willman, 1973). The Gower Formation is erosionally truncated beneath Devonian strata to the west and south in eastern Iowa. Gower sections up to 100 to 150 feet (30-45 m) thick occur within parts of the outcrop belt, the maximum thicknesses noted for the formation. Gower thicknesses are complementary with those of the Scotch Grove Formation on a local scale; where mounds of the Palisades-Kepler Member are present the Gower is thinnest, but above the Waubeek Member (Scotch Grove inter-mound) the Gower achieves it greatest preserved thicknesses. The contact between the Gower and Scotch Grove formations is characteristically sharp and apparently conformable at most localities, but the contact appears to be locally disconformable. The contact is locally gradational over a few meters, as indicated by local interbedding of laminated Gower rocks and Waubeek-like strata (Witzke, 1981b) and by apparent gradation above the flanks of some Palisades-Kepler mounds (see previous section). At the Gower and Anamosa type localities, the formation is unconformably overlain by Middle Devonian carbonates of the Otis Formation.

Available biostratigraphic evidence is not sufficient to constrain the exact age range of the Gower Formation. Its position above early mid-Wenlockian strata suggests that the basal Gower is approximately mid Wenlockian in age. Upper Gower strata may range into the Ludlovian. Occurrences of *Harpidium* (*Isovella*) sp. aff. *H. luckeyensis* and *Harpidium* (*Lissocoelina*) sp. in the Gower are consistent with a late Wenlockian-Ludlovian age assignment (Boucot and Johnson, 1979). Conodont collections (Witzke, 1981a) are of limited value. *Pseudooneotodus* sp. in the Gower includes elements with a single pointed denticle or two completely fused denticles with a blunt tip, and differs from *P. bicornis* with two discrete denticles. If descended from a form like *P. bicornis*, the Gower species must post-date its oldest occurrence, suggesting a Wenlockian or Ludlovian age. Pb and symmetry transition elements from an indeterminate *Ozarkodina* apparatus (*O.* sp.) were recovered from LeClaire strata. The S elements are closely comparable to those of *O. confluens* (Ludlovian-Pridolian), but the Pb elements do not belong to that species (Klapper, 1981, pers. comm.). The biostratigraphic significance of *O.* sp. is unclear, and must await the discovery of associated Pa elements. Other Gower conodont collections include the long-ranging *Panderodus unicostatus* and nondiagnostic symmetry transition elements (aff. *Kockelella*). Additional processing is encouraged.

Anamosa Member

The "Anamosa stage" was named by Calvin (1896) for a sequence of laminated dolomite quarrystone and dense dolomite at Stone City, west of Anamosa, in Jones Co., Iowa (STOP 4; Loc. STC, Fig. 1). These laminated rocks were termed the "Anamosa facies" by later workers (Hinman, 1968; Philcox, 1972; Henry, 1972; Witzke, 1981a). The rock unit was given formal status as a member within the Gower Formation by Witzke (1985). The Anamosa Member forms the dominant facies of the Gower Formation in eastern Iowa. The Anamosa is erosionally truncated beneath

Devonian and Quaternary units, but thicknesses of 100 to 150 feet (30-45 m) are noted. It is 100 feet (30 m) thick at Stone City, at least 108 feet (33 m) at the type Gower section (see above), and up to 150 feet (45 m) at LeClaire, Scott County.

The Anamosa Member is the most distinctive of all Silurian units in Iowa, being characterized by prominently laminated dolomite. Laminae range from less than a millimeter to about 1 cm in thickness, and wavy and planar laminations are recognized. Laminae through much of the Anamosa sequence are laterally continuous and uninterrupted at the scale of individual quarries, and perhaps extend over even broader distances. Some wavy laminated beds resemble stromatolitic mats, but the planar laminations appear varve-like in character. Laminae are marked at a microscopic level by variations in crystal size. Although laminae are generally uninterrupted through most of the Anamosa, laminations become faint and irregular in the highest stratigraphic levels, and upper Anamosa strata locally display disrupted or truncated laminae and intraclastic lithologies. "Birdseye"-like voids are locally present in the upper strata.

The laminated dolomites are commonly interbedded with dense featureless microcrystalline dolomite (to more than 1 m in thickness), some of which can be correlated over several miles (Henry, 1972). Small brown chert nodules are scattered within some sections of the Anamosa Member, as at the type locality. The Anamosa Member displays some lithic changes eastward, as seen in the LeClaire area along the Mississippi River, where thicker intervals of non-laminated dolomite mudstones become more common and a characteristic petroliferous odor is noted in some beds. Molds of gypsum crystals and possible anhydrite nodules (lumpy voids) have been described in the Anamosa Member (Henry, 1972).

Skeletal fossils are rare to absent in the characteristic laminated Anamosa beds, but some bedding surfaces contain scattered brachiopods (rhynchonellids, protathyrids) or bivalves. Units of brachiopod-rich wackestones and packstones (locally with scattered to common corals) interbed with the laminated dolomites, and many of these beds represent distal debris flows from nearby carbonate mound facies. However, brachiopod-rich beds are regionally noted near the base of many Anamosa sections, and some skeletal beds appear to be unrelated to any nearby carbonate mound facies. Wavy and planar laminated beds are remarkably unfossiliferous, but concentrations of the enigmatic "rods" are seen along some laminae (see earlier discussion). Burrowing is absent through most of the Anamosa Member, but minor burrow mottling is noted in some beds. Large leperditiid ostracodes are noted in upper Anamosa strata. Of note is the total absence of many normal-marine fossil groups, including echinoderms, bryozoans, and trilobites.

Brady Member

Philcox (1970b, 1972) introduced the term "Brady facies" for a distinctive rock unit described from the Brady Quarry (McGuire Quarry), Cedar Co., Iowa (Loc. BRQ, Fig. 1). Witzke (1985) accorded this unit member status within the Gower Formation. The Brady Member is recognized at localities in Cedar, Jones, Linn, Johnson, and Jackson counties, eastern Iowa. The member is represented by carbonate mound facies with moderately to steeply dipping beds, intimately interbedded with laminated Anamosa strata around the mound margins. The Brady Member "represents a discontinuous mounded, brachiopod-rich carbonate facies surrounded by laminated dolomites of the Anamosa Member" (Witzke, 1985, p. 39). The Brady Member mounds are everywhere bevelled beneath Devonian or Quaternary units, and the exact vertical dimensions of the mounds are thereby obscured. However, exposures in the type area and elsewhere indicate preserved thicknesses of 100 feet (30 m) or greater for the member. The Brady Member mounds achieve lateral dimensions estimated to range between about 300 and 5000 feet (90-1500 m). The Brady Member abruptly overlies upper Scotch Grove strata, and it most commonly overlies or flanks mounds of the Palisades-Kepler Member. The deepest portion of the type Brady Quarry shows crinoidal dolomites of the Palisades-Kepler Member below the central Brady mound, and Brady strata flank and overstep the Palisades-Kepler mounds at Palisades-Kepler State Park. Locally the Brady mounds overstep the edges of the Palisades-Kepler mounds to overlie upper Scotch Grove intermound strata (Witzke, 1987). The Brady Member is overlain by Middle Devonian strata of the Otis Formation at the type locality, and the contact is an angular unconformity.

Brady lithologies are typically dominated by distinctive brachiopod-rich dolomites displaying wackestone to packstone (minor grainstone) fabrics. Brachiopods are variably preserved as moldic or dolomitized shells. These brachiopod-rich rocks are best represented in flanking positions around the Brady mounds, and display wedge-shaped beds interfingering down-dip with laminated inter-mound strata. Some Brady beds contain scattered to abundant corals, generally clusters of small tabulates or the distinctive rugosan *Pycnostylus*. Fossiliferous beds in the member are interbedded with dense unfossiliferous to sparsely fossiliferous dolomite mudstones within the mounds, and laminated dolomites are interbedded in the mound flanks. The central portions of the Brady mounds are dominated by skeletal mudstone to wackestone, locally with pentamerids and other brachiopods not seen in flanking beds. Domal stromatolites (to 1 m diameter) are locally present in proximal flank and mound crest positions. Relict submarine cements are abundant in many Brady mounds, displayed as isopachous rims around grains or as fibrous lumpy to botryoidal coatings.

Dips within most of the mounds reflect original depositional topography, slightly enhanced by later compaction, and typically range between 0 and 45° aligned radially around the central mounds. Wedge-beds in the mound flanks thin and flatten out down-dip into laminated inter-mound strata (Philcox, 1970b). If bedding surfaces are traced along exposures displaying mound-flank/inter-mound relations, Brady Member strata are noted to be two to more than five times greater in thickness than equivalent beds in the Anamosa Member (Witzke, 1985), demonstrating that the Brady mounds were developed as topographic highs during deposition. Perhaps the most remarkable aspect of the Brady mounds is the local development of steeply dipping to overturned beds in the mound flanks, unrelated to original depositional slopes (Hinman, 1963; Smith, 1967; Philcox, 1972; Witzke, 1981a). Steeply dipping units (60-90°) generally occur in interbedded sequences of laminated and brachiopodal dolomites in the mound flanks, and are interpreted to result from post-depositional compaction of flanking sediments around the rigid cemented central mounds. S-shaped slump folds, locally overturned, further indicate the plastic nature of the flanking sediments during compaction.

LeClaire Member

The "limestone of LeClaire rapids," Scott Co., Iowa, was described by Hall (*in* Hall and Whitney, 1858) as a horizontally bedded to "folded" sequence of hard dolomite, in part with abundant brachiopod and rugose coral molds. Worthen (1862) recognized that the "LeClaire Limestone" is intercalated with laminated rocks (i.e., Anamosa Member) in the type area. Termed the "LeClaire facies" in many subsequent reports, Witzke (1985) formally designated the LeClaire as a member within the Gower Formation and excluded its use as a stratigraphic term for strata not correlative with laminated Anamosa rocks. Because many previously described exposures of LeClaire strata around the town of LeClaire have been destroyed by construction and no specific type locality was ever designated, Witzke (1985) proposed a primary reference section north of LeClaire (Bud Creek). However, upon further study it was recognized that existing exposures within the town of LeClaire (along Cody Road between Davenport and Chestnut streets), although of limited vertical extent, are sufficient to characterize the member, and these are designated herein the type locality to preserve historic intent (Loc. LC, Fig. 1).

The LeClaire Member includes flat-lying and mounded dolomites that interbed laterally with laminated strata of the Anamosa Member, analogous to stratigraphic relations between the Brady and Anamosa members. The LeClaire in some respects resembles the Brady Member, although geographic separation, bedding characters, and lithic and faunal distinctions serve to separate the two members. The LeClaire Member is presently recognized only in southern Clinton and Scott counties, Iowa, and nearby Whiteside, Henry, and Rock Island counties, Illinois. However, the terms "Gower" and "LeClaire" have never been formally used as stratigraphic nomens by the Illinois Geological Survey, who previously labelled equivalent strata the Port Byron Formation (Port Bryron lies across the river from LeClaire) or Racine Formation (Willman, 1973). Because of historic precedence and lithostratigrapic utility, Witzke (1985, p. 41) suggested the terms "Gower," "LeClaire," and "Anamosa" be used in northwestern Illinois. The subsurface extent of the LeClaire Member in Illinois is not known, but the term should apply only to skeletal and mounded strata laterally co-extensive with laminated carbonates. Thicknesses of the LeClaire Member in the type area approximate 150 feet (45 m), and lateral dimensions of contiguous LeClaire strata locally exceed 2 miles (3.2 km). It is probable that individual bodies of LeClaire strata, which form extensive interconnected carbonate mound and bank facies, achieve even greater dimensions than noted. Laminated Anamosa facies interbed in a complex manner along the margins of the LeClaire stratigraphic bodies, in both dipping and horizontal strata (as seen in Scott Co., Iowa, and Whiteside Co., Ill.).

The LeClaire Member includes a variety of distinctive dolomite lithologies, some of which resemble those seen in the Brady and Palisades- Kepler members. Fossil-moldic wackestones are generally dominant, but barren to sparse mudstones and skeletal-replaced packstones are also common. Sparse mudstones locally dominate, but unlike the Anamosa, these are not laminated. Vugular porosity is common in many wackestone units. The dolomites vary from extremely fine to coarse crystalline, and some dense lithologies break with glassy fracture. Wackestones variably display molds of fine crinoid debris, brachiopods, and/or corals. Packstone lithologies, graded in part, include coarse crinoidal or brachiopod-rich rocks. Mounded LeClaire rocks commonly show relict submarine cement fabrics (botryoidal void linings and isopachous rims). Dolomitized internal sediment fills also occur, and nautiloid-rich "pockets" are known.

Strata of the LeClaire Member are displayed in a bewildering array of dipping and horizontal beds. Horizontal to low-angle dips (0-10°) characterize most exposures. More pronounced dips (20-45°) occur marginal to some mounds, and steeply dipping beds are locally observed (to 60°). Overturned beds in slump folds have been noted at one locality where LeClaire strata interbed with laminated Anamosa beds marginal to mound facies, similar to overturning noted in the Brady Member. Cut-and-fill features (to about 2 m deep) are locally observed in the type LeClaire area, suggesting that local sediment removal (erosional channeling or mass wasting) punctuated mound sedimentation. Probable radial dipping patterns in the type LeClaire area suggest a mounded feature about 3000 feet (900 m) in diameter, although horizontal LeClaire strata adjoin the feature to the south and laminated Anamosa strata replace it to the north.

Northward along the Mississippi River in the Cordova-Princeton area, moderate to steep dips delineate the locations of mounded features, but horizontally bedded LeClaire strata adjoin these and form the dominant representation of the member. About 2 miles (3.2 km) west of this area, laminated Anamosa and skeletal LeClaire lithologies are interstratified in horizontal to gently dipping beds (alternations 10-150 cm). Quarrying in western Scott County (New Liberty) has exposed radially-dipping (0-50°) mounded features about 200 feet (60 m) in diameter composed of barren to sparsely skeletal mudstones, which grade laterally into horizontally-bedded strata of similar lithology.

The LeClaire Member contains both crinoid-rich open-marine faunas (similar to those of the Palisades-Kepler Mbr.) as well as coralline or brachiopod-rich restricted-marine faunas (similar to those of the Brady Mbr). Crinoidal lithologies locally contain articulated dorsal cups of inadunate (*Crotalocrinites*) and camerate (especially *Eucalyptocrinites*) crinoids. Diverse nautiloid faunas are recognized, especially in nautiloid-rich pockets, and trilobites locally occur. Corals and stromatoporoids are noteworthy in some sections, especially isolated or fasciculate *Pycnostylus* and tabular to ramose *Favosites*. Brachiopod faunas in the LeClaire show considerable variation; some brachiopod associations resemble those in the Brady Member, primarily low-diversity rhynchonellid- dominated faunas, locally with protathyrids or meristellids. Other brachiopod faunas are associated with crinoidal lithologies, and these typically include *Atrypa*, *Dinobolus*, and *Harpidium*.

GENESIS OF CARBONATE MOUND FACIES

Eastern Iowa contains the westernmost exposures of carbonate buildups within the great Midwestern Silurian "reef" belt, a region noted for prominent development of mound or reef-like facies. This region is best represented by strata around the margins of the Michigan Basin and across the Wabash Platform, but similar facies rim the proto-Illinois Basin as well. The carbonate buildups of eastern Iowa provide additional regional perspectives on the genesis of these remarkable features. Although no specific claims about precise analogies between the Iowa facies and those developed in the Michigan Basin area will be advanced here, personal acquaintance with Silurian facies in Wisconsin and Illinois underscores the general similarity of the Iowa buildups with those seen elsewhere in the Midwest. Because the genesis of carbonate buildups in the Michigan Basin area has been contentious (see summaries by Shaver, 1991; Shaver and Sunderman, 1989), further discussion of similar features in the Iowa area seems warranted.

Terminology and Concepts

Silurian carbonate buildups in Iowa and elsewhere in the Midwest have been categorized under a variety of labels by different workers, primarily "reefs," "mounds," "bioherms," and "clinothems." While it may seem that a discussion of the appropriateness of these various terms amounts to semantic quibbling, our choice of terms has direct bearing on how we perceive these features and how we attempt to classify them. The term "reef" through various usage has achieved a duality of meaning, one descriptive and one genetic (Heckel, 1974). In the context of Dunham's (1970) "stratigraphic reef," simply a laterally restricted mass of carbonate rock, the Silurian buildups of the Midwest could clearly be termed reefs. However, this descriptive usage conveys little practical meaning (Heckel, 1974). On the other hand, a genetic meaning has been implied by most workers using the term which incorporates specific criteria of topographic relief, potential wave resistance, and organic construction or framework, and has been variously labelled as an "ecologic reef," "organic reef," or "framework reef" (Heckel, 1974; Shaver, 1991). This use of the term "reef" has been championed by Shaver (1991) and others for the Silurian carbonate buildups of the Midwest. He (ibid., p. 117) stated that the accumulated evidences from various studies of Silurian buildups ". . . obviate the need for doubts of ecologic-reef status." For other workers, this conclusion has been less than obvious.

Less restrictive or non-genetic terminology has been used by some workers to describe the Silurian carbonate buildups of the Midwest. "Bioherm," a term applied to mound-like structures of organic origin (Heckel, 1974), has also been widely utilized by various Silurian workers, including studies in Iowa (Hinman, 1968; Philcox, 1970a). It is probably a less restrictive term than "organic reef" in that it does not imply wave resistance or the presence of a rigid organic framework. However, certain sedimentary distinctions become significant in a strict definition of the term "bioherm," particularly the ability to distinguish current formed structures from organically formed structures and organically generated carbonate mud from inorganically precipitated mud (Heckel, 1974). Shaver (1991, p. 117) expressed his feelings about the term: "[Various] studies also suggest that the conservative modern-period claims by some persons of nonresistance to waves for given buildups are questionable, and, therefore, that designations by the noncommittal term 'bioherm' have been grossly overworked."

"Clinothems" have been used to describe some Silurian carbonate buildups (e.g., Devaney et al., 1986), and the term basically implies constructive deposition on inclined surfaces below a carbonate-producing shelf or platform. As noted by Shaver (1991), Sam Calvin (1896, p. 426) first proposed a similar idea for some Iowa buildups, suggesting that "strong currents . . . swept" material into "obliquely bedded masses." Sediments derived from higher energy settings would be transported and deposited on laterally prograding inclined slopes (clinoforms). If the Silurian carbonate buildups are categorized as clinothems, an organic framework core and wave resistance are not essential ingredients as required in an ecologic reef classification. In fact, it is wave currents that are responsible for the entrainment of sediments in the clinothem model, and wave resistance may interfere with bulk sediment transport. The flanking inclined beds of these buildups commonly display grading, consistent with their interpretation as clinothems. The interpretation of a massive "reef core" with an organic boundstone framework, which is integral to the ecologic reef classification, has vexed some workers who have failed to recognize such a core in some buildups, or who find core regions that form a volumetrically insignificant part of individual buildups. As noted by Delaney et al. (1986, p. 1381) for many large Silurian buildups, "the bulk of the complex consists largely of inclined flank units that, given this geometry, must have extended upslope to the bench surface rather than to a massive complex core," suggesting that ". . . core areas were insignificant sources of carbonate debris for clinothem progradation." Promoters of an ecologic reef model for Silurian buildups were not persuaded by clinothem concepts, which Shaver and Sunderman (1989, p. 941) categorized as "the New Surrealism."

Some workers, including this author, have preferred to categorize these features by descriptive but non-genetic terminology. The terms "carbonate buildup" and "carbonate mound" carry no genetic baggage that constrains our depositional interpretations. Instead, these terms merely describe the general form of these interesting Silurian features, and no duality of meaning can be construed, unlike the term "reef." It seems highly inappropriate to categorize individual mounded features seen in the Silurian of the Midwest using a genetic terminology until various criteria have been clearly established. This becomes a cart-before-the-horse issue. For example, we should not classify all Silurian buildups in the Midwest as ecologic or organic reefs until they have all been examined for definitive criteria. Our classification and interpretations must proceed on a case-by-case basis, carefully evaluating various questions. 1) What is the central or interior portion of these buildups? 2) Is there an organic framework? 3) What depositional and diagenetic fabrics can be recognized? 4) What is the relationship between flanking strata and the central mounds? 5) What bedding and facies geometries can be recognized within individual buildups? 6) What is the distribution of various biotic groups in the mounds, and what is their importance as sediment contributors or sediment binders? 7) What role have inorganic processes such as cementation and compaction played in the depositional and post-depositional history of these features? Many additional questions may be equally pertinent.

The identification of a central "reef core" or organic framework is essential if an ecologic reef classification is employed. Lowenstam's (1950) influential study of Silurian "reefs" in the Great Lakes area delineated central framework cores for these structures, displaying prominent "*Stromatactis*-like framebuilders" of undetermined biologic affinities. However, modern carbonate petrographic studies have recognized that most stromatactis structures are not of biologic origin at all, but represent void-filling internal sediments and calcite cements developed within lime-*mud* buildups (Wilson, 1975, p. 79). In fact, many of Lowenstam's (1950) illustrations of supposed framebuilders and encrusters clearly resemble void fillings and botryoidal marine cements, not evidence of an organic framework. Although frameworks have been described from various Silurian buildups by later workers, and some may be correctly identified as ecologic reefs, it does not follow that all Silurian buildups necessarily must observe the same pattern. In fact, numerous workers have failed to identify a framework within individual Silurian buildups in the Midwest. In addition, the mere presence, if confirmed, of organic boundstone within a mounded feature does not, by itself, make the entire feature a reef.

A boundstone framework must be an integral and volumetrically significant part of a buildup if it can be logically considered a reef. In situations where flanking strata volumetrically overwhelm a framestone core (by up to a few orders of magnitude or greater), it becomes increasingly difficult to consider the integral role of such a framework in the construction of the entire feature. Likewise, occurrences of corals or stromatoporoids within portions of mounded features do not necessarily constitute evidence of an organic framework; these coralline elements must show clear evidence that they form a rigid framework or display boundstone fabrics. These points are not trivial, as it is apparent that some workers have inferred the presence of an organic framework even in the absence of clear rock fabric indicators. For example, Shaver and Sunderman (1989, p. 944) stated: ". . . Silurian buildups in the Great Lakes area have been questioned as being true ecologic reefs because some of the so-called structural cores are dominated by bluish-gray carbonate mud and commonly lack ready evidence of framework organisms. The ravages of dolomitization and other diagenetic processes, however, have probably obscured a prominent role of algae and other frame-builders or binders in the construction of the reef cores." This proposition is, of course, completely ad hoc. If a framework cannot be recognized, we should not infer its presence, nor should our classifications and terminology necessitate its inclusion.

Carbonate mounds are exposed at numerous localities in eastern Iowa, at three general stratigraphic positions. A review of sedimentary and paleontologic observations for mounds within each stratigraphic interval is presented in order to evaluate various criteria for their classification and genesis. The geometries of various carbonate mounds are largely derived from field observations, supplemented with subsurface information and integrated with regional stratigraphic syntheses. The central or "core" regions of individual mounds are defined by bedding relations in the field, identified centrally to the radial dips of the surrounding flanking units. In some areas, bedding relations are complex and do not display obvious radial patterns, where the central regions of the mounds are not as clearly delineated. The absence of obvious framebuilders in most Silurian mounds in Iowa cannot be relegated to the "ravages of dolomitization" nor can it be slated to inadequate exposures. Many mounds and mound complexes in Iowa are cross-cut by quarry faces or natural exposures, clearly exposing their central areas. Sedimentary and skeletal fabrics are well preserved within the Iowa dolomites, and no clear case can be made for the diagenetic destruction of original framebuilders. Sedimentary fabrics are preserved with remarkable clarity in many mounds, including skeletal grains and inorganic cements. No special pleadings are needed to invoke "phantom" framebuilders when none are seen.

Mounds of the Johns Creek Quarry Member

Mounds of the Johns Creek Quarry Member are known from only four localities in Iowa (Philcox, 1970a; Johnson, 1977b; Witzke, 1981a), although the general features of this mound facies are well displayed at these exposures. The mounds reach thicknesses to about 50 feet (15 m) and individual features achieve lateral dimensions of up to about 500 feet (150 m). The central areas of the mounds consistently display skeletal mudstone and wackestone fabrics, indicating a general dominance of carbonate mud for the initial stages of mound growth. Philcox (1970a, p. 970) recognized that "the core consists predominantly of unbedded fine dolomite, probably originally biomicrite," and Johnson (1977a, p. 38) identified a "core of unbedded micritic dolomite." The micritic character is dominant at all exposures of the central mounds, but my observations indicate that faint to locally prominent bedding is usually observed within the central mounds.

Large well-preserved colonial corals are seen locally within the central mounds, and these are conspicuous in some exposures. The corals variably occur in life position or as overturned masses. Coral sizes most commonly range between 10 and 40 cm, although large forms locally reach sizes of 1 to 2 m. For some observers, these corals would constitute evidence of an organic framework. However, an interconnected framework or organic boundstone fabric does not characterize these occurrences. Rather the corals are enveloped within carbonate mud fabrics, and the discontinuous

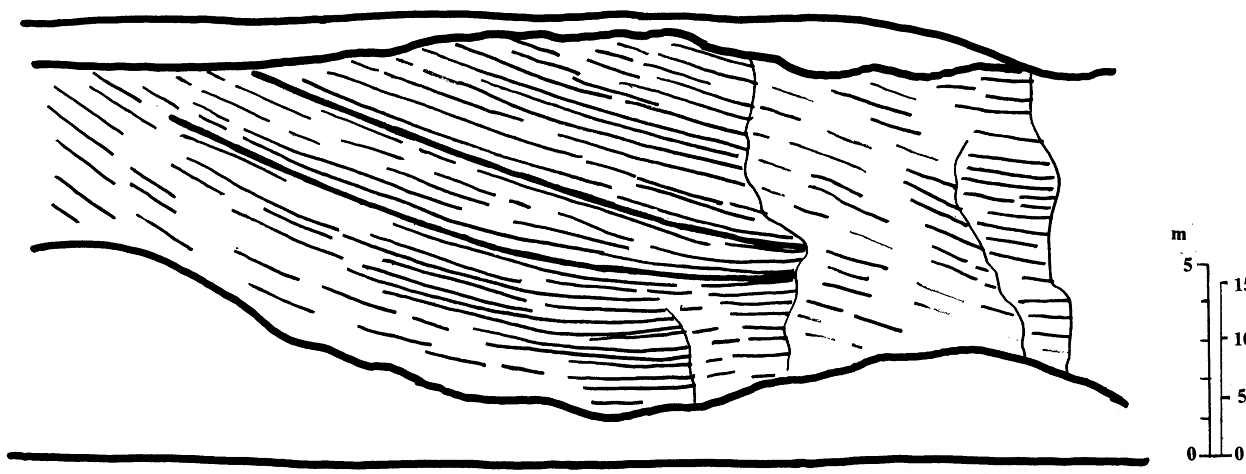

Figure 8. Dipping flank beds, carbonate mound facies of the Johns Creek Quarry Member. This is a photo tracing of a segment of the quarry wall at the Monticello-Manternach Quarry (Stop 2). Lithologies are dominantly crinoidal packstones. Note wedge-shaped geometries in middle part.

coral colonies do not display a central framework. Likewise, common overturned colonies indicate that they were not bound into a rigid framework. Coral distribution is spotty within the central mounds, and their total volume is insignificant when compared to the matrix mud which envelopes them. The open corallite framework of individual colonies is filled with carbonate mud sediment, and it is likely that the corals acted as local sediment bafflers promoting mound growth. Nevertheless, coral growth forms noted by Philcox (1970a) show that coral growth was strongly influenced by rates of mud deposition, not that mud deposition was controlled by the corals. In addition, muddy sediments (skeletal mudstones to wackestones) characterize the central mounds at all localities even where corals are rare to absent. Therefore, corals are locally significant but not apparently essential for mound growth.

Mound growth in the Johns Creek Quarry Member was initiated primarily as accumulations of carbonate mud, probably of biogenic origin. It is proposed that the mounds were a locus for mud-precipitating organisms, probably algal. However, the muds show no evidence of organic binding or encrusting, and the mounds may simply represent in-place biogenic mud accumulations that provided suitable habitat for certain coral taxa and other invertebrates. The muddy sediments contain scattered to common skeletal material, primarily crinoid debris and corals. Additional skeletal grains in the central mounds include scattered brachiopods, stromatoporoids (tabular to hemispherical forms 10-30 cm), and gastropods. No current-formed sedimentary structures have been identified in the central mounds, and it seems unlikely that mud deposition was related to hydrodynamic processes on the seafloor. Instead, the mud-rich lithologies in the central mounds indicate that water turbulence was insufficient to winnow the muddy sediments (Philcox, 1970a). By contrast, the flanking strata in these mounds are skeletal rich and graded, dominated by wackestones and packstones, locally including grainstone units.

Flanking strata commonly show coarse crinoidal lithologies arranged in graded increments about 10 to 30 cm thick, and wedge-shaped beds are identified (Fig. 8). Philcox (1970a, p. 970) noted a faunal transition but a "rather abrupt change in sediment grade" between the central mounds and flanking strata; the boundary is gradational in some exposures with individual packstone beds interbedded with muddy coral-bearing strata. Graded packstone beds in the mound flanks are dominantly crinoidal, with articulated cups of *Manticrinus* and *Eucalyptocrinites* locally noteworthy; additional fauna includes bryozoans, molluscs, solitary corals, small tabulates (favositids, *Halysites*), brachiopods, and trilobites. The common occurrence of packstones, and locally grainstones, indicates that water currents operated to winnow sediments during deposition.

However, the graded character of many beds further indicates that flanking sedimentation was episodic, marked by down-slope movement of carbonate sediments in debris flows. The mound crests and flanks were not subjected to continuous water agitation, and a position generally below fairweather wavebase is inferred. Episodic grain winnowing and debris flows are reasonably inferred to have been triggered during storm events, and the crests of the mounds are interpreted to have occupied a position above storm wavebase. Graded beds and packstone lithologies become less common down-dip in the flanking strata, to merge with horizontally-bedded skeletal wackestone lithologies.

Dipping and graded flank strata indicate that the mounds were topographic features on the seafloor, although later compaction may have accentuated the apparent dips. Indications of syndepositional lithification of the mound sediments are suggested by occurrences of relict submarine cement fabrics and internal sediment fills. These features are less conspicuous than in upper Scotch Grove and Gower carbonate mound facies in Iowa, but lumpy and botryoidal relict cements preserved as grain coatings and void linings are observed in mounds of the Johns Creek Quarry Member (STOP 2), both in the central mounds and in flanking strata. Because no organic sediment binding or framework has been identified in these mounds, it is reasonable to suggest that the mounds became rigid features primarily through submarine cementation processes.

The geometries of the Johns Creek Quarry mounds include both isolated mounds and coalesced mound complexes. A single mound is well exposed at the type locality of the member, revealing a central mud-rich region of horizontal to gently dipping strata surrounded by more steeply radially dipping and skeletal-rich flanking strata (Johnson, 1977b). Relationships with horizontal inter-mound strata of the member are portrayed at the type locality, and flanking beds grade upward and laterally into strata of the Welton Member. By contrast, central mound and flank strata near Monticello (STOP 2) and Elwood (see Philcox, 1970a) display more complex relations, with multiple mud-rich loci each flanked by skeletal-rich strata in complexly dipping geometries (see illustration for STOP 2). Flanking beds locally show wedge-shaped geometries, variably thickening up-dip or down-dip in different positions. As originally noted by Philcox (1970a, p. 970), these localities each expose a group or cluster of small mounds. The variable dips of the skeletal-rich flanks suggest that the mound facies locally coalesced into an intergrown complex. Horizontally-bedded units between non-coalesced mounds grade upward into lithologies of the Welton Member containing *Costistricklandia* (STOP 2).

Mounds of the Palisades-Kepler Member

Mounds of the Palisades-Kepler Member are well exposed at many localities in eastern Iowa (Witzke, 1981a), and a discussion of the general lithic, paleontologic, and stratigraphic characteristics of the member is given earlier in this guidebook.

Therefore, this section will focus on interpretations of mound genesis. Contrasting views of the nature of the central mounds ("bioherm core") were put forward by different workers, and discrepencies in interpretation need to be addressed. Two studies in the 1960s identified a rigid organic framework for the bioherm cores of strata now assigned to the Palisades-Kepler Member. Smith (1967, p. 108) recognized "framework builders" (algae, tabulate corals, stromatoporoids), and suggested "ample evidence of frame-building, sediment-binding, and detritus-catching elements" to erect "rigid, wave-resistant structures." Hinman (1968, p. 40) also wrote: "The core of the bioherm is an essentially unstratified mass built by the accumulation of sessile rock-forming coelenterates. The stromatoporoids and the rugose and tabulate corals, together with the pelmatozoans and perhaps algae, produced a rigid framework rising above the sea floor." These ideas were in direct conflict with interpretations put forward in subsequent studies.

Philcox (1971, p. 345-6) saw mound genesis in a very different way: "Early growth . . . was apparently almost entirely the result of localized crinoid growth and accumulation of skeletal debris; few colonial coelenterates were present. . . There is little evidence from the distribution, orientation and mutual relationships of the coelenterate colonies . . . that these organisms were important in the sediment-binding or other protective, reef-building role. . . they were more important as sediment contributors than in constructing any form of 'rigid, wave-resistant frame.'" Similar observations were made by Witzke (1981a, p. 455): "There is no evidence of any rigid skeletal framework within the upper Scotch Grove mounds, and the mounds became rigid primarily through submarine cementation processes." Considering that these various workers looked at many of the same exposures, how can such discrepencies in observation and interpretation be reconciled? I would suggest that earlier models of Silurian reefs (e.g., Lowenstam, 1950) influenced certain workers to such an extent that model-driven expectations clouded empirical observations. What observations, then, are pertinent to resolving the apparent

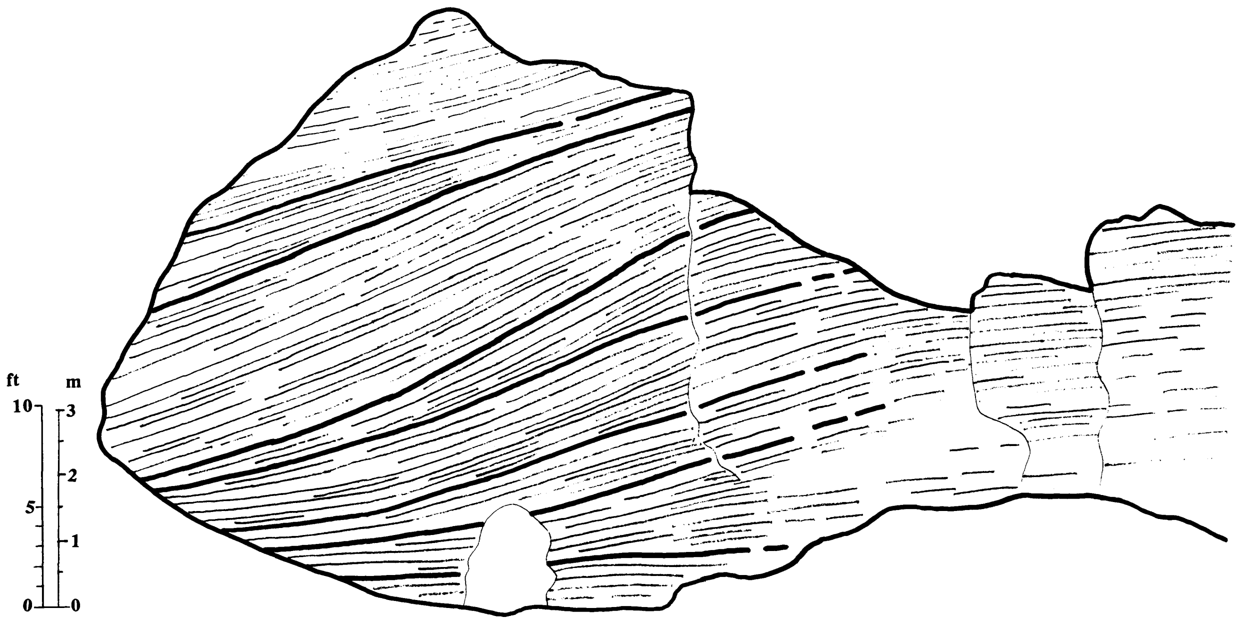

Figure 9. Dipping flank beds with some wedge-shaped geometries, Palisades-Kepler Member. This is a photo tracing along the north edge of Mound B, Palisades-Kepler State Park (Stop 5). Lithologies are crinoidal packstones and wackestones.

discrepencies in understanding the genesis and structure of the central mounds?

The central mounds of the Palisades-Kepler Member are characterized by skeletal-moldic mudstones and wackestones, with minor packstone, at all exposures. "The matrix is usually a carbonate mud" (Philcox, 1970b, p. 174). Skeletal grains are dominated by crinoidal debris. Tabulate corals and stromatoporoids are notably rare, although coralline material becomes locally prominent at a few localities at certain positions within the mounds. Scattered flat stromatoporoids and halysitids occur in the lowest portions of some mounds, but these do not typically form organic frameworks (ibid.). Philcox (1971, p. 339) described beds with domed favositids (9-12 inches) and large stromatoporoids marginally capping some central mounds, but he noted that beds more centrally located in the mounds contain fewer corals "and are dominantly crinoidal." Philcox's observations largely parallel my own. To be sure, corals are notable components in areas within the central mounds, but they are rare to absent through much of the central mound complexes. Hinman (1968, p. 25) illustrated a "typical core facies" with molds of corals, stromatoporoids, brachiopods, and crinoid debris, but the rock shown is a skeletal wackestone dominated by carbonate mud (verified by my own field observations). Brachiopod wackestones are locally noteworthy in the central mounds, and clusters of large pentamerids (*Rhipidium*, *Harpidium*), some in life position, have been identified at several localities. There are a few localities where coralline elements are clustered in moderate abundance, both in life position and in random orientation within a mud-rich matrix, and it is possible that corals may have locally served as sediment binders. Nevertheless, corals and stromatoporoids form a volumetrically trivial component of the central mounds, and their local abundance is more an exception than the rule. An inordinate amount of attention has been paid to the importance of coralline elements in some previous studies to the extent that the dominance of carbonate mud and non-coralline skeletal debris in the central mounds remained largely undiagnosed.

By contrast, the great abundance of crinoid debris in the flanking beds has been noted by all who have studied mounded facies of the Palisades-Kepler Member. Flanking strata typically include crinoidal packstones (minor grainstone), and crinoidal wackestones and mudstones also occur. Graded beds are common, displaying wedge-shaped geometries in places (Fig. 9). Grading is variably expressed as alternations of coarse to fine skeletal packstone or as packstone-wackestone couplets. Crinoid stem segments are oriented in some beds. Small tabulate corals (<8 cm), fenestrate bryozoans, and molluscs are locally observed in the graded units. Flanking strata commonly grade down-dip from packstones to wackestones, and the flanking margins of some mounded complexes are locally dominated by skeletal wackestones (generally crinoidal, but locally brachiopod-rich).

Two general lines of evidence for early lithification of the mounds are noted. First, relict submarine cements are seen in both central mound and flanking facies. Isopachous rims are observed in grain-rich units, which display relict fibrous or bladed microstructure. In addition, lumpy and botryoidal coatings with similar microstructure are observed to line void spaces and coat larger skeletal grains in places within the mounds. Second, evidence for early lithification of mud-rich units is more indirect, but is clearly displayed within the central mounds at several localities where internal sediment fills and fractured to brecciated lithologies are observed. Large internal sediment fills within the mounds were first observed by Philcox (1970b, p. 179), who termed them "pene-contemporaneous cave deposits"; these "show cross-cutting relationships with the host rock, and some can be traced to the tops of the mounds." The largest of these features at Palisades-Kepler extends for about 9 m and displays an infilling sequence of three general lithologies: 1) a lower skeletal-rich unit with nautiloids, 2) a middle irregularly-stratified mud-rich unit, and 3) an upper unit of unstratified mud with coated grains or pisoliths (ibid.). Skeletal-rich fills, commonly one meter or more in size, have been observed at this and other localities (Witzke, 1981a) which variably contain abundant current-segregated straight-shelled nautiloids or convex-side-down trilobite molts (primarily *Bumastus ioxus*). Similar sediment fills are known from other mounded facies in the Great Lakes area (Mikulic, 1979).

Occurrences of internal sediment fills strongly suggest that the mound sediments were lithified in order to have remained open during infilling, and the presence of marine fossils within them indicates that infilling was, at least in part, penecontemporaneous with mound deposition. Philcox (1970b, p. 179) suggested that these features "may have been initiated as fissures by gravitational collapse during mound building." Further evidence of early lithification of mud-rich sediments in the central mounds is seen in places where fractured masses are enclosed within the mound sediments, and brecciated mudstones locally fill fissures or occupy jumbled masses within the mounds (Smith, 1967). Fracturing and fissuring apparently accompanied mound growth, probably induced by gravitational stresses within the rigid and erect mounds.

Carbonate buildups of the Palisades-Kepler Member display complex geometries that are categorized for simplicity into two basic forms: 1) isolated single mounds, and 2) coalesced mound complexes or mound clusters. The latter category is the dominant geometry across eastern Iowa, although a few examples of isolated single mounds are known. Isolated single mounds display simple hay-stack geometries, with a bedded central area surrounded by radially dipping flanking units (e.g., Fig. 10, Loc. CVR; Fig. 11, Loc. AP). The central areas of some large single mounds are essentially horizontally bedded, but dips increase away from the central region in all directions. The central areas are dominated by crinoidal wackestones and packstones, with scattered corals and stromatoporoids. Flanking strata dip relatively symmetrically away from the central region, and merge down-dip with horizontal inter-mound or skeletal bank facies. Grain-rich crinoidal flanking strata commonly show graded bedding in part, but the exterior flanking portions of some large Palisades-Kepler mounds (e.g., Fig. 10, Loc. CVR) are dominated by mud-rich lithologies and probably represent more distal facies.

Coalesced mound complexes or mound clusters are typical of most exposures of the Palisades-Kepler Member, including the type area. These complexes display a bewildering array of dipping and horizontal strata, with considerable variations in the vertical dimensions of individual mounded features and dips of the contained beds. Strata are horizontal or display low-angle dips through the bulk of these complexes, but flanking units locally show dips of 25 to 40° (e.g., Fig. 9). The assymmetry of these complexes is particularly well displayed in the type Palisades-Kepler area (Fig. 10); the steepest dips are seen in the northern region of the complex but dips flatten out to the southeast and merge with horizontal inter-mound strata in that direction (also see discussion by Philcox, 1970b). Based on relative dips, the best exposed portion of the complex shows six or more mounded features reaching vertical dimensions to over 100 feet (30 m) (labeled A through F on Fig. 10), and "numerous small mounds" adjoin this complex to the southeast of mound F (ibid.). Another mounded complex is seen about 2 km (1.3 mi) to the southeast of the complex at Palisades-Kepler, separated by intervening inter-mound strata of the Waubeek and Buck Creek Quarry members (Witzke, 1981a).

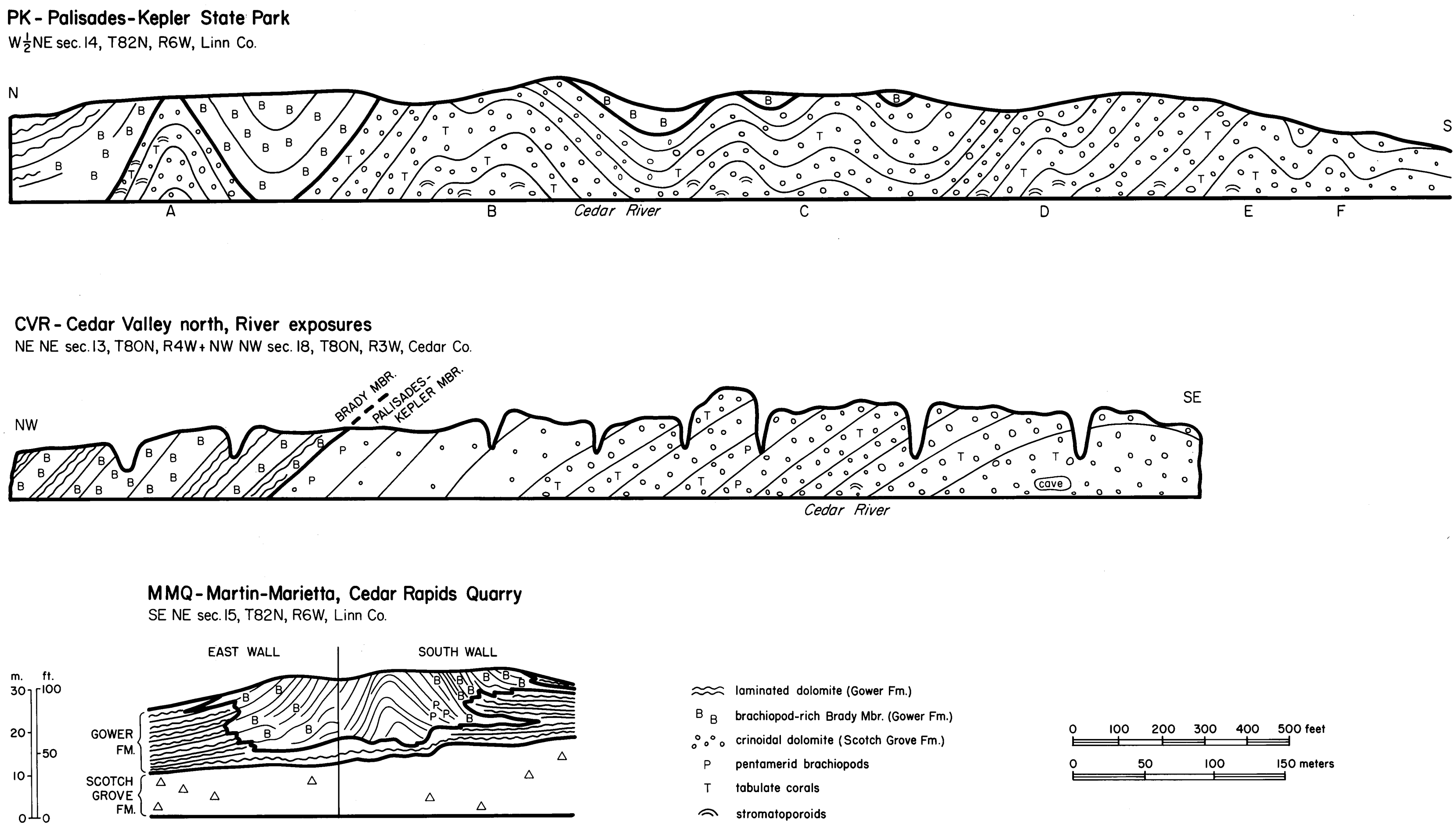

Figure 10. Exposures of carbonate mound facies, Palisades-Kepler and Brady members. Localities PK (Stop 5) and CVR are sketched from general field observations; Locality MMQ derived from photo tracing. All figures are vertically exaggerated approximately 3:1.

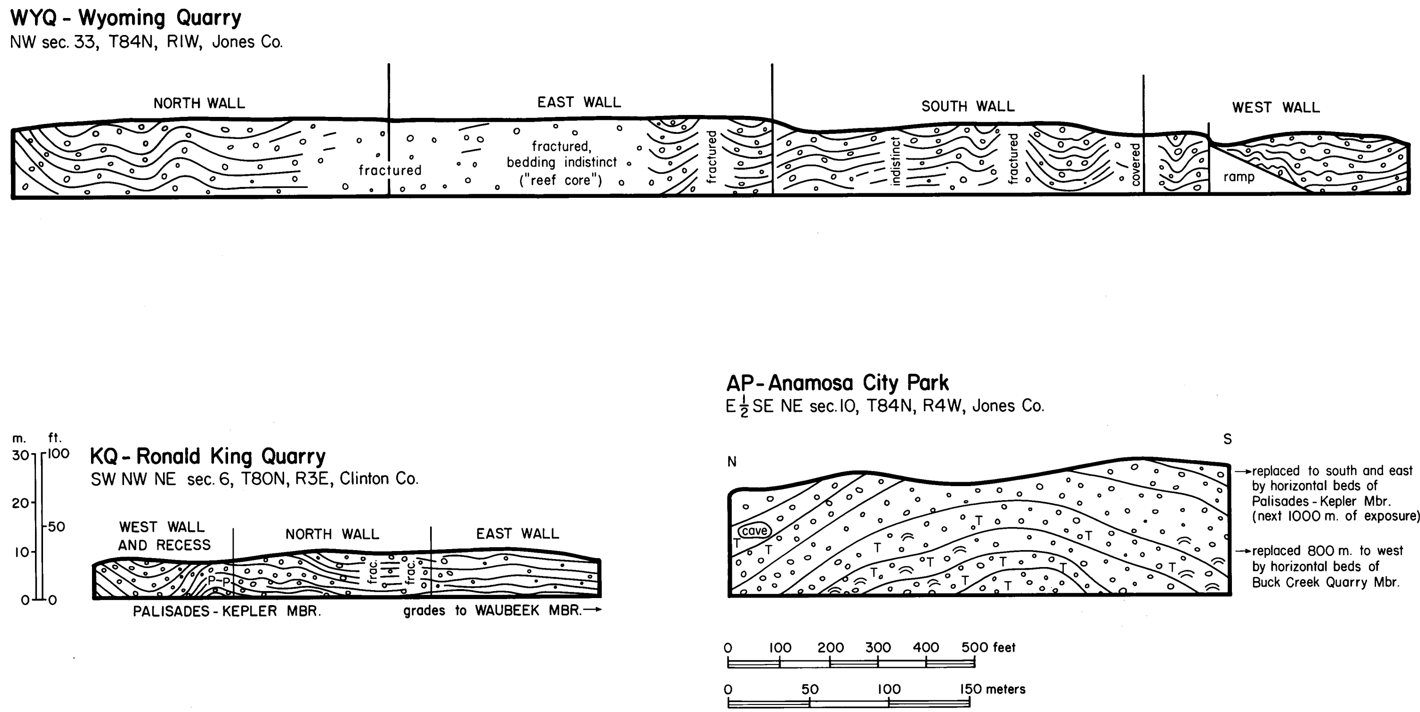

Figure 11. Exposures of carbonate mound facies, Palisades-Kepler Member. Figures are derived from photo tracings; all are vertically exaggerated 3:1. Same scale and symbols as Figure 10.

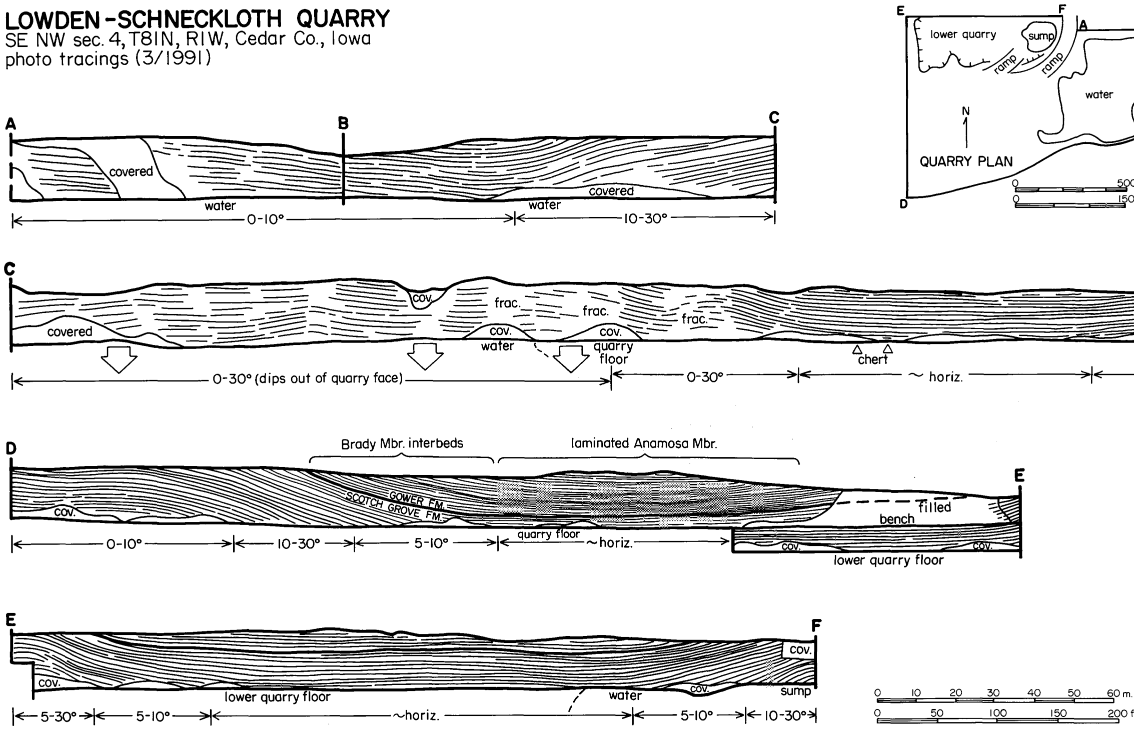

Figure 12. Carbonate mound facies of the Scotch Grove and Gower formations exposed at the Lowden-Schneckloth Quarry, Cedar County (Loc. SNQ, Fig. 1). Figure displays bedding as seen along quarry walls, traced from photo mosaic. General dips are noted (arrows show where beds dip out of quarry face). No vertical exaggeration.

Additional mounded complexes or mound clusters display similar features, but show variations in scale and lithology. A grouping of four mounds at the Lowden-Schneckloth Quarry (Loc. SNQ, Fig. 1) shows features about 50 feet (15 m) thick (Fig. 12). Bedding is visible through the mounds, except where disrupted by fracturing and karst. These mounds display hay-stack to platform-like geometries, with the interior of some showing horizontal or low-angle dips flanked marginally by steeper dipping beds (south half of segment DE, Fig. 12). The mounds and their flanks are dominated by vuggy skeletal mudstones and wackestones, and packstones, although present, are notably less well developed than at other localities. Coalesced complexes or clusters of smaller and more closely spaced mounded features are seen at other localities, where mounds display complex dips with less than about 50 feet (15 m) of vertical closure on individual features (e.g., Beverly Quarry, Loc. BVQ, Fig. 13; Locs. WYQ and KQ, Fig. 11). Based on dipping bed geometries, the Wyoming Quarry (Fig. 11) shows a coalesced clustering of at least seven mounded features of relatively low relief. These types of mounded geometries form the dominant aspect of the Palisades-Kepler Member in eastern Iowa.

As noted, some mounds show platform-like geometries with horizontally bedded strata in the central regions, flanked by dipping beds. At some localities, horizontally bedded, vuggy skeletal wackestone strata of the Palisades-Kepler Member adjoin mounded features (Loc. AP, Fig. 11), but these horizontal units are not flanked by dipping beds and grade into typical inter-mound strata (also seen southeast of type Palisades-Kepler complex). This reflects the asymmetry of many mounded complexes, and further indicates that carbonate mounds can pass gradationally into strata properly considered to be carbonate banks. A carbonate bank should display little or no topographic relief, show no organic binding, and have horizontal to low-angle slopes; banks can show a gradational spectrum with various types of carbonate buildups (Heckel, 1974). The Palisades-Kepler Member in eastern Iowa shows a complex array of carbonate mound, platform, and bank geometries, and it seems clear that constrain-

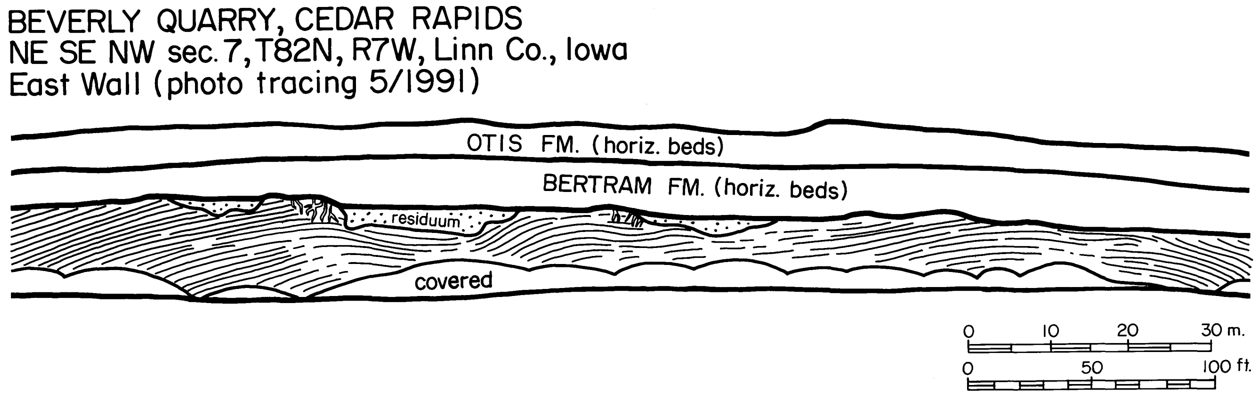

Figure 13. Small-scale carbonate mounding within the Palisades-Kepler Member, Beverly Quarry (Loc. BVQ, Fig. 1). Bedding is traced from photo mosaic. Lithologies include skeletal mudstones to packstones. Otis and Bertram formations are Devonian units. No vertical exaggeration.

ing the genesis of these complex carbonate facies within a simplistic ecologic reef model is inappropriate.

Potential wave resistance is conditional to categorization as ecologic reefs, and the Iowa mounds have been interpreted as "wave-resistant structures" by some workers (Smith, 1967, p. 108). The question as to whether or not the crests of the Iowa mounds grew above wavebase can be approached by evaluating the sedimentary evidence. Any portion of the mound that occupied a position above normal fairweather wavebase would presumably be subjected to continuous water agitation (shoaling), and muds and other fine sediments would thereby be winnowed leaving boundstone or grainstone lithologies. However, no such sedimentary associations have been identified in the Iowa mounds. Instead, the central mounds are mud-rich, suggesting a position below normal wavebase where mud accumulation is more likely. Nevertheless, evidence for current activity is seen in both the central and flanking facies of the Iowa mounds, but the sedimentary features indicate episodic, not continuous, sediment transport.

Episodic current activity is indicated by several lines of evidence: 1) graded flank beds (debris flows), 2) preservation of crinoid cups and other readily disarticulated fossils, 3) burial of overturned corals and stromatoporoids, and 4) coral morphologies. Graded flank beds clearly indicate episodic deposition of debris, possibly triggered by storm events in up-slope or mound crest positions. Likewise, the preservation of crinoid cups and other articulated fossils is suggestive of rapid burial events, as these would normally disarticulate even in the absence of current activity due to decay of connective tissues. Occurrences of overturned corals further suggest that strong currents capable of dislodging and transporting relatively large sediment masses were operant during mound deposition. In addition, Philcox (1971, p. 345) described coral growth forms in relation to their enclosing sediments and concluded that there were "marked fluctuations in sedimentation rates." He (ibid.) further indicated that "sedimentation was an irregular process," with episodes of "temporarily increased turbulence" in environments in which "turbulence was normally limited." Two general conclusions seem warranted: 1) mound sedimentation was primarily, if not entirely, below normal fairweather wavebase, and 2) mound crests and proximal flanks were subjected to periodic current activity, suggesting a position above storm wavebase. It is further suggested that the vertical limits to mound growth may have been marked by the position of fairweather wavebase, as the abundance of carbonate mud and the general absence of an organic framework in the central mounds generally precludes sediment preservation in shoal-water settings.

Mound growth in the Palisades-Kepler Member seems to be related to the in-place accumulation of skeletal debris and carbonate mud (of uncertain origin). It is possible that the mounds represent some sort of self-perpetuating process of biologic origin whereby the initial accumulations of carbonate material created topographic features which were favored sites of attachment for successive generations of organisms, especially crinoids and mud-producers (algae). Philcox (1971, p. 345) suggested that the early growth of these mounds "was apparently almost entirely the result of localized crinoid growth and accumulation of skeletal debris." The mounds became rigid features primarily through inorganic submarine cementation, not through any process of organic binding or frame-building. The mounds grew laterally, coalescing with adjacent carbonate buildups to form complex geometries. Storm currents affected the mounds as they grew vertically, episodically sweeping sediment from their tops and depositing grain-rich beds in the flanking positions.

Mounds of the Brady Member

Two members within the Gower Formation contain carbonate mound facies, each displaying differences in overall geometry and lithology. Mounds of the Brady Member are commonly developed in association with, either marginally or above, older mounds of the Palisades-Kepler Member, suggesting that the older mounds served as loci for subsequent Brady mound growth. These relationships are clearly displayed in areas where wedge-shaped Brady beds interstratify with Anamosa strata down-dip from the now-eroded crests of the older Palisades-Kepler mounds (e.g., Lowden Quarry, Loc. SNQ, Fig. 12; Locs. PK, CVR, Fig. 10). As noted previously, mounds at the type Brady locality overlie probable mounded facies of the Palisades-Kepler Member. However, some Brady mounds show no clear relationship to underlying Palisades-Kepler strata, and overlie, instead, inter-mound strata of the upper Scotch Grove Formation (e.g., Loc. MMQ, Fig. 9). Therefore, Brady Member mounds can develop either above pre-existing mounds, or they can originate as independent features not directly related to the older buildups.

Central regions of Brady mounds were well displayed during investigations at the Cedar Rapids Martin Marietta Quarry (Loc. MMQ, Fig. 10) and the type Brady locality. There is no ambiguity in identifying the location of the interior ("core") of these mounds, as the central areas are clearly outlined by radial dipping patterns (see Fig. 10). The central mounds at both localities are characterized by dense to slightly vuggy dolomitized mudstones and skeletal wackestones, with no indications of any organic framework. Skeletal material is present but scarce, including brachiopods (rhynchonellids, strophomenids, atrypids), scattered small corals (favositids, halysitids, *Pycnostylus*), and molluscs. Large pentamerids (*Harpidium*) and trimerellid brachiopods occur in the central mound at Loc. MMQ (Fig. 10, note symbols) near the transition into flanking rhynchonellid-rich strata. Skeletal molds are well preserved, and it is unlikely that diagenetic alteration has removed some pre-existing framework. Instead, the central mounds are more reasonably interpreted to have been deposited primarily as carbonate mud, with some skeletal benthos inhabiting the surface of the developing mounds. Mud in the early stages of mound growth apparently accumulated without the benefit of sediment-binding or -baffling organisms, suggesting relatively quiet-water environments. Domal stromatolites locally occur in marginal areas of the central mound at the Brady Quarry (Philcox, 1972) and in the upper portions of a Brady mound exposed northwest of the type Gower section. Such stromatolitic occurrences indicate that algal sediment binding occurred in the upper portions of some Brady mounds.

The presence of slump folding and over-steepened beds in the flanking strata strongly suggests that the underlying central mounds were rigid features not subject to the same compaction and soft-sediment deformation seen in the enclosing sediments. That is, the central mounds were lithified prior to the deposition of the flanking strata. Lithification of the carbonate muds was probably accomplished by microspar or micritic cementation, which would be largely indistinguishable from the surrounding dolomitized muds (in fact, such cements are commonly indistinguishable in better preserved limestones facies). Megascopic botryoidal relict isopachous cements are noted to line some void spaces in the central mounds, clearly indicating that submarine cementation occurred within the mounds. However, relict cements become more noteworthy in the grain-rich flanking facies, primarily because these facies were more porous and capable of preserving larger volumes of megascopic cements.

Flanking strata of the Brady Member are among the most distinctive facies noted in the Iowa Silurian, typically characterized by interbedded brachiopod-rich packstones, sparse mudstones, and laminated strata (e.g., Fig. 14). Allochthonous accumulations and interwoven "thickets" of rugose corals (*Pycnostylus*) also occur in the flanking strata, and small favositids and halysitids are sometimes present (Philcox, 1970b, 1972). "Rod" wackestones and rare packstones also occur in some laminated flanking units. Skeletal-rich beds show demonstrable wedge shapes at many localities, typically thinning distally into gently dipping to horizontal laminated strata of the Anamosa Member. Flanking beds of the Brady Member generally show dips of 5 to 40°, but, as discussed previously, over-steepened dips of 60 to >90° and Z- and S-shaped slump folds are not uncommon (Hinman, 1968; Philcox, 1972). The laminated flanking sediments show the

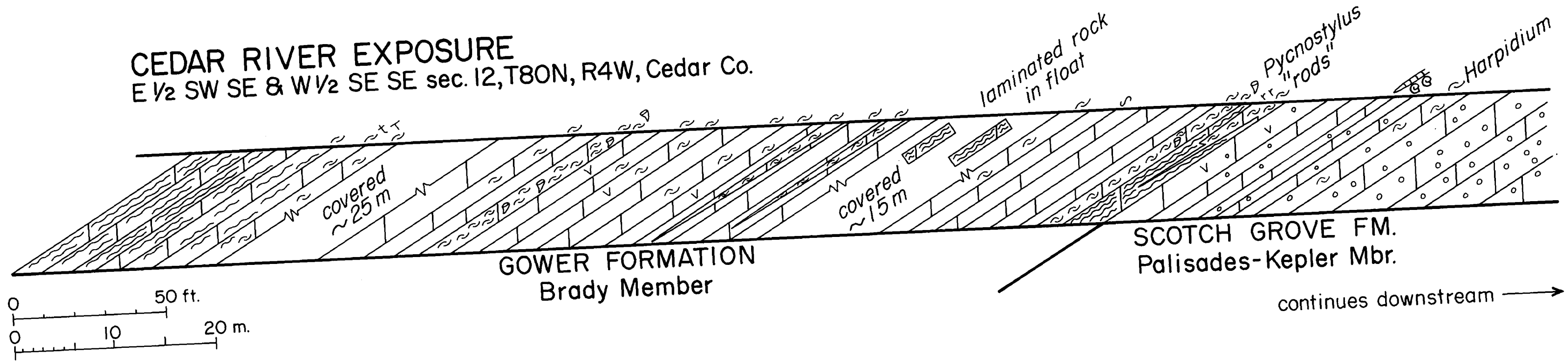

Figure 14. Generalized Gower and uppermost Scotch Grove stratigraphic section exposed along northwest margin of large carbonate mound, Locality CVR (see Figure 10). Datum is Cedar River. Note brachiopod-rich wedge-beds within Brady Member. See page 86 for lithologic symbols.

highest degree of folding and over-steepening, whereas some of the interbedded skeletal-rich units were apparently lithified before the onset of compactional deformation of the laminated sediments (see Philcox, 1972). The porous skeletal-rich units commonly show abundant evidence of megascopic relict submarine cements.

The Brady mounds display relatively simple mound-shaped geometries at most localities, clearly outlined by radial dipping patterns. Compound mounds and coalesced mound complexes seem less common in the Brady Member than in the Scotch Grove Formation, although some compound geometries are indicated where the Brady overlies coalesced mounded units of the Palisades-Kepler Member (e.g., Loc. PK, Fig. 10). In addition, individual Brady mounds (separated by intervening laminated Anamosa strata) are known to occur in regional clusters. For example, at least three Brady mounds are located within a one-half mile (0.8 km) stretch northwest of the type Gower quarry, and at least two Brady mounds are known within a 0.8 mile (1.3 km) stretch in the vicinity of Palisades-Kepler State Park.

Deposition of the Brady mounds was aptly summarized by Philcox (1972, p. 697): "The grain size of the original principal sediment (lime mud), the even laminated bedding, and the presence of stromatolites indicate a quiet shallow-water environment. . . Wedge-bedding indicates that the [mounds] . . . continued to expand laterally by accretion of flank beds consisting mainly of laminated muds. The upper surface of each mound expanded at the same time as a flat platform analogous to a delta top-set, on which little sediment accumulated. The mud, possibly of algal origin . . ., was generated mainly on the platform, from which it was swept off to form flanking wedge-beds, as were vast numbers of brachiopods at intervals. Some of the sediment accumulating on the mound flanks became unstable and slumped." Like the Palisades-Kepler mounds, the flanking strata of the Brady mounds probably accumulated episodically in response to storm events that swept the top surface of the mounds. The high dominance, low diversity communities that inhabited the mound crests indicate probable benthic stresses related to salinity fluctuations, and the general absence of many typical stenohaline marine organisms in the Brady mounds supports general hypersaline conditions (Witzke, 1983). The absence of skeletal benthos in the laminated Anamosa strata further indicates that even greater stresses were present in the inter-mound environments.

Mounds of the LeClaire Member

The LeClaire Member displays considerable variations in lithology, fauna, and facies geometry, as shown by exposures within a 30 mile (50 km) radius of the type locality. The member is displayed in simple mounded geometries with gentle to steep radial dips at a few localities. The central mounds are dominated by barren mudstone to skeletal wackestone lithologies, and, except for occurrences of crinoid debris, resemble the central mounds of the Brady Member. No organic framework has been noted, but tabulate corals are locally scattered in the central mounds. Flanking strata, which typically display dips of 5 to 45°, include interbedded mudstones, wackestones, and packstones, but mudstone lithologies dominate at some localities. Packstones are graded in part, and are variably dominated by brachiopods (resembles Brady Member) or crinoid debris (resembles Palisades- Kepler Member). Corals and stromatoporoids, overturned in part, are scattered to common in the flank beds, including tabulates and *Pycnostylus*. As noted earlier, periodic sediment removal locally punctuated mound sedimentation. Relict botryoidal and isopachous cements and internal sediment fills are displayed in some flanking units (e.g., Fig. 15, see symbols), supporting early submarine cementation within the mounded facies.

Isolated mounds with radially dipping flanks, however, are not the dominant geometry within the LeClaire Member. The bulk of exposures of the member are characterized by horizontal to gently dipping strata. Scattered crinoid debris, locally with corals and brachiopods, occurs in vuggy skeletal mudstone to wackestone fabrics in the horizontally-bedded facies (Figs. 15, 16). Such facies are observed to interfinger laterally with laminated Anamosa strata and also merge with mounded LeClaire facies (Fig. 16). These non-mounded facies of the LeClaire are interpreted as mud-dominated bank facies, which merge in complex geometries with mound facies to form laterally extensive stratigraphic bodies locally reaching dimensions to at least 2 miles (3.2 km). These bodies are observed to interfinger at their margins with laminated Anamosa strata in low-angle down-dip positions, indicating that the mound-bank complexes occupied a slightly higher depositional position than the inter-mound laminated and mud-rich strata of the Anamosa Member. This resembles mound and inter-mound relationships in the Brady Member, but the lateral extent and complex geometries of LeClaire stratigraphic bodies differ from anything seen in the Brady Member. In addition, occurrences of "normal-marine" benthos in the LeClaire

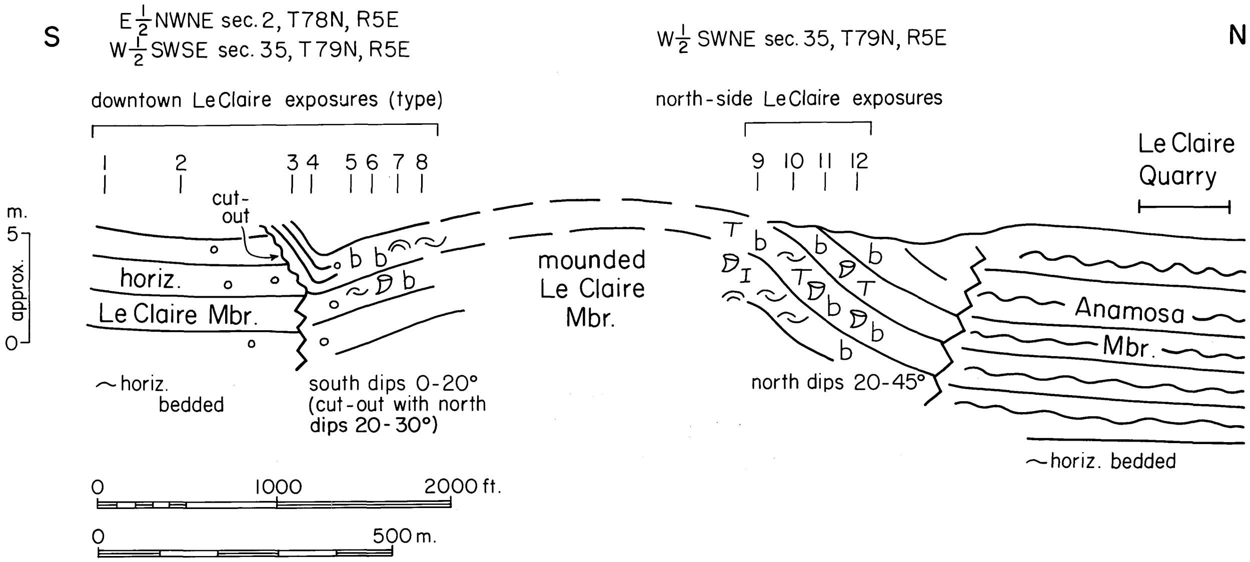

Figure 15. General stratigraphic relations in Gower Formation, type LeClaire area, Scott County, Iowa. See page 86 for lithologic symbols.

(including crinoids and trilobites) are in marked contrast to faunas of the Brady (Witzke, 1983). What depositional factors were responsible for differences between coeval carbonate mound facies in the LeClaire and Brady members?

Witzke (1987, p. 245) proposed that Gower deposition occurred within a salinity-stratified embayment of the epicontinental seaway under general antiestuarine circulation (AEC) (see Fig. 17). It was suggested that the LeClaire stratigraphic bodies formed a partial circulatory barrier to bottom water exchange within the embayment (ibid.): "Salinity stratification under AEC is most easily maintained when an entrance sill or barrier is present at the mouth of a semienclosed basin or embayment. Complex mound and bank facies of the LeClaire member are developed in the interpreted mouth area and apparently formed an effective barrier to bottom circulation. Pronounced salinity stratification within the basin accounts for absence of shelly benthos in bottom areas; shelly faunas thrived only in the upper water layer in areas where carbonate mounds rose above the halocline." The laminated Anamosa facies, therefore, is interpreted to have been deposited in subtidal hypersaline environments in back-barrier settings. "Rates of bottom outflow from the basin apparently were sufficient to prevent evaporite saturation during most of Gower deposition. Surface inflow from the adjacent open-marine seaway permitted stenohaline faunas to live in the entrance area of the basin [i.e., LeClaire Member]. Westward decrease in faunal diversity in the mound facies [Brady Member] indicates a gradient of increased salinity toward shore within the oxygenated upper water layer, as expected in AEC" (ibid.). Carbonate mound facies within the Gower Formation formed through the in-place accumulation of mud and skeletal debris in biotic oases within an otherwise inhospitable hypersaline arm of the seaway. Episodic storm and current activity transported material from the crests of the mound into mound flank and inter-mound settings. The movement of hypersaline and marine phreatic waters through the topographically elevated mounds promoted active cementation during deposition.

Summary Statement

Although carbonate mound facies occur at three general stratigraphic positions assigned to four members within the Iowa Silurian sequence, the various members share some basic similarities in their geometries and genesis. 1) The central mounds in all cases are dominated by carbonate mud fabrics, with no evidence yet found for an organic framework. Corals are noted in each member, but these are volumetrically insignificant components, subordinate to the mud and usually secondary to other skeletal grains as well

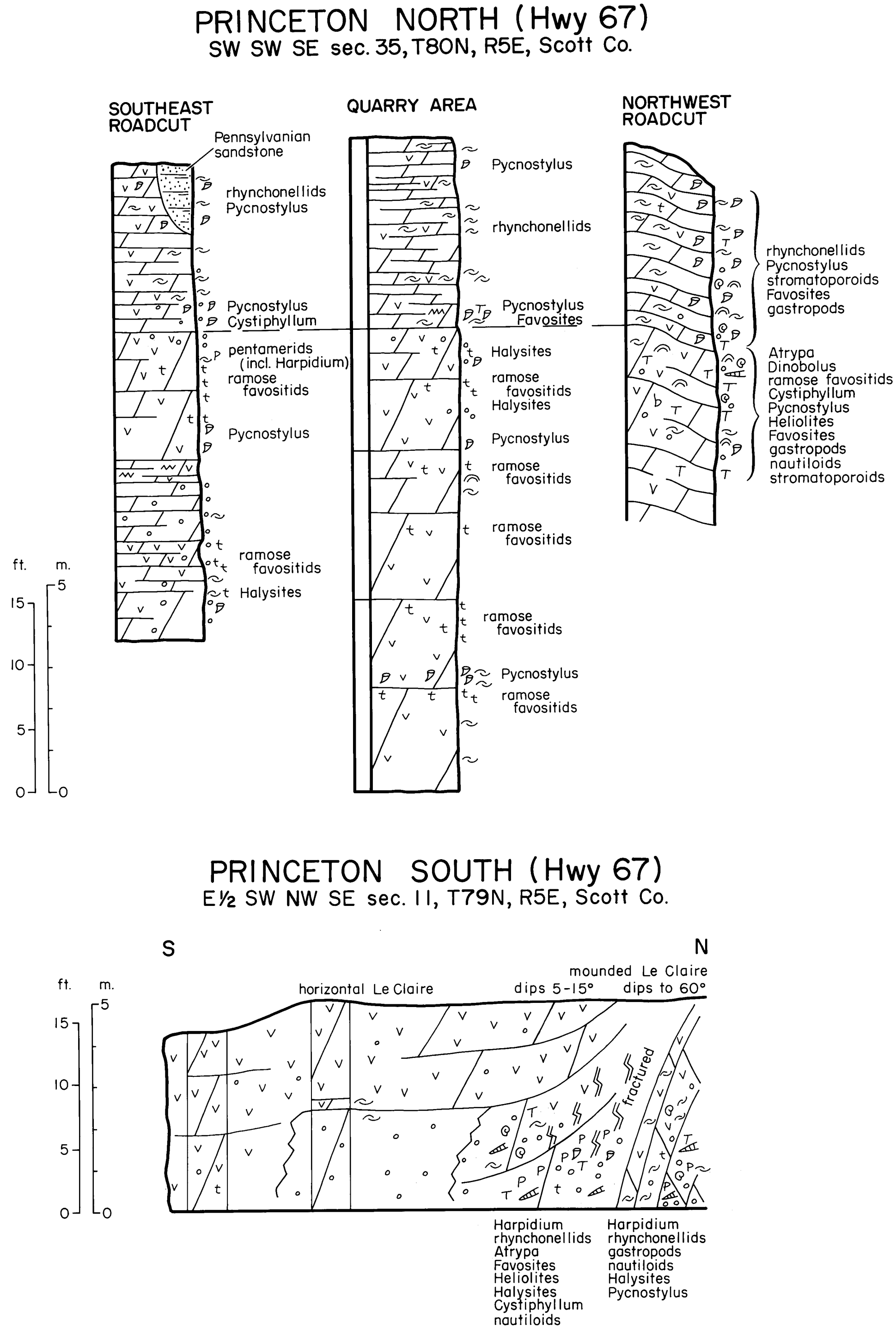

Figure 16. Exposed sections of LeClaire Member, Gower Formation, Princeton area, Scott County, Iowa. See page 86 for lithologic symbols.

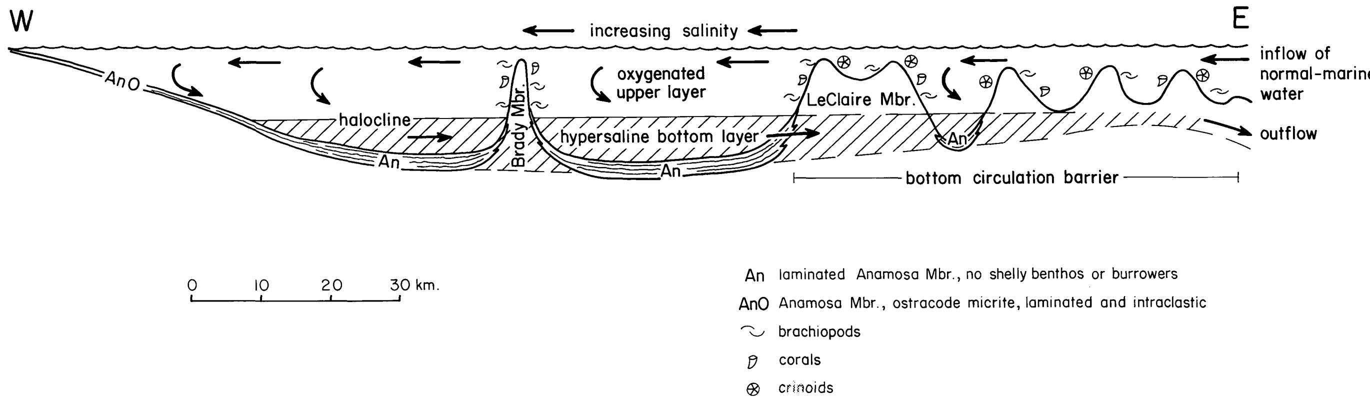

Figure 17. Gower depositional model showing antiestuarine water circulation, salinity stratification, and carbonate mounding (from Witzke, 1987b).

(especially crinoids or brachiopods). 2) Flanking strata around the mounds contain skeletal packstones, graded in part, in all four members. Occurrences of graded wedge-beds in each member are interpreted to represent episodic debris flows from the mound crests, probably triggered by storm events that swept the mounds. No evidence of shoaling has been noted in any of the mounds, and they are all interpreted to have formed primarily below fairweather wavebase. 3) All mound facies show evidence of submarine cementation, especially prominent botryoidal and isopachous relict fibrous or bladed cements. In addition, occurrences of penecontemporaneous fracturing and internal sediment fills further indicate that the mound facies became lithified during their deposition. 4) Similar geometries are noted in all four members, including isolated single mounds with radial dipping flanks, and mounded complexes representing coalesced groups of mounds with complicated internal geometries. Three members (excluding the Brady) further show lateral relationships between mounded facies and horizontally-bedded skeletal/mud bank facies.

Although the similarities are striking, certain differences between mound facies in the various members are of special note. 1) The vertical and lateral scale of carbonate mounds varies significantly in the four members. Mounds of the Johns Creek Quarry Member do not exceed about 50 feet (15 m) in height, but mounds of the other three members reach vertical dimensions of 150 to 200 feet (45-60 m). The lateral extent of single mounds (that is, those displaying simple radial dipping patterns in their flanks) varies between members (maximum diameters noted): Johns Creek Quarry (500 ft; 150 m); Brady (2000 ft; 600 m); Palisades-Kepler and LeClaire (4000 ft; 1200 m). Mound/bank complexes in the Palisades-Kepler and LeClaire members reach lateral dimensions of 2 miles (3.2 km) or greater. 2) The mounds are buried by rocks deposited in several different environments. The Johns Creek Quarry mounds are buried beneath deeper-water open-marine subtidal facies of the Welton, but the Palisades-Kepler mounds are capped by restricted-marine units of the Gower Formation. The Brady and LeClaire mounds were terminated by subaerial exposure or burial within restricted nearshore facies. 3) Some noteworthy lithologic and paleontologic differences contrast the various mounded units. Large upright colonial rugosans are noted only in the Johns Creek Quarry Member, and rhynchonellid-protathyrid packstones in flanking wedge-beds are characteristic of the Brady Member. Crinoidal packstones are absent in the Brady, but are especially well developed in the Scotch Grove mounds. Stromatolites and laminated sediments within mounded facies have been observed only in the Brady Member.

What controlled the limits to mound growth during the various stages of carbonate mounding? In general, the vertical limits to growth would be controlled in part by the balance between three general factors: 1) rates of carbonate production, 2) rates of subsidence (tectonic accomodation), and 3) rates of sea-level change. As interpreted earlier, the vertical limit to mound growth at any given time may have been marked by the position of normal fairweather wavebase, as suggested by the abundance of mud and the absence of wave-resistant structures. In addition, various environmental parameters not directly related to these factors may also have played a role, such as, changes in productivity, circulation, salinity, or climate. The first stage of mound growth in Iowa, the Johns Creek Quarry Member, was seemingly aborted at a much earlier stage than subsequent mounding. It is buried by the deepest-water sediments interpreted for the Iowa Silurian (Witzke, 1981a; Johnson, 1977b), and it seems reasonable to suggest that the rates of sea-level change during transgression outstripped rates of carbonate production within the mounds. By contrast, the flourishing normal-marine biotas that built the Palisades-Kepler mounds were elimated coincident with a significant environmental crisis associated with changes in seaway circulation and increasing salinity stress, probably brought on by a relative drop in sea-level (Witzke, 1983). Lastly, although the record has been erosionally removed, growth of the Gower mounds was probably terminated by relative changes in sea-level that led to the complete withdrawal of the seaway from eastern Iowa.

To conclude, I unequivocally maintain that an ecologic reef model fails to explain the development of carbonate mound facies in the Silurian of eastern Iowa. It is inappropriate to attempt to constrain these remarkable features to models more aptly applied to Neogene coral reefs. In fact, there are no clear modern analogs for the carbonate mounds seen in the Silurian of Iowa and other areas of the Midwest. Nevertheless, there are some potential late Paleozoic analogs that share much in common with these Silurian mounds, namely the Waulsortian mounds of the Lower Carboniferous of Europe and North America. Many features of Waulsortian mounds (see Wilson, 1975, p. 148-168) are shared with the Silurian mounds of Iowa and elsewhere, summarized as follows: 1) central mounds dominated by lime mud (skeletal mudstones to wackestones); 2) skeletal material dominated by crinoid and bryozoan debris, with scattered corals, brachiopods, and other fossils; 3) mounds lack an organic framework; 4) mounds surrounded by dipping flank beds (5-50°, locally slumped), including abundant crinoidal packstones/grainstones at many localities ("flanking encrinites"); 5) occurrences of submarine cements, especially void-filling stromatactoid structures; 6) early lithification suggested by presence of muddy internal sediment fills and fissure- and cavity-filling skeletal material; 7) mounds show wide variations in scale, ranging between about 10 and 500 feet (3-150 m) in height with lateral dimensions of 200 feet to 2 miles (60 m-3 km); 8) mounds commonly display simple hay-stack geometries, but asymmetric and complex growth morphologies and bank facies are also known; and 9) mounds have been generally interpreted as layered accumulations formed below fairweather wavebase. The similarities are striking, and it may be fruitful to pursue further comparitive studies.

RELATIVE SEA-LEVEL CHANGES IN THE IOWA SILURIAN SEQUENCE

Philosophical Considerations

As first proposed by Johnson (1975), the vertical succession of paleocommunities and sedimentary environments in the Iowa Silurian sequence provides evidence for relative changes in sea-level during deposition. In any given region or locale, such as eastern Iowa, temporal changes in relative sea level during deposition can result from the interplay of several factors: 1) rates of sediment accumulation, 2) rates of tectonic or crustal subsidence, and 3) global eustatic changes in sea level. For example, if the rate of tectonic subsidence exceeds that of sediment accumulation, there will be an apparent local deepening of relative sea level. On the other hand, global eustatic rises in sea level will produce synchronous changes in depth-related sedimentation and paleocommunities that should be recognized both interregionally and between continents. Eustatic sea level changes should be apparent at all localities where the rate of eustatic change exceeds that of sediment accumulation and subsidence. There has been consternation among some workers who have been vexed by our ability to distinguish eustatic sea-level changes from those induced by local or regional tectonic processes. Other workers have been concerned with the absence of a clear-cut mechanism for eustatic change during non-glacial times, and some have opted to model sedimentary cycles as a consequence of tectonic subsidence largely independent of eustatic change.

Nevertheless, our ability to recognize eustatic change is approachable through regional and interregional stratigraphic and sedimentologic study. It begins with the establishment of local paleobathymetric curves interpreted from the stratigraphic record. Synchronous and parallel changes in relative paleobathymetry across widely separated areas and between continents provide compelling evidence of probable eustatic controls on sedimentation. Of course, detractors will point out that synchroneity cannot be conclusively demonstrated within the limits of our biostratigraphic resolution for the Paleozoic. Such will always be the nature of the problem, even within tight biozonal or radiometric constraints. Unless a major new breakthrough in precision age-dating occurs, the spectre of time-transgressive boundaries will never completely go away. It seems better to apply Occam's razor to our observations and suggest that parallel changes in relative sea level coincident with biozonal boundaries are most reasonably interpreted as synchronous. Processes become less likely and more cumbersome when we need to invent some interregionally time-transgressive parallelism between biostratigraphy, tectonic subsidence, and sequential changes in sedimentation in order to dismiss the potential for synchroneity.

There seems to be two basic approaches for sorting out eustatic and tectonic signals. First, there are those who initially attempt to apply various tectonic subsidence models to a stratigraphic sequence (e.g., backstripping models), and evaluate the sequence in terms of locally or regionally driven processes. However, eustatic signals become more difficult to approach if original assumptions drive the sedimentary cycles ad hoc by local tectonic processes. On the other hand, there are those who seek to identify potential global eustatic signals through interregional stratigraphic and sedimentologic studies. Those signals that are apparently synchronous and recognized over broad regions in different tectonic regimes and on different continents are the most likely to be eustatic in origin. Other non-correlable changes in relative sea level, by contrast, are more likely to reflect a local sedimentary or tectonic signal.

At present, the two approaches appear philosophically at odds. It seems to this writer that once eustatic events are identified (or proposed), it then becomes possible to identify (or propose) local tectonic processes with greater precision than possible before. By contrast, driving our models of sedimentary cyclicity by means of local tectonic processes provides limited independent means of contrasting eustatic and tectonic signals. Sequence stratigraphy, particularly within cratonic settings, seems to provide a better way of evaluating local tectonic processes than the other way around, that is, explaining the sequence stratigraphy by means of local tectonic processses. Sorting out the eustatic signal provides a powerful means for evaluating changes in subsidence and sediment accumulation and, hence, for more accurately calibrating local tectonic influences. In addition, eustatic components recognized in local sea-level histories can potentially complement the biostratigraphy to provide increased chronostratigraphic resolution.

Background

The Silurian sequence in Iowa provided the starting point for a long-term and widespread evaluation of potential eustatic events for the Early Silurian by Markes Johnson and colleagues, based on relative changes in benthic paleocommunities and sedimentation patterns. Beginning with his seminal work in Iowa (Johnson, 1975, 1977b), a relative sea-level curve for Iowa was subsequently compared to curves generated at various localities in North America (e.g., Johnson and Campbell, 1980; Johnson et al., 1981; Johnson and Lescinsky, 1986), South China (Johnson et al., 1985), and Baltica (Johnson et al., 1991). There proved to be a remarkable coincidence of several high-stand events for these three continents, both in relatively stable cratonic settings and in more tectonically active areas. A simple and elegant conclusion to these findings was given: "The mass of detailed Silurian data now available from three paleocontinents strongly argues in favor of practical eustacy" (ibid., p. 335).

Following the lead of Johnson, Witzke (1981a, 1983) proposed a relative sea-level curve for the post-Llandoverian sequence in the Iowa area. Additional workers, notably Shaver et al. (1978), recognized probable synchroneity of Silurian carbonate mound/reef generations in the Great Lakes area and Iowa. The eustatic signal for the Wenlockian-Ludlovian has not yet been as clearly established as that for the Llandoverian, but it is only through the integration of local bathymetric histories that we can ever hope to address the issue of eustacy. It is to that end that a relative sea-level curve for the Iowa Silurian is presented (Fig. 18). The illustrated curve is intended to reflect relative, not absolute, changes in water depth. It is not assumed, *a priori*, that this curve is exclusively eustatic in origin. Relative changes in paleobathymetry are reflected both by changes in depth-related sedimentary factors as well as changes in benthic community structure. Shallowing and deepening trends during deposition are used to bracket a series of transgressive-regressive (T-R) sedimentary cycles, for convenience given alphanumeric designations (Fig. 18).

Assemblages (recurrent groupings) of Silurian benthic paleocommunities, as documented through various studies, are known to be arranged in general onshore-offshore trends, which provide a means of ordering relative changes in bathymetry during deposition. Boucot (1975) gave these benthic assemblages (B.A.) numerical rankings reflecting their relative positions, onshore-to-offshore, B.A 1 to B.A. 6. While not without its problems and detractors, the general

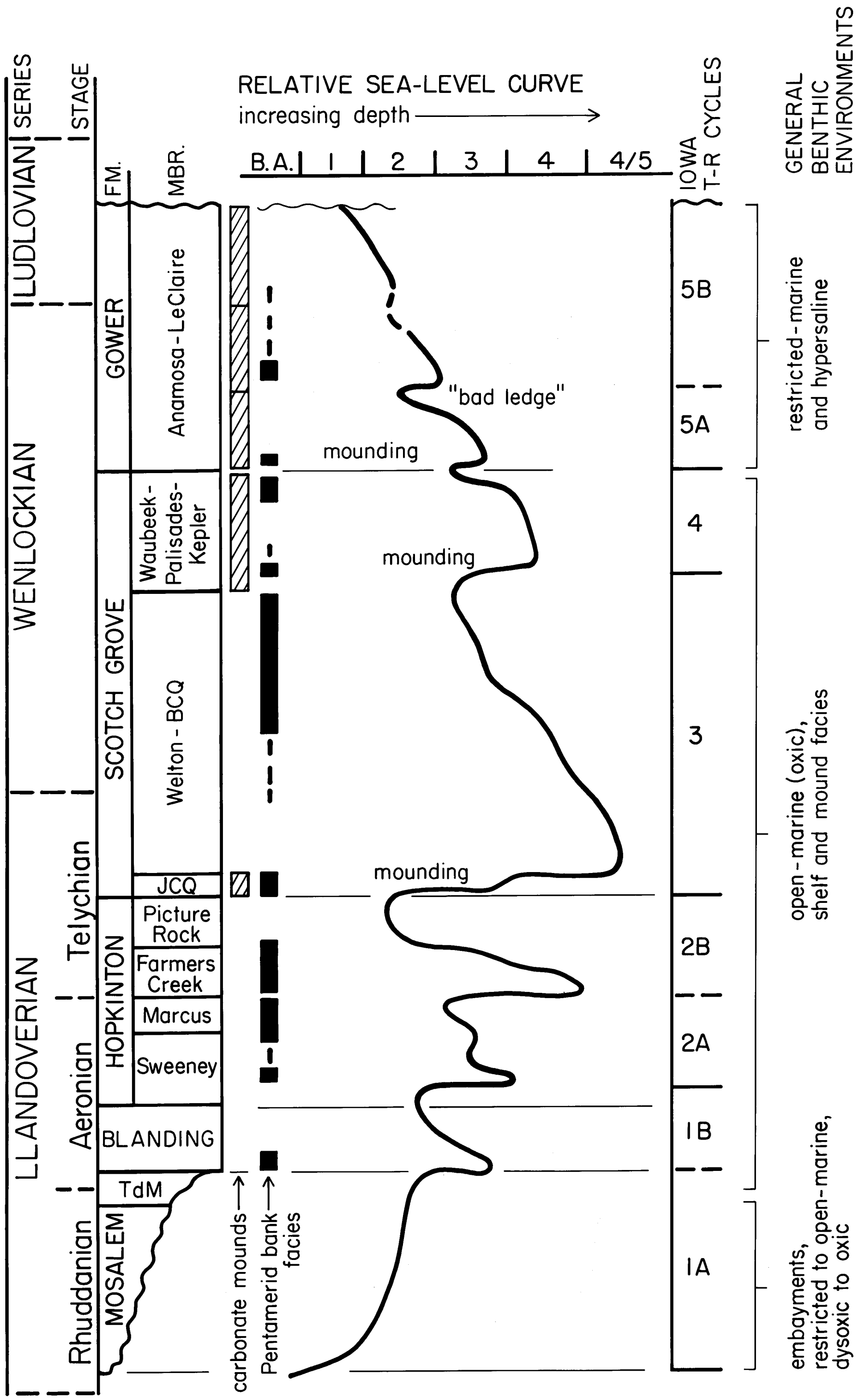

Figure 18. Proposed relative sea-level curve for Silurian sequence of eastern Iowa. Numbered benthic assemblages (B.A.) are used to interpret relative paleobathymetric position, based on recurring fossil associations and sedimentary criteria.

ranking of these fossil assemblages has proven to remarkably consistent at various localities around the world. It is not assumed, however, that a B.A. 3 community in Iowa occupied the same exact range of water depths as another B.A. 3 community in New York or China, and the numerical designations are used only as a useful concept for recognizing *relative* changes. A broader discussion of the B.A. positions assigned to the various Iowa paleocommunities is given by Witzke and Johnson (1992). A general discussion of the sedimentary and paleontologic evidence for sea-level changes in the Iowa Silurian follows.

Mosalem through Lower Sweeney Sedimentary Cycles

The first episode of Silurian transgression during the Rhuddanian (early Llandoverian) encroached westward up valleys incised in the Maquoketa shales in eastern Iowa. Infilling of paleovalleys proceeded during Mosalem deposition, upward through a progression of nearshore B.A. 1 and more offshore B.A. 2 communities in restricted and partially oxygen-stressed embayments. Relative deepening of sea-level continued as open-marine carbonate shelf facies of the Tete des Morts Formation spread westward to a shoreline trending across eastern Iowa. The Rhuddanian sequence in eastern Iowa records a gradual deepening of relative sea-level, with progressive westward inundation by shallow marine environments. No shallowing (regressive) episodes are recognized within this interval.

An abrupt shift in sediment grade and paleocommunities is noted at the base of the Blanding Formation, suggesting significant depositional deepening during the early Aeronian. Basal Blanding strata (and equivalent basal Waucoma strata) contain B.A. 3 and B.A. 4 faunas, sharply overlying B.A. 2 coral-stromatoporoid faunas of the Tete des Morts (or Ordovician strata). A shift from higher-energy wackestone/packstone facies to lower-energy mud-dominated facies also marks the contact. In addition, the Blanding oversteps the Mosalem-Tete des Morts edge, and occurs westward to central Iowa. The faunal and sedimentary shifts, in combination with notable geographic expansion, indicates that the early Aeronian was a time of significant deepening of the seaway. The abruptness of the basal Blanding expansion is interpreted to reflect a significant increase in the rate of transgression.

Upper Blanding and lower Sweeney strata are interpreted as a general shallowing-upward sequence, showing a progressive upward increase in higher-energy facies (wackestone/packstone) and a return to tabulate coral-stromatoporoid communities (B.A. 2). In general, the Rhuddanian-early Aeronian sequence in Iowa forms a general transgressive- regressive cycle of sedimentation (T-R Cycle 1 on Fig. 17). However, a prominent sea-level highstand event is clearly marked at the base of the Blanding, and this transgressive inflection provides a convenient position to subdivide T-R Cycle 1 into two parts (1A, 1B, Fig. 18).

Hopkinton Sedimentary Cycles

In general, the Hopkinton sequence forms a third-order(?) transgressive-regressive sedimentary cycle with maximum sea-level highstand in the lower Farmers Creek (T-R Cycle 2, Fig. 18). However, the sequence can be subdivided into two smaller-order cycles (2A, 2B). The first is marked by a transgressive event in the mid Sweeney, recorded by a noteworthy change in sediment grade to dense mudstones and wackestones, similar to lithologies seen in the lower Blanding. Roughly coincident with this sedimentary change, B.A. 3/4 stricklandiid communities are noted. The magnitude of this transgressive event is reflected regionally, as the lower Hopkinton Formation oversteps the Blanding edge to overlie Ordovician rocks from central Iowa into Nebraska (Witzke, 1981a). A return to coral-stromatoporoid and pentamerid communities (B.A. 2/3), accompanied by a notable increase in higher-energy packstones, in the upper Sweeney-Marcus suggests an overall shallowing trend.

A second Hopkinton sedimentary cycle (T-R Cycle 2B) was initiated by deposition of lower Farmers Creek strata, which abruptly overlies the Marcus and marks the maximum highstand for the formation. A notable change in sediment grade in the lower Farmers Creek, dominated by crinoidal mudstones and wackestones, as well as a prominent shift in benthic communities (B.A. 4 stricklandiid association) suggest significant depositional deepening. Progressive shallowing through the remainder of the Farmers Creek and Picture Rock members is indicated by the upward shift from B.A. 4 through B.A. 2 communities and an increase in packstone interbeds (pentamerid tempestites and crinoidal packstones). The Hopkinton sequence is interpreted to record two transgressive highstands (mid Aeronian and early Telychian, Fig. 18), with maximum lowstand marked at or near the top of the formation. The Hopkinton-Scotch Grove contact is conformable through the outcrop belt, but possible local disconformable relations in the subsurface further supports a regressive maximum in the uppermost Hopkinton.

Scotch Grove Sedimentary Cycles

The Scotch Grove Formation abruptly overlies the Hopkinton, marking significant depositional deepening in the sedimentary sequence. The appearance of skeletal mudstones and wackestones in the Johns Creek Quarry Member above Hopkinton skeletal wackestones/packstones is interpreted to record a shift to lower-energy depositional conditions. Likewise, the change from B.A. 2 to B.A. 3 communities across the boundary further supports a relative deepening. The Johns Creek Quarry Member is interpreted to include only the initial stages of an overall transgression, and overlying lower Scotch Grove strata show evidence of continued depositional deepening. Pentamerid- rich inter-mound units (B.A. 3) and mud-rich carbonate mound facies of the Johns Creek Quarry are succeeded by skeletal wackestone and mudstone units containing diverse deeper-water B.A. 4 and 4/5 benthic communities, including common stricklandiids. A transgressive highstand is interpreted for the lower Welton (and correlative strata in the lower Scotch Grove), which marks the highest relative sea-level for the entire Silurian sequence of Iowa (approximately C_6 late Telychian).

Progressive shallowing is interpreted for the remainder of the lower to middle Scotch Grove sequence, marked by a transition from B.A. 4 to B.A. 3 communities. Pentamerid packstone tempestites are prominent in the mid Scotch Grove, reflecting impingement of storm currents on the bottom on a more regular basis coincident with depositional shallowing. Coralline communities were developed locally during this shallowing episode. The lower to mid Scotch Grove sequence is interpreted as a single transgressive-regressive cycle of deposition (T-R Cycle 3, Fig. 18), spanning the late Llandoverian (late Telychian) into the early Wenlockian.

A relative bathymetric deepening is suggested for the upper Scotch Grove, coincident with the second phase of carbonate mound growth in the Silurian sequence (T-R Cycle 4, Fig. 17). A relative deepening phase is suggested by shifts in sediment style and benthic paleocommunities. Tempestite packstones with B.A. 3 communities are common in the mid Scotch Grove, but upper Scotch Grove inter-mound facies commonly lack packstones and contain outer B.A. 3 to B.A. 4 benthic faunas. Upper Scotch Grove sedimentation was abruptly terminated by significant environmental change, marked by a shift from open-marine to restricted-marine sedimentation. This shift, in combination with possible subaerial exposure of Palisades-Kepler mound crests and a change to shallower B.A. 2-3 communities at the Gower contact, is interpreted as evidence for depositional shallowing at the close of Scotch Grove deposition. As such, the upper Scotch Grove is interpreted to form a discrete T-R cycle of approximately early Wenlockian age.

Gower Sedimentary Cycles

The base of the Gower Formation marks a third generation of carbonate mound growth in the Iowa Silurian sequence. Following a brief shallowing phase (and local subaerial exposure?) at the close of Scotch Grove deposition, environmental conditions changed significantly within the East-Central Iowa Basin at the onset of Gower sedimentation. Circulatory restrictions established salinity stratification within the basin, probably coincident with the growth of mound and bank facies at the mouth of the enclosed basin. A general shift from B.A. 3 to B.A. 2 benthic communities between the Palisades-Kepler and Brady mounds suggests relative shallowing, but basal Brady mounds and much of the LeClaire Member contain B.A. 3 faunas. Although the Gower contains generally shallower-water faunas than the upper Scotch Grove, a minor deepening phase seems to be coincident with the onset of Gower mounding. This is followed by general and progressive shallowing. The Gower is capped by an erosional unconformity which formed following withdrawal of Silurian seas from the area, and the type Anamosa sequence also shallows upward to intraclastic peritidal facies (with B.A. 1 faunas) below the sub-Devonian unconformity. Therefore, the Gower sequence is interpreted as a general transgressive-regressive sequence (T-R Cycle 5, Fig. 18). This cycle began in the late early or mid Wenlockian, but its upper age limit is not known with certainty (probably spans into the Ludlovian).

However, there is evidence that the larger Gower cycle can be subdivided into two or more smaller-scale subcycles (tentatively labelled T-R Cycles 5A and 5B, Fig. 18). The bulk of Gower strata was deposited in subtidal inter-mound environments, but the monotonous laminated sequence is known to be interrupted at several localities by irregularly laminated to intraclastic strata low in the sequence (e.g., "bad ledge" at Stone City). This interval is interpreted to reflect short-term shallowing into peritidal or shallow subtidal environments, followed by a return to relatively deeper subtidal laminated carbonate deposition. Likewise, alternations in the Anamosa sequence in the LeClaire area between horizontally laminated units and stromatolitic or skeletal mudstone intervals may be related to minor fluctuations in relative sea level (which may influence circulatory parameters). The relative stratigraphic relationships between various carbonate mound and bank facies in the LeClaire Member are not known with certainty, and it remains a possibility that more than one phase of carbonate mound growth may be represented in the Gower Formation. Occurrences of coralline mudstone/wackestone strata above the thick Anamosa sequence at LeClaire, and erosional cut-outs within LeClaire mound facies, lends support to the idea that multiple shallowing-deepening phases may be present in the Gower sequence. The details remain to be worked out.

Evaluation of Eustacy

The relative sea-level curve for Iowa provides a basis for comparison with other Silurian sections at widely separated geographic locations. The Iowa Llandoverian sequence has been compared to sequences at localities in North America, Europe, and China by Johnson and his co-workers (see summary in Johnson et al., 1991), and a remarkable coincidence in three or four sea-level high-stands was noted over broad areas. These high-stands have been correlated as follows: earliest Aeronian (lower Blanding), mid Aeronian (mid Sweeney), early Telychian (Farmers Creek), and late Telychian/earliest Wenlockian (lower Scotch Grove). Intervening and correlable sea-level low-stands, especially for the late Aeronian and mid Telychian, are marked inter-regionally by depositional shallowing, siliciclastic progradation, or unconformity surfaces.

Significant depositional deepening is recorded in the lower Scotch Grove sequence of Iowa, and this late Telychian event is apparently recognized in many Silurian stratigraphic sequences around the world. This transgressive episode was probably initiated late in *celloni* Zone time in Iowa, but maximum transgressive deepening in Iowa and elsewhere is marked in the subsequent lower *amorphognathoides* Zone. At some localities in the Midcontinent and Great Lakes area, transgressive strata encompassing the lower *amorphognathoides* Zone disconformably overlie *celloni* Zone strata (e.g., Oklahoma, Barrick et al., 1990), and the initial stages of late Telychian transgression may not be captured if unconformities or hiatuses occur in the sequence. Coincident depositional deepening is recognized at many localities throughout North America, including such distinctive stratigraphic units as the basal Bainbridge-Wayne-Clarita sequence across the southern Midcontinent (Barrick, 1978), lower Joliet and Cordell formations in the Michigan Basin area, lower Cedar Lake Formation of the Williston Basin (Johnson and Lescinsky, 1986), Sauquoit Shale of New York-Ontario, and the Jack Valley Member in the Great Basin (Sheehan, 1980).

Subsequent inter-regional Wenlockian sea-level history is not as clearly established. Progressive upward shallowing through the lower and mid Scotch Grove Formation is constrained entirely to the *amorphognathoides* Zone. This same early Wenlockian shallowing trend is apparently recognized by "gradual shoaling" in Ohio (Kleffner, 1987) and a regressive maximum in the northern Appalachian Basin (Brett and Goodman, 1991). A minor transgressive phase is indicated in the upper Scotch Grove Formation coincident with the initiation of carbonate mounding (Palisades-Kepler Member). It remains to be seen if this deepening trend, beginning at or near the base of the *ranuliformis* Zone, is of inter-regional scope. Of note, the base of the *ranuliformis* Zone in southern Ohio (Bisher Fm.) occurs within the "first deposits of a transgressing sea"(Kleffner, 1987, p. 85), and the lower Rochester Shale of New York-Ontario (lower *ranuliformis* Zone) likewise marks a possibly coeval transgressive episode (Brett, 1983; Kleffner, 1991).

Progressive shallowing is indicated through the remainder of the upper Scotch Grove sequence in Iowa (*ranuliformis-amsdeni* Zone). As noted, this shallowing trend continues into the Gower Formation, briefly interrupted by a minor deepening phase near the base of the formation (T-R Cycle 5, Fig. 18). The general shallowing-upward sequence represented by the upper Scotch Grove and lower Gower interval apparently corresponds to a late early to middle Wenlockian eustatic sea-level fall documented by Kleffner (1990, p. 322) at North American and European localities. Salina Group sedimentation (A-0 carbonate and anhydrite) in the Michigan Basin was also apparently initiated during the closing stages of this regressive interval (Droste and Shaver, 1982). A general regressive sequence continues through the late Wenlockian at many Midwestern localities (Kleffner, 1990), although a minor mid Wenlockian deepening phase may be locally recognized. For example, the local spread of deeper carbonate facies during *stauros* Zone deposition (=lower *O. sagitta sagitta* Zone of Kleffner, 1989) marks the base of a transgressive phase in the Wayne Formation (Lego Mbr.) of Tennessee (Barrick, 1983), which is roughly coeval with the minor transgressive interval at the base of the Gower Formation in Iowa.

The late Wenlockian was apparently marked by progressive shallowing at most localities (prominent regressive maximum noted in the late Wenlockian of the northern Appalachian Basin by Brett and Goodman, 1991), although the base of the Ludlovian may mark a eustatic deepening (M. Johnson, 1991, pers. comm.). One or more minor deepenings during Gower deposition have been noted, possibly suggesting that the sequence ranges into the Ludlovian.

The Silurian depositional sequence in eastern Iowa is remarkably similar to that noted in the Michigan Basin area. The Llandoverian sequence in Michigan shows the same pattern of relative sea-level rises and falls seen in Iowa (Johnson and Campbell, 1980), and the Wenlockian sequence likewise shows close parallels. Early Wenlockian shallowing is identified in the Michigan Basin area (ibid.; Johnson, 1981; Colville and Johnson, 1982). As in Iowa, the early Wenlockian shallowing episode is succeeded by a probable deepening event and the initiation of carbonate mounding (reef generation 3 of Shaver et al., 1978) in units variously assigned to the Engadine, Amabel, lower Racine, and upper "Niagara." Of note is the recovery of stricklandiid brachiopods in the lower Engadine and Racine (Boucot and Ehlers, 1963), suggestive of depositional deepening and a shift from B.A. 2 (Cordell) to B.A. 3/4 (Engadine) paleocommunities in Michigan.

Although the lower Racine and Engadine in the Michigan Basin area have been assigned a late Wenlockian age in some previous studies (e.g., Berry and Boucot, 1970), it seems more likely that these units are of early to mid Wenlockian age. This is suggested by their position above early Wenlockian strata and because the initiation of apparently coeval carbonate mounding around the southern edge of the basin (northern Indiana) occurred during the early Wenlockian. I find the stratigraphic observations of Droste and Shaver (1976; 1977; 1982) in the southern Great Lakes area compelling, namely that carbonate mound growth around the Michigan Basin began in strata coeval with the Laurel and Waldron to the south. This is particularly significant in that these latter strata are biostratigraphically constrained by Barrick (1983) to span the *ranuliformis-amsdeni* Zone of early Wenlockian age (Kleffner, 1989). The development of these mounded facies led to dramatic changes in water circulation and the beginning of Salina Group sedimentation within the central Michigan Basin. As noted by Droste and Shaver (1982, p. 23), "restricted circulation and excessive evaporation, partly effected by an early barrierlike bank system, brought about chemical precipitation of salts by middle Wenlockian time."

The petrographic similarities of Salina Group carbonates in the Michigan Basin and the Gower carbonates of eastern Iowa is striking (Witzke, 1981a). It is also noteworthy that widespread deposition of the A-1 carbonate (lower Salina Gp.) in Michigan was apparently contemporaneous with lower Anamosa sedimentation in Iowa, suggesting similar inter-regional depositional controls, most likely eustatic in origin. As noted by Witzke (1981a, p. 471), "... the simultaneous appearance of similar depositional conditions in both areas suggests that a major external control influenced sedimentation in these geographically separated areas. A widespread drop in sea level in the [mid] Wenlockian is reasonably invoked as an external mechanism that influenced the establishment of restricted circulation and elevated salinities in the partially enclosed basins in both areas." Although some workers have denied contemporaneity of later-stage carbonate mound growth (pinnacle reef crests) and laminated A-1 carbonate deposition in Michigan (e.g., Gill, 1977), recent studies have noted the coeval development of A-1 "reef crest" sediments and laminated carbonates in a salinity stratified basin (Leibold et al., 1991). The A-1 mound crests include stromatolitic mudstones and protathyrid brachiopod packstones (ibid.), closely analogous to the Brady mounds of Iowa. In fact, exposures of upper Scotch Grove and Gower Formation strata in Iowa provide a realistic working model for understanding stratigraphic relations in the Michigan Basin (also see Shaver, 1991, p. 121), with parallels as follows: upper "Niagara" and Palisades-Kepler mounds, laminated A-unit carbonates and the Anamosa Member, protathyrid-bearing A-unit pinnacle reef crests and Brady mounds, and basin marginal mound/bank facies (e.g., Fort Wayne Bank; Pleasant Mills and "Guelph" formations) and the LeClaire Member.

In summary, all relative sea-level changes recognized in Iowa have direct parallels in other regions, and most, if not all, are reasonably considered to be eustatic in origin. In particular, the four sea-level highstands recorded within the Iowa Llandoverian sequence are widely correlable, and eustacy is strongly implicated (Johnson et al., 1991). Parallel Wenlockian sea-level changes are also recognized between Iowa and much of the cratonic interior/eastern basins of North America, but additional global synthesis is needed. The general Wenlockian pattern of relative sea-level changes seen in Iowa is suggested to be widely correlable as follows: early Wenlockian shallowing, mid early Wenlockian deepening, late early to mid Wenlockian shallowing, mid Wenlockian deepening, and general late Wenlockian shallowing (minor cycles may be present).

The Ludlovian pattern is attenuated and obscured in Iowa due to pre-Middle Devonian erosion.

Six significant sea-level highstand events are recognized in the Llandoverian-Wenlockian sequence of Iowa. Most recent estimates of the duration of this time interval range between about 14 and 20 million years (Menning, 1989), suggesting an average duration of about 2.3 to 3.3 million years per cycle. The Llandoverian is known to be a time of continental glaciation in Gondwana (Caputo and Crowell, 1985), although the duration of the Silurian sea-level cycles is considerably longer than typical Milankovitch-style glacial cycles (0.02-0.4 Ma) of the Carboniferous and Quaternary. In addition, no evidence is known for glaciation in the Wenlockian (ibid.). The absolute magnitudes of depth changes within individual sea-level cycles in Iowa is difficult to quantify, but estimates of about 10 to 70 m (30-230 ft) seem reasonable (Johnson, 1987). Although glacial eustacy may be involved in some of the Early Silurian sea-level cycles, it seems likely that other mechanisms for raising and lowering sea-level were also operant.

Local Tectonic Signals

The widespread correlation of the Iowa Silurian sea-level cycles is most simply interpreted as a consequence of eustatic changes, but eustatic signals can be dampened or accentuated by local tectonic processes. Bunker et al. (1985) demonstrated that differential basinal subsidence in eastern Iowa was contemporaneous with Early Silurian deposition. Apparent rates of subsidence, as estimated from relative thickness changes, increased by a factor of about four from west-to-east into the center of the East-Central Iowa Basin during deposition of the Blanding and Hopkinton formations (ibid.; see also Fig. 6). However, the patterns of relative sea-level change are clearly displayed by variations in carbonate facies and paleocommunities along the entire extent of this transect (Witzke, 1983), and the eustatic signal does not seem to be masked by the differential rates of subsidence. However, the relative magnitude of the local sea-level changes may have been influenced by basinward variations in the rates of subsidence, as the Hopkinton cycles vary between B.A. 2 and 3 communities in the western margin area (estimated maximum 30-40 m change) and between B.A. 2 and 4 communities in the basin center (estimated maximum 40-60 m change). Maximum subsidence during Blanding-Hopkinton deposition was focused in a broad multi-county area 10-70 miles (15-110 km) north of the Plum River Fault Zone (Bunker et al., 1985).

By contrast, maximum subsidence during Scotch Grove deposition occupied a general parallel trend superimposed on the axis of the present-day Plum River Fault Zone, with abrupt thickening noted coincident with the trend of the fault itself (ibid.). In fact, a transect across the Fault Zone in Johnson and Linn counties showed that the formation more than doubles in thickness northward over a distance of only 10 miles (16 km), with a probable eight-fold increase noted in the lower part of the formation (ibid.). This thickening occurs within non-mounded facies of the Scotch Grove, although it is of interest that all large-scale mounds (> 300 m diameter) of the Palisades-Kepler Member occur north of the present-day fault. Of note, ". . . maximum structural subsidence in western Linn County apparently occurred coeval with maximum marine transgression" (ibid., p. 87). Even though wide variations in subsidence rates during Scotch Grove deposition are apparent, the eustatic signal is consistently identified, with a shift from B.A. 2 to B.A. 4 (or 4/5) communities recognized from the top of the Hopkinton through lower Scotch Grove interval at all sections. It seems reasonable to suggest that the rates of sea-level change in the late Telychian far exceeded even the highest rates of subsidence and sediment accumulation in the East-Central Iowa Basin.

The Gower Formation is erosionally bevelled across eastern Iowa, and subsidence rates are more difficult to constrain in the area due to this truncation of thickness. Nevertheless, facies architecture within the East-Central Iowa Basin clearly demonstrates the profound influence tectonically-driven subsidence played in the development and location of the restricted Gower embayment. In fact, laminated Gower sediments are largely restricted to the central area of the basin. The abrupt and synchronous onset of restricted sedimentation in the basin bespeaks the probable influence of eustatic change. Parallels in sedimentation in the Michigan and East-Central Iowa basins during the mid Wenlockian through Ludlovian reinforce the idea that circulatory restriction was a complex response to variations in subsidence, sediment accumulation, and sea-level change between the central basins and their margins. The East-Central Iowa Basin can truly be considered a "little sister" of the larger Michigan Basin, sharing the same familial signature of unique sedimentary facies, sea-level history, and basinal subsidence.

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THE DEVELOPMENT OF INDIAN BLUFF CAVE

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Indian Bluff Cave, located in Picture Rock Park (STOP 1), Jones County, is developed in the Farmers Creek Member of the Hopkinton Formation. This cave illustrated in Figure 1 (Hedges, 1974) has a main passage with a known length of about 500 feet, and short side passages. The cave varies in height from less then two feet to about seven feet. Much of the floor consists of a "clay" fill of unknown depth. The main passage forks a short distance beyond Gietkowski's Grotto and both branches become impassable after a short distance due to this fill. The cave is currently dry, except for occasional seepages of vadose water and thus is hydrologically inactive.

This cave was developed under phreatic (at or just below the water table) conditions, with development largely controlled by the presence of favorable strata and pre-existing rock fractures (Bounk, 1983a).

Bretz (1942) suggested that caves of shallow phreatic origin are usually linear in form, often with side passages. Picknett et al. (1976) suggested that this form is related to: 1) the increased solutional potential of descending ground water, where it mixes with water in

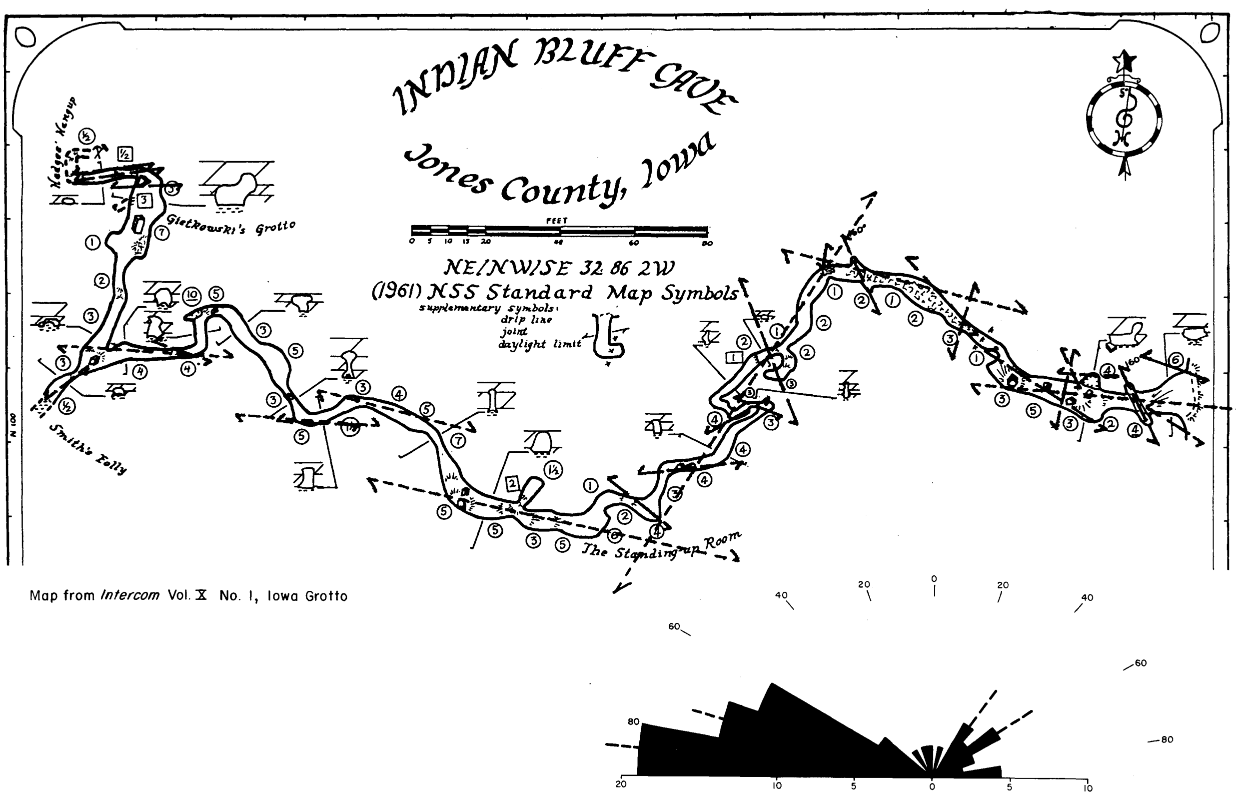

Figure 1. Map of Indian Bluff Cave (Hedges, 1974), with cave fractures shown as solid lines, and prominent surface fracture trends (illustrated by the rose diagram), which control passage trends shown as dashed lines.

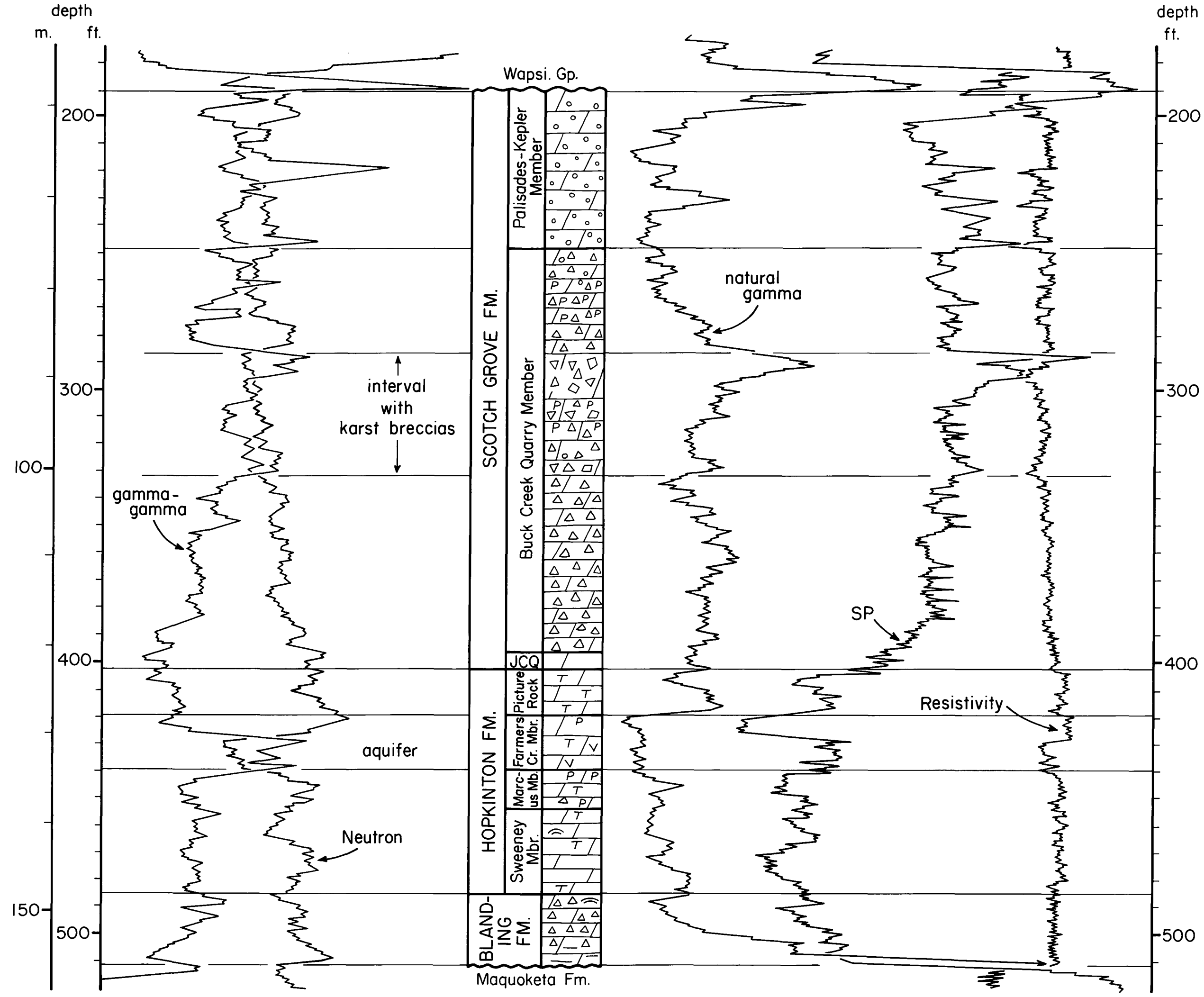

Figure 2. Stratigraphic section from Robins corehole, Linn County, shown with a suite of geophysical logs (modified from Wahl and Bunker, 1986; Witzke, 1991, pers. comm.)

the saturated zone; and 2) the uppermost part of the saturated zone, being the first locus of prolonged contact between this descending water and any stratum. Solutional openings tend to develop at or near the top of the phreatic zone, resulting in relatively rapid water movement at this level. This increases solution at this level at the expense of deeper levels (Davis, 1960).

The Farmers Creek Member, formerly the *Cyclocrinites* Beds of Markes Johnson (1975), is characterized by the green calcareous algae *Cyclocrinides dactioloides*, as well as a diverse marine fauna. This unit is commonly vuggy, with greater permability to groundwater movement, and thus under favorable conditions it has a greater susceptibility to karst development than does much of the Silurian strata of Iowa (Bounk, 1983a).

Figure 2 shows a suite of geophysical logs from a core hole in central Linn County. The natural gamma log indicates a relative lack of argillaceous material while the gamma gamma log indicates a high porosity and the neutron log indicates a high water content in the Farmers Creek Formation. Together, these indicate a relatively pure dolomite, with a high porosity and permeability; all conditions favorable to cave and karst development. This unit has been traced throughout the

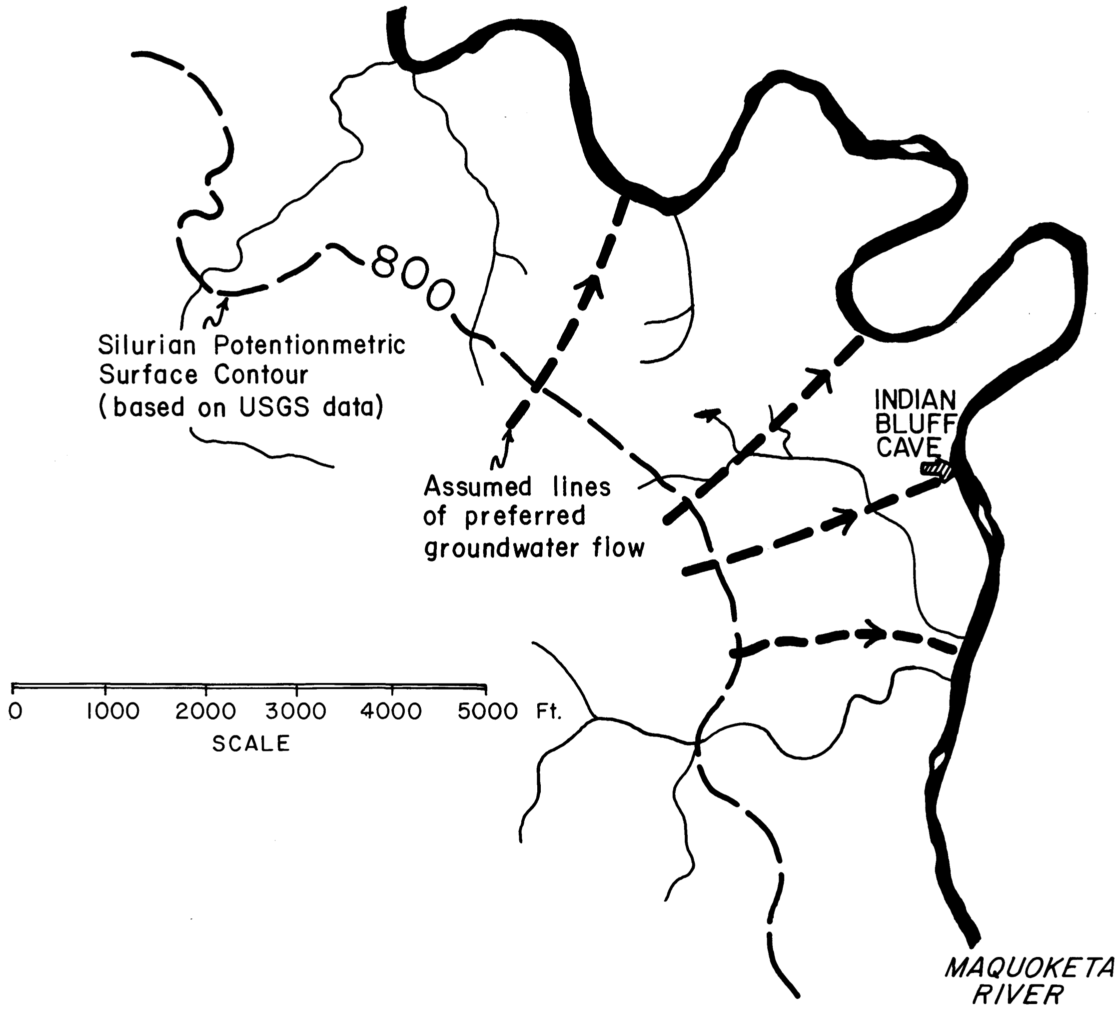

Figure 3. Map of the area around Indian Bluff Cave, showing the relationship between the cave and present day interpolated groundwater flow. (from Bounk, 1983a).

subsurface of Linn Co. (Wahl and Bunker, 1982) and is a major water-bearing zone throughout most of eastern Iowa. An unusually high percentage of Silurian caves in Iowa occur in this unit, thus indicating that under phreatic conditions this unit is relatively susceptible to cave development (Bounk, 1983a). A similar situation is present in the upper part of the Scotch Grove Formation, where short caves and rock shelters are common.

While the known part of Indian Bluff Cave trends in a east-west direction subparallel to the present direction of ground water flow (Fig. 3), much of the cave appears to be controlled by a series of fractures and fracture trends (Fig. 1). Many of these fractures, while conspicuous in the cave by means of passage shape and/or ceiling slots, are numerally inconspicuous on the surface. Other fracture trends which control passage development do not appear as fractures in the cave and are only detectable by means of additional fracture studies. These fractures and fracture trends, under the influence of a west-to-east ground-water flow, were selectively dissolved in a pattern which best facilitated groundwater flow to what is now the Maquoketa River.

In conclusion, the relationship between the present potentiometric surface and the orientation of this cave indicates that Indian Bluff Cave developed in a ground-water system similar to that at present, although under a higher water table (Bounk, 1983a). This suggests that the cave system was formed after the last glaciation in this area, in the pre-Illinoian (Hallberg, 1980) and after development of drainage on the drift surface. Although landscape development on the pre-Illinoian deposits has been complex, present day drainage has evolved since at least Late Sangamon time, with extensive erosion in Wisconsin time (Hallberg, 1979, personal communication). This brackets cave development somewhere between Late-Sangamon time and a lowering of the phreatic zone to a level below the known cave (Bounk, 1983a).

The fractures and fracture trends at Indian Bluff Cave appear to exercise three or four types of control over cave development. The first, most local of these, is in the development of cross-fractures, such as seen near the cave entrance. In this case, as elsewhere in the upper Midwest, solution occurs along a fracture which crosses the passage trend. This fracture, while highly susceptible to solution, does not facilitate down head water flow, and the fracture experiences only local solution which makes it highly visible as a cross slot. The second type of control occurs where the passage trends along a fracture or series of parallel fractures, which are visible in the ceiling. This presumably occurs where a relatively small number of fractures facilitate down head flow. This grades into the third where a length of passage follows a known fracture trend, although no fractures of this trend can be seen in this part of the cave. The fourth type of control occurs where a passage trend averages around a known fracture trend, but does not occupy the trend. Often, the passage will be locally controlled by other fractures or trends, which cause it to zigzag around a numericaly prominent trend which often is only detected by fracture studies (Fig. 4). This type of control can be seen in the SW-NE trending segment of passage near the entrance of Indian Bluff Cave. Another example of this is seen in Duttons Cave near West Union, where alternate segments of passage follow near N-S and near E-W fracture trends, averaging around a N60°E to N70°E trend, which is seen nowhere in the cave (Fig. 5). This trend, if extended to the southwest, runs close to Middlestadt Cave and under the north edge of an area of closed drainage. Both of these have been dye traced to Dutton's Cave.

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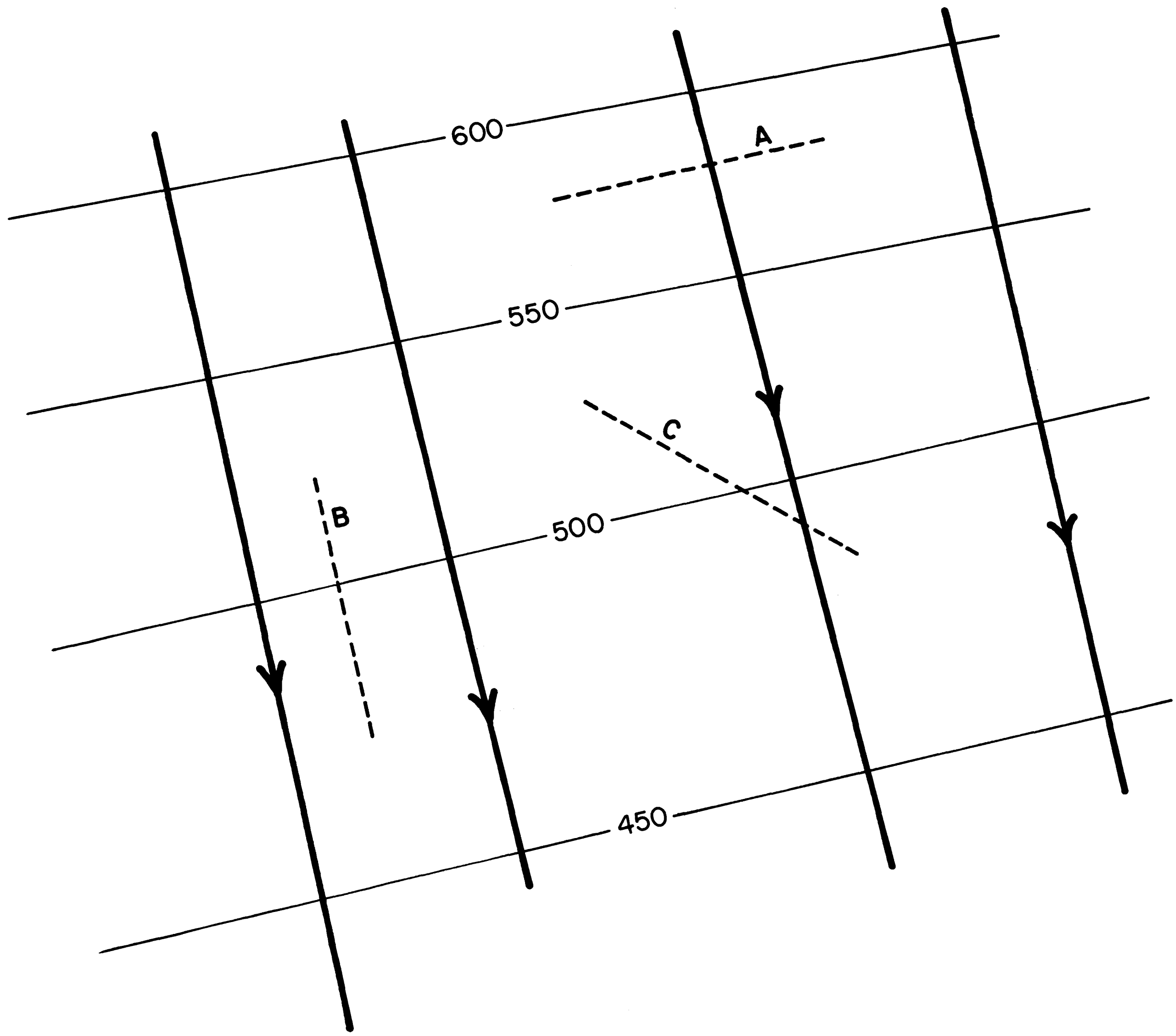

Figure 4. Diagram showing the relationship between the hydraulic gradient (lines with numbers indicating elevation above datum) and fractures (dashed lines). Heavy lines indicate preferred groundwater flow, such as would occur in a hydraulically isotropic rock. In a rock where major permeability occurs along fractures, a fracture in the orientation of fracture A will carry little or no water unless it connects two fractures with flows. Therefore, water in this fracture will become rapidly saturated with carbonate which will not be removed, thus little solution will occur. A fracture in the orientation of B will carry a high groundwater flow, constantly removing dissolved carbonate allowing solution to occur relatively rapidly. Fracture C is an intermediate example. (from Bounk, 1983a).

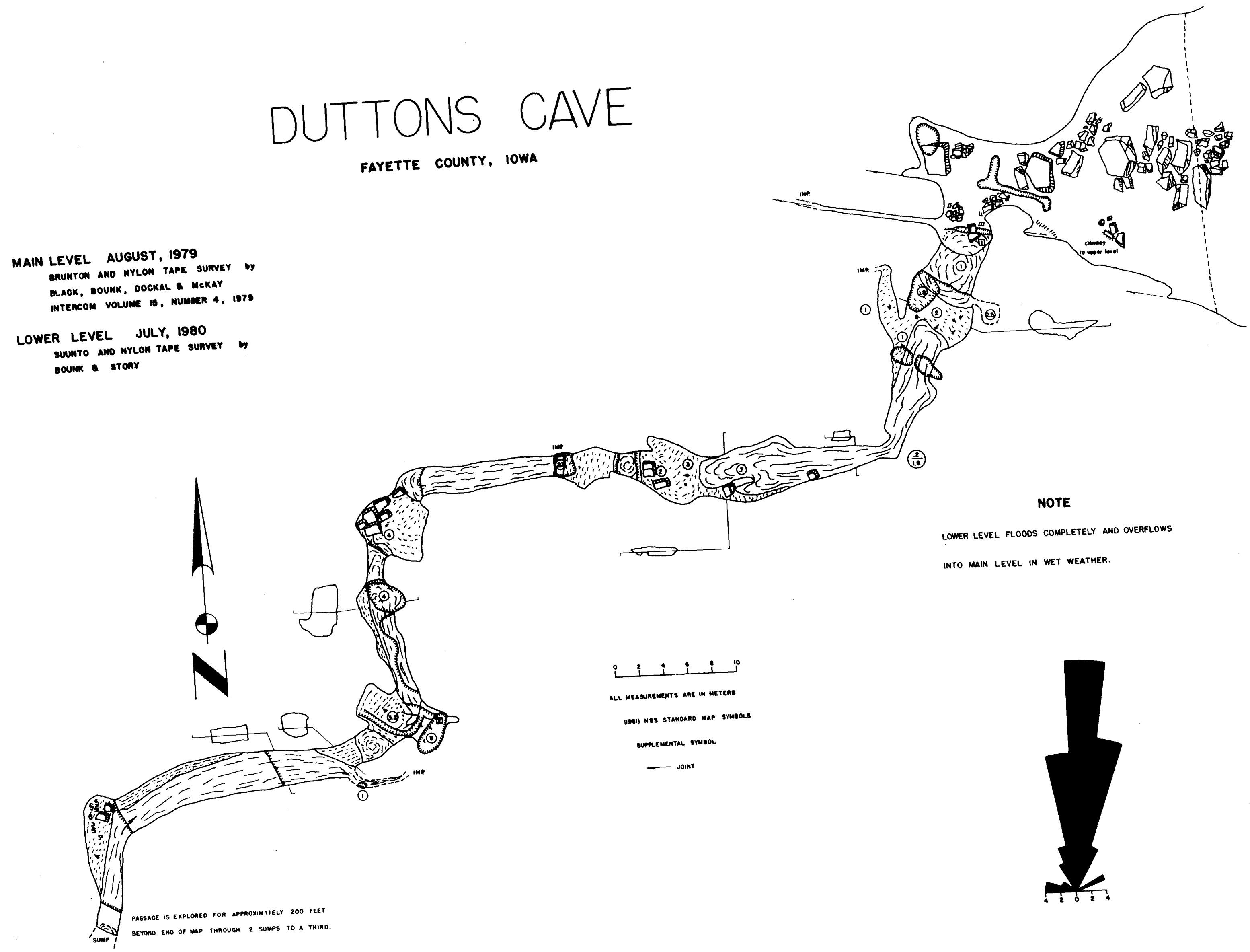

Figure 5. Map of Duttons Cave, Fayette County, with rose diagram of surface fractures. (from Bounk, 1983b).

OBSERVATIONS ON THE DIAGENESIS AND STABLE ISOTOPIC COMPOSITIONS OF SILURIAN CARBONATES IN IOWA

by

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INTRODUCTION

Silurian strata in Iowa provide a wealth of new research opportunities for carbonate specialists. It is our hope that discussions of Silurian stratigraphy, facies architecture, and carbonate petrology, presented in this guidebook will provide a foundation and impetus for renewed research activity on these rocks. Although the Silurian outcrop belt in the Upper Mississippi Valley provides a great number of well-exposed sections suitable for detailed studies, only a handful of modern workers have devoted their attention to these strata.

The close juxtaposition of coeval limestone and dolostone units in the Iowa outcrop belt is especially noteworthy. These diagenetic facies transitions could logically serve as foci for analytically-oriented petrologic investigations of the fluid-rock interactions involved in regional dolomitization. Although no systematic efforts have yet been undertaken to develop such studies, a modest set of carbon and oxygen isotopic data has already been collected on these rocks, and these data show interesting relationships that are worthy of comment.

SILURIAN CARBONATE FACIES ARCHITECTURE

Witzke (in prep.) has mapped the distribution of Silurian limestones in Iowa, and established their lateral equivalence to dolostone units in the standard Silurian sequence of Iowa. The limestone units consist of the LaPorte City Formation of central Iowa, and the Waucoma Formation of northeast Iowa (Fig. 1). Petrographic and depositional interpretations of the LaPorte City Formation are reported in Witzke (1981b), and those for the Waucoma Formation in Bowman (1985). Both limestone units are preserved near the updip erosional limits of their respective dolostone correlates (Fig. 2), immediately beneath the major cratonic erosional unconformity bracketing the Tippecanoe and Kaskaskia sequences of Sloss (1963).

Silurian limestones in Iowa (Waucoma, LaPorte City) and their dolostone correlates (Blanding, Hopkinton, Scotch Grove; Fig. 2) all record open marine deposition in subtidal environments below fairweather wave base. Shoaling or peritidal facies are not preserved in the rock record, but complex lateral facies variations, including carbonate mound development, do occur in the Scotch Grove Formation (see discussion elsewhere in this guidebook). A rapid upward change in benthic environments, from open marine skeletal wackestones/packstones to unfossiliferous laminated mudstones, is recorded at the Scotch Grove-Gower formational contact. This lithic transition records significant restriction of marine circulation, with laminated carbonates in the unit signifying development of a salinity-stratified water column. This change in benthic environments is attributed to mid-Silurian sea level fall, and the development of antiestuarine circulation in the silled East-Central Iowa Basin (Witzke, 1987).

The large scale limestone-dolostone facies archi-

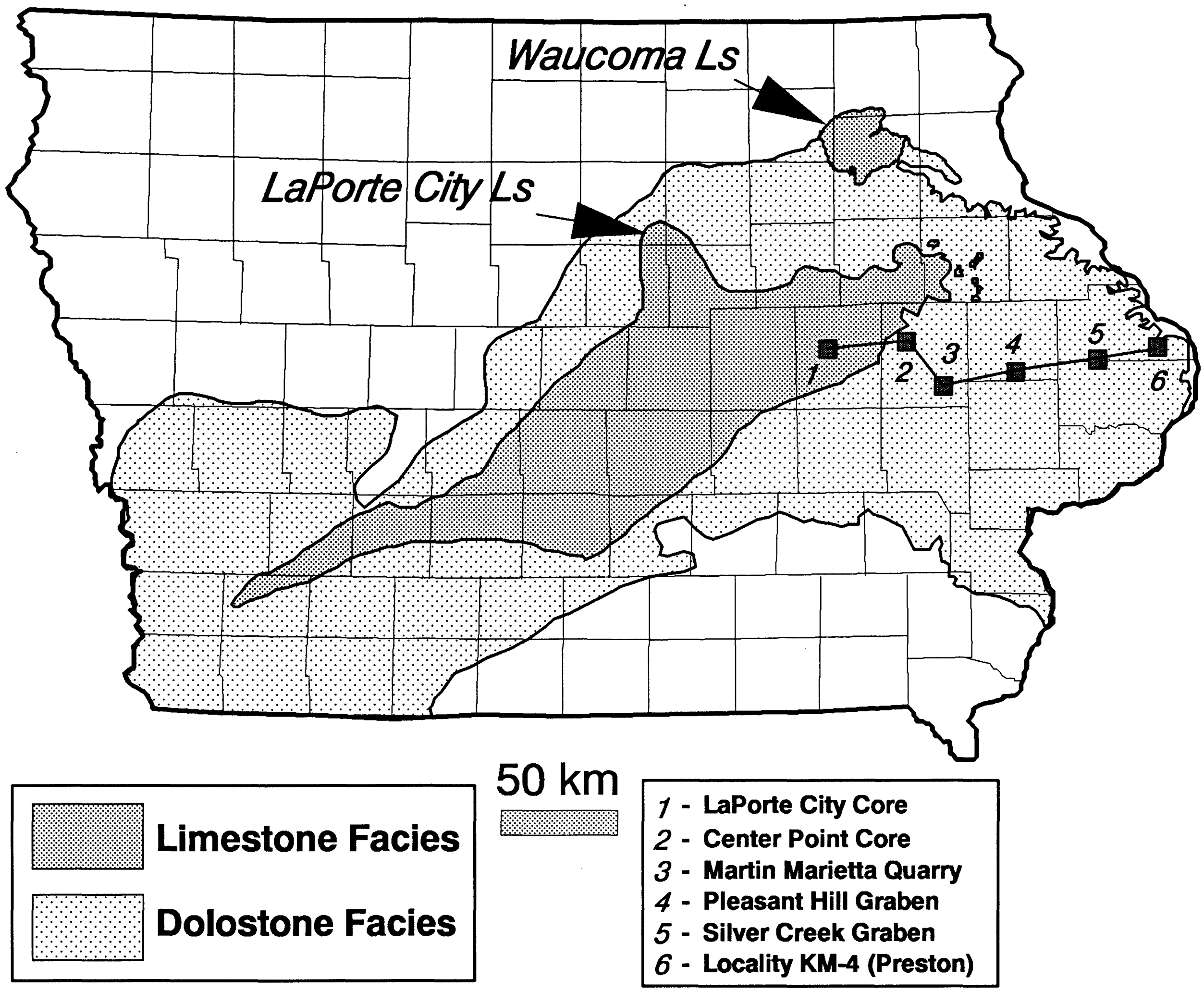

Figure 1. Distribution of the Silurian System in Iowa. The distribution of limestone facies versus completely dolomitized sections is also shown. The line of section refers to localities illustrated in Figure 6.

tecture of the Iowa Silurian sequence, as well as interpretations of the Silurian depositional sequence, led Witzke (1981b) to propose a generalized model for the diagenetic regime(s) in which the carbonate succession was regionally dolomitized (Fig. 3). The preservation of limestone facies around the periphery of the East-Central Iowa Basin was proposed to be a result of paleohydrologic responses to Silurian depositional events (Witzke, 1981b). Specifically, the limestones are believed to coincide with areas affected by the early flushing and displacement of original marine pore fluids by meteoric phreatic fluids around the subaerially-exposed basin margin. This fluid displacement occurred in response to a rapid mid-Wenlockian regression recorded by the sudden onset of hypersaline subtidal marine deposition of laminated Gower strata within the basin interior (Witzke, 1981b; Fig. 3). Accordingly, the Silurian marine limestones are viewed as localized products of meteoric phreatic diagenesis, although the presence of scattered dolomite rhombs in matrix positions within the LaPorte City and Waucoma formations (Witzke, 1981b; Bowman, 1985) indicates that dolomitizing fluids passed through these strata.

Lateral variations in the extent of dolomitization might have been kinetically controlled by slow rates of dolomite crystallization, coupled with variations in the residence times of dolomitizing chemical environments within different parts of the Silurian paleoaquifer system. The marine-meteoric phreatic mixing zone environment has been widely implicated in the re-

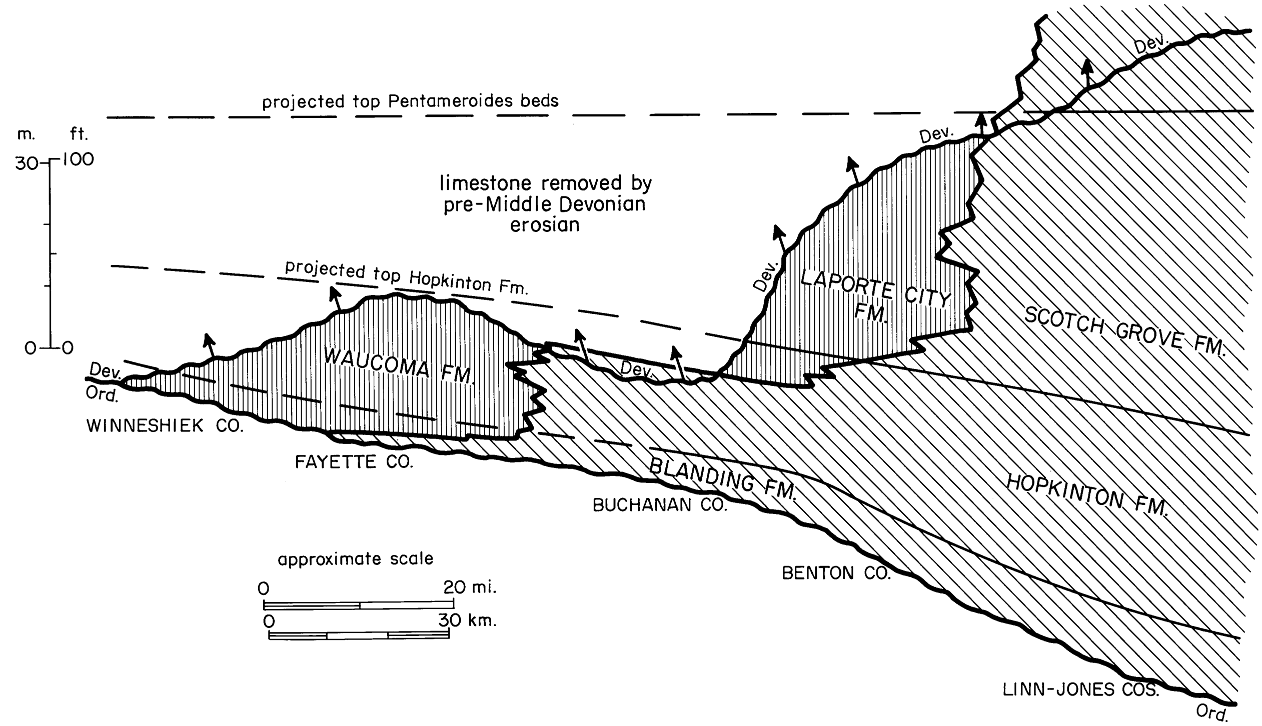

Figure 2. Stratigraphic cross section showing the positions of limestones in the Waucoma and LaPorte City formations, and their stratigraphic relationships to dolomitized Silurian sequences in Iowa.

gional dolomitization process (Ward and Halley, 1985; Humphrey, 1988; Humphrey and Quinn, 1989), and is proposed as a dominant chemical environment for the dolomitization of Silurian carbonates (Fig. 3). Limestone-dolostone boundary relationships (figs. 2 and 3) are interpreted to record the limits of an ancient meteoric phreatic lens that developed in response to rapid marine regression in the East-Central Iowa Basin. Subsequent marine offlap is proposed to have proceeded slowly, leading to a more gradual basinward advance of the mixing zone, and complete dolomitization of the remaining carbonate succession (Fig. 3).

The production of denser, hypersaline brines within the bottom layers of the stratified Gower seaway could have provided an additional source of possible dolomitizing fluids that would have migrated through the underlying rock sequence (Fig. 3). It is difficult, however, to envision this mechanism as a volumetrically significant process regarding the dolomitization of the entire sequence. Integrated petrographic and geochemical studies of these rocks could provide additional insights into the processes by which they were dolomitized.

STABLE ISOTOPIC DATA

A preliminary set of carbon and oxygen isotopic data has already been collected on Silurian carbonates in eastern Iowa. These data were largely gathered to interpret the formation of late diagenetic carbonate cements in the Paleozoic Plum River Fault Zone, including those associated with Mississippi Valley-Type sulfide mineral deposits. The petrologic contexts of some typical microsample sites from this data set are illustrated in Figure 4. Early exploratory analyses of crushed whole-rock samples (several grams each) of coeval limestone and dolostone strata of the Laporte City and Scotch Grove formations from two drillcores are also included, and provide additional perspectives. All current sample collection localities in Iowa are shown in Figure 1, and data are compiled in Ludvigson (1988). Preliminary interpretations of isotopic variations in Silurian carbonates in Iowa were discussed by Ludvigson et al. (1991).

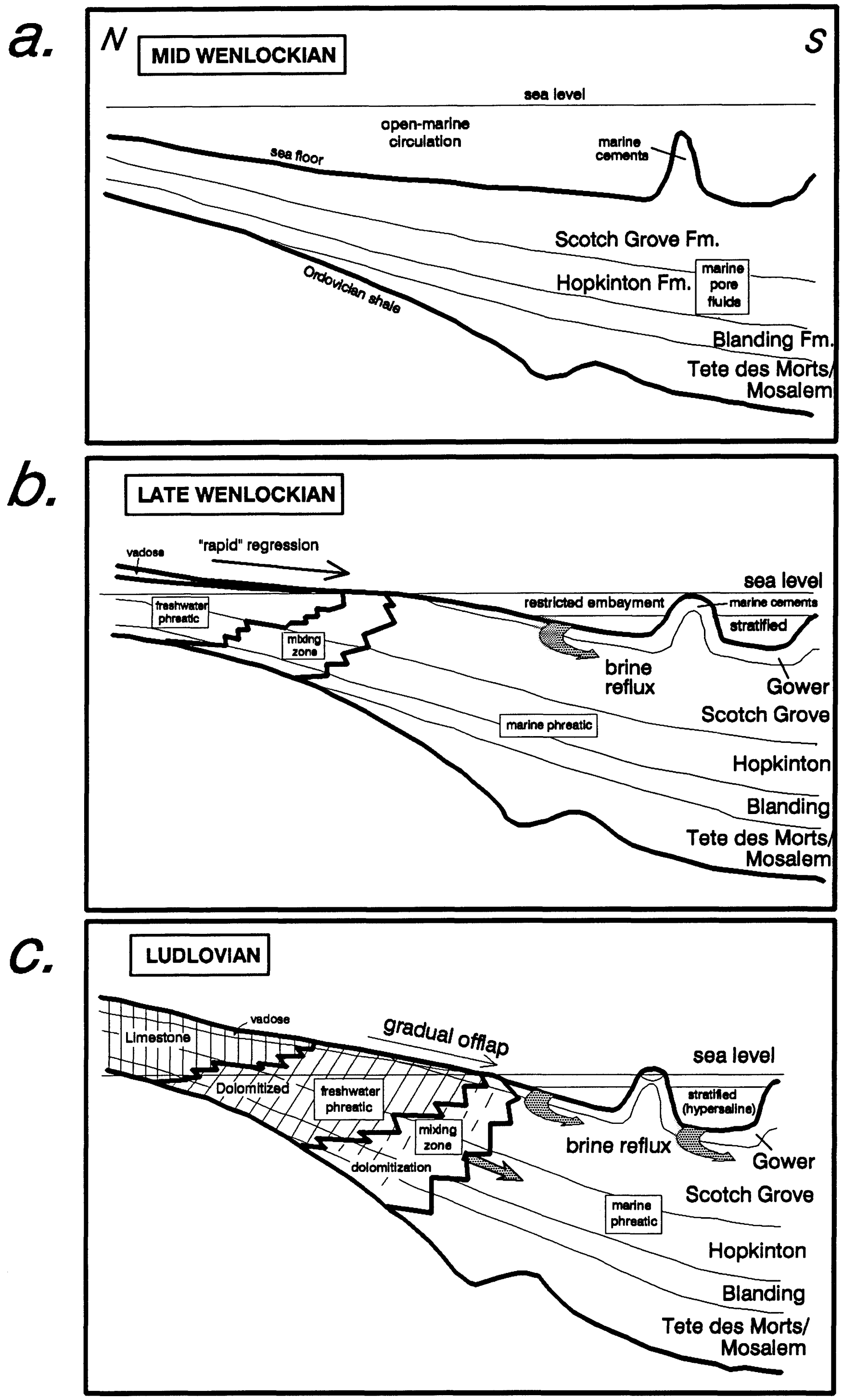

Figure 3. Schematic drawing of Wenlockian to Ludlovian depositional, hydrologic, and diagenetic events in the Silurian sequence of Iowa. Limestones at the perimeter of the East-Central Iowa Basin were rapidly flushed of marine pore fluids, and were diagenetically altered by meteoric phreatic fluids. Subsequent gradual basinward migration of the marine-meteoric mixing zone may have led to the pervasive dolomitization of the remaining sequence.

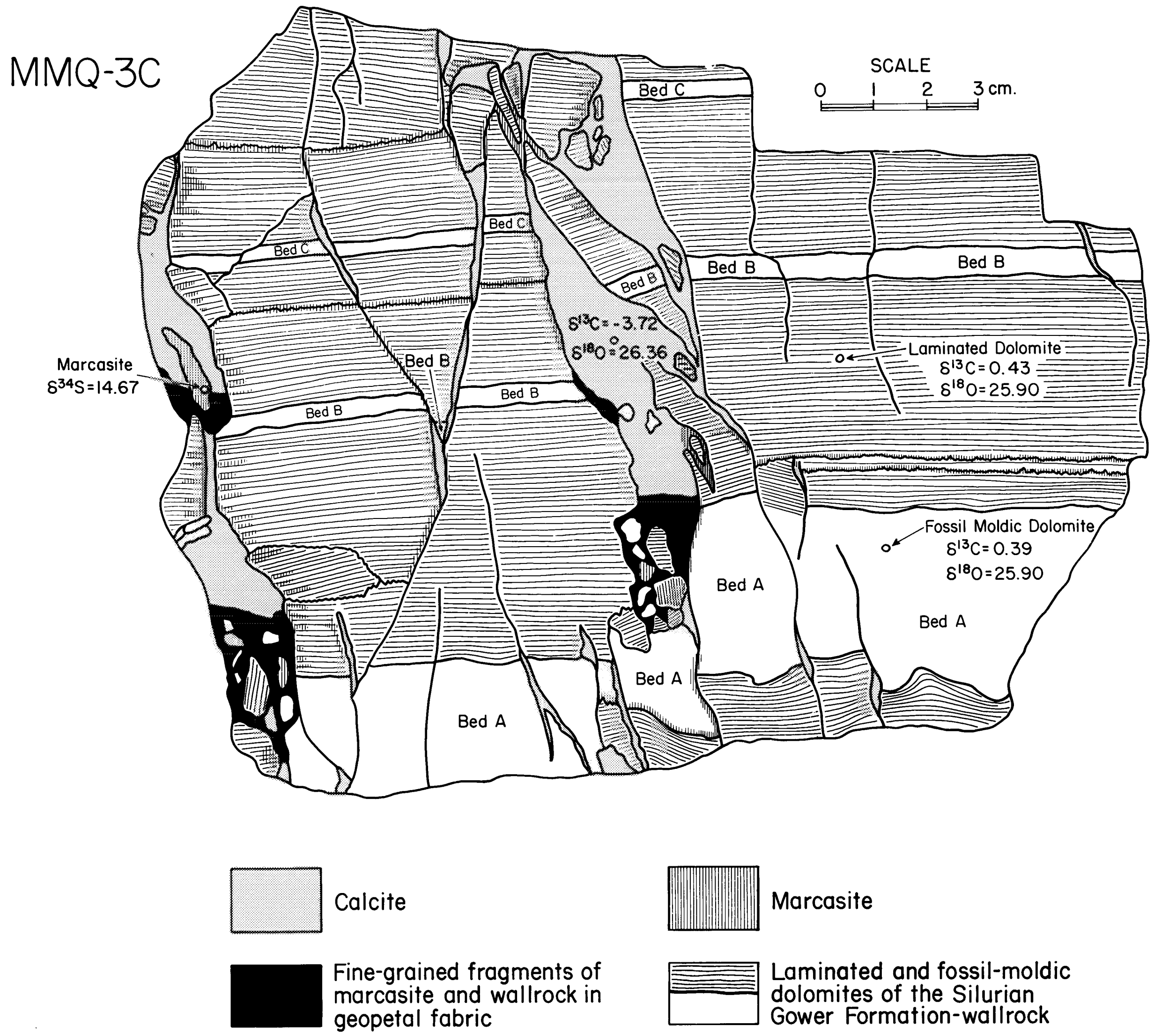

Figure 4. Drawing of polished slab surface of rock sample MMQ-3C (Gower Formation, Cedar Rapids Martin Marietta Quarry), showing microstructural detail of isotopic sampling sites. Note samples in laminated and fossil moldic dolostones. Oxygen isotopic data in this diagram were reported relative to the SMOW standard. From Garvin et al., 1987.

Methods

Powdered microsamples of fine-grained rock matrix (2 to 3 mg) were drilled with dental bits from polished rock slabs (Fig. 4) using a technique modified from Prezbindowski (1980). Analyses were performed at the laboratory of Dr. E.M. Ripley, Department of Geosciences, Indiana University, using a Finnigan Delta E stable isotope ratio mass spectrometer. Analytical precision is better than ± 0.05 ‰, and sample-to-sample reproducibility is 0.1 ‰ for both carbon and oxygen. All data are reported as per mil deviations relative to the PDB standard.

Results

Figure 5a shows the carbon and oxygen isotopic compositions measured from the LaPorte City Formation and dolomitic strata of the Gower and Scotch Grove formations. Although $\delta^{18}O$ varies between -3 to -6.5 ‰, most of the values are confined to a narrow range between -3.5 to -5.5 ‰. The $\delta^{13}C$ values cluster

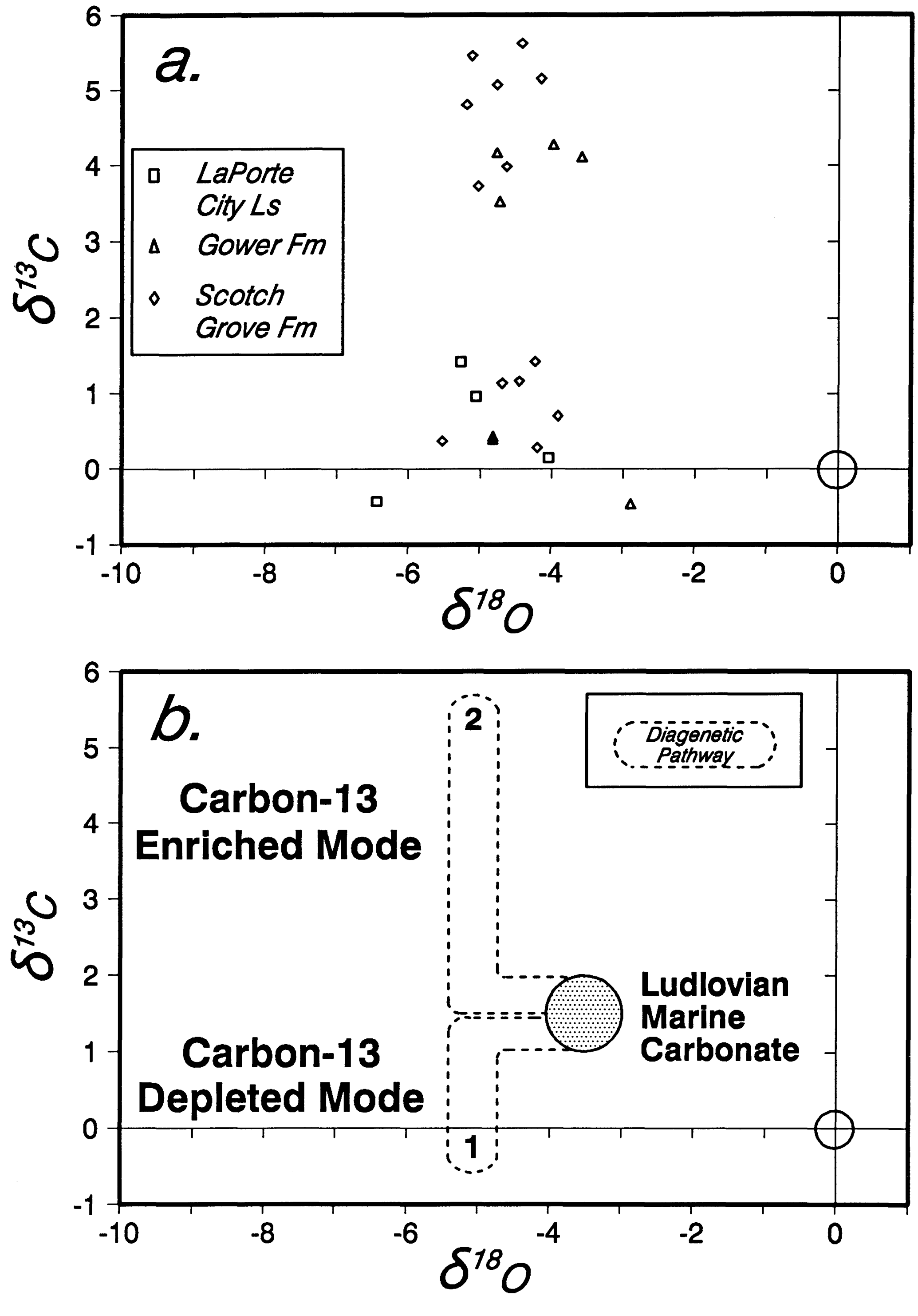

Figure 5. Carbon and oxygen isotope plot of Silurian marine carbonates in Iowa. a. Plot of samples from the LaPorte City Limestone, Gower Formation, and Scotch Grove Formation. b. Comparison of data from above with an estimate of the composition of Silurian marine carbonate (from Lohmann, 1988), and possible diagenetic pathways affecting different portions of the Iowa Silurian sequence.

in two distinct modes, one ranging between -0.5 ‰ to 1.5 ‰, and the second between 3.5 ‰ to 5.6 ‰. All of the analyses from the LaPorte City Formation plot within the depleted $\delta^{13}C$ mode, whereas the dolomitic formations plot in both. For additional perspective, the estimated composition of Ludlovian marine carbonate (Lohmann, 1988) is plotted in Figure 5b.

The most notable feature of the data is the wide range of $\delta^{13}C$ values. Geographic relationships between the sampling localities (Figure 1) permit an assessment of possible lateral isotopic variations along an east-west line of section. Figure 6 shows $\delta^{13}C$ compositions of Silurian carbonates as a function of distance along the east-west section line. Although these interpretations must be regarded as provisional because of the small size of the data set, the average $\delta^{13}C$ composition of LaPorte City-Scotch Grove carbonates apparently increases from about 0.5 ‰ in the western area to about 4 ‰ in the eastern area. The apparent increased variability in the eastern samples might result from differences in sampling methods, in that data from the two western localities were obtained from crushed whole-rock samples. That is, their apparent isotopic uniformity might result from homogenization. Nevertheless, it should be noted that the line of section traverses from the western margin of the East-Central Iowa Basin eastward into the interior of the basin.

Discussion

One of the most peculiar characteristics of the stable isotopic compositions of Silurian marine carbonates in North America is their bimodal distribution of $\delta^{13}C$ values. Brand and Veizer (1981) reported bimodal $\delta^{13}C$ values from the Upper Silurian Read Bay Formation, and data from Silurian marine carbonates in Michigan (Cercone, 1984; Cercone and Lohmann, 1985, 1986) and Iowa (this report) also show this enigmatic feature. What is the genetic significance of these two $\delta^{13}C$ modes?

Significance of Carbon-13 Depleted Modes

Carbonates in the depleted $\delta^{13}C$ mode, although diagenetically altered, have isotopic compositions that are depleted by 1 to 2 ‰ relative to the estimated original composition of Silurian marine carbonate (Lohmann, 1988; Fig. 5b). The limestones in the depleted $\delta^{13}C$ mode (Fig. 5) can be reasonably inferred to be marine rocks that were diagenetically altered in meteoric phreatic environments, following diagenetic pathway 1 in Figure 5b, although systematic sampling of limestone components is needed to more firmly establish this conclusion. The relatively small depletions in ^{13}C along this proposed meteoric phreatic diagenetic pathway, up to about 2 ‰, are smaller than those observed in some Paleozoic limestones (eg. Meyers and Lohmann, 1985), but still are compatible with meteoric phreatic diagenesis by rock-dominated pore fluids. In addition, Silurian weathering zones predated the colonization of land areas by vascular plants, and the lack of respiring root systems in early Paleozoic soils might have dictated that meteoric diagenetic systems in exposed carbonates were dominated by original rock chemistry.

Dolostones plotting in the ^{13}C-depleted mode might be interpreted to have formed in mixed marine-meteoric fluids, although as noted in following discussions, their relatively invariant $\delta^{18}O$ compositions may pose some problems. Note also that the highest $\delta^{18}O$ values, slightly enriched with respect to the estimated composition of marine carbonate, occur in dolostones of the Gower Formation.

Significance of Carbon-13 Enriched Modes

Explanations for enriched $\delta^{13}C$ modes in Silurian marine carbonates have been controversial. Various arguments have been advanced to interpret ^{13}C-enrichments in Silurian rocks as: (1) original depositional signals, and/or (2) diagenetic overprints.

Cercone (1984) observed that ^{13}C-enriched carbonate components are confined to the upper portions of Middle Silurian pinnacle reefs in the northern Michigan reef trend, and suggested a nondiagenetic origin attributed to enhanced marine surface productivity related to salinity stratification. More recently, Leibold et al. (1991) noted systematic $\delta^{13}C$ enrichments in reef crest limestones compared with coeval off-reef laminated carbonates from the Michigan Basin, further supporting the view of a stratified basin with high productivity surface waters.

Lohmann et al. (1985) suggested that ^{13}C enrichments in micritic components in some Paleozoic mounded carbonates could be attributed to early methanogenic diagenesis (anaerobic fermentation). Cercone and Lohmann (1985, 1986) noted that ^{13}C enrichments are not common to all the pinnacle reefs of the northern Michigan trend, and suggested that ^{13}C-enriched carbonates are of diagenetic origin, resulting from interactions with refluxed hypersaline fluids affected by anaerobic fermentation in overlying

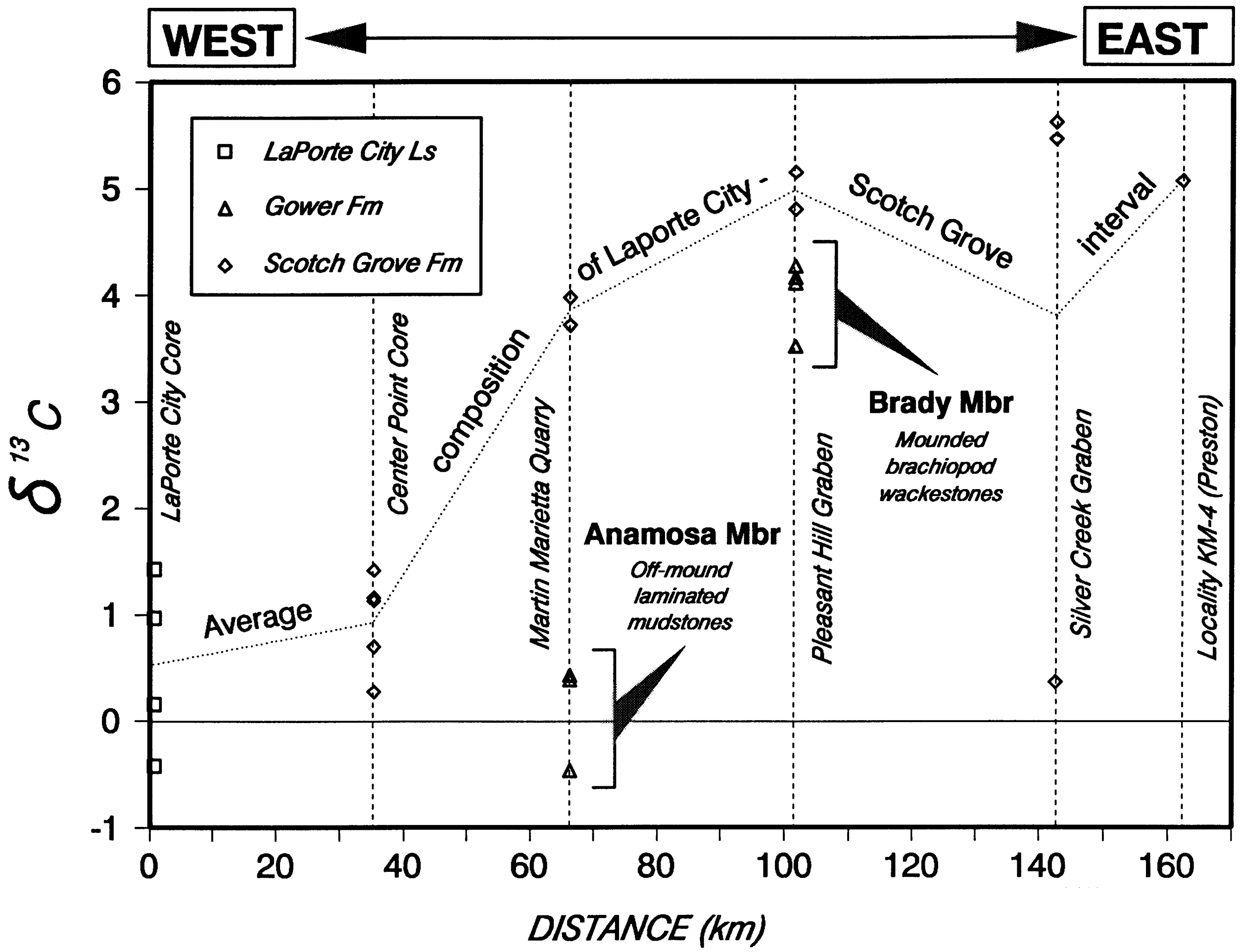

Figure 6. East-west transect showing the $\delta^{13}C$ compositions of Silurian carbonates along the line of section. Note that the average $\delta^{13}C$ in the LaPorte City-Scotch Grove interval increases from west to east. Compositions of Gower dolostones may be facies dependent (note differences between the Anamosa and Brady members), but the two units have not been sampled together at a single locality.

organic-rich strata. More recently, Lohmann and Zachos (1991) reported that $\delta^{13}C$ enrichment (up to +8 ‰) in the Silurian Pipe Creek Jr. limestone reef in eastern Indiana resides in clear blocky equant diagenetic spars of meteoric phreatic origin. The ^{13}C-enriched mode in the Iowa Silurian succession might be similarly interpreted, with diagenetic overprinting in anaerobic meteoric-phreatic environments, following diagenetic pathway 2 in Figure 5b.

Do Gower Carbonate Mounds in Iowa Provide Evidence for a High-Productivity Surface Layer?

Definitive geochemical interpretations of the Silurian sequence in Iowa must be deferred until more systematic research efforts have been completed, although several intriguing data relationships are noteworthy. Of the two sampling localities from the Gower Formation, one occurs in off-mound laminated subtidal hypersaline facies (Anamosa Member; Ludvigson, 1988) and the other in mounded brachiopod-moldic facies (Brady Member; Ludvigson, 1988). Note that data from the Anamosa Member (Martin Marietta Quarry, Fig. 6) plot in the depleted $\delta^{13}C$ mode, whereas data from the Brady Member (Pleasant Hill Graben, Fig. 6) plot in the enriched $\delta^{13}C$ mode.

Whether or not the relationship shown in Figure 6 can be replicated at a single locality traversing from mounded to off-mound facies is not presently known. Isotopic sampling illustrated in Figure 4 evidently includes components that originated from mounded and off-mound positions. Lateral relationships between the mounded Brady Member and the off-mound Anamosa Member are clearly displayed at the Cedar Rapids Martin Marietta Quarry (Witzke, 1981a), and graded beds of fossiliferous dolostone interbedding with laminated Anamosa beds are debris flows that were sourced from mound crest positions. Beds A, B, and C in Figure 4 are examples of these graded beds, and the closely similar isotopic compositions of the laminated and fossil-moldic dolostones could indicate similar original $\delta^{13}C$ values for mounded and off-mound facies, unless differences in original rock compositions were completely overprinted during dolomitization.

If the isotopic differences in Gower facies shown in Figure 6 really are indicative of original isotopic variations between mounded and off-mound facies in the salinity-stratified Gower seaway, such results would be compatible with the suggestions of Cercone (1984) and Leibold et al. (1991) that mound crests built upward into high-productivity surface waters. Differences in the $\delta^{13}C$ compositions of Brady and Anamosa dolostones apparently range between 3 to 5 ‰ (Fig. 6), whereas Cercone (1984) documented differences of about 3 ‰, and Leibold et al. (1991) described differences of about 1 to 2 ‰.

Is There Really a Carbon Isotopic Gradient in the Silurian Rocks of Eastern Iowa?

The only stratigraphic interval consistently sampled along the transect in Figure 6 is that of the LaPorte City Limestone and the laterally equivalent dolomitized Scotch Grove Formation. This interval, recording open marine subtidal deposition throughout, clearly shows a pronounced increase in $\delta^{13}C$ compositions from western to eastern sampling localities. While apparently real, the significance of this $\delta^{13}C$ gradient is open to question. We currently regard this feature as diagenetic in origin, since there is no independent evidence for seaway stratification and resultant decoupling of surface and benthic water masses during Scotch Grove deposition, as would seem to be required if the feature were of nondiagenetic origin.

Ludvigson et al. (1991) suggested that the ^{13}C enrichment in the Iowa Silurian rocks originated from diagenetic overprinting of original marine carbonate compositions. They (ibid.) further suggested that basinward groundwater flow and eastward migration of the marine-meteoric mixing zone coincident with Ludlovian marine offlap (Fig. 3) led to the development of dolomitizing chemical environments in eastern downflow areas that were increasingly influenced by anaerobic methanogenesis. Authigenic carbonates that form in methanic environments are known to be enriched in ^{13}C (Gautier and Claypool, 1984), providing a plausible mechanism to explain enrichment of $\delta^{13}C$ in dolomitized marine carbonates, following diagenetic pathway 2 in Figure 5b.

Without question, there are many interpretive issues that need to be resolved before the scenario outlined above can be accepted. Marine sediments typically pass through a zone of anaerobic sulfate reduction before the onset of significant methanogenesis (Gautier and Claypool, 1984), so that paragenetic relationships between ^{13}C-enriched carbonate components and sedimentary sulfides in the rock need to be established. Further, although the difference between calcite-water and dolomite-water $\delta^{18}O$ fractionations at sedimentary temperatures are not known with precision, there is some question as to whether dolomites in the Iowa rocks are sufficiently depleted in ^{18}O with respect to coeval marine carbonate (Fig. 5) to be reasonably interpreted as having formed from mixtures of meteoric and marine fluids. If not, dolomitization by marine-derived (refluxing brines?) pore fluids might be inferred. Morrow and Kerr (1977) showed that dolomites in the Early Silurian Allan Bay Formation have $\delta^{18}O$ compositions ranging from -4.3 to +0.6 ‰, increasing in offshore and upsection directions. These data were interpreted to record the increasing influence of infiltrating meteoric waters in the nearshore direction, and could be illustrative of $\delta^{18}O$ variations that might be expected from Silurian dolostones that originated from mixing zone settings.

CONCLUSIONS

Silurian marine carbonates in Iowa record the enigmatic widely varying $\delta^{13}C$ compositions that have previously been reported from Silurian pinnacle reefs in Michigan. Arguments can be advanced for interpretation of ^{13}C-enriched carbonates as original nondiagenetic components recording salinity stratification and high productivity surface waters, or as secondary components recording anaerobic methanic diagenesis. Regional dolomitization of the Iowa Silurian sequence has been related to the basinward migration of the marine-meteoric mixing zone during Late Silurian offlap from eastern Iowa. Stable isotopic data from the Iowa sequence may be compatible with this scenario, although the comparatively narrow range of $\delta^{18}O$ compositions could be suggestive of marine dolomitization processes, possibly involving brine reflux mechanisms.

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PART II

STOP DISCUSSIONS AND DESCRIPTIONS

KEY

	dolomite (dolostone)
	dolomitic limestone
	limestone
	shale
	sandstone
— —	argillaceous
	shaly
·—— - -	silty
· · ·	sandy
△ or	chert
▽	brecciated chert or residuum
	brecciated
	fractured to brecciated
V	vuggy
Q	quartz-lined vugs
cV	calcite-lined vugs
c	calcite spar/masses
	intraclasts
	laminated
	coarsely laminated
	stromatolitic
	stylolites
	argillaceous stylolitic streaks/swarms
b	botryoidal/isopachous cements (dolomitized)
I	internal sediment fills
g	glauconite
x	pyritic
	crossbeds
	ripples
° ° °	skeletal-moldic rock (small crinoid debris)
O O O	skeletal-moldic rock (medium-large crinoidal)
m	burrow-like mottling
	brachiopod
	brachiopod in life-position
p	pentamerid brachiopod
S	stricklandiid brachiopod
	solitary rugose coral
ø	colonial rugose coral (or indeterminate colonial coral)
T	tabulate coral (>2 cm)
t	small tabulate coral
#	bryozoan
	sponge spicule
	laminar stromatoporoid
	domal stromatoporoid
	nautiloid
	gastropod
	bivalve (pelecypod)
	trilobite
	ostracod
	graptolite
G	calcareous green alga
r	"rods"
	burrows
⁒	indeterminate skeletal debris

FIELD TRIP GUIDEBOOK TO SILURIAN EXPOSURES IN JONES AND LINN COUNTIES

Stop descriptions and discussions by Brian J. Witzke

The trip will depart from Iowa City and proceed to Stop 1 via Highways 1 and 151 (see map on back cover). The morning will be spent at Stops 1 and 2, and a lunch break is planned as part of Stop 3. Stops 4 and 5 are scheduled for the afternoon. We will be visiting natural exposures, old quarries, and operating quarries, all of which may be potentially hazardous. ***Hard hats are required in the operating quarries.*** Please be courteous and cautious at all times. Trip participants are advised that some strenuous hiking and slope climbing will be included on Stops 1 and 5. Please note that cited references in the stop descriptions are given in the list of references following the article by Witzke in this guidebook.

STOP 1. PICTURE ROCK COUNTY PARK, JONES COUNTY

Bluff exposures at Indian Bluff Cave; supplemental stops in the uplands west of the park

The deeply incised Maquoketa River Valley at Picture Rock County Park exposes an instructive sequence of Silurian dolomite strata of the Hopkinton and Scotch Grove formations. The steep cliff faces within the park are formed primarily within the Picture Rock Member of the Hopkinton, and the park serves as the type locality of the member (Johnson, 1983). Strata of the Farmers Creek Member are exposed beneath the imposing cliff faces of the Picture Rock Member, and are host to local cave developments in eastern Iowa. The area south of the entrance to Indian Bluff Cave (see paper by Bounk in this guidebook) provides an accessible route through the middle and upper Hopkinton interval (near center of Block A on Fig. 1). Depending on slope conditions, we will ascend the rock sequence beginning below the prominent rock shelter and proceed to the mouth of the shelter through a crevice to the top of the bluff. Alternatively, the sequence can be accessed along the south edge of the main buttress immediately south of the rock shelter. Strata of the lower Scotch Grove Formation are exposed above the top of this cliff section as one approaches the crest of the bluffs. The interval of Silurian rocks accessed in this area is graphically illustrated in Figure 2 (main section). A written description of the various rock units is included.

The Farmers Creek Member is a massive to thick bedded interval of porous vuggy dolomite. The member exhibits fossil-moldic porosity throughout, and molds of small crinoid debris and brachiopods are especially common. Fossils may be difficult to distinguish on the steep exposure faces, although the roof to Indian Bluff Cave nicely displays a variety of typical Farmers Creek fossils. The green alga *Cyclocrinites*, which resembles a golf ball, is characteristic of the member. Lower strata yield *Stricklandia laevis*, and globular pentamerids (*Harpidium maquoketa*) become noteworthy upward through the remainder of the member. The overlying Picture Rock Member contains more coarsely crystalline fabrics and a conspicuous fauna of tabulate corals and lamellar stromatoporoids, silicified in part. *Pentamerus oblongus* occurs near the base of the member, and nice specimens in life position can be seen near the base of the cliffs about 350 feet (100 m) to the south. Although a prominent cliff former at this locality, the Picture Rock Member displays an irregular wavy-bedded aspect on the weathered faces. The top beds the member are mostly covered in the wooded slopes above the top of the cliffs.

Lower Scotch Grove strata are exposed along the slopes above the cliff face, and an interesting array of facies variations are seen within the park. An atypical porous and fossiliferous phase of the Johns Creek Quarry Member is found above the main section (Fig. 2), displaying scattered *Pentameroides* and *Petalocrinus* in the lower part and prominent large corals in the upper part. Colonies of *Halysites* and large rugosans ("*Diphyphyllum*") reach sizes to 90 cm in diameter, and

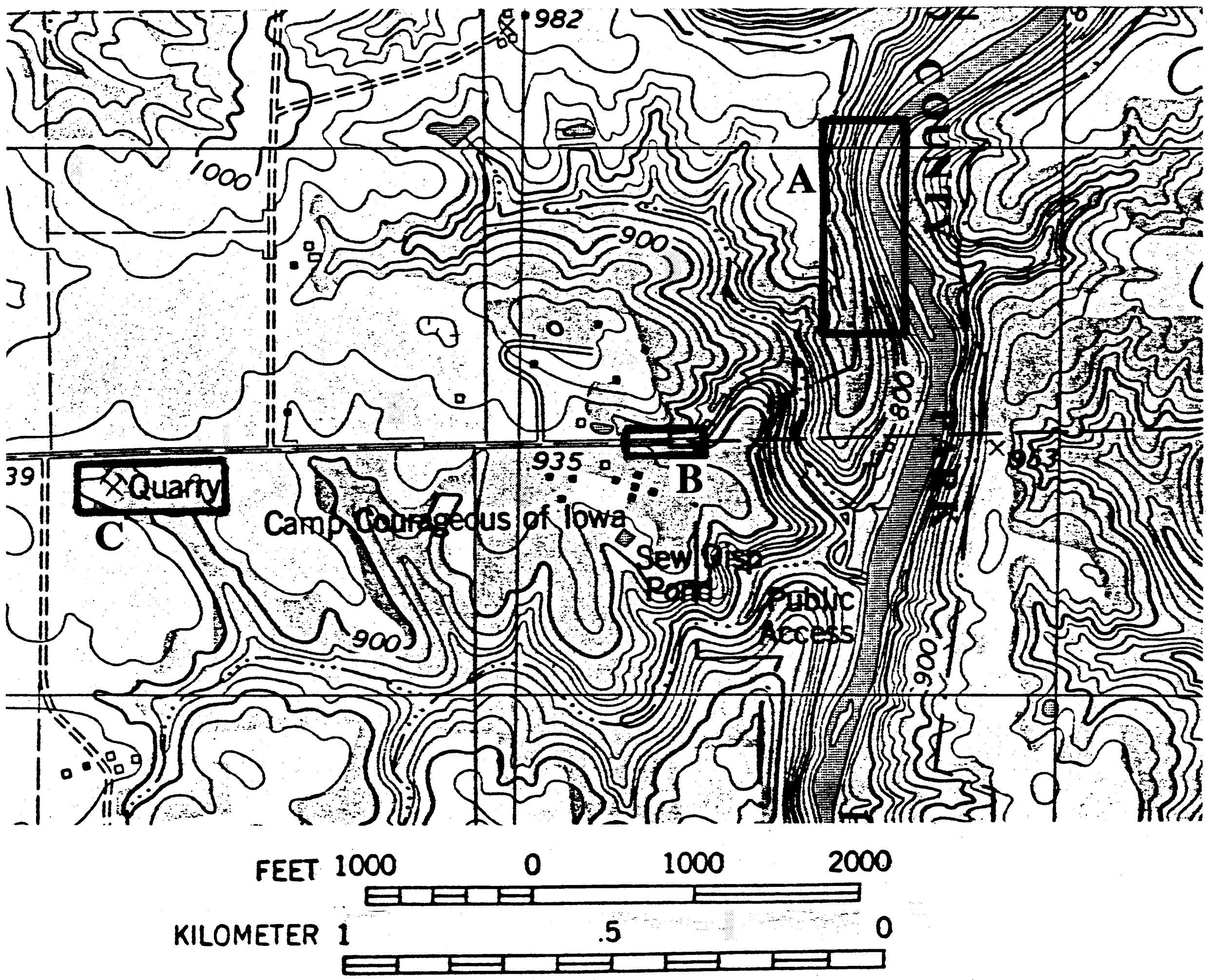

Figure 1. Location map for area around Picture Rock County Park, Jones County (from USGS Scotch Grove quadrangle).

some specimens are clearly overturned. Although the Johns Creek Quarry Member does not appear to be mounded at this locality, the large coral colonies resemble forms seen in the central portions of carbonate mounds at the same stratigraphic position elsewhere in the county (e.g., Stop 2). Progressing northward along the ridge crest in the county park, basal Scotch Grove strata become less skeletal-moldic and denser, and corals become less common and of smaller size. The Johns Creek Quarry Member is also exposed at the type Scotch Grove quarries 2 miles (3.2 km) to the south, where more characteristic inter-mound lithologies of the member are seen, primarily sparse skeletal mudstone fabrics.

If time and conditions permit, we will proceed northward about 700 feet (200 m) along the ridge crest following scattered lower Scotch Grove outcrops until a bluff face exposing Welton Member strata is reached (Fig. 2, supplemental section). This exposure displays moldic wackestone fabrics with scattered to commmon small crinoid debris, typical of the Welton. However, the contact between Welton and Johns Creek Quarry strata appears to be laterally and vertically gradational within the park (the contact is sharp at most localities in eastern Iowa). Characteristic lower Welton fossils are seen at this locality, including brachiopods (*Costistricklandia castellana*, *Atrypa*),

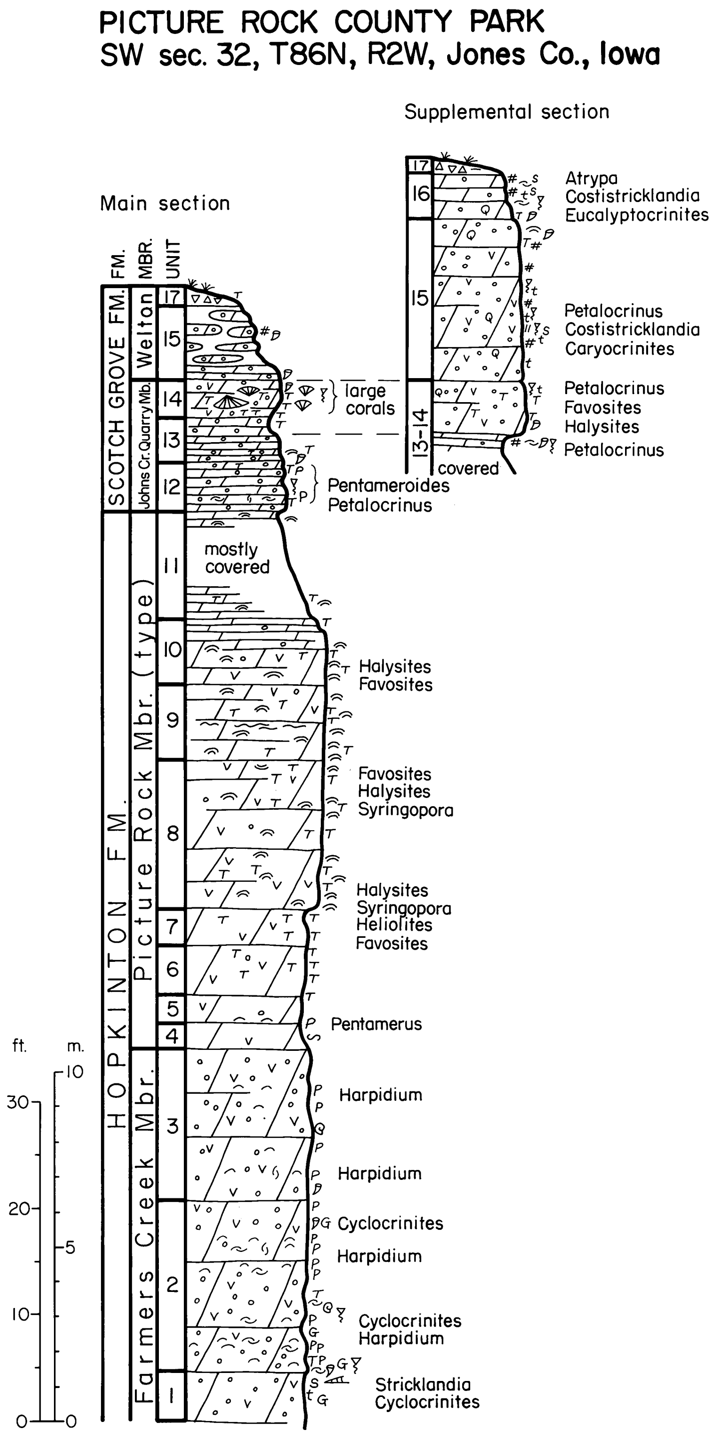

Figure 2. Graphic stratigraphic section, exposures in Picture Rock County Park (Stop 1A). Unit descriptions are provided in text. See page 86 for lithologic symbols.

echinoderms (*Eucalyptocrinites*, *Petalocrinus*, *Caryocrinites*), small tabulate corals, and bryozoans (fenestellids and branching forms). The Welton sequence generally becomes less skeletal moldic and more bryozoan-rich upward within the bluff face exposure. These bryozoan-rich strata have also yielded *Dicoelosia* and trilobites in ditch exposures to the west (near Block B, Fig. 1), where the interval grades upward into cherty Buck Creek Quarry facies.

Fossiliferous chert residuum, or geest, is scattered along the ridge crest and as transported pieces down the steep slopes. Chert geests are best developed on upland surfaces in and around the vicinity of Camp Courageous (Fig. 1, between blocks B and C), where a variety and abundance of silicified fossils derived from the Buck Creek Quarry Member have been collected over the years (Fig. 3). Tabulate and rugose corals and *Pentameroides subrectus* are particularly common in the geests. In addition, these upland geests serve as the type locality for the corrugated pentamerid, *Pentameroides (Callipentamerus) corrugatus*, and the unusual inadunate crinoid, *Petalocrinus mirabilis*. Although very rare, most Iowa specimens of the square coral, *Goniophyllum pyramidale*, have come from this locality.

A brief stop along the roadside near the upland entrance to the county park is planned (Fig. 1, Block B). Ditch-side exposures of cherty Buck Creek Quarry strata in this area display a 25 to 35 cm-thick (10-14 in) bed containing a laterally continuous colony of a silicified fasciculate rugosan ("*Diphyphyllum*") that can be traced for at least 50 feet (15 m). These gigantic sheet-like masses of coral (Fig. 3) were first noted by Thomas (1917) in this area. A bed with a laterally traceable mass (>50 ft) of the same taxon of silicified coral appears again in ravine exposures between Blocks B and C (Fig. 1) at the same stratigraphic position. If the "*Diphyphyllum*" bed is continuous between these two areas, the sheet-like mass of coral possibly may extend laterally for 1500 to 2000 feet (450-600 m).

Depending on our time constraints, it may be possible to visit the upland quarry west of the park (Fig. 1, Block C). This is the Hanken Quarry described by Johnson (1977b), which exposes characteristic lithologies and faunas of the Buck Creek Quarry Member in the mid Scotch Grove Formation. The quarry displays about 35 feet (10 m) of medium to thick-bedded cherty dolomite, crowded with molds of *Pentameroides subrectus* in the upper part. *Pentameroides* is scattered in the lower and mid parts of the quarry, with small to large silicified specimens of tabulate corals (*Favosites*, *Halysites*) observed.

The trip will proceed westward to Highway 38, and then northwestward into Monticello. From downtown Monticello, we will travel 3.5 miles west on County Road D62 (W. 1st St.), and then a short distance north on the gravel road (just east of the bridge over Wet Creek) to the east gate of the Manternach Quarry (Stop 2).

PICTURE ROCK COUNTY PARK

Main section in buttress area adjacent to large rock shelter south of Indian Bluff Cave entrance, NW NE SE SW sec. 32, T86N, R2W, Jones Co., Iowa.
Supplemental section of Scotch Grove Fm. in upper bluff 200 m (650 ft) to north, SE SW NE SW sec. 32.
Measured by B.J. Witzke and G.A. Ludvigson, 6/12/91.
Note: Entire section is dolomite (dol.). Dolomite crystal sizes as follows: extremely fine crystalline (xf xlln), very fine (vf), fine (f), medium (m), coarse (c). All rocks are light gray to buff in color (locally oxidized and weathered to pale brown gray) unless noted. Thicknesses are in metric.

POST-SILURIAN (Transported clasts of Scotch Grove Fm., Buck Creek Quarry Mbr.)

Unit 17.

Cherty colluvium and geest. Uplands above Picture Rock bluffs locally display chert geests derived from strata of the Buck Creek Quarry Member. Chert clasts from these geests and

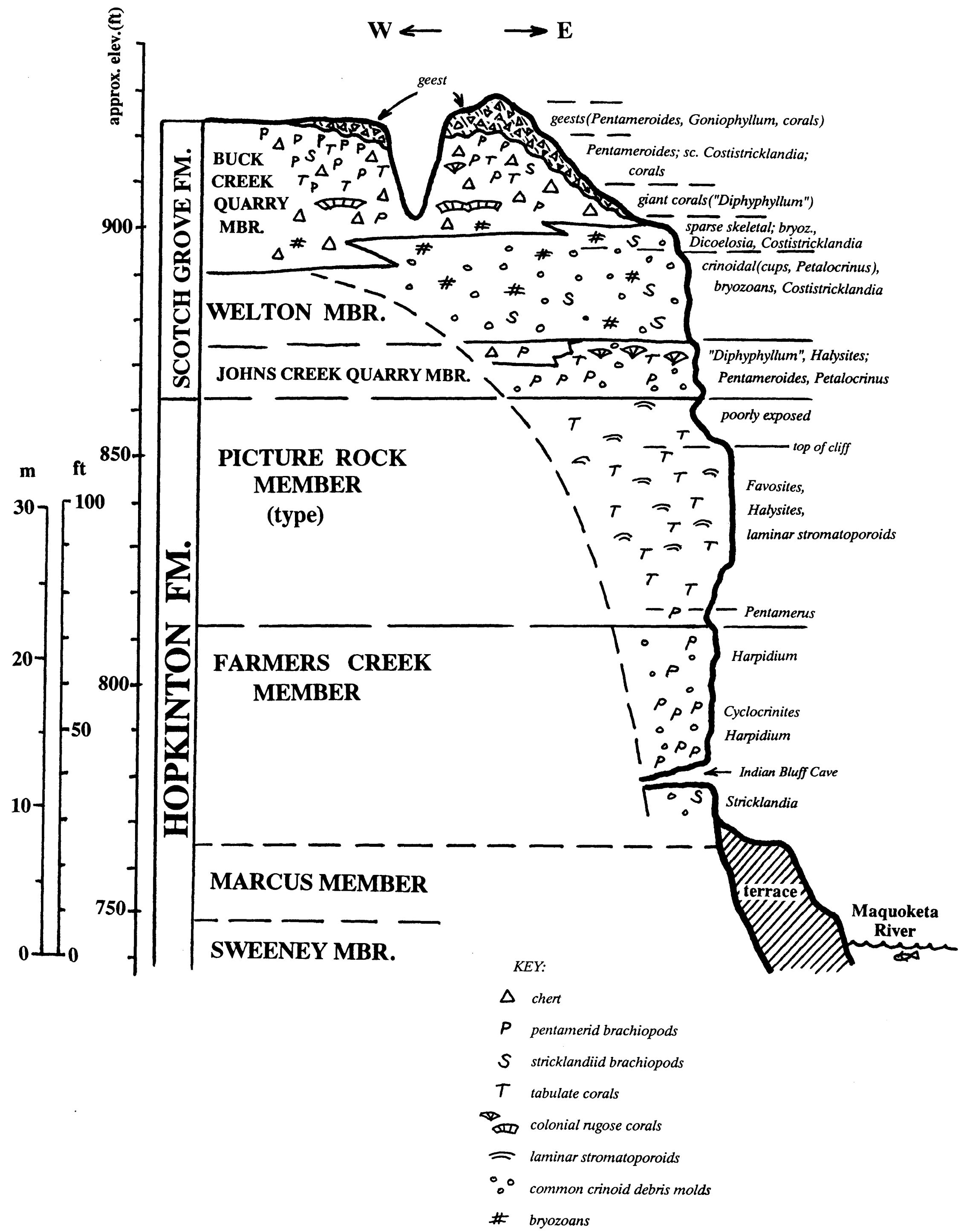

Figure 3. General stratigraphic section in the Picture Rock area, extending from county park (Block A, Fig. 1) on the east to the Hanken Quarry (Block C, Fig. 1) on the west.

subjacent cherty dolomite strata were transported downslope in the colluvium and soil to overlie units 11-16 along the bluff slope. Silicified fossils noted on slopes in the park include brachiopods (*Pentameroides*, *Atrypa*), crinoid debris (including *Petalocrinus*), bryozoans, tabulate corals (*Favosites*, *Halysites*, alveolitids), rugose corals (*Arachnophyllum*, "*Diphyphyllum*", cup and horn corals).

SILURIAN
SCOTCH GROVE FORMATION
WELTON MEMBER

Unit 16.
Dol., dense, xf-vf xlln, sparse fine skeletal-moldic wackestone, less skeletal than below, in three beds; lower bed (55 cm) with quartz void linings, scattered corals (*Alveolites* to 15 cm, silicified *Halysites*, cup corals), brachiopods (*Atrypa*, *Costistricklandia*), crinoid debris (including *Eucalyptocrinites* cups); middle and upper beds (each 40 cm) with sparse crinoid debris (including *Petalocrinus* arm-fans), corals (small favositids and cup corals), bryozoans (including large fenestellid fronds and branching forms), brachiopods (*Resserella*, *Eoplectodonta*, *Costistricklandia*, others); 1.35 m.

Unit 15.
Dol., more porous than above, common small crinoid debris molds through, scattered vugs (some lined with megaquartz), thick bedded to massive; scattered small tabulate corals (<5 cm, *Halysites*, favositids); middle part of unit with bryozoans, *Costistricklandia*, crinoids (*Petalocrinus* arm-fans, indet. cups), cystoids (*Caryocrinites*); top 1 m with scattered tabulate corals 5-15 cm diameter (*Favosites*, *Alveolites*, *Halysites*), cup corals; top 30 cm with scattered laminar to domal stromatoporoids; 3.4 m.

JOHNS CREEK QUARRY MEMBER

Unit 14.
Dol., ledge former, more resistant than below, xf-vf xlln, fine skeletal-moldic wackestone, scattered large vugs; very coralline unit in main section, includes large *Halysites*(up to 80 x 20 cm) at base; large colonial rugosans ("*Diphyphyllum*") in upper 90 cm, overturned or rotated in part, two largest specimens measure 50 x 15 cm and 90 x 40 cm, calcite spar filled in part; additional corals (5-25 cm) include *Favosites*, *Alveolites*, cup corals; upper 40 cm with scattered *Petalocrinus* arm-fans; NOTE: this unit becomes progressively less coralline northward along bluff crest, with maximum coral sizes diminishing (5-20 cm diameter), tabulate corals (*Favosites*, *Halysites*, *Alveolites*), fine crinoid debris molds (including *Petalocrinus* in upper part), colonial rugosans absent; northward sections are lithologically inseparable from the Welton Member, suggesting that coral-rich strata of the upper Johns Creek Quarry Member may be a lateral facies of the basal Welton; 1.1 m.

Unit 13.
Dol., fine skeletal-moldic wackestone, crinoidal, porous, scattered vugs, recessive in part, thin to medium bedded (2-15 cm); scattered tabulate corals 5-20 cm diameter (*Halysites*, *Favosites*), small cup corals; interval becomes less skeletal-moldic and denser northward, fossils include small cup corals, *Petalocrinus*, brachiopods (*Resserella*, spiriferid), branching bryozoans; 1.25 m.

Unit 12.
Dol., xf-vf xlln, fine skeletal-moldic wackestone, crinoidal, thin to medium bedded (5-15 cm); scattered corals, especially in upper part (*Halysites*, *Favosites*, *Alveolites*), large *Syringopora* (45

x 7 cm) near top; *Pentameroides* common in two bands (40 cm and 120 cm above base), disarticulated shells (2-5 cm), some in life position; scattered small cup corals, *Petalocrinus* arm-fans; 1.45 m.

HOPKINTON FORMATION
PICTURE ROCK MEMBER (TYPE SECTION)

Unit 11.
Dol., xf-vf and vf-f xlln; interval is mostly covered; silicified laminar stromatoporoids (35 cm diameter) noted near top; *Favosites*, fine skeletal molds; 3.0 m.

Unit 10
Dol., xf-vf xlln, top of main cliff section; irregularly bedded, bedding more distinct upward; more skeletal moldic (fine crinoid debris) than below, scattered vugs; common lamellar features and laminar stromatoporoids (silicified in part; 10-50 cm diameter); scattered tabulate corals 5-20 cm diameter (*Favosites*, *Halysites*), silicified *Heliolites* at top; mostly covered above; 1.9 m.

Unit 9.
Dol., vf-f and f-m xlln, irregular wavy-bedded aspect but prominently exposed as massive cliff former; scattered fine skeletal molds, becomes more finely skeletal moldic (crinoidal) upward; scattered vugs, especially in upper part; lamellar features scattered through, some (most?) are laminar stromatoporoids (10-50 cm), prominent lamellar horizon (10 cm thick) 1.0 m above base; scattered tabulate corals, small (2-7 cm) *Favosites* in lower 60 cm, more prominently coralline above with *Favosites* and *Halysites* 5-30 cm diameter; some corals are overturned; 2.15 m.

Unit 8.
Dol., vf-f and f-m xlln, prominent cliff-forming interval, base of interval forms overhang, massive but with wavy-bedded aspect in part; large vugs scattered through, especially near top; thin lamellar features and laminar stromatoporoids scattered to common (10-40 cm diameter), part silicified in upper half, overturned stromatoporoids in upper 65 cm; scattered tabulate corals, *Halysites* dominant (5-15 cm diameter), *Favosites* and *Syringopora* also present; 4.25 m.

Unit 7.
Dol., vf-f and f-m xlln, some fine skeletal molds as below; lithic change, denser and less vuggy than below, small vugs (1-3 cm); thin dense xf-vf dol top 5 cm; scattered tabulate corals, larger corals in top 45 cm, *Syringopora*, *Halysites* and *Favosites* (to 25 cm, part silicified), *Heliolites*; 1.05 m.

Unit 6.
Dol., vf-f xlln, massive, cliff former, scattered vugs; lower 40 cm with scattered small *Favosites* (2-8 cm); middle part with *Favosites*, *Halysites* (to 20 cm); upper part contains *Heliolites* (3-15 cm), *Halysites* (5-25 cm), *Favosites*(5-10 cm, two species, corallite diameters <1 mm and 2.5 mm); 1.4 m.

Unit 5.
Dol., vf-f and f-m xlln; scattered small vugs, skeletal moldic porosity, crinoid debris; scattered small *Pentamerus*(1-3 cm) at base; top 25 cm with scattered small tabulate corals (3-8 cm), *Favosites*, *Halysites*; 85 cm.

Unit 4.
Dol., vf-f xlln, f-m xlln near top, more coarsely crystalline and less porous than unit below, slightly recessive; dense, no skeletal fossils evident, burrow fabrics in part; scattered small vugs;

gradational above; 75 cm.

FARMERS CREEK MEMBER

Unit 3.

Dol., xf-vf xlln, fine skeletal moldic, crinoidal wackestone fabrics common, porous, scattered vugs, thick bedded to massive, vague bedding surfaces laterally discontinuous; scattered to common *Harpidium* in thin bands, life-position at 90 cm above base; scattered gastropods, horn coral in lower part; 4.3 m.

Unit 2.

Dol., xf-vf xlln, massive to thick bedded, fine to coarse skeletal moldic, more finely skeletal in middle part, crinoidal, slightly more skeletal moldic than below, porous, scattered small to large vugs; *Harpidium*-rich layers (to 5 cm thick), primarily disarticulated valves displayed in moldic packstone fabrics, graded in part, noted at 60-85 cm, 1.45 m, 2.8-3.1 m, 3.5 m (some in life-position), 3.7-3.9 m, and 4.85 m above base; *Coolinia*, isolated *Harpidium* at base; *Cyclorinites*, gastropods, small cup corals, small tabulate corals (favositids, halysitids), nautiloids, bryozoans scattered through unit; camerate crinoid cups (*Dimerocrinites*, *Theleproktocrinus*) noted in lower to middle part; fossiliferous *Harpidium*-rich beds with *Cyclocrinites* well displayed on rock overhang at entrance to Indian Bluff Cave; 4.85 m.

Unit 1.

Dol., xf-vf xlln, fine skeletal moldic wackestone fabrics, crinoidal, scattered vugs, massive with indistinct bedding break at top, local overhang above; scattered brachiopods, *Cyclocrinites*, small nautiloids, small favositids; *Stricklandia* noted 30 below top; base of cliff about 7 m above Maquoketa River; 1.45 m.

STOP 2. MONTICELLO-MANTERNACH QUARRY

Norbert Manternach owner (operated by Vulcan Materials Corp.)
SW NE sec. 24, T86N, R4W, Jones Co.
Carbonate mound facies of the Johns Creek Quarry Member

Hard hats are required at this stop. Please stay clear of any unstable quarry faces. A sketch of the quarry plan and photo mosaic tracings of the quarry walls (Fig. 4) provide an overview of complex carbonate mound facies in the lower Scotch Grove Formation, Johns Creek Quarry Member. Bedding is traced directly from photos, although fracturing makes bedding indistinct at several places within the quarry. Talus and quarry piles obscure the base of most quarry faces, as shown on Figure 4. Access to quarry segments A-B-C-D-E and G-H-I will provide a general overview of facies architecture for the first phase of carbonate mounding in the Iowa Silurian sequence. Mounding in the basal Scotch Grove Formation is locally developed as isolated single mounds, or, as seen at this locality, complex coalesced developments of multiple mounds. Two generalized mound and mound-related facies are recognized in this quarry: 1) central mound facies, primarily characterized by skeletal-moldic mudstone and wackestone fabrics in horizontal to gently dipping beds (0-10°), and 2) flanking facies, characterized by crinoidal wackestone and packstone (some grainstone) fabrics in more steeply dipping beds (5-35°).

Philcox (1970a) first described carbonate mounds ("bioherms") from this quarry, and he was also the first to recognize carbonate mounds at this general stratigraphic position. The quarry exposures at the time of Philcox's study were considerably more limited, and subsequent quarrying has significantly enlarged the exposures. Nevertheless, Philcox (1970a, p. 970) recognized a "group of small bioherms" consisting of a "central core of fine, unbedded dolomite containing abundant coral colonies and well-bedded, coarse, crinoidal dolomite forming outward-dipping flanks and draping cover deposits." The central mound facies at the time of Philcox's investigation included isolated coral colonies reaching sizes to 6.5 feet (2 m), and I also observed corals of this size in the mid 1970s. However, I have been unable to relocate these huge individual coral colonies, and the largest corals presently observed range from 20 to 40 cm (8-16 in). Most are smaller, and nowhere have the corals been observed to form an organic framework. Corals include tabulates (*Halysites*, *Syringopora*, *Favosites*, *Heliolites*, alveolitids) and rugosans ("*Diphyphyllum*", *Arachnophyllum*). As noted by Philcox (ibid.), the central mounds are dominated by carbonate mud fabrics, with scattered small to large corals floating in the muddy matrix. Corals are not ubiquitous to the central mounds, and are best displayed in the area immediately north of point E (Fig. 4). Large corals are also seen in central mound facies along quarry segment I-J-K. Additional fossils in the central mounds include scattered stromatoporoids, brachiopods (*Atrypa*, *Plectatrypa*, *Ferganella*, *Dalejina*, *Protomegastrophia*), nautiloids, gastropods, solitary rugosans, and trilobites (*Kosovopeltis*).

Flanking facies are are easily examined along quarry segment C-D and near point H (Fig. 4). Dolomitized coarse crinoidal packstones characterize much of the flanking facies, commonly in thin graded packstone-wackestone couplets. Graded bedding and wedge-bed geometries suggest deposition of flanking strata as episodic downslope debris flows. Crinoid debris forms the dominant skeletal component of the flanking facies, and crinoid cups are locally preserved (especially *Manticrinus*). Some graded beds contain scattered to common fenestellid bryozoans, and additional fossils are locally noted (small tabulate corals, solitary rugosans, gastropods, nautiloids, and rhynchonellids). Inter-grain porosity in the flanking units is commonly lined with lumpy botryoidal and isopachous coatings displaying fibrous or bladed microstructure; these are best interpreted as relict submarine cements. Identical botryoidal coatings are also seen in the central mound facies at places within the quarry (see Fig. 4).

The stratigraphic position of the carbonate mound complex seen in the Manternach Quarry is constrained by lateral and vertical facies relations displayed along quarry segment H-I and by capping residuum north of point E (Fig. 4). North of point H, flanking strata grade from packstone to wackestone fabrics, and dips flatten out to horizontal. Horizontally-bedded units, which laterally

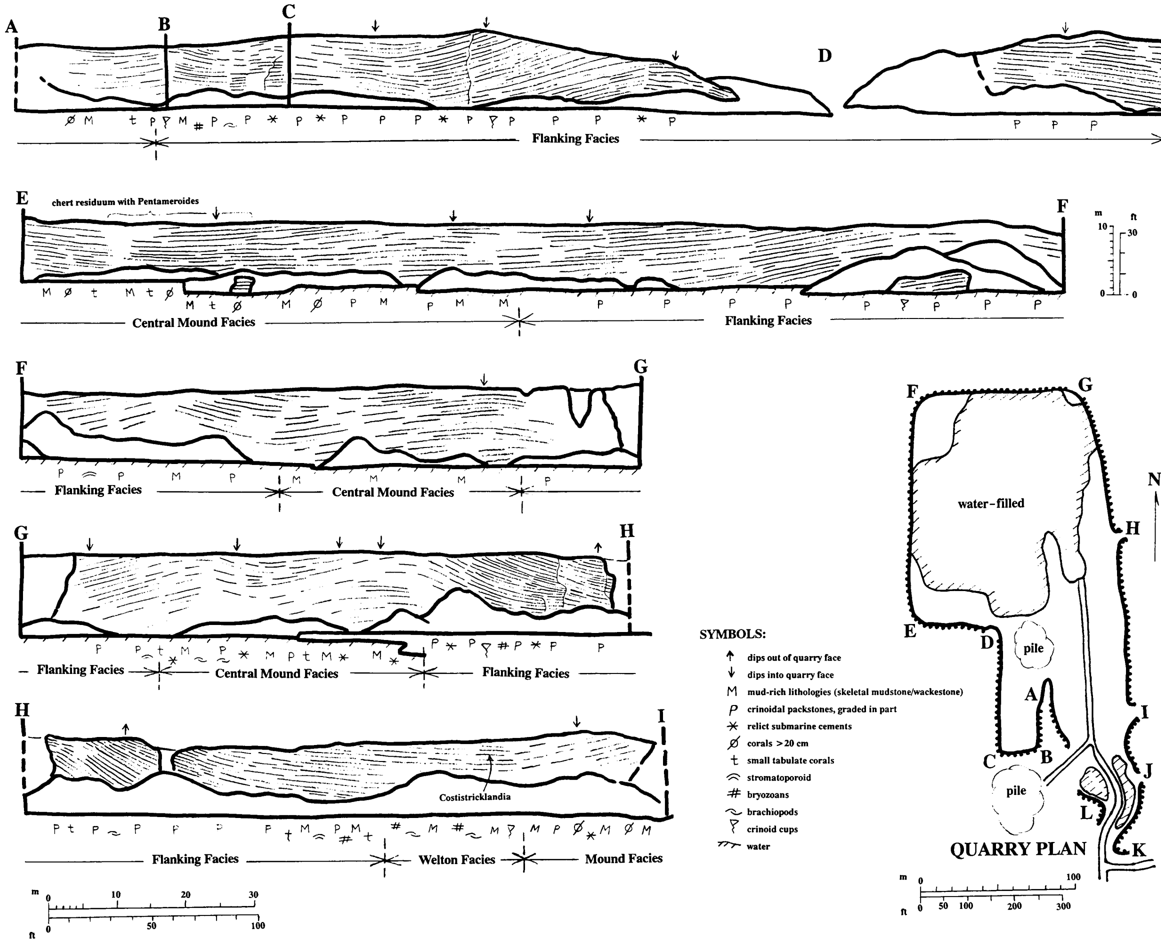

Figure 4. Bedding within carbonate mounds of the Johns Creek Quarry Member as seen along the walls of the Manternach Quarry, Jones County (Stop 2). Traced from photo mosaic; no vertical exaggeration. White areas show where walls are covered by talus or quarry piles. General extent of central mound and flanking facies is delineated.

replace and cap the dipping flanks, display skeletal-moldic wackestone fabrics characteristic of the Welton Member. These strata also contain typical lower Welton faunas, including *Costistricklandia castellana*, gypidulinids, *Eospirifer radiatus*, *Atrypa*, abundant fenestellid bryozoans, and echinoderm cups (*Eucalyptocrinites*, *Caryocrinites*). As such, the mounded complex clearly occupies a position in the lower Scotch Grove Formation and displays lateral and vertical relations with the lower Welton Member. In addition, chert residuum of the Buck Creek Quarry Member occurs above the mounded facies north of point E (Fig. 4), containing silicified corals and *Pentameroides subrectus* (first noted by Johnson, 1977b). This further constrains the carbonate mounds of the Manternach Quarry to a position below the middle Scotch Grove Formation.

We will return to county road D62, and proceed 1.4 miles west on the gravel road beginning immediately west of the bridge over Wet Creek. Turn south on gravel road, continue across county road E16, and bear west to intersection with paved county road X31. Follow X31 south into Anamosa. Turn west on Main Street and then south on S. Elm Street (E34). Cross Wapsipinicon River, and turn east into Wapsipinicon State Park (lower road).

STOP 3. WAPSIPINICON STATE PARK AND ANAMOSA CITY PARK, JONES CO.

NE SE sec. 10 and E 1/2 SE NE sec. 10, T84N, R4W
Facies of the Palisades-Kepler Member, Scotch Grove Fm.

Bluff exposures along the Wapsipinicon River in Wapsipinicon State Park (Fig. 5) will serve as the backdrop for our lunch stop. Exposed strata are assigned to the upper Scotch Grove Formation, and include horizontally-bedded porous and vuggy dolomite. The lithologies are dominantly skeletal-moldic crinoidal wackestones with scattered small tabulate corals. Near the entrance to the park, these strata display lateral gradation into less skeletal wackestones with a diverse fauna of brachiopods (*Hedeina, Atrypa, Plectatrypa, Gypidula, Ferganella, Dalejina, Coolinia,* others), corals (dominantly small solitary rugosans), bryozoans, bivalves, gastropods, trilobites (*Bumastus, Proetus,* calymenids, encrinurids), crinoid cups (especially *Dimerocrinites* with *Eucalyptocrinites, Manticrinus, Cyathocrinites,* others), and blastozoans (*Hallicystis, Lysocystites*)(see Witzke and Johnson, 1992). The horizontally-bedded exposures in the park were formerly designated the type locality for the Fawn Creek Member (Witzke, 1985), but this nomen has been dropped and the definition of the Palisades-Kepler Member modified to include non-mounded strata (see section on stratigraphy). These horizontally bedded strata are observed to merge with a large carbonate mound displaying radially dipping beds on the opposite side of the river (Block A, Fig. 5), demonstrating lateral continuity of horizontal and mounded strata in the member.

Strata of the Palisades-Kepler Member are exposed along the Wapsipinicon River for 4500 feet (1350 m), with a prominent mounded development along the northwestern edge (Block A, Fig. 5). Probably not coincidentally, carbonate mounds at the type locality (Palisades-Kepler State Park, Stop 5) are also best developed along the northwestern edge of the complex. The Palisades-Kepler Member in the vicinity of Wapsipinicon State Park is replaced laterally, both to the east and west, by cherty sparsely fossiliferous strata of the Buck Creek Quarry Member. These cherty strata are exposed at nearby localities shown in Figure 5 (three blocks designated "B"), but these will not be visited as part of this field trip. The lateral transition between the dense sparsely skeletal facies of the Buck Creek Quarry Member and the porous skeletal-moldic facies of the Palisades-Kepler Member occurs near the entrance to the state park (see previous paragraph), within sight of cherty strata on the opposite side of the road (Fig. 5).

We will exit the state park and re-cross the river, then turn right (east) after passing the sewage treatment area. Vehicles will park along Walworth Avenue near the ball diamonds. We will walk a short distance to the old quarry face designated block "A" on Figure 5. Please be cautious at the base of this face, as the area has served as a dump, and there is broken glass and jagged metal obstacles. This quarry face exposes a single carbonate mound of the Palisades-Kepler Member, displaying low-angle radial dips (Fig. 6). This mound is about 1000 feet (300 m) in diameter and at least 75 feet (23 m) high, considerably larger than the small mounds seen at Stop 2. It merges at its southern extremity with flat-lying facies of the Palisades-Kepler Member (as seen in Wapsipinicon State Park). The flanks of this mound show prominent beds of dolomitized crinoidal packstones, many displaying graded bedding. Small tabulate corals and stromatoporoids are scattered in the flanks. The central area of the mound contains fewer crinoidal packstones, and is generally dominated by skeletal wackestone fabrics. The mud-rich central mound contains scattered to common tabulate corals and stromatoporoids, and bryozoans are locally noted. An organic framework is not observed, although coral growth was clearly favored within the central mound facies. Nautiloids occur in loose aggregates, possibly representing current-related accumulations.

We will return to S. Elm Street, and proceed north to Main Street. Turn left (west) on Main (County road E28), and continue about 3.2 miles to the Stone City turn-off. Proceed southwest 0.7 miles to the Weber Quarry entrance adjacent to the Wapsipinicon River bridge (Fig. 7).

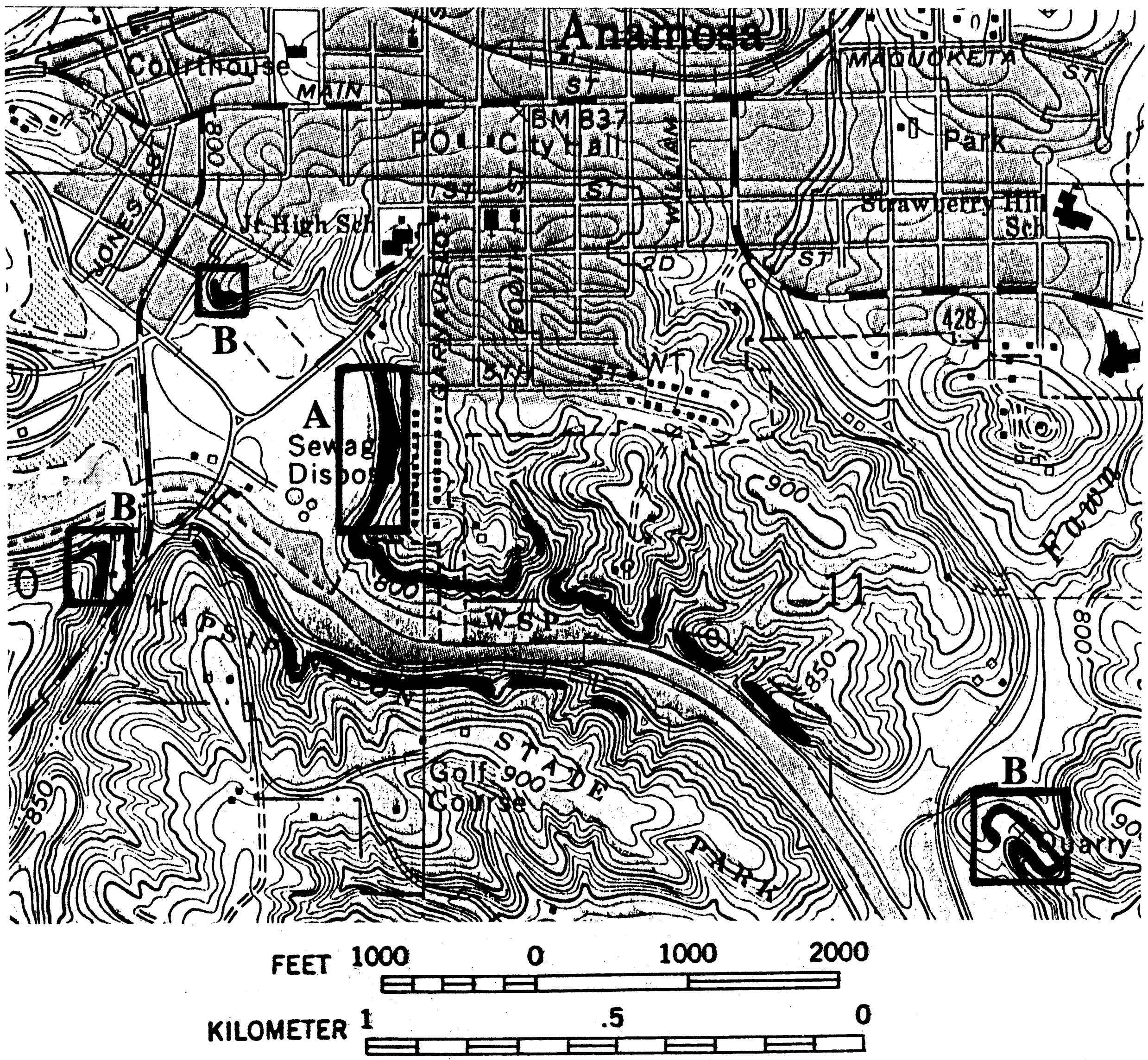

Figure 5. Location map for Wapsipinicon State Park and south edge of Anamosa, Jones County (from USGS Anamosa quadrangle). Black pattern denotes general area of bedrock exposure.

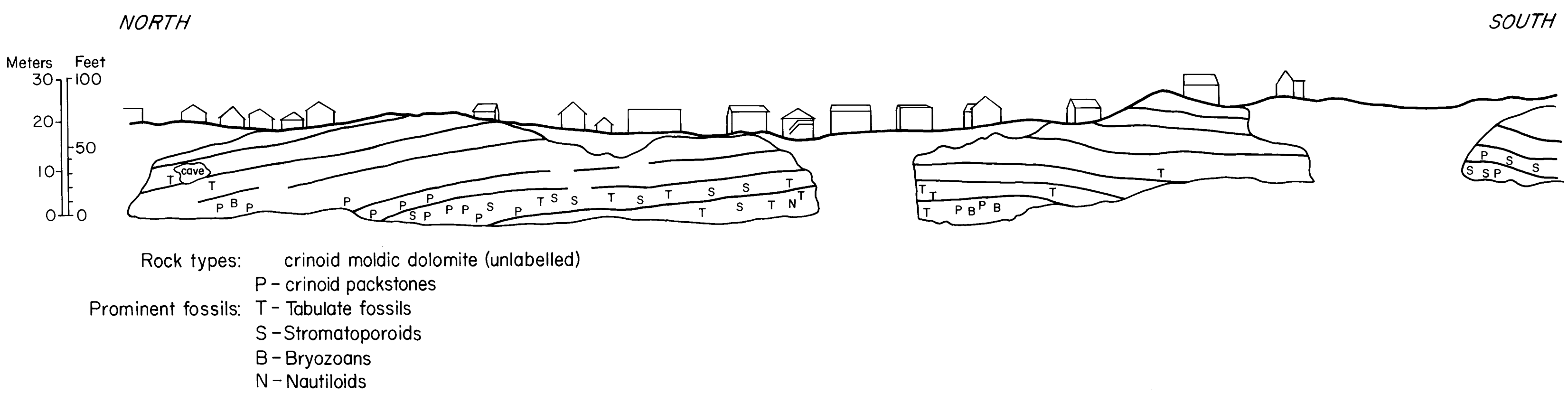

Figure 6. General bedding within carbonate mound facies of the Palisades-Kepler Member as seen along old quarry face, Anamosa City Park (Stop 3). Derived from photo mosaic; no vertical exaggeration.

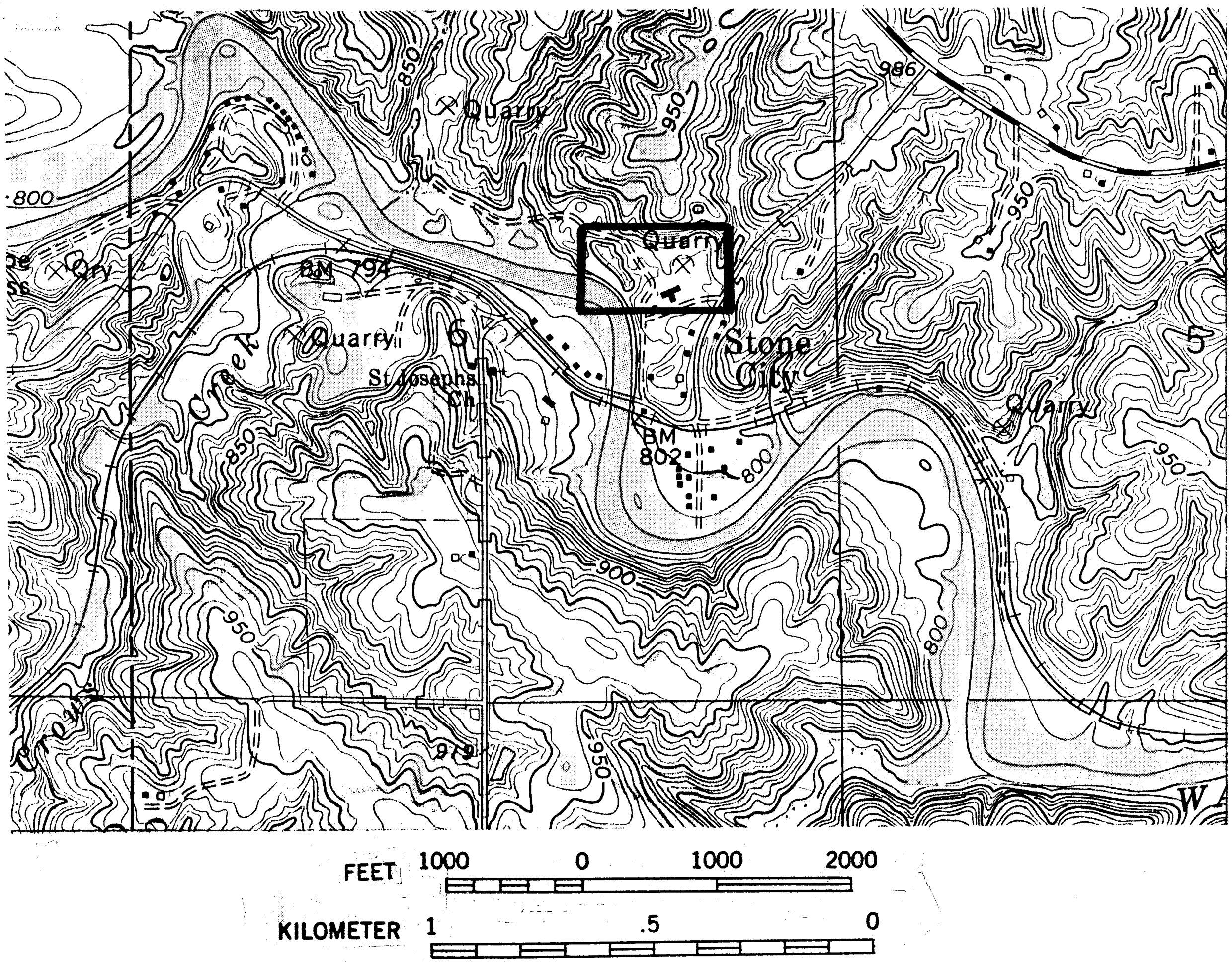

Figure 7. Location map for area around Stone City, Jones County (from USGS Anamosa quadrangle). Rectangle outlines old Weber Quarry workings (Stop 4).

STOP 4. STONE CITY QUARRY, WEBER STONE CO., INC.

SE SW NE sec. 6, T84N, R4W, Jones County
Type locality, Anamosa Member, Gower Formation

The picturesque valley of the Wapsipinicon River in the Stone City area was home to Grant Wood's art colony, and the rolling countryside is reflected in many of Grant Wood's paintings. Exposures of Silurian quarrystone are well displayed in the Stone City area, and the quarry industry has been thriving in this area since the 1840s. Military engineers were among the first to recognize the utility of this stone, constructing bridge piers from local quarried stone during the 1840s. The completion of railheads to Stone City in 1859 dramatically increased markets for this durable stone. The well-bedded laminated dolomites of the Anamosa Member have been quarried for about 150 years in a broad belt paralleling the Wapsipinicon River stretching from the Linn-Jones county line (Mt. Hope County Park) to the old Penetentiary quarries near the northwest edge of Anamosa. We will be visiting the classic Stone City quarries, first opened in 1869. This quarry area has been progressively enlarged to meet the demand for quality dimension stone, and Weber Stone Company has significantly expanded their quarrying operations to the east along the north bank of the river in recent years. They presently are the primary supplier of decorative and dimension stone in the state.

The Stone City quarries have been designated the type locality for the Anamosa Member (Calvin, 1896; Witzke, 1985), although the quarries lie about 3 miles (4.8 km) west of Anamosa. The historic reason for this seeming geographic discrepency is explained by Calvin (1896, p. 56-57): "When they [Stone City quarries] were first opened Anamosa was the nearest railway station and on this account they became generally known as the Anamosa quarries. The stone was shipped to many points in Iowa and even beyond the limits of the state, and became known among architects and engineers as Anamosa limestone. Under this name it was discussed in trade journals . . . While therefore the beds of this particular stage do not occur at Anamosa, it seems best to retain a name so long and so firmly established." The dolomite continues to be known as "Anamosa stone" to this day. The stratigraphic position of the Anamosa Member is well displayed in the Weber quarries, with the lower contact accessible at this stop (Fig. 8). Although not noted in previous studies, the Anamosa Member is capped by Devonian strata (Otis Fm.) in the expanded quarry area east of the road. Due to time constraints, we will not examine the upper units of the Anamosa Member or the contact with the Devonian. However, unit descriptions of the composite stratigraphic section are included for completeness. We will begin our field examination in the lowest quarry levels. The cooperation of Weber Stone Company in visiting their property is gratefully acknowledged. Hard hats are required; please be cautious along the quarry faces.

The lowest stratigraphic interval in the quarry is assigned to the Waubeek Member of the Scotch Grove Formation. The Waubeek is the dominant inter-mound facies of the upper Scotch Grove, and is laterally equivalent to mounds of the Palisades-Kepler Member. The Waubeek Member is characterized by thick-bedded skeletal-moldic wackestone and mudstone fabrics. Fine crinoid debris is typically dominant, but additional subtidal open-marine fauna also occurs including brachiopods (orthids, resserellids, gypidulids), trilobites (*Encrinurus*), and corals (small tabulates and solitary rugosans). The Waubeek Member is vuggy to varying degrees, and vugs are commonly lined with quartz crystals. No chert is present, although Waubeek strata are replaced by cherty Buck Creek Quarry strata 2 miles (3.2 km) to the northwest (Matsall Bridge) and 3.5 miles (5.6 km) to the east (Stop 3 area). The top of the Waubeek Member at Stone City shows a marked faunal change, containing faunas identical to those seen elsewhere in lower Gower strata (e.g., Stop 5). Brachiopods are particularly numerous, including rhynchonellids, *Spirinella*, *Protathyris*, and *Coolinia*. The characteristic Gower rugosan, *Pycnostylus*, also occurs in this bed. The contact with the overlying Anamosa Member is drawn at the base of the first laminated bed (unit 6). Waubeek-like lithologies (unit 7) occur above this laminated unit, suggesting possible interfingering between the two members.

Characteristic laminated Anamosa quarrystone ledges occur above, forming the so-called "gray

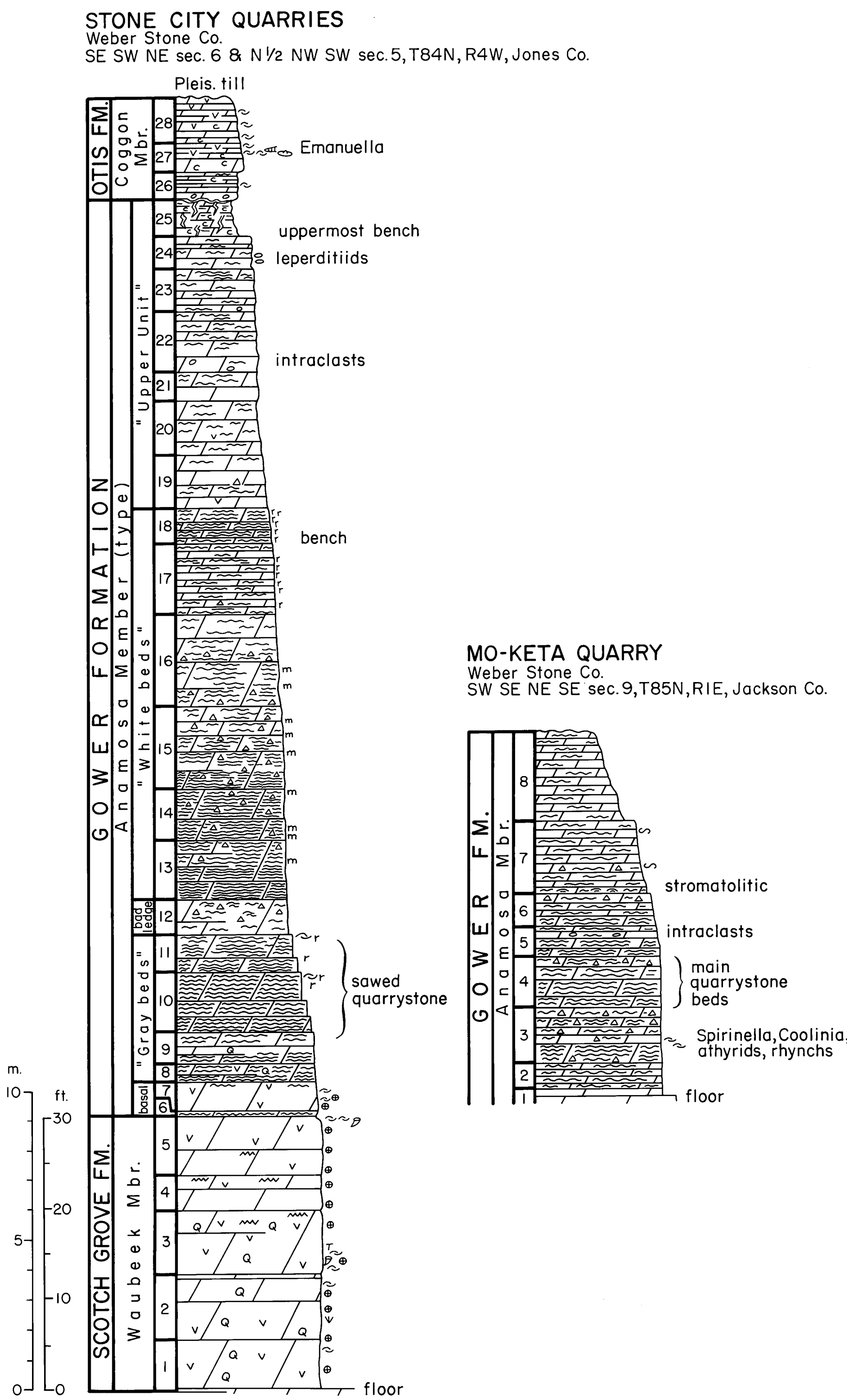

Figure 8. Graphic stratigraphic section of exposed interval seen in the Stone City quarries (Stop 4). Unit descriptions are provided in text. Mo-Keta quarry section is provided for comparison. See page 86 for lithologic symbols.

beds"(Fig. 8). These beds are prominently laminated, displaying wavy laminations with interlaminar porosity, and some appear stromatolitic in origin. Minor thin mudstones intercalate with the laminated strata. The laminae are uninterrupted along the quarry walls, and thin mudstone units are laterally persistent over distances of at least 2 miles (3.2 km)(Henry, 1972). Such continuity of laminae and mudstones would not be expected in peritidal settings, and a subtidal setting is indicated. The laminated beds are generally unfossiliferous, although scattered skeletal debris is noted along some laminae (including rhynchonellids, pterioid bivalves, leperditiids). The enigmatic "rods" are noteworthy in the upper beds.

The so-called "bad ledge" (unit 12) does not split into uniform slabs and is generally unsuitable for dimension stone. The unit is laminated in part, but unlike the underlying quarrystone beds, the laminae are faint and laterally discontinuous and mudstones pinch and swell. In addition, intraclasts occur locally along the discontinuous surfaces, stromatolitic heads are noted, and numerous small vugs and pores are present. As first noted by Henry (1972), some of the tabular pores probably represent evaporite crystal molds, most likely after gypsum. The change from "gray unit" deposition to "bad ledge" deposition probably represents a short-lived shallowing, possibly marked by peritidal or higher-energy sedimentation in part.

Overlying laminated strata of the "white beds" (Fig. 8) differ in several fundamental ways from quarry ledges of the "gray beds": 1) planar vs wavy laminations, 2) presence of faint irregular burrow-like mottling (labelled "m" on Fig. 8), and 3) occurrences of small nodular chert bands. Laminations and thin mudstones are continuous along the quarry walls, suggesting subtidal deposition. Laminae become faint upward in the "white bed" interval, although wavy laminations with scattered to abundant "rods" are prominent in the upper part of the "white beds." We will not examine strata of the "upper unit" (Fig. 8), which are exposed in the eastern quarries of the Weber operation. Like the "bad ledge," the "upper unit" displays irregular and discontinuous faint laminations, and intraclasts and numerous small vugs and pores are present. Large leperditiid ostracodes are common in some beds of the "upper unit." In general, the "upper unit" displays features suggestive of peritidal to nearshore deposition, and the Anamosa Member is interpreted as a general shallowing-upward sequence.

Lateral stratigraphic relationships within the Gower Formation of eastern Iowa clearly show contemporaneity of carbonate mound growth (Brady Member) and laminated Anamosa deposition. The Anamosa Member is considered as a subtidal inter-mound facies deposited in stressed hypersaline environments that were inhospitable for most benthic organisms. We will visit correlative fossiliferous Brady mound facies at Stop 5.

The trip will cross the bridge at Stone City and turn south on the county road a short distance to the west. Continue southward about 2.1 miles to intersection with county road E34, then east on E34 about 1.5 miles. Turn south to intersection of Highways 1 and 151, and continue south on Highway 1 through Martelle and Mt. Vernon. At south edge of Mt. Vernon, turn west on Highway 30 and proceed about 3.6 miles to turn-off into Palisades-Kepler State Park. Bear right and continue along road to the lower area along the Cedar River (Fig. 9). Proceed northwest (right) along lower road until it dead-ends in a parking area. A trailhead leads upstream along the Cedar River bluffs.

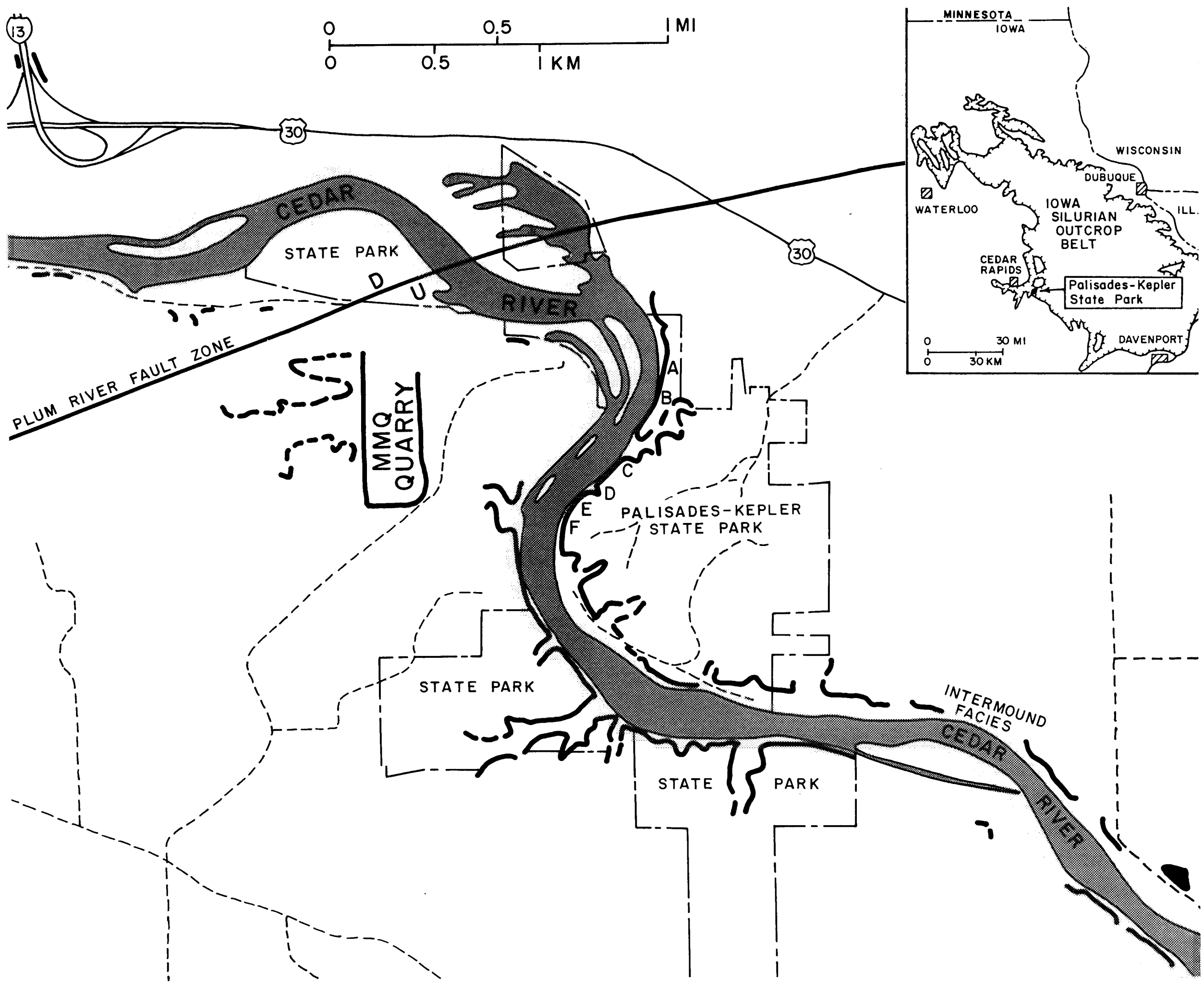

Figure 9. General location map for Palisades-Kepler State Park area, Linn County (from Witzke, 1987a). Bedrock outcrop shown in black; secondary roads are dashed. Localities A through G correspond to mound crest locations of Philcox (1970b).

STONE CITY; Weber Quarries
secs. 5 & 6, T84N, R4W, Jones Co.
units 1-12 described in old west quarry area, SE SW NE sec. 6
units 10-28 described in east quarry workings, N 1/2 NW SW sec. 5
measured B.J.Witzke & G.A.Ludvigson, Aug. 24, 1989

MIDDLE DEVONIAN
WAPSIPINICON GROUP
OTIS FORMATION, COGGON MEMBER

Unit 28
Dol., xf xlln, upper half is porous and vuggy, poikilotopic calcite cements common; bedded ledges 10-20 cm, becomes recessive upward, overlain by unoxidized to oxidized Pleistocene glacial till; contains scattered brachiopods (*Emanuella*); 1.5 m.

Unit 27
Dol., xf xlln, xf-f xlln in lower half, common brown poikilotopic calcite nodular cements through, calcite spar void fills; basal 47 cm is overhanging ledge-former, single bed; upper 50 cm in 4-6 beds, top 35 cm is very porous and skeletal moldic with scattered to abundant brachiopods (*Emanuella*), rare bivalves, gastropods, nautiloids (unit processed for conodonts; barren); 97 cm.

Unit 26
Dol., xf xlln, dense, scattered small pores, sparse skeletal molds (brachiopods) in middle part, small (mm-scale) intra- or lithoclasts 20 cm above base, irregular base (up to 7 cm relief); lower 41-48 cm is well bedded in 4-5 beds; upper 48 cm is thin flaggy bedded (beds 1-5 cm), poikilotopic calcite nodules and cements in part, some faint laminations at top; 89-96 cm.

SILURIAN
GOWER FORMATION
ANAMOSA MEMBER

"Upper Unit"
Unit 25
Dol., xf-f xlln, locally c-vc saccharoidal, abundant calcite cements and vein fills, light gray, highly weathered, rubbly, highly fractured to brecciated, fractures and voids locally filled with green clay, green clay locally in seams and along stylolitic surfaces; top 10-20 cm becomes very shaley, green silty shale fills in irregularities on dolomite surface; unit represents pre-Middle Devonian weathering surface developed on Silurian bedrock; 1.25 m.

Unit 24
Dol., vf-f xlln, some xf (appears "sublithographic" in part), dense, hard, more coarsely crystalline and harder than below, bedded 10-20 cm, faintly and irregularly laminated in part, green clay along partings in upper beds; leperditiid ostracods common in lower interval; top of unit is uppermost quarry bench; thickness approximate, 1.1 m.

Unit 23
Dol., xf-vf xlln, in 4 to 6 beds; lower half irregularly and faintly laminated, very porous, abundant small rounded pores (most < 1 mm), quartz sand rare, small intraclasts noted; upper half if irregularly wavy laminated with common small pores in lower part, becomes more finely laminated (planar to wavy) in upper part, laminae are in part discontinuous, some are obliquely truncated; 1.4 m.

Unit 22

Dol., xf-vf xlln, in 5 to 6 beds, very porous in lower 1.05 m, abundant small rounded pores (most < 1 mm, some 3-4 mm), scattered rounded elongate intraclasts to 3 cm, interval faintly laminated internally, prominent lithologic break at base; upper 95 cm less porous, faint irregular wavy laminae, some laminae are disrupted (or truncated), top 10 cm interlaminated on cm-scale with porous rock similar to unit 23; 2.0 m.

Unit 21

Dol., xf-vf, dense, in two beds; lower half has common small molds and pores as above; upper half has faint planar to wavy laminations; 1.0 m.

Unit 20

Dol., xf-vf, faint irregular laminations, laminae are locally disrupted and discontinuous, laminae scattered in upper part; 40-56 cm above base is non-laminated mudstone with small vugs; 1.75 m.

Unit 19

Dol., xf, rubbly weathered in 4 to 6 beds, primarily non-laminated, porous, common small indeterminate molds and voids; scattered smooth chert nodules 80 cm above base; faintly laminated 40-80 cm above base; 1.8 m.

"White beds"

Unit 18

Dol., xf-vf, in 4 beds, faint to prominent wavy laminations, small pores along some laminae; 5 cm thick dense mudstone 70 cm above base; "rods" common to abundant along laminae; lithologic break at top; 1.18 m.

Unit 17

Dol., xf, dense, pale gray to white, in 11-14 beds, faint planar laminae through most of unit; wavy laminations in basal 3 cm, 1.8-1.9 m above base, and scattered in top 47 cm; becomes non-laminated near top; nodular chert band (nodules to 7 cm) 47 cm above base; "rods" scattered along some laminae at 30 cm and 1.0-1.9 m above base; top of unit is major quarry bench; 2.37 m.

Unit 16

Dol., as above, faint planar laminations (mm to cm separation), laminations become fainter in top 1.0 m; dense except microporous laminations 1.35-1.5 m above base and small pores or molds (<1 mm) in top 75 cm; small nodular chert bands 8 cm, 45 cm, 1.68 m, 1.8 m above base; burrow-like mottling noted 63 cm and 1.1-1.2 m above base; 3.08 m.

Unit 15

Dol., as above, faint planar laminations through most of unit, laminae become more indistinct in upper 75 cm; non-laminated mudstone 67-88 cm above base; burrow-like mottling noted 1.22 m, 1.87 m, 2.24 m, 2.51 m above base; bands of scattered nodular chert noted 28 cm, 49 cm, 53 cm, 72 cm, 97 cm, 1.12 m, 1.35 m, 1.38 m, 1.61 m, 1.79 m, 2.06 m, 2.14 m, 2.31 m, 2.36 m, 2.42 m above base; 2.79 m.

Unit 14

Dol., as above, fine planar laminae throughout; burrow-like mottling along some laminae 7 cm, 10 cm, 13 cm, 19 cm, 22 cm, 29 cm, 42 cm, 84 cm, 1.06 m, 1.1 m, 1.58 m, 1.65 m above base; sparse nodular chert bands noted 25 cm, 48 cm, 91 cm, 1.05 m, 1.11 m, 1.23 m, 1.33 m, 1.42 m, 1.51 m, 1.65 m above base; 1.69 m.

Unit 13

Dol., as above, fine planar laminae throughout, laminae locally form small monoclinal flexure in lower 65 cm; burrow-like mottling noted 1.35 m above base; sparse nodular chert bands noted 1.32 m, 1.71 m, 1.86 m, 2.0 m above base; 2.03 m.

"Bad Ledge"

Unit 12

Dol., xf-vf, laminated in part, lower 45 cm with wavy laminae and small pores or molds along laminae; upper 70 cm irregularly laminated in part, laminae faint to prominent, laminae form stromatolitic-like heads to 7 cm high x 25-30 cm wide in upper part, some laminae disrupted, local small vugs and pores (evaporite molds in part), intraclasts noted locally; brown to white nodular chert in bands at 45 cm and 65 cm above base, chert nodules scattered 85 cm-1.0 m above base; contact appears gradational above; 1.15 m.

"Gray beds"

Unit 11

Dol., light gray, xf-vf, wavy laminations through most, laminae become fainter upward; top 15 cm becomes more porous; dense mudstones ("flints" 1-4 cm thick) noted 27 cm and 83 cm above base and near top; "rods" scattered along laminae in middle to upper part, "rods" abundant near top; scattered rhynchonellid brachiopods and bivalves present at top; unit sawed into two benched cuts; 1.32 m.

Unit 10

Dol., as above, prominent wavy laminations, scattered to common small interlaminar pores; "rods" scattered along laminae in upper part; 2 cm thick mudstone ("flint") layer 1.05 m above base and near top; rhynchonellid brachiopods, leperditiids, and bivalves locally noted in upper part; unit sawed into three benched cuts; 1.98 m.

Unit 9

Dol., xf-vf, lower 15 cm with prominent wavy laminations, some laminations form stromatolitic doming to 25 cm wide x 1 cm high; 15-37 cm above base is faintly laminated to non-laminated mudrock, scattered vugs (some quartz-lined); 30-60 cm above base is non-laminated, scattered small pores or molds; 60-72 cm above base is dense, hard mudstone ("flint"), bedding split in middle; upper 36 cm faintly laminated in part, scattered indeterminate small pores or molds; 1.08 m.

Unit 8

Dol., xf-vf, very light gray, finely laminated throughout, laminae slightly wavy, alternating (<1 mm-2 mm) denser and more porous layers; in 1 to 3 beds; lower 15 cm has small indeterminate pores or molds; small vugs (some quartz-lined) noted 4 cm and 21 cm below top; 60 cm.

basal Gower interval

Unit 7

Dol., xf-vf, single massive bed; sparse to common molds and small vugs (increasing upward); scattered to common skeletal molds (especially small crinoid debris) through most, decreasing in abundance upward, indistinct m gray mottlings (burrows?) in upper part; brachiopods (?gypidulids) in band 42 cm above base, indeterminate brachiopods 35 cm below top; upper 7 cm is faintly laminated with indeterminate small molds scattered through; 98 cm.

Unit 6

Dol., in 2 beds, lower 12 cm is finely planar laminated; top 3 cm is finely skeletal moldic (small crinoid debris); 15 cm.

SCOTCH GROVE FORMATION
WAUBEEK MEMBER

Unit 5

Dol., xf-vf, pale gray, sparse to common skeletal molds, scattered vugs (1-4 cm diameter, some calcite lined) especially 40-50 cm and 1.1-1.5 m above base; prominent stylolite at 90 cm above base forms bedding surface; skeletal molds dominantly small crinoid debris, molds slightly more common upward; top 5 cm splits off as separate bed, very fossiliferous, brachiopods (rhynchonellids, *Protathyris*, *Spirinella*, *Coolinia*, others), corals (small tabulates, cup corals, *Pycnostylus*); 2.0 m.

Unit 4

Dol., xf-vf, sparse skeletal moldic (small crinoid debris), denser and fewer vugs than below; prominent stylolites at top and 12 cm down; in 2 beds (top bed 35 cm); 1.1-1.3 m.

Unit 3

Dol., xf-vf, sparse skeletal moldic (dominantly small crinoid debris); vugs scattered throughout, especially common 40-90 cm below top, some vugs lined with calcite, chalcedony, or megaquartz; in 1 to 2 beds, bedding break 1.45 m above base, prominent bedding break at top, stylolites at top and 35 cm down; scattered indeterminate brachiopods (orthids), silicified cup coral 60 cm above base, small silicified *Favosites* (5 cm) 1.0 m above base; 2.15 m.

Unit 2

Dol., xf-vf, as above; vugs most common in lower 70 cm, some with calcite and quartz (chalcedony, megaquartz), vugs scattered above; in 2 thick beds (lower is 1.3 m), top 12 cm is ledge with prominent bedding breaks; small crinoid debris molds throughout, trilobite (*Encrinurus*) in lower part, scattered brachiopods (including gypidulids, resserellids); 2.15 m.

Unit 1

Dol., xf-vf, sparse skeletal molds; scattered vugs (2-8 cm), some with chalcedony and megaquartz vug linings; basal 10 cm seen in quarry floor adjacent to sump, remainder forms single bed, prominent bedding plane at top; small crinoid debris molds throughout, scattered indeterminate brachiopods in upper part; 1.75 m.

STOP 5. PALISADES-KEPLER STATE PARK, LINN COUNTY

NE SW and W 1/2 NE sec. 14, T82N, R6W
Type locality of Palisades-Kepler Member
Carbonate mound facies in upper Scotch Grove and lower Gower formations

The trip will proceed along the trail to visit the main portion of the carbonate mound complex originally described by Philcox (1970b). He indicated the location of six general "mound crests," labelled A through F on Figure 9, which form a broad coalesced complex of mounded features that has been designated the type locality of the Palisades-Kepler Member. A generalized cross-section of the exposed interval in this area is presented in a previous portion of this guidebook (see fig. 10, Witzke this guidebook, in the section entitled "Genesis of carbonate mound facies"), which will provide a useful reference along the trail. In general, mound crests C, D, E, and F are marked by relatively subtle changes in dip within the complex, and the included mounds are characterized by relatively low-angle dips (0-20°). Exposures downstream grade into less skeletal horizontally-bedded units, and merge eastward into characteristic inter-mound units of the upper Scotch Grove Formation (Fig. 9). By contrast, mounds A and B are marked by steeper dipping flanks which are capped by brachiopod-rich units of the Gower Formation, Brady Member.

We will examine various dolomite lithologies in the area of mounds D, E, and F, observing both skeletal-moldic wackestones and crinoidal packstones. Crinoid debris forms the dominant skeletal component, although the matrix material, which is interpreted to have originally been carbonate mud, volumetrically dominates the mound facies. Scattered small tabulate corals (primarily *Favosites*) and stromatoporoids are present, but are nowhere abundant. Large crinoid debris is conspicuous at places along the trail, locally displaying current-oriented fabrics or graded bedding. Careful inspection will reveal occasional crinoid cups, primarily *Eucalyptocrinites*. The trail continues onto the upland bluff margin, and then drops down into the prominent drainage between mounds B and C further to the northeast. We will proceed up this drainage for a short side-trip to examine brachiopod-rich Brady strata in the topographic low between mounds B and C of the Palisades-Kepler Member. Similar Brady facies are exposed in stratigraphic position above the flanks of mound A, which we will visit if time and conditions permit.

After the side-trip, we will return to the main trail which parallels the river, and proceed northward (upstream) towards mounds A and B. These two mounds form the most instructive part of the mounded complex exposed within the park, marking the northwestern edge of the complex. We will descend a steep drainage from the upland trail position to visit the northern flanks of mound B. Some of the steepest dips exposed in the Palisades-Kepler Member within the state park are seen at this position, where flanking beds reach dips up to 35°. A photo mosaic tracing of this exposure is found in an earlier portion of this guidebook (see "Genesis of carbonate mound facies," fig. 9 of Witzke, this guidebook). Crinoidal packstones are prominently displayed at this location, in part forming graded increments of packstone-wackestone about 10 to 20 cm in thickness. Some wedge bedding is apparent on the exposure, and the flanking strata are interpreted as episodic debris flows, probably triggered by storm events.

If conditions permit, we will proceed northward from this area to examine the northern edge of the Palisades-Kepler complex around mound A, where upper Scotch Grove strata are buried beneath brachiopod- and mud-rich facies of the Gower Formation, Brady Member. Exposures of the Palisades-Kepler Member within mound A locally display common tabulate corals and stromatoporoids, and mud-rich internal-sediment and cavern fills are developed in places. The uppermost beds of the Palisades-Kepler Member are, in general, less skeletal and more mud-rich than underlying units, and crinoid debris becomes notably less common upward. The Scotch Grove-Gower contact is drawn at the base of the lowest rhychonellid- or protathyrid-rich bed in the mound flank areas, but sedimentary changes appear gradational across the contact. Higher along the mound crests, the development of cave- and fissure-fills is suggestive of subaerial exposure of the topographically higher portions of the mounds at the close of Scotch Grove deposition.

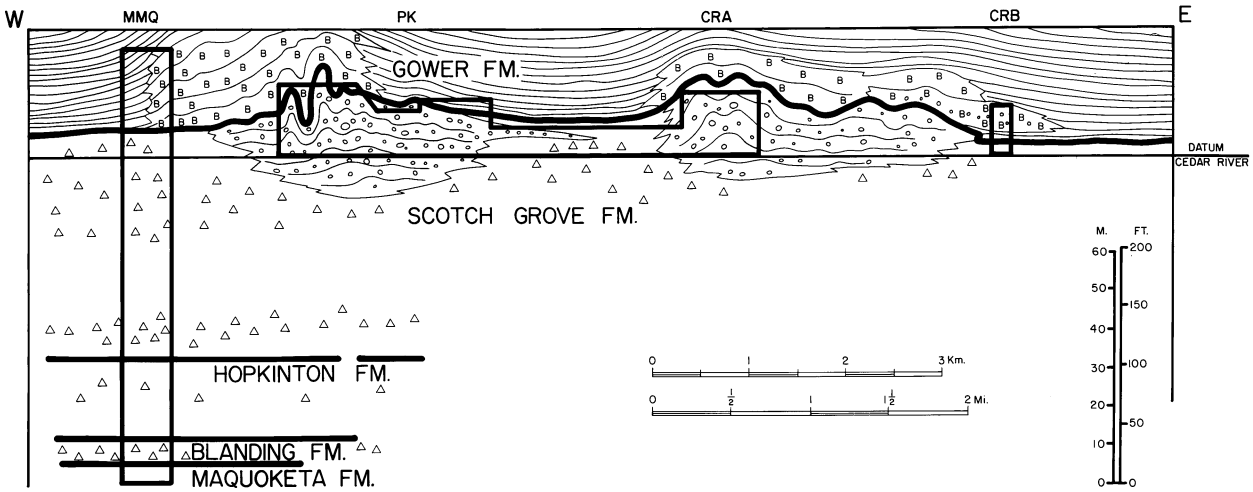

Figure 10. Interpretive Silurian cross-section along Cedar River in vicinity of Palisades-Kepler State Park (from Witzke, 1987a). Boxed areas correspond to outcrop or core sections; remainder is reconstructed. See page 86 for lithologic symbols.

Dipping Brady strata along the northern flanks of mound A include sparsely skeletal mudstones interbedded with brachiopod-rich wackestones and packstones. Brady faunas in this area are generally dominated by rhynchonellid brachiopods, but additional brachiopods (protathyrids, meristellids, *Spirinella*, *Reticulatrypa*, *Fardenia*, *Leptaena*), gastropods, bivalves (*Plethomytilus*, *Cyrtodonta*), and corals (*Pycnostylus*) are identified. The skeletal-rich beds are wedge-shaped in part. Dipping Gower beds flatten out northward to merge and interfinger with horizontally-bedded mudstones and laminated strata of the Anamosa Member.

Although we will not have time to visit the exposures west of the state park at the Cedar Rapids Martin-Marietta Quarry (Loc. MMQ, Fig. 9), the stratigraphic relations exposed there expand our perspective of the Palisades-Kepler mound complex. The Scotch Grove Formation exposed at MMQ is exclusively developed in inter-mound wackestone and mudstone facies (Waubeek and Buck Creek Quarry members), providing constraints on the westward limits of the Palisades-Kepler mound complex. Gower strata at MMQ are primarily horizontally-bedded laminated strata of the Anamosa Member, although a prominent and well developed mound of the Brady Member is seen in the southern portion of the quarry (see fig. 10, Witzke this guidebook, in the section entitled, "Genesis of carbonate mound facies"). This shows that westward or northwestward overstepping of the Palisades-Kepler mound complex by later-stage Brady mounding had taken place.

The general stratigraphic relations of upper Scotch Grove and Gower strata as seen in the area around Palisades-Kepler State Park are displayed in an interpretive cross section (Fig. 10). The crinoidal mound complex at Palisades-Kepler (area PK, Fig. 10) displays the steepest dips along its northwestern margins. The complex is replaced to the east and west by inter-mound strata of the Waubeek and Buck Creek Quarry members. An additional mound or mound complex is developed further downstream along the Cedar River (Loc. CRA, Fig. 10). Mounded and flanking units of the Brady Member bury and overstep the Palisades-Kepler mound complex primarily along its northwestern margin, and Brady strata interfinger down-dip with laminated Anamosa strata to the north.

We will return on the trail to the parking area, and depart from the state park back to Highway 30. Depending on time constraints and conditions, we may be able to visit inter-mound facies downstream from the state park exposed along the Cedar River in the Harold Ennis County Preserve

(SE NE sec. 24, T82N, R6W). Proceeding east on Highway 30 from the state park, we will take the first gravel county road south. The parking area for the county preserve is 1.3 miles south of the highway. Trails lead to the Cedar River, where natural exposures of the Waubeek Member stretch along the river banks (Fig. 9, "intermound facies"). Waubeek strata are dominated by skeletal mudstones and wackestones, commonly vuggy, and contain scattered to common small crinoid debris molds. Additional fauna is noted at this locality, including a fauna similar to that seen near the entrance to Wapsipinicon State Park (Stop 3): brachiopods (*Hedeina*, *Atrypa*, gypidulids, *Dalejina*, etc.), corals (small tabulates, solitary rugosans), bryozoans, gastropods, trilobites (illaenids, proetids), and camerate crinoid cups (Witzke and Johnson, 1992). Exposures of inter-mound strata (cherty in part) are scattered upstream along the river, becoming gradually more crinoidal as Waubeek strata merge with the eastern edge of the Palisades-Kepler mound complex near the boundary with the state park about one mile (1.6 km) upstream from the county preserve (see Philcox, 1970b). Carbonate mound facies also re-appear about 0.3-0.6 (0.5-1.0 km) miles downstream from the county preserve Loc. CRA, Fig. 10).

We will leave the parking area, and proceed east along the gravel road that parallels the Cedar River to its intersection with Highway 1. Follow the highway southeastward to return to Iowa City. Thank you for your cooperation.

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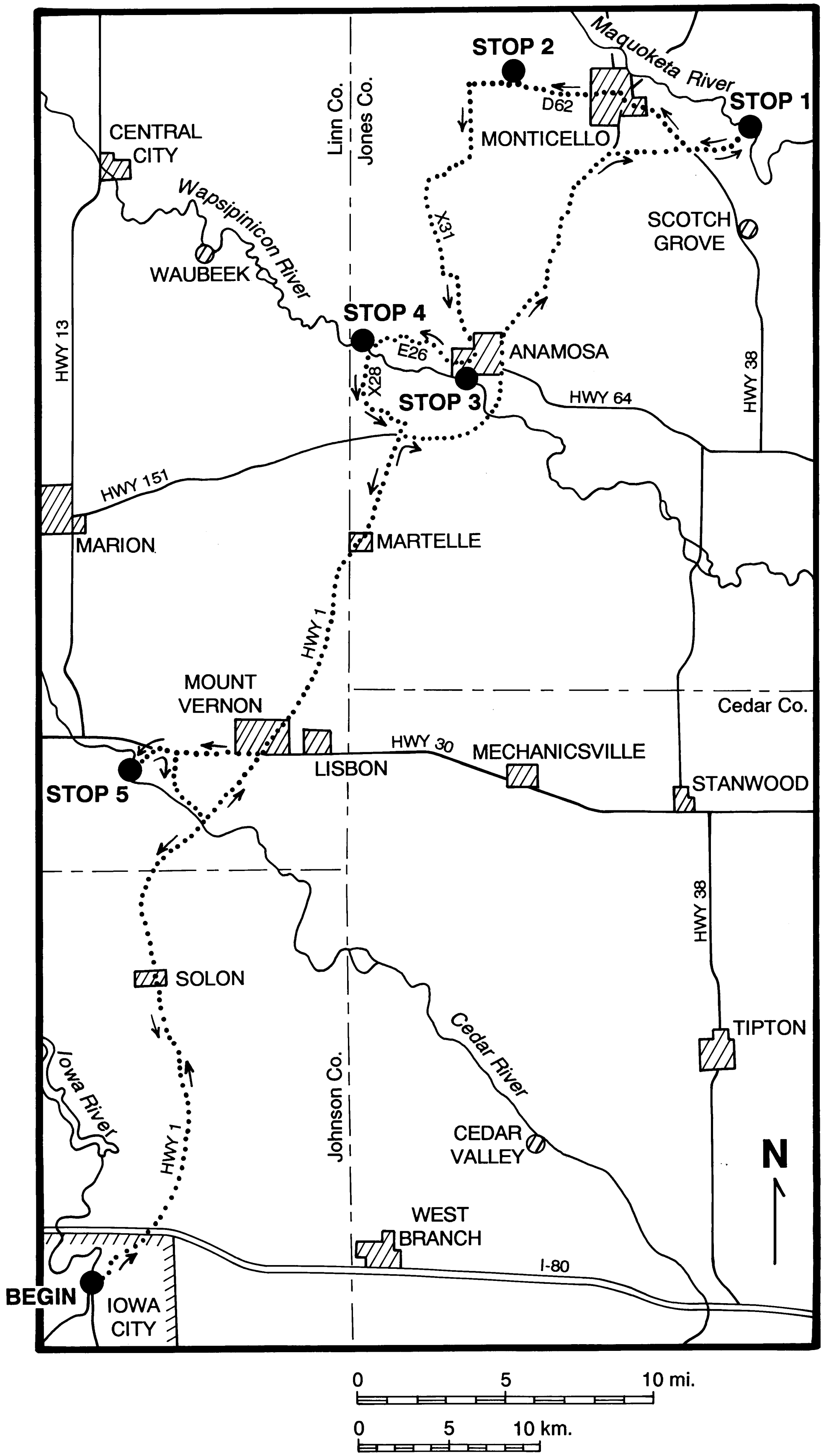
STOP 2
Maquoketa River
STOP 1
D62
MONTICELLO
CENTRAL CITY
Linn Co.
Jones Co.
Wapsipinicon River
X31
SCOTCH GROVE
WAUBEEK
STOP 4
E26
ANAMOSA
X28
STOP 3
HWY 13
HWY 64
HWY 38
HWY 151
MARION
MARTELLE
HWY 1
MOUNT VERNON
Cedar Co.
HWY 30
MECHANICSVILLE
LISBON
STANWOOD
STOP 5
HWY 38
SOLON
TIPTON
Iowa River
Cedar River
Johnson Co.
HWY 1
CEDAR VALLEY
N
WEST BRANCH
I-80
BEGIN
IOWA CITY
0
5
10 mi.
0
5
10 km.