NEW PERSPECTIVES ON THE PALEOZOIC HISTORY OF THE UPPER MISSISSIPPI VALLEY:

AN EXAMINATION OF THE PLUM RIVER FAULT ZONE

Guidebook No. 8

Guidebook for the 18th Annual Field Conference
of the Great Lakes Section,
Society of Economic Paleontologists and Mineralogists

Co-sponsored by the Geological Society of Iowa

Iowa Department of Natural Resources
Larry J. Wilson, Director
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of the Great Lakes Section,
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PART 1

Stratigraphic and Paleogeographic Perspectives

INTRODUCTION

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Welcome to the 18th Field Conference of the Great Lakes Section of SEPM! We hope to make this field excursion a rewarding one for you.

Given the immodest title of this conference, you might like to know in a nutshell just what is so truly new that we have to show you. Even though the rock sequences we're looking at have been well-studied since David Dale Owen's pioneering work in the early 1800's, a surprising amount of new, provocative work has been done on the Paleozoic rocks in the Upper Mississippi Valley.

One of the more exciting recent developments in the study of sedimentary rocks has been the flowering of broad-scale paleogeographic synthesis. Examples include publication of the Paleozoic, Mesozoic, and Cenozoic paleogeography volumes in the first half of this decade under the aegis of the Rocky Mountain Section of SEPM. Valid paleogeographic synthesis depends, however, on correctly reconciling rock and time stratigraphy. The pace of future developments, however, will be limited by the expansion of the data base. Happily, the papers in this guidebook by McKay and Nations contribute to that data base, and should be of value to workers with interests in the Paleozoic.

The concept of sequence stratigraphy, as controlled by eustatic cycles in sea level, is being applied to many Paleozoic successions in the region. Markes Johnson (1975, 1977a, 1977b, 1979, 1980, 1983) has a long publication record on depth-related benthic communities in the Silurian of the Upper Mississippi Valley, and his paper in this guidebook compares the community succession in the field trip area with that from several other cratonic basins in North America. Papers by Witzke and Kolata, and Bunker in this guidebook also discuss some new sequence stratigraphic interpretations for the Ordovician and Devonian systems in the Upper Mississippi Valley.

In recent years there have been several attempts to analyze the crustal dynamics of cratonic basins through the use of subsidence curves. Examples include the Michigan (Sleep et al., 1980) and Illinois basins (Heidlauf et al., 1986). One of the most vexing problems in the North American Midcontinent, however, is that large regions <u>including the area of this field conference</u> are not easily reconciled with simple well-defined regional patterns of basinal subsidence. Hence, the frequent cryptic references to "platforms" or "shelves" that lack a coherent paleogeographic framework. In fact, localized basinal subsidence <u>has</u> occurred in this region, but was of limited duration in comparison to more well-defined cratonic basins with histories of unidirectional subsidence. Areas of maximum basinal subsidence in the central midcontinent have shifted during Phanerozoic time, and <u>reversals</u> in vertical crustal movements through time are apparent in several areas. These phenomena have been outlined recently by Bunker et al. (1988), and will be briefly summarized on the following pages. The tectonic processes responsible for these complex patterns are poorly understood. Specialists with interests in modelling the crustal dynamics of cratonic sedimentary basins have not yet addressed these phenomena--in part because they seemingly are not yet aware of them!

This field trip will examine stratigraphic successions whose depositional patterns were influenced by the subsidence of the East-Central Iowa Basin (Fig. 1), a Late Ordovician to Middle Devonian feature that was destroyed by late Paleozoic regional uplift. This mid-Paleozoic feature was first identified by the pioneering work of Wallace Lee (1946), and is discussed at length by Bunker et al. (1985). While the Silurian carbonates that accumulated in this basin only

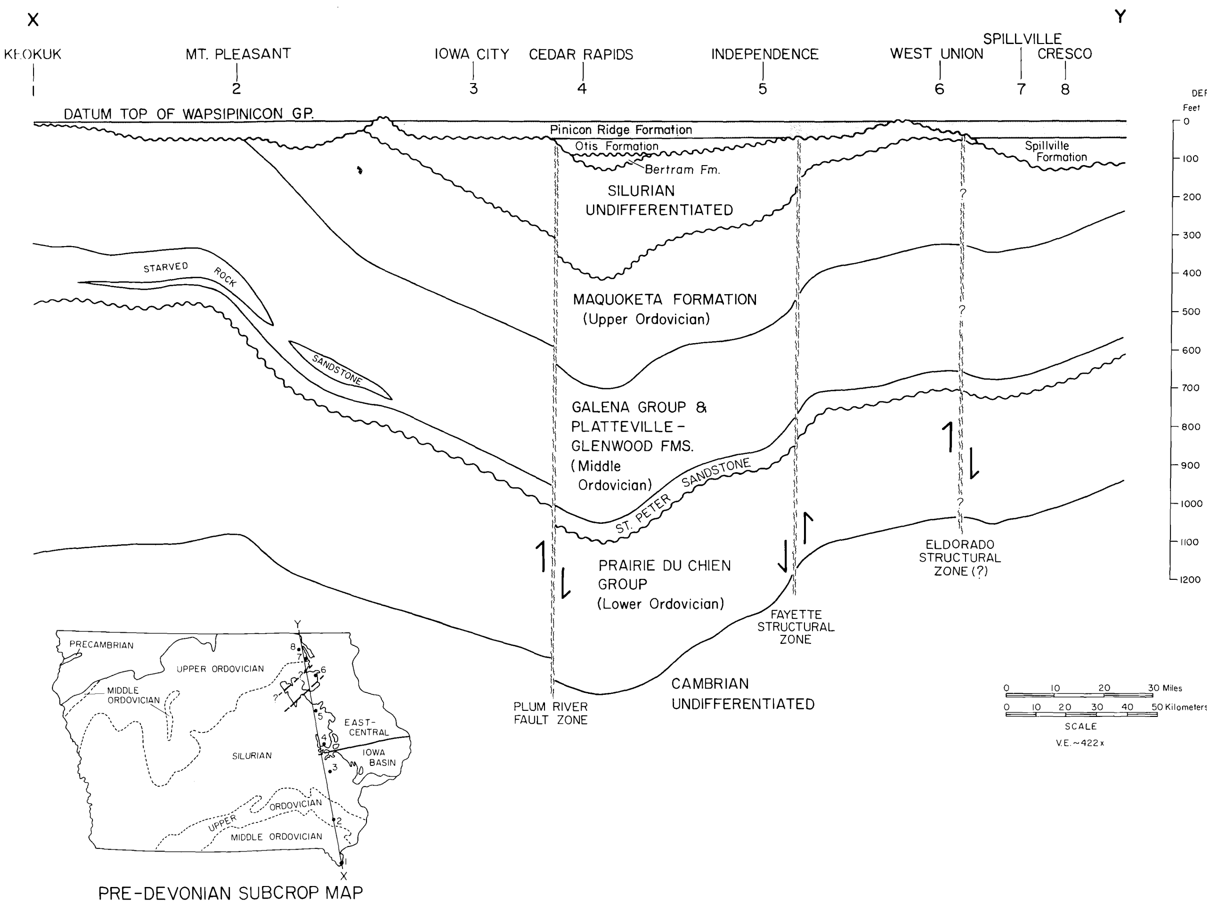


Figure 1. North-south stratigraphic cross-section of the East-Central Iowa Basin, using the top of the Wapsipinicon Group as the datum. The basin is evident as a pre-Kaskaskia structural feature. From Bunker et al. (1985).

achieve an aggregate thickness of perhaps 600 feet (183 m.; data from critical localities is lacking), the paper by Markes Johnson in this guidebook shows that Silurian deposition in this area was consistently deeper than in equivalent strata from the Williston and Michigan basins, where much thicker accumulations are present. It is tempting to speculate about implications for inter-basinal comparisons of rates of basinal subsidence and depth-controlled rates of carbonate sedimentation. Witzke (1983) showed that the Silurian deep water Stricklandiid community is confined to the central portions of the East-Central Iowa Basin, and is replaced laterally by shallower water communities on the basin's flanks. The basin was a structural, stratigraphic, and bathymetric feature, important distinctions that are outlined by Witzke and Kolata (paper in this guidebook).

While the Plum River Fault Zone (Fig. 2) is the central focus of this field conference, we will make an effort to emphasize the regional depositional patterns of the rock sequences that we are looking at, and to clarify for you the regional significance of the fault zone. This ancient structure influenced the pattern and timing of basinal development in this region. Rock fabrics from brittle fault systems have only recently been seriously studied by structural specialists, and for many of you, this trip could represent the first opportunity to look at and think about fault rock development in sedimentary carbonates.

Many cratonic structures expressed in Paleozoic rock sequences correspond to Precambrian basement features, indicating that

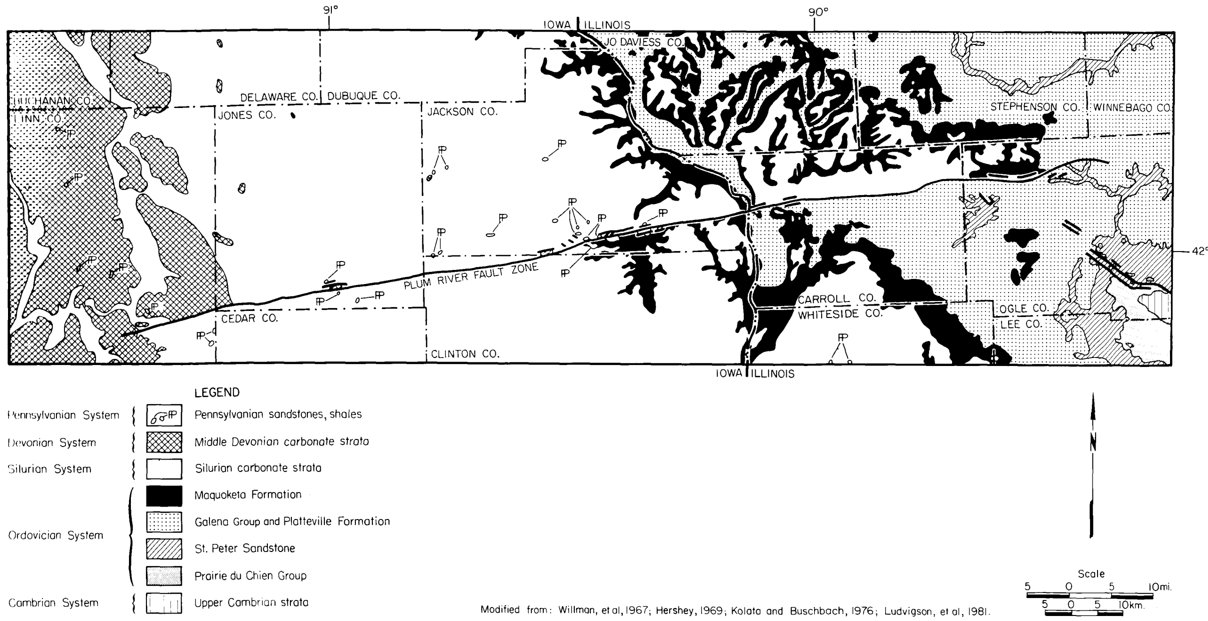


Figure 2. Bedrock geologic map of the Plum River Fault Zone. From Bunker et al. (1985).

Phanerozoic midcontinent tectonism has operated largely through the reactivation of earlier crustal structures. While structural elements in the near-surface rocks are known in some detail (Kolata and Bushbach, 1976; Kolata et al., 1978; Treworgy, 1981; Bunker et al., 1985), there is a need to integrate that understanding into the larger framework of crustal evolution in the North American Midcontinent. There have been many significant advances in recent years regarding the history of Proterozoic continental accretion and crustal evolution in the Lake Superior region and contiguous areas. The relationships of various Precambrian lithologic packages to plate tectonic processes have been clarified by Hoffman (1988) and Nance et al. (1988). Proterozoic evolution of the Midcontinent is now much better defined thanks to isotopic age determinations and interpretations by Van Schmus and Bickford (1987, 1988), and Nelson and DePaolo (1985), and studies by Sims (1985), and Anderson (in prep.). The Penokean Orogeny that produced the continental crust in the area of this field conference has been studied by Larue and Ueng (1985), Laberge and Meyers (1987), Wunderman (1988), and Southwick and others (1988). Subsequent Proterozoic thermo-tectonic modifications and the addition of supercrustal rocks have been described by Hoppe et al. (1983), Anderson and Ludvigson (1986), and Van Schmus et al. (in press). Ray Anderson's paper in this guidebook is illustrative of the kind of combined Proterozoic-Phanerozoic synthesis that is needed to further our understanding of Paleozoic tectonism.

The following pages present a very brief synopsis of a summary paper by Bunker et al. (1988) prepared for the Geological Society of America's Decade of North American Geology Sedimentary Cover of the Craton volume (Sloss, 1988). Analysis is organized within the context of Sloss' (1963) cratonic sedimentary sequences, and lesser intervals where appropriate. Importantly, the cratonic sequences are bracketed by the development of major interregional unconformities. In the North American Midcontinent, these unconformities not only record periods of major continental erosion, but also herald changes in the structural grain of epeirogenic tectonism and cratonic sedimentation (Ham and Wilson, 1967; Bunker et al., 1988). Studies of the fracture-filling minerals in the Plum River Fault Zone have led Ludvigson (1988, and this guidebook) to suggest that deeply circulating groundwaters that infiltrated during episodes of continental exposure actually accelerated the

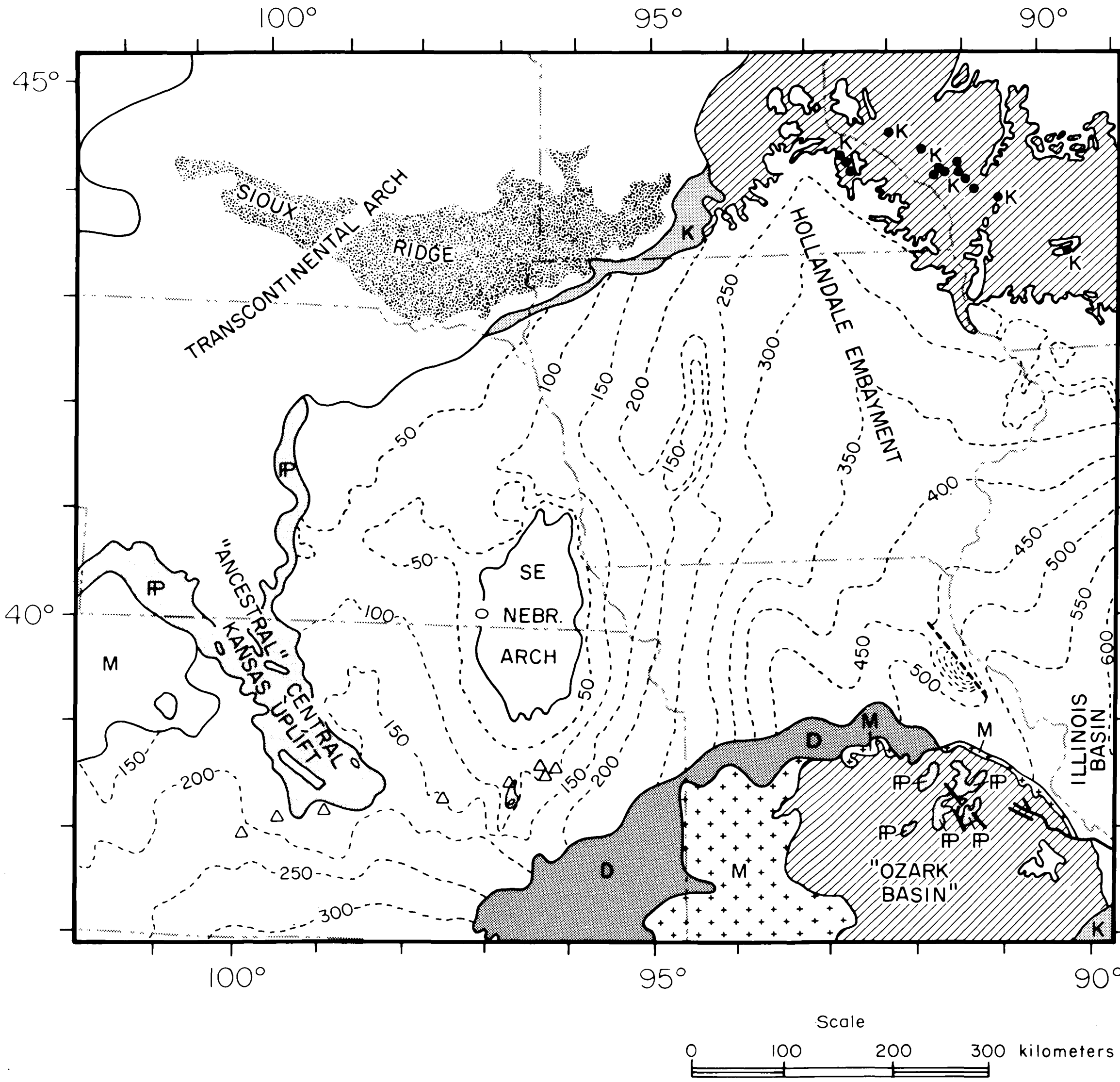


Figure 3. Isopach map of the Sauk Sequence, excluding the Mt. Simon Sandstone and correlative units. Contour interval is 50 m. Triangles denote paleotopographic highs where Precambrian rocks are directly overlain by post-Sauk strata. From Bunker et al. (1988).

rates of deformation in cratonic fault systems. Papers by Witzke and Kolata, and Bunker (this guidebook) indicate that changes in the regional structural grain between the Sauk-Tippecanoe and Tippecanoe-Kaskaskia sequences actually occurred during the early phases of the succeeding episode of predominantly marine deposition. Although the broad outlines of this synthesis were originated many years ago by Wallace Lee (1943, 1946, 1956) at the Kansas Geological Survey, they still are not widely appreciated by many geologists. They are briefly summarized here to help frame some of the themes that we hope to develop for you during this field conference.

SAUK SEQUENCE

Figure 3 is an isopach map of the upper part of Sloss' (1963) Sauk Sequence (Upper Cambrian-Lower Ordovician). The basal

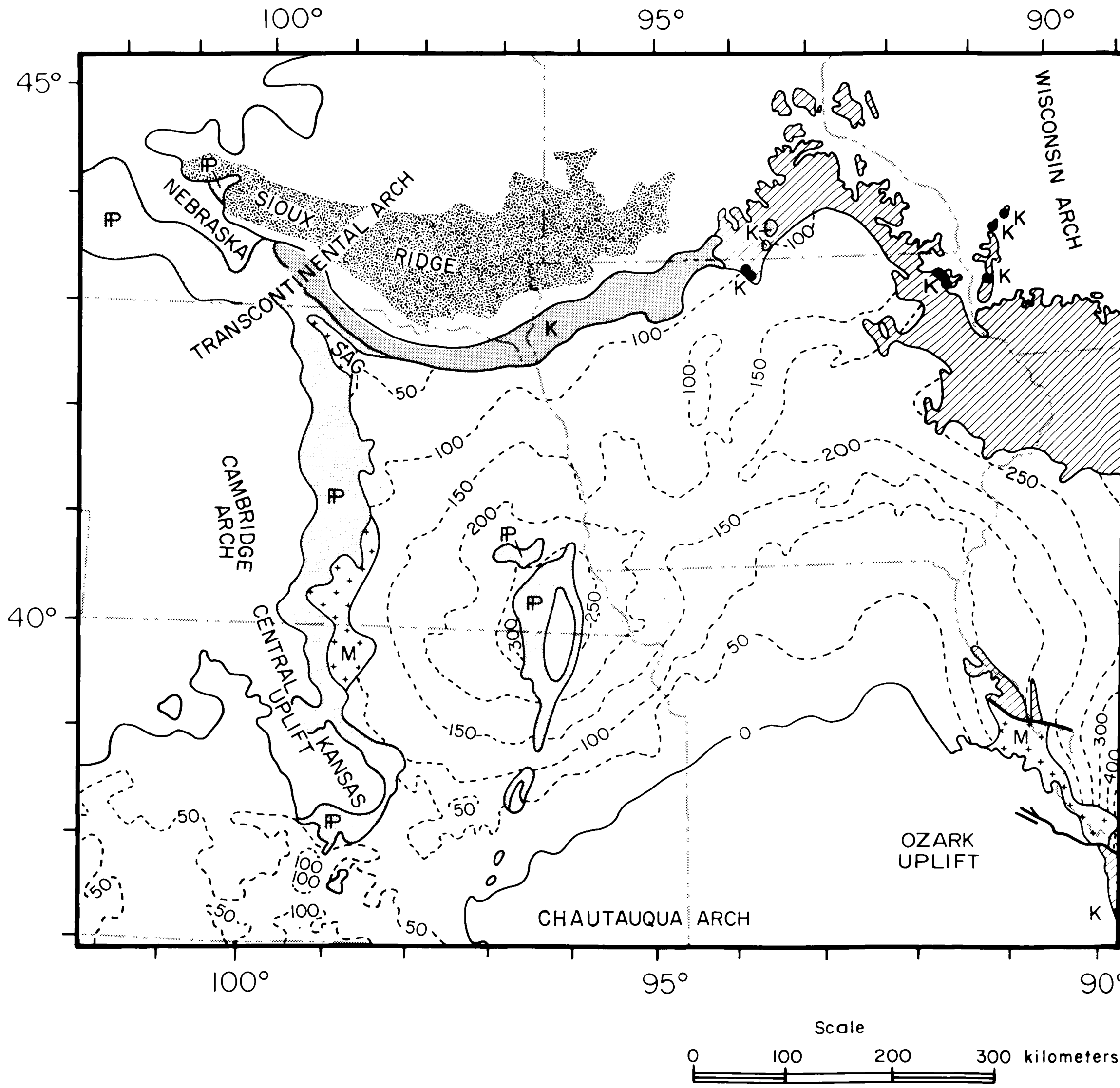


Figure 4. Isopach map of the Tippecanoe Sequence, excluding the St. Peter Sandstone and correlative units. Contour interval is 50 m. From Bunker et al. (1988).

sandstones (Mt. Simon/Lamotte/Reagan) are excluded here because of interpretive difficulties distinguishing them from underlying Keweenawan sedimentary rocks along the Midcontinent Rift System. The major feature controlling upper Sauk depositional patterns in eastern Iowa is the south-southeastward plunging trough named the Hollandale Embayment. The paper by McKay in this guidebook presents information on Upper Cambrian stratigraphy and deposition in the eastern portion of the Hollandale Embayment, and other areas. Please note the thinning of Sauk strata toward the feature identified as the Southeast Nebraska Arch.

TIPPECANOE SEQUENCE

Figure 4 is an isopach map of Sloss' (1963) Tippecanoe Sequence (Middle Ordovician-Silurian). Comparison between Figures 3 and 4 reveals that the depocenters for Sauk and Tippecanoe strata have remarkably different configurations. The Tippecanoe

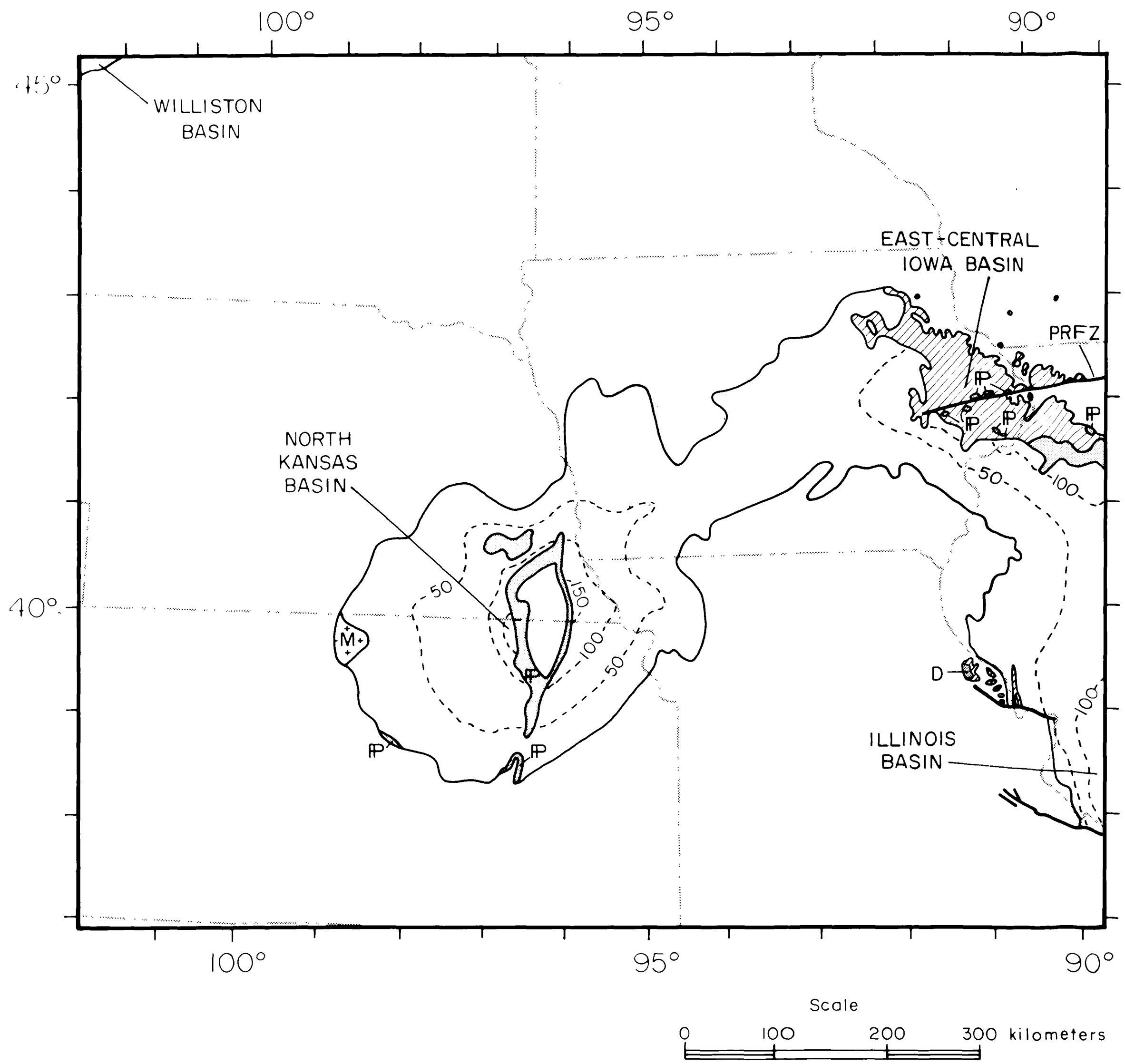


Figure 5. Isopach map of the Silurian System. Contour interval is 50 m. PRFZ - Plum River Fault Zone. From Bunker et al. (1988).

Sequence of eastern Iowa accumulated in the eastward-plunging trough termed the East-Central Iowa Basin (Fig. 1). Note also that the area of the Southeast Nebraska Arch (Fig. 3) became a regional depocenter during Tippecanoe sedimentation. Silurian carbonate rock successions (Fig. 5) in the central midcontintent region are confined to the interior portions of these Tippecanoe depocenters, namely the East-Central Iowa and North Kansas basins (Bunker et al., 1988). Witzke (1981) compared the Silurian stratigraphy in the two basins, and reported that basinal subsidence was initiated earlier in the East-Central Iowa Basin. The Ordovician sedimentary history of these two basins is addressed in the paper by Witzke and Kolata in this guidebook.

KASKASKIA SEQUENCE

Regional isopach patterns of Sloss' (1963) Kaskaskia Sequence (Devonian-Mississippian) are influenced by a confusing combination of

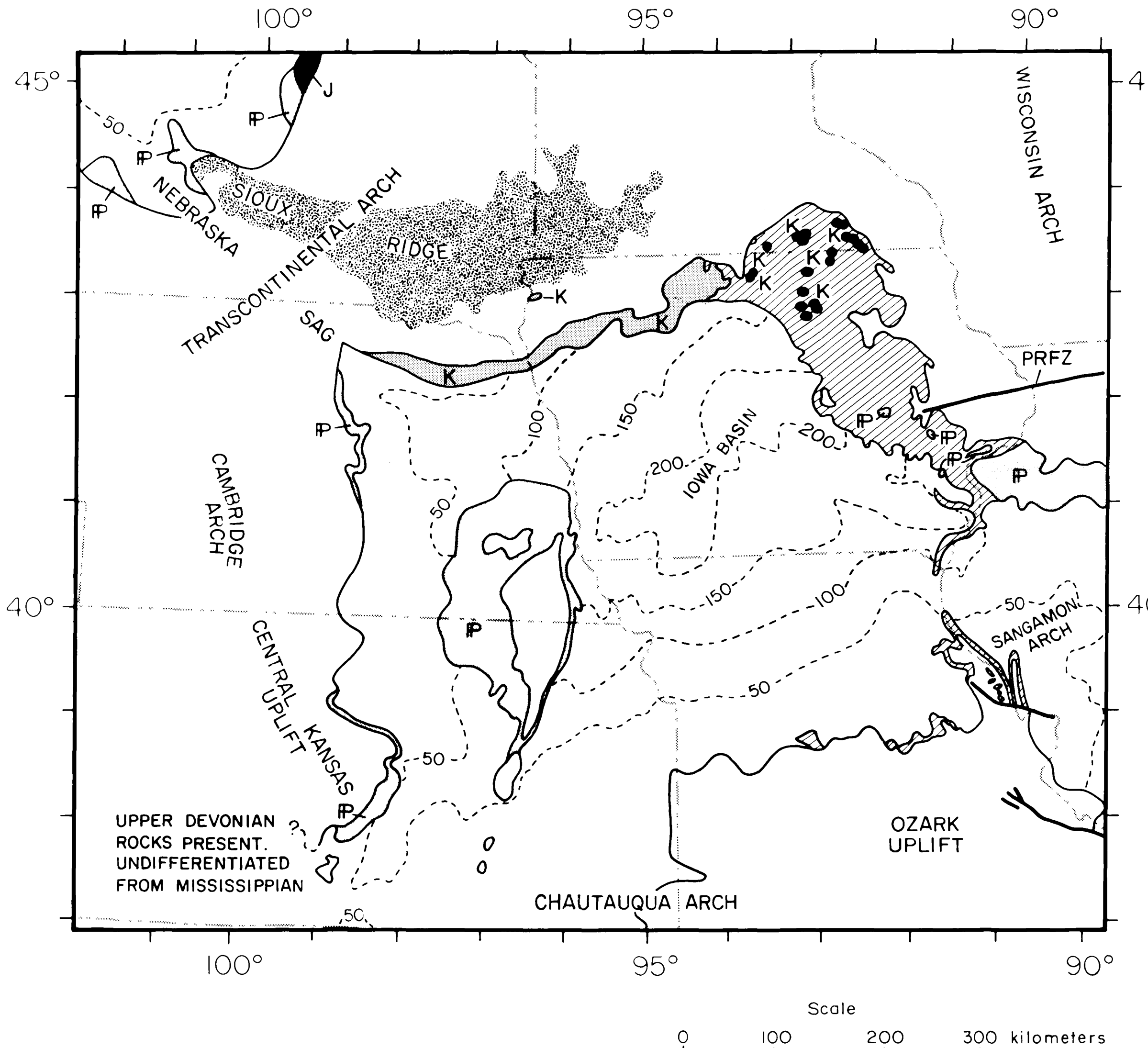


Figure 6. Isopach map of the Devonian System. Contour interval is 50 m. PRFZ - Plum River Fault Zone. From Bunker et al. (1988).

depositional patterns and by a major episode of tectonism and erosion prior to the deposition of the succeeding Absaroka Sequence (Pennsylvanian). An isopach map of the Devonian System (Fig. 6) illustrates changes in depositional patterns. As discussed by Bunker et al. (1985), Witzke et al. (1988), and the paper by Bunker in this guidebook, earliest Middle Devonian deposition was influenced by a Tippecanoe-like pattern of subsidence, but also included asymmetric half-graben basinal development along the Plum River Fault Zone (Fig. 1). During Middle to Late Devonian deposition in the region, the axis connecting the former areas of the North Kansas and East-Central Iowa basins merged into a single elongate basin, termed the Iowa Basin (Fig. 6). Witzke et al (1988) and the paper by Bunker in this guidebook show that the rocks in the center of the Iowa Basin consistently accumulated in shallower marine settings than their correlates to the southeast. The Iowa Basin evidently reflects the southeastward progradation of a shallow marine carbonate shelf outward from the

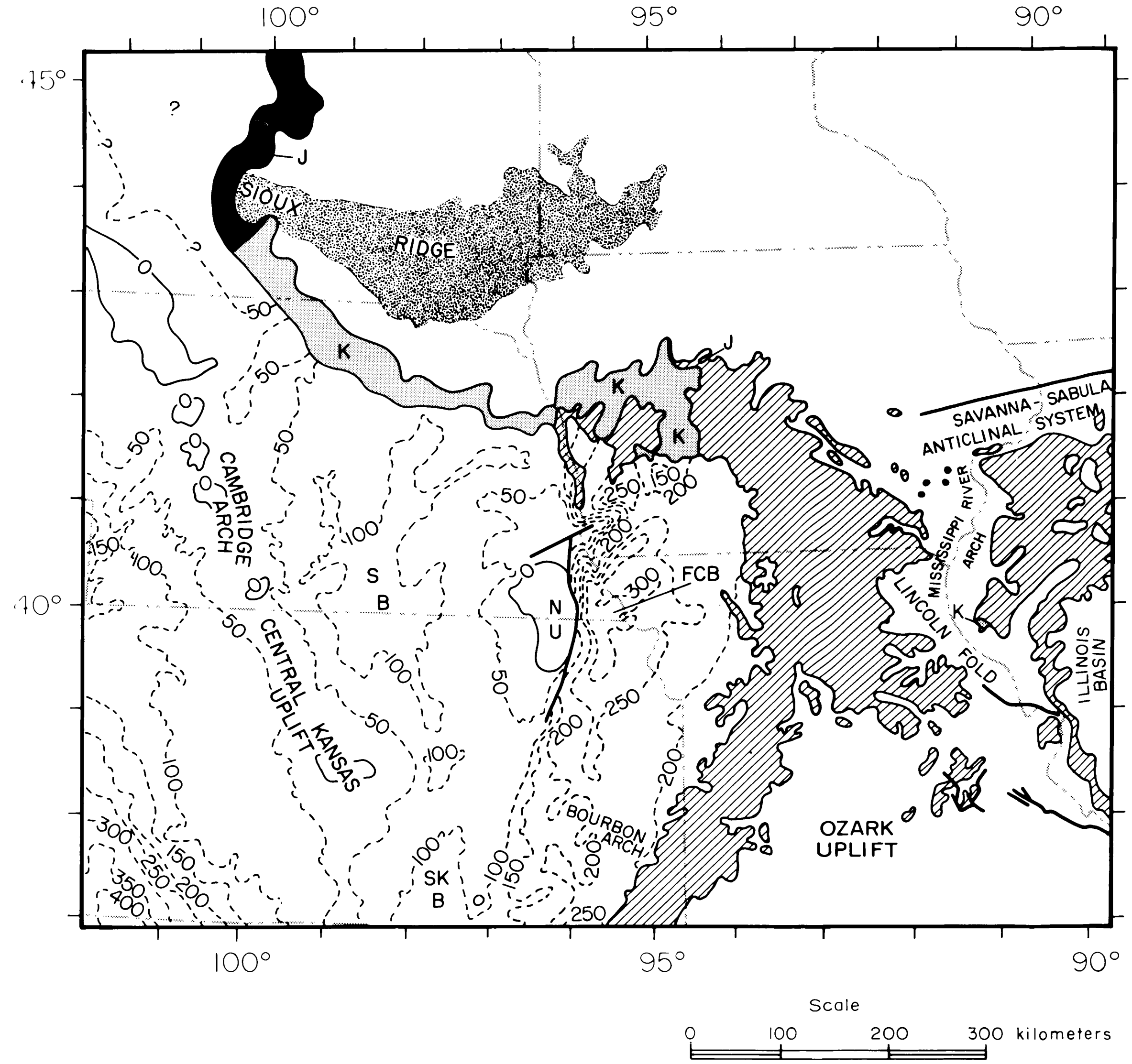
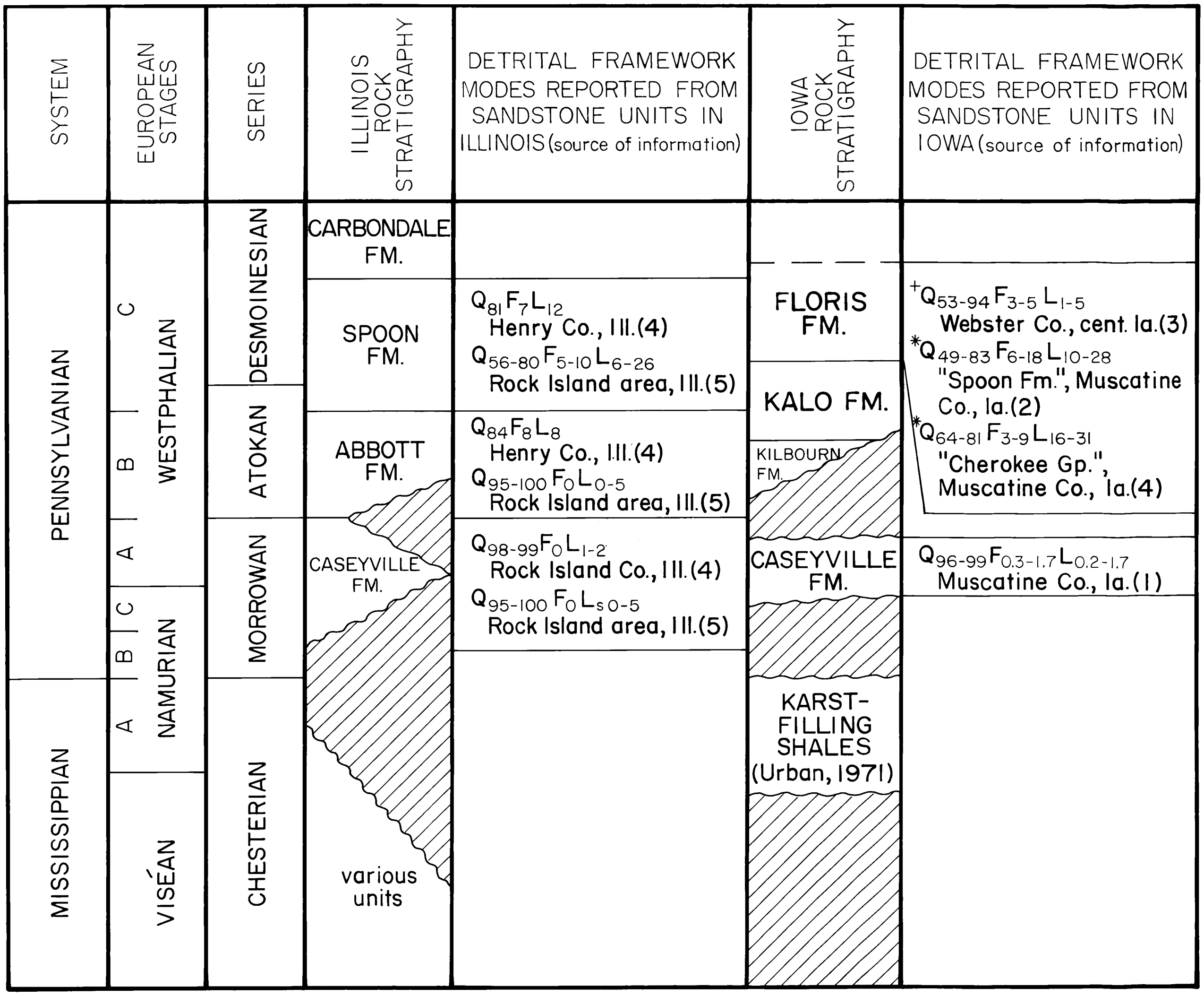


Figure 7. Isopach map of Morrowan, Atokan, and Desmoinesian rocks, including basal Missourian units. Contour interval is 50 m. FCB - Forest City Basin, SB - Salina Basin, NU - Nemaha Uplift, SKB - Sedgewick Basin, HE - Hugoton Embayment. From Bunker et al. (1988).

Transcontinental Arch. This basin was thus not a bathymetric low like the East-Central Iowa Basin. The paper by Bunker in this guidebook discusses the influence of sea level eustacy on cyclic carbonate deposition during the progradation of this southeast-facing shelf. At STOP 6, field trip participants will have a chance to observe deformational features in the Middle Devonian Cedar Valley Group that Ludvigson (1988) has interpreted as evidence that the Plum River Fault Zone was active during the deposition of these mid-Paleozoic rocks.

ABSAROKA SEQUENCE

Prior to the deposition of Pennsylvanian rocks in the midcontinent region, a major structural reorganization occurred in the craton, and hundreds of meters of earlier Paleozoic strata were eroded from areas of regional uplift. In the field trip area, Paleozoic strata were bevelled to



Sources of information on detrital framework modes:

1) Fitzgerald, 1977, p.35
2) Fitzgerald, 1977, p.48
3) Burrgraf *et al.*, 1981, p.43
4) Anderson *et al.*, 1982, p. 16-17
5) Isbell *et al.*, 1984, p.490

+Correlation suggested in Ravn *et al.*, 1984, p.28
*Correlation suggested in Ravn *et al.*, 1984, p.7

Figure 8. **Stratigraphy of the basal Pennsylvanian rocks in Iowa and Illinois, with QFL (quartz-feldspar-lithic grains) serial designations, based on petrographic studies in Iowa and Illinois. From Ludvigson (1985).**

the northeast, and local uplift occurred along the Savanna-Sabula Anticlinal System to the south of the Plum River Fault Zone (Fig. 7; Bunker et al., 1985). To the southwest, the fault-bounded Nemaha Uplift developed in the areas formerly occupied by the Southeast Nebraska Arch (Fig. 3), the North Kansas Basin (Fig. 5), and the Iowa Basin (Fig. 6).

The structural relationships of Pennsylvanian rocks straddling the Plum River Fault Zone (see STOP 5) clearly show that their deposition post-dates most of the vertical movement along the fault system. Ludvigson (1988, this guidebook, see Sunday STOP 2) suggested that the ubiquitous ferric oxide and calcite cements found in the Plum River Fault Zone were precipitated from meteoric waters that infiltrated from the ancient sub-Pennsylvanian erosion surface.

Pennsylvanian rocks in the central Midcontinent region accumulated in sedimentary basins whose configurations were strikingly dissimilar to earlier Paleozoic depocenters. In the Upper Mississippi Valley region, up to 100 feet (30 m) of Early Pennsylvanian (Morrowan) siliciclastic sediments of the Caseyville Formation accumulated in the Quad Cities Illinois-Iowa areas, immediately to the south of the Savanna-Sabula Anticlinal System (Fig. 7). As noted by Bunker et al. (1985), the presence of these rocks seems to be at odds with conceptions of a long-lived Mississippi River Arch. The Caseyville in this area is physically separated from its closest known correlates in southern Illinois by over 275 km. This small, isolated, apparently nonmarine basin remains enigmatic in many respects (Ludvigson and Swett, 1987), although Bunker et al. (1985) suggested that it may be related to local tectonism related to uplift along the Savanna-Sabula Anticlinal System.

A substantial body of relatively new data has been assembled on the framework grain mineralogy of the Pennsylvanian rocks of the Upper Mississippi Valley. Fitzgerald (1977, 1985), Anderson et al. (1982), Isbell et al. (1984), Isbell (1985), Ludvigson (1985), and Ludvigson and Swett (1987) all noted that sandstones in the Caseyville Formation are quartzarenites interpreted to be composed of sediments recycled from older Paleozoic rocks (Fig. 8). Sandstones contained in the overlying Atokan and Desmoinesian units, however, are conspicuously micaceous feldspathic litharenites (ibid.) with a major component of metamorphic lithic detritus (Fig. 8). These sediments are interpreted to have been derived from a freshly exposed crystalline rock terrane, possibly the Canadian Shield or the Northern Appalachian region (Ludvigson, 1985; Ludvigson and Swett, 1987). Pennsylvanian sandstones straddling the Plum River Fault Zone were analyzed by Ludvigson (1985), and interpreted to include remnants of both the Caseyville and younger units on the basis of their framework grain mineralogy. Corroborating evidence for the Morrowan age of the quartz-pebble conglomeratic sandstones along the Plum River Fault Zone is presented by Nations in this guidebook. We will be visiting exposures of these rocks at STOP 5 and Sunday STOP 2.

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PROTEROZOIC GEOLOGY AND GEOLOGIC HISTORY OF THE AREA AROUND THE PLUM RIVER FAULT ZONE, EASTERN IOWA AND NORTHWESTERN ILLINOIS

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INTRODUCTION

The Plum River Fault Zone was defined and mapped in Paleozoic rocks of eastern Iowa and northwestern Illinois. In that Era it was apparently most active during the Ordovician through Pennsylvanian period when it was involved in the subsidence of the East-Central Iowa Basin (Bunker et al., 1985). At that time the Paleozoic section included approximately 2800 feet (850 m) of sedimentary rocks, dominated by a 1000 foot-thick (300 m) basal fluvioclastic sequence overlain by coarse-to fine-grained marine clastics and carbonates. This sequence represented only a thin skin on a crust that totaled about 150,000 feet (45,000 m) in thickness. Also, the continental margin lay about 700 miles (1100 km) to the east, where the Taconic, Acadian, and Alleghenian orogenies (cumulatively known as the Appalachian Orogeny) were in progress, and about 500 miles (800 km) to the south. It would be logical to assume that structural activity in the upper 1.9% of the crust in the interior of a craton, 700 miles (1100 km) away from an active structural margin, probably represents reactivation of pre-existing, more deeply-rooted structural features. These pre-existing structural features would have been developed in the Proterozoic basement that underlies the Phanerozoic sediments in the area of the Plum River Fault Zone.

The basement rocks in the eastern Iowa and northern Illinois area were initially accreted to the Archean age, greater than 2500 Ma (million years), Superior Craton during the Penokean Orogeny, about 1880 to 1790 million years ago. They included island arc sequences, dominated by mafic to intermediate composition volcanic and plutonic rocks and associated sediments, oceanic crust and associated sediments, and late-stage felsic plutons. These rocks were later modified by the intrusion of anatectic felsic plutons during episodes of anorogenic igneous activity about 1760 Ma, 1450 Ma, and 1360 Ma ago, continental rifting and passive margin sedimentation between about 1760 to 1650 Ma and about 1350 Ma ago, and orogenic stresses about 1650 Ma and 1250-1000 Ma ago.

WILSON CYCLES AND THE PRECAMBRIAN HISTORY OF THE MIDCONTINENT

Wilson Cycles were originally described by J. Tuzo Wilson (1963) as the opening and closing of an ocean basin by plate tectonic processes. This definition was expanded by Nance et al. (1988) and Hoffman (1988a) to include anorogenic igneous and extensional tectonic activity prior to the opening of the oceans. From their definitions, a typical Wilson Cycle can be characterized by three litho-tectonic periods that I call (1) anorogeny, (2) taphrogeny, and (3) orogeny (Fig. 1). Rocks representing each of the three periods are not always observable. Individual units may be erosionally removed or modified or obscured by subsequent cycles, and specific lithologies and volumes of rock produced may vary from cycle to cycle.

Anorogeny

The anorogenic period begins with an assembled supercontinent that includes all or most of the major continental crustal blocks surrounded by an external ocean. This assembly of continental crust acts as a thermal insulator,

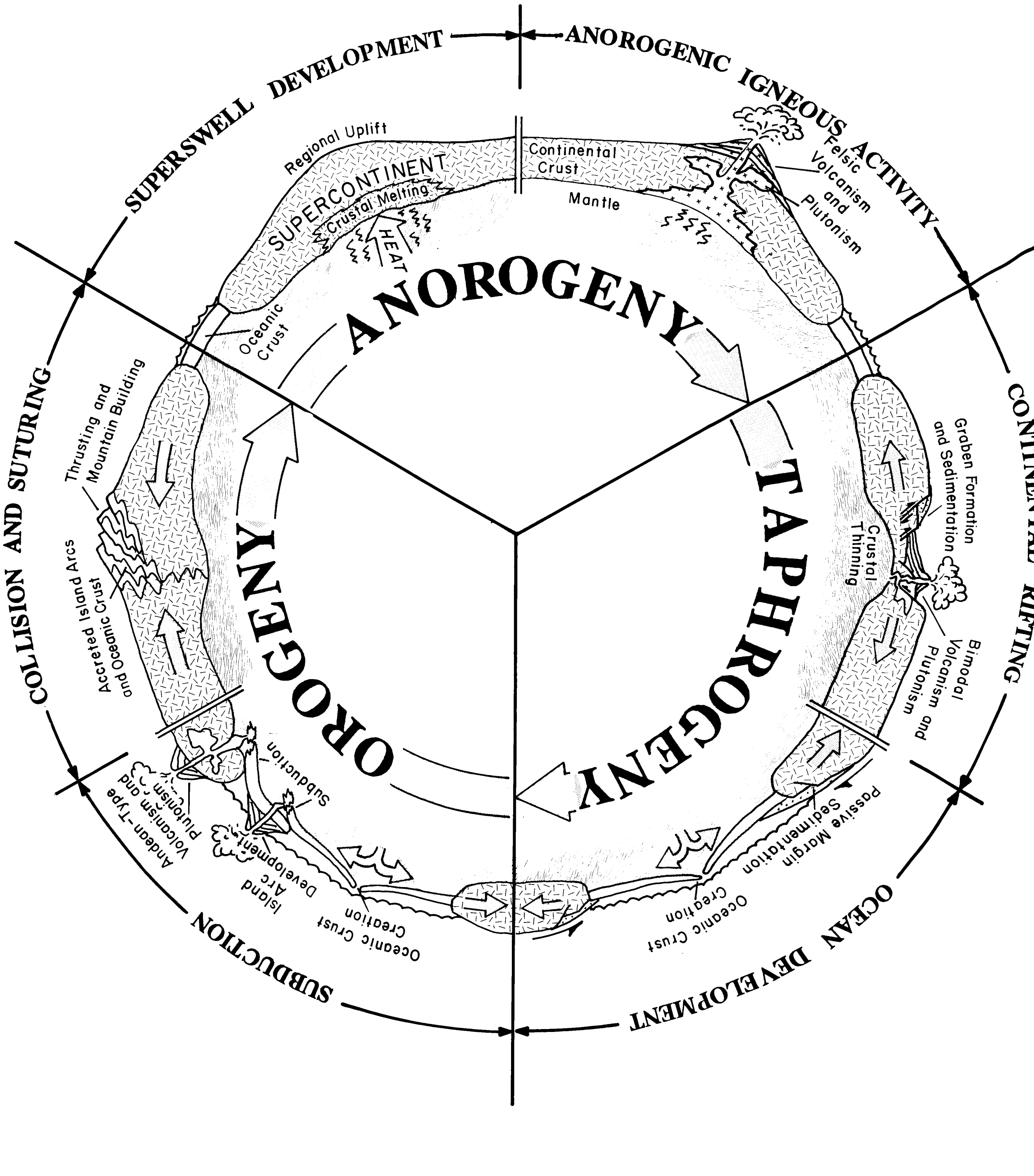

Figure 1. Components of a complete Wilson Cycle and some characteristic geological processes.

greatly diminishing the ability of the mantle to dissipate the heat that it convects to the crust/mantle boundary (Hoffman, 1988a), heat that is most effectively dissipated at mid-ocean spreading centers. The accumulating heat induces partial melting in the lower crust, crustal doming, and ultimately so-called "anorogenic" volcanism and plutonism. These anorogenic magmas are dominated by subalkalic and marginally paraluminous granites and their extrusive equivalents (Anderson, 1983).

Taphrogeny

The doming and elevated heat flow eventually leads to the initiation of rifting within the supercontinent, rifting which may reach fruition with the opening of one or more interior ocean basins. This is the taphrogenic period and is initially characterized by production of a suite mafic dikes and bimodal volcanic and plutonic rocks in the areas of active rifting. Early rift volcanic rocks are commonly interbedded with immature, locally-derived clastic sedimentary sequences, eventually giving way to more mature sedimentary sequences. The new oceans develop with the formation of oceanic crust, generated at spreading centers as the severed continental crustal masses move away from one another. Mature passive margin sediments frequently accumulate on the foundered rift margins at this time.

Orogeny

The interior ocean basins continue to grow, with detached continental blocks drifting apart until the newly-formed oceanic crust has sufficient time to cool and increase in density, eventually gravitationally detaching from the continental crust and subducting beneath it, or beneath oceanic crust near the continental margin. This marks the beginning of the third phase of the Wilson Cycle, the orogenic phase. Driven by gravity sliding from the thermally elevated mid-ocean spreading ridge, the oceanic crust is consumed by subduction and recycled back into the mantle, moving the continental crustal blocks back together and ultimately closing the oceans. The melting of water-saturated oceanic crust and associated sediments in the subduction zone leads to intermediate composition volcanism and plutonism along the active continental margin, or in island arcs. New island arcs, associated volcanogenic sediments, and slices of the oceanic crust and associated sediments are frequently caught between colliding continental crustal blocks and welded to the continent. This accretionary process accounts for about 90% of the rocks of the Precambrian craton of North America. These newly accreted terranes, passive margin sediments, and associated pre-existing rocks usually are subsequently deformed and altered by the compressive forces generated by the collisions of island arcs and continents.

PROTEROZOIC HISTORY OF THE PLUM RIVER FAULT ZONE REGION

Proterozoic (2500-570 Ma) rocks constitute about 70% of the Precambrian craton of North America. They appear to be the product of six Wilson Cycles (Table 1); the Huronian Cycle (2500-2200 Ma), the Penokean/Trans-Hudson Cycle (2200-1790 Ma), the Mazatzal Cycle (1790-1500 Ma), the Grenville Cycle (1500-1000 Ma), the Avalonian Cycle (950-600 Ma), and the Iapetan Cycle (600-300 Ma). The primary terrane in which the Plum River Fault Zone developed was apparantly accreted to the Archean Superior Craton during the Penokean Orogeny (1880-1790 Ma). This period was the most active period of terrane accretion in the Proterozoic history of the North American Craton (Nelson and DePaolo, 1985) . Newly-formed island arcs and oceanic crust were welded onto all sides of the older Superior Craton; along the southern margin the Penokean Volcanic Belt (called the Wisconsin Magmatic Terrane by Sims, 1985); the Trans-Hudson Belt (Hoffman, 1981) (previously referred to as the Churchill Belt) on the west; the Laborador Belt on the north; and apparently also on the east, as seen in the rocks of the Otish and Mistassini basins of Quebec (Hoffman, 1988b). Subsequently, most of terrane accreted at that time along the eastern margin was removed by subsequent rifting, most notably during the Grenville Cycle.

Table 1. Late Archean and Proterozoic Wilson Cycles identified in North America and their component periods.

CYCLE	AGE (Ma)	ANOROGENY	TAPHROGENY	OROGENY
	4600	**ARCHEAN**		
Lumpy Lake	3300	????		
	3200		????	
	3000			Lumpy Lake
Kenoran	2900	????		
	2800		????	
	2750-2680			Kenoran
	2500	**PROTEROZOIC**		
Huronian	2500	Algoman		
	2400-2200		Huronian	
	2200			????
Penokean/ Trans-Hudson	2100	????		
	2000-1950		Nipissing	
	1950-1760			Penokean/ Trans-Hudson
Mazatzal	1780-1700	Fox River Valley		
	1760-1680		Baraboo	
	1650-1630			Mazatzal
Grenville	1500-1310	Wolf River/ St. Francois		
	1340		Sibley	
	1250-1000			Grenville/ Llano
Avalonian	1000	????		
	950-850		????	
	850			Avalonian
Iapetus	820-730	Blue Ridge		
	650-600		Iapetus	
	570	**PALEOZOIC**		
	500-300			Appalachian/ Ouachata

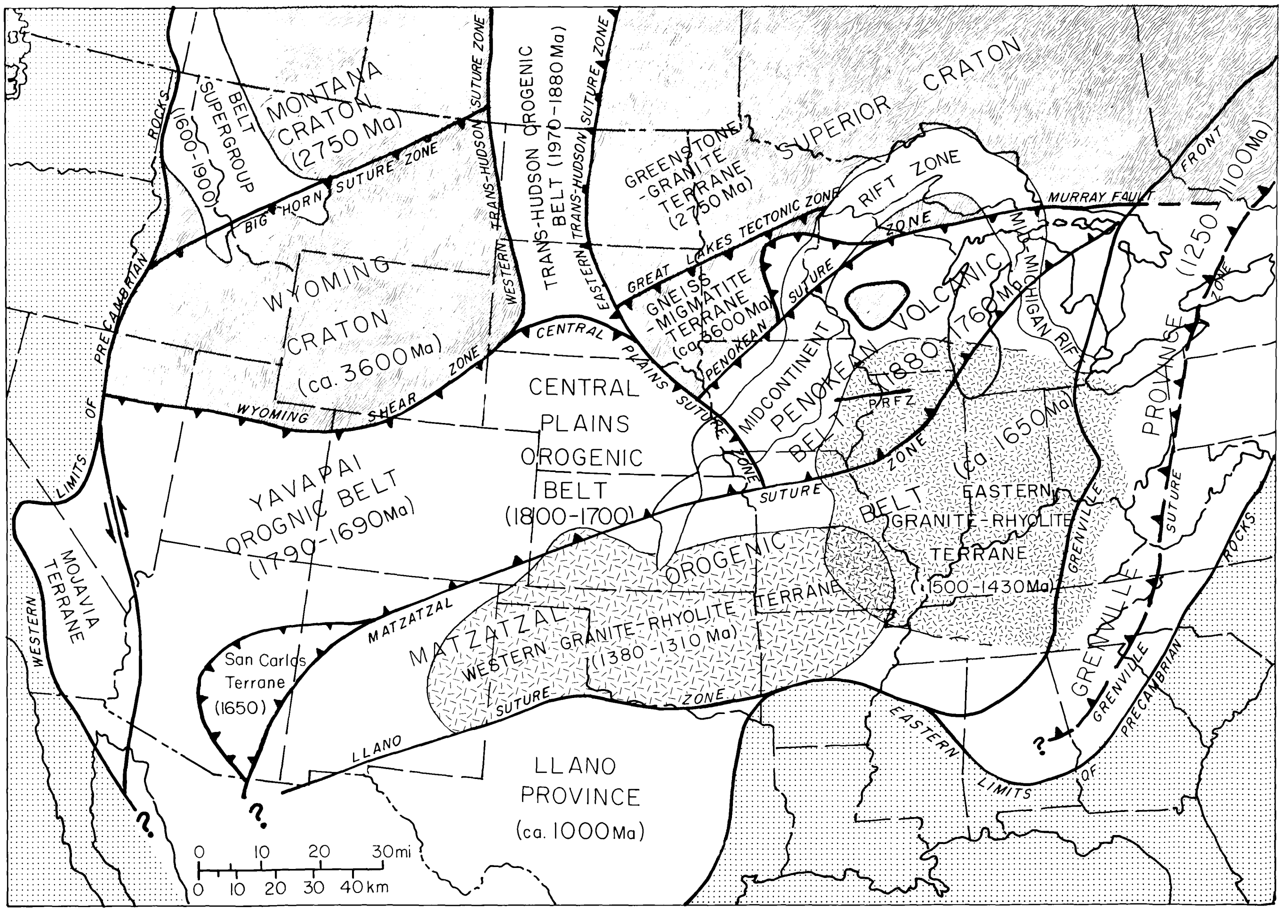

Figure 2. Precambrian terranes of central North America.

Penokean Cycle

The Penokean Volcanic Belt (Fig. 2) is best studied in central Wisconsin where it is exposed between the Niagara Fault Zone, which represents its zone of suturing to the Archean Superior Craton to the north, and the Phanerozoic outcrop belt of southern Wisconsin. In this area it is a complex of island arcs and associated rocks representing at least two periods of subduction and arc collision (LaBarge and Myers, 1984), and including at least one exotic Archean terrane (Van Schmus and Anderson, 1977). While the Penokean Volcanic Belt has been extensively studied in its central Wisconsin outcrop belt, little is known about it in the subsurface. Its suture zone with the Superior Craton has been interpreted to extend southwest from western Wisconsin, across southeastern Minnesota and northwestern Iowa to its contact with the younger Central Plains Orogenic Belt by Anderson and Black (1983), Van Schmus and Bickford (1986, 1987), Sims (1985), and others. Recently Southwick et al. (1988) and Wunderman (1988) have suggested that the suture may continue westward from Wisconsin into central Minnesota and then trend sharply southward before joining the southwest trend as previously proposed in northwest Iowa. The southern extent of the Penokean Volcanic Belt is not known, but it has been proposed, based on interpretation of gravity and magnetic data, to be bounded by the Mazatzal Suture Zone (also known as the Baraboo Suture Zone) by Anderson and Ludvigson (1987) and Van Schmus and Bickford (1986, 1987). This zone trends southwesterly

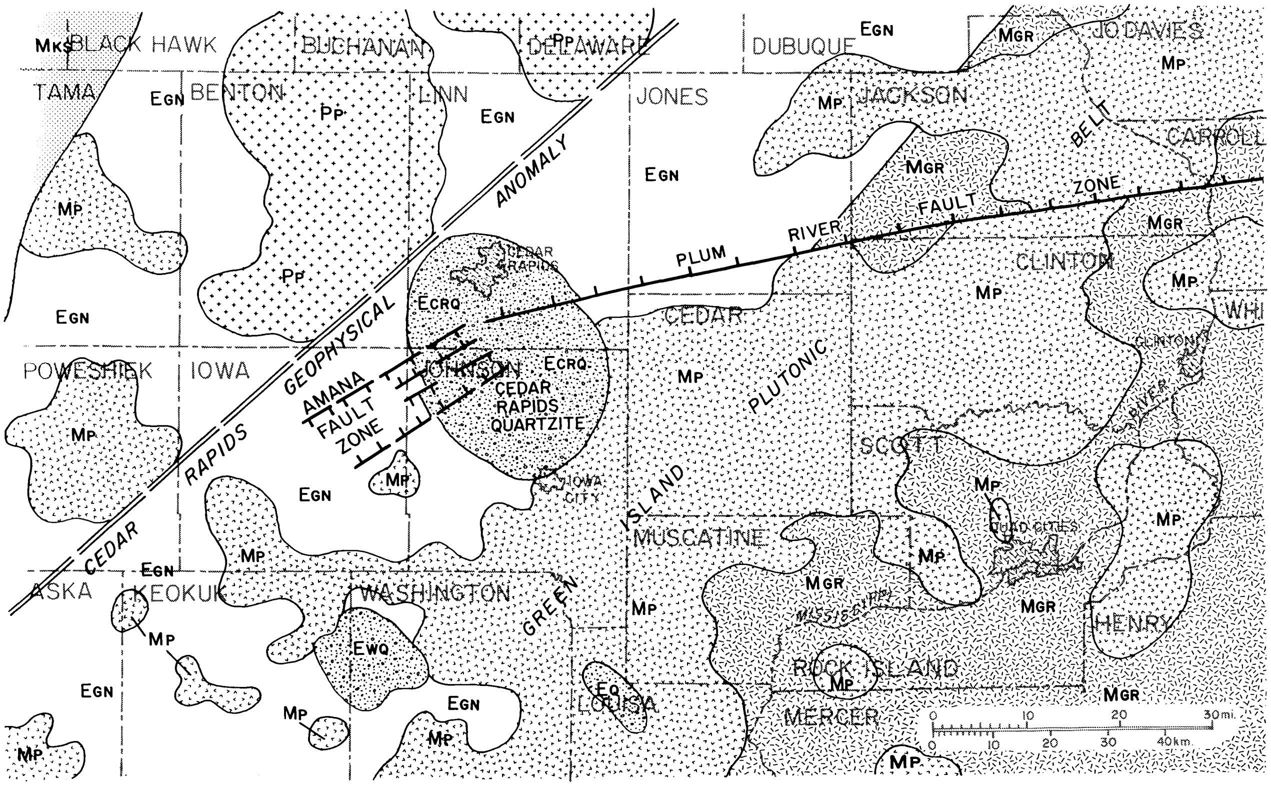

Figure 3. Precambrian geology in the area of the Plum River Fault Zone (Anderson, 1988a). Egn-Early Proterozoic (Penokean) gneiss-dominated terrane, Eq-Early Proterozoic (Baraboo Interval) quartzite, Ecrq-Early Proterozoic (Baraboo Interval) Cedar Rapids Quartzite, Ewq-Early Proterozoic (Baraboo Interval) Washington County Quartzite, Mgr-Middle Proterozoic (Wolf River Period) granite and rhyolite terrane, Mp-Middle Proterozoic (Wolf River Period) felsic pluton, Mks-Middle Proterozoic (Keweenawan) sedimentary rocks, Pp-Proterozoic plutons of unknown lithology or age.

across Lake Michigan, passing through the Chicago area and crossing from Illinois into Missouri in the Hannibal area, and then continuing nearly due west, encountering the eastern margin of the Central Plains Orogenic Belt somewhere near Chillicothe, Misssouri (see Fig. 2).

Mazatzal Cycle

The Penokean Volcanic Terrane was first modified by the emplacement of a belt of felsic volcanic and plutonic rocks during the Fox River Anorogeny (about 1780 to 1740 Ma ago) of the Mazatzal Cycle. A number of felsic igneous rocks of this age are known in Wisconsin, including the Fox River Valley granites and rhyolites (Van Schmus, 1978), and several plutons of this age have been reported in southern Ontario just east of the Grenville Front by Van Breemen and Davidson (1988), and in Nebraska (Van Schmus and Bickford, 1986). In northwest Iowa, the Hull Rhyolite (keratophyre) has been dated at about 1782 Ma (W.R.. Van Schmus personal communication, 1988) and rocks of similar age have been dated in adjacent South Dakota. The rhyolites (and keratophyres) have apparently been protected from erosional removal by overlying quartzites in the northwestern Iowa area and Wisconsin. It is possible that plutons emplaced during the Fox River Anorogeny exist in the vicinity of the Plum River Fault Zone,

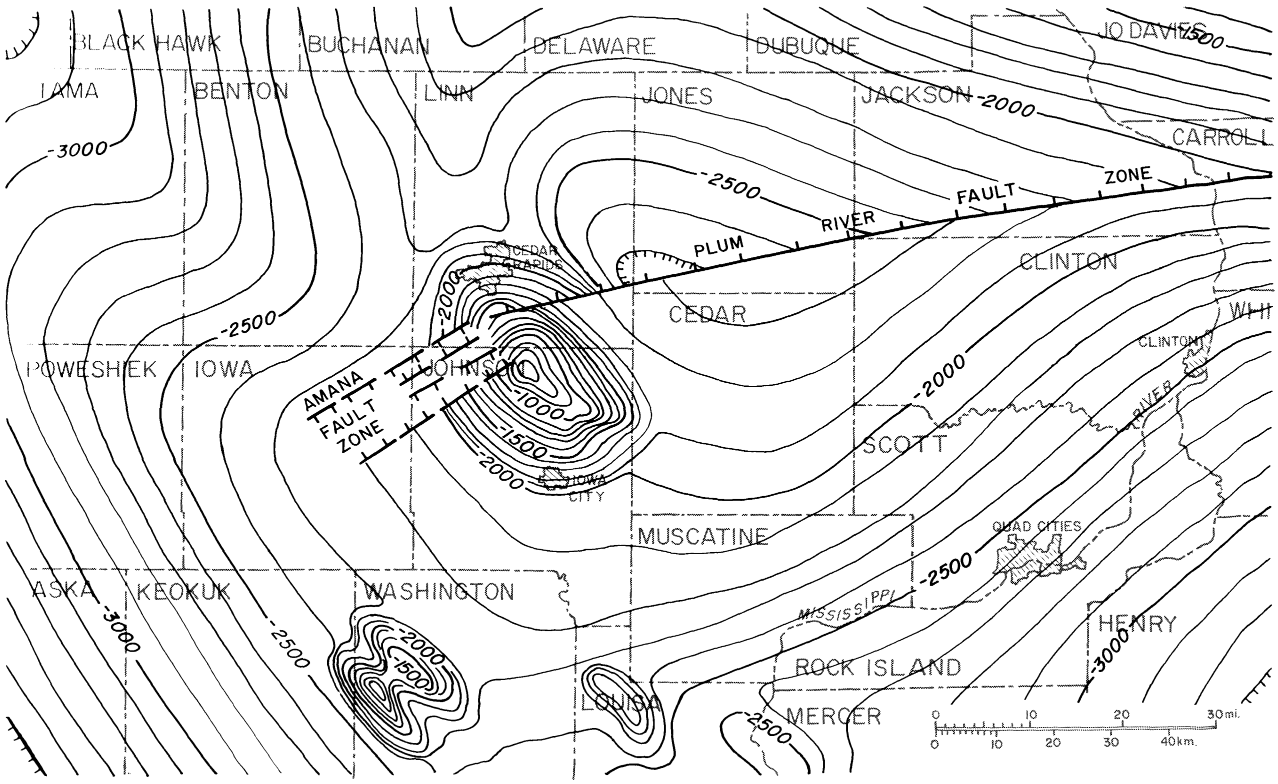

Figure 4. Structure on the Precambrian surface in the area of the Plum River Fault Zone (Anderson, 1988b). Contour interval-100 feet, datum-mean sea level.

although none have yet been identified.

The elevated heat flow that produced isostatic structural uplift and anorogenic volcanism about 1760 Ma was relieved with the successful rifting and ocean development south of the Mazatzal Suture as mapped on Figure 2. A suite of passive margin sediments were apparently deposited during the Mazatzal Taphrogeny along and inboard from the continental margin and are preserved as a discontinuous belt of quartzite-dominated erosional remnants from eastern Wisconsin to New Mexico. In Wisconsin these remnants include the Waterloo, Baraboo, Barron, Flambeau quartzites, and possibly many smaller outliers that have been described by Brown (1986). In Iowa they include the Washington County Quartzite (Anderson and Ludvigson, 1986), the Cedar Rapids Quartzite (Anderson, 1988a), and other probable outliers near Columbus City and Cairo, all in the subsurface of southeastern Iowa, and the Sioux Quartzite exposed in extreme northwest Iowa and adjacent areas of Minnesota and South Dakota. A number of possibly related quartzite bodies have been encountered at the Precambrian surface in wells drilled in southern Nebraska (Sam Treves personal communication, 1987). In New Mexico the Ortega Quartzite described by Soegaard and Erikkson (1985, 1986) was also deposited on this passive margin, as was the Mazatzal Quartzite most recently studied by Conway et al. (1981). These rocks have been interpreted as predominantly fluvial sequences with probable marine intervals near the top of the thicker sequences (Weber, 1981; Dott, 1983). Dott (1983) described the non-marine, quartzite-dominated outliers as "remnants of a vast sedimentary wedge that blanketed at least the southern edge of the post-Penokean craton of Proto-North America." At Baraboo, Wisconsin, a synclinorium preserves a sequence of marine sediments that lie conformably above the basal quartzites. These units are not exposed, but have been sampled by core drilling and were briefly described by Dalziel and Dott (1970). Directly above the Baraboo Quartzite, the Seeley Slate is a banded gray and green slate. It is overlain by the Freedom Formation, a suite of dolomite, slate, chert, and banded iron-formation. Unconformably above the Freedom Formation, the Dake Quartzite was described by Leith (1935) as a coarse-grained quartzite with a high proportion of chlorite and sericite as a matrix. The upper most unit in the synclinorium is the Rowley Creek Slate, a gray quartz-chlorite-sericite slate.

One of the buried Baraboo Interval erosional outliers, the Cedar Rapids Quartzite, lies at the western end of the Plum River Fault Zone (Fig. 3). The Cedar Rapids Quartzite was first encountered during the drilling of a deep well in northeastern Cedar Rapids in 1888. Unfortunately no samples of the rock were preserved, however the driller reported quartzite at the base of the well at about 2150 feet (655 m) below land surface. No subsequent Precambrian wells have been drilled in the area, but recent extrapolative mapping of the Precambrian surface (Anderson, in prep.) by downward continuation of structures mapped on shallower datums by Wahl et al. (1978) has revealed a significant positive structure on the basement surface (Fig. 4), interpreted as a quartzite outlier. The mapped body is elliptical in shape with its major axis trending approximately N45^{o}W and covering an area of about 80 square miles (205 km^{2}), primarily in Johnson and Linn counties. Its total vertical relief on the Precambrian surface is about 1500 feet (455 m), displacing about 1/2 of the cumulative Phanerozic section as is preserved off the structure in that area. The Plum River Fault Zone cuts across the northern 1/4 of the Cedar Rapids Quartzite, with the north side of the fault zone structurally lowered several hundred feet.

A second, poorly defined structure informally known as the "Amana fault zone" (Bunker, this guidebook) apparently intersects the Plum River Fault Zone in the area of the Cedar Rapids Quartzite. Located on Figure 3, the "Amana Fault Zone" trends about N55^{o}E, on strike with the grain of the Penokean basement terrain and parallel to the Penokean Suture Zone, which lies about 200 miles (320 km) to the northwest. Two of a series of eight gravity traverses across the Plum River Fault Zone by Svoboda (1980) crossed the "Amana fault zone" in the area of the Cedar Rapids Quartzite. In 1986 and 1987 Jerpbak (in prep.) collected seismic reflection data in the "Amana fault zone", between the gravity profiles, and near its intersection with the Plum River Fault Zone south of Cedar Rapids. Initial interpretations of the seismic data are consistent with Svoboda's gravity models.

The seismic data collected by Jerpbak across the "Amana fault zone" over the Cedar Rapids Quartzite just south of Cedar Rapids (Fig. 5) show the fault zone to be about 1/2 mile (0.8 km) wide and composed of two faults. The northern most of these faults appears as a zone of disrupted strata about 700 feet (215 m) wide, dipping about 12^{o} to the north, and showing normal displacement, with the surface of the Cedar Rapids Quartzite south of the fault displaced upward about 2 seconds, or about 1000 feet (300 m). The surface of the Cedar Rapids Quartzite north of the fault is approximately horizontal, lying about 1900 feet (580 m) below the land surface. The uplifted block south of the fault is about 2000 feet (610 m) wide and its upper surface apparently dips southward towards the second fault zone. The second (southernmost) of the two fault zones is also about 700 feet (215 m) wide. The Precambrian surface in this fault zone lies about 1400 feet (425 m) below the land surface, and the fault appears to dip only slightly to the north. To the south of this southern fault the surface of the Cedar Rapids Quartzite is upthrown about 700 feet (215 m), and it dips irregularly to the south from the fault. The fault was probably active during the Proterozoic, and fault-related relief was probably present on the surface of the quartzite during the initial Phanerozoic marine transgressions.

It appears that some orogenic activity was occurring during the taphrogenic period of the Matzatzal Cycle. During the Central Plains/Yavapai Orogeny a series of island arcs were accreted to the proto-North American Craton between about 1790 and 1690 Ma. This should probably be considered a very late stage of the Penokean/Trans Hudson Orogeny, although there appears to have been sufficient crustal accumulation at that time to confine mantle heat and begin the anorogenic period.

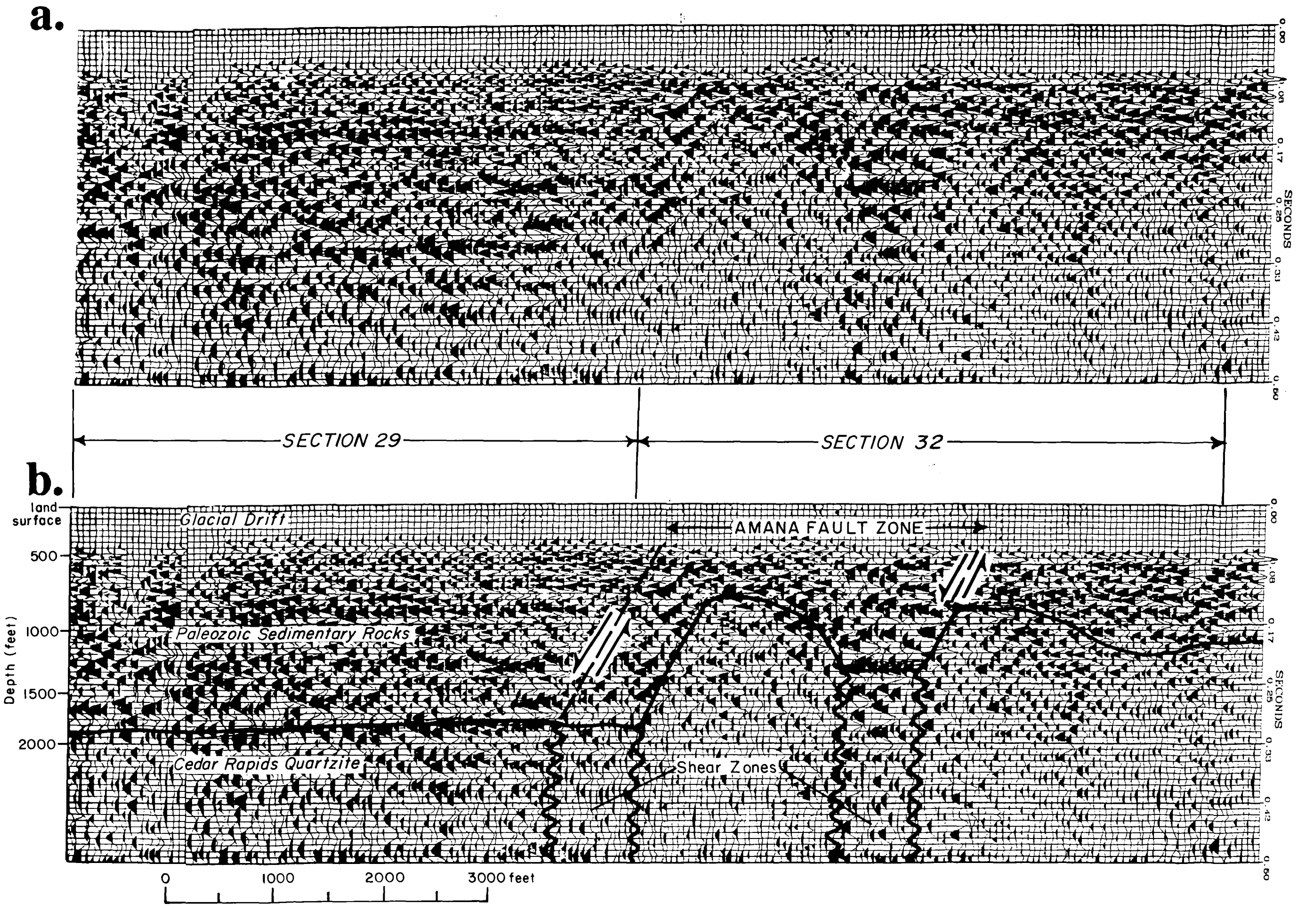

Figure 5. Reflection seismic profile across the "Amana fault zone" and preliminary interpretations (Jerpbak, in prep.). A-uninterpreted section, B-interpreted section.

No rocks associated with the Mazatzal Orogeny are known in the region, possibly because of the nature of the orogeny and distance to the Mazatzal Suture Zone (see Fig. 2), as well as the lack of drill penetrations of overlying Phanerozoic rocks. Metamorphism of clastics associated with the Baraboo, Washington County, and Waterloo quartzites may be the product of the orogeny (Anderson and Ludvigson, 1986).

Grenville Cycle

The Wolf River-St. Francois Anorogeny of the Grenville Cycle led to the emplacement of large volumes of granites and rhyolites during two periods, the Wolf River period (about 1500 to 1430 Ma) and the St. Francois period (about 1380-1310 Ma). During the Wolf River period numerous anorogenic plutons were emplaced along a belt stretching from southern California to Laborador, south of the old Archean terranes. The Wolf River Batholith of east-central Wisconsin is the best-known of these plutons. These plutons and associated extrusive rocks, known as the Eastern Granite-Rhyolite Terrane (Bickford and Van Schmus, 1985), cover older Proterozoic rocks in a large region of the eastern Midcontinent and eastern Iowa. In the area of the Plum River Fault Zone the Green Island Plutonic Belt, trending northeasterly across eastern Iowa and into adjacent Wisconsin, can best be delineated on the aeromagnetic map of Iowa (Zietz et al., 1976). The plutonic belt has been defined by its magnetic anomaly and was named for the location of a drill core near the town of Green Island about 4 miles (6.5 km) north of the Plum River Fault Zone. This drill hole encountered granite at the Precambrian surface that yielded a U-Pb zircon age of 1485 $\pm$10 Ma (Hoppe et al., 1984). In northwest Illinois, west of Rockford, three cores drilled by Consolidated Edison over a positive magnetic anomaly also encountered granite. Samples from one core, UPH-3, yielded a U-Pb zircon age of 1465 $\pm$8 Ma (Hoppe et al., 1984). In northwest Iowa, a core was obtained from a positive magnetic feature called the Quimby Anomaly. The drill encountered a similar granite that yielded a U-Pb zircon age of 1433 $\pm$6. Ma (Van Schmus et al., in press). Many other small, positive magnetic anomalies identified in Iowa and surrounding states are presently interpreted as Wolf River period granitic plutons.

An ancestral structure to the the Plum River Fault Zone apparently played some role in controlling the emplacement of the Green Island Plutonic Belt. The belt trends in a northeasterly direction across southeastern Iowa until it approaches the Plum River Fault Zone just east of Cedar Rapids. It then trends in a more easterly direction paralleling the fault zone for about 30 miles (48 km) before resuming its northeasterly trend and crossing into Wisconsin roughly following the trend of the present-day Apple River.

The second period of anorogenic volcanism, the St. Francois period, is best known from exposures of rhyolite and granite in the St. Francois Montains of southeastern Missouri. The St. Francois exposures are near the eastern end of a wide, east-west trending belt of 1380-1310 Ma felsic igneous rocks, shown in Figure 2 and called the Western Granite Rhyolite Terrane (Bickford and Van Schmus, 1985). At the western end of this belt, numerous felsic igneous rocks are exposed in Colorado and nearby areas including rocks in the Wet Mountains (Thomas et al., 1984) and the Silver Plume and St. Vrain Batholiths (Anderson and Thomas, 1985). No plutons of this age have been encountered or interpreted in eastern Iowa, northern Illinois, or northeastern Missouri, so it is unlikely that this period of anorogenic igneous activity affected the area of the Plum River Fault Zone.

The opening of the Grenville Ocean is best known from the rocks of the Grenville Supergroup in eastern Ontario and adjacent Quebec. These rocks include a series of bi-modal volcanic and plutonic rocks and immature clastics and grade upward into fine-grained marine clastics and finally quartz arenites. Also preserved within this sequence are glaciogenic rocks of the Gowganda Formation. This taphrogenic period has been named the Sibley Taphrogen in order to avoid confusion with the better known Grenville Orogeny. It is named for the other rock sequence associated with the taphrogeny in the Midcontinent, the Sibley Group of the north shore of Lake Superior in the Thunder Bay-Nipigon region of Ontario. The Sibley Group consists of a fining-upward sequence of three clastic-dominated formations with minor carbonate beds.

Orogenic activity along the North American cratonic margin of the Grenville Ocean began about 1250 Ma ago and continued until about 1000 Ma. Extensive mountain building and overthrusting was associated with the orogeny along the Grenville Suture, with the western margin of associated tectonism and metamorphism known as the Grenville Front. At about the same time as the Grenville Orogeny was in its waning stages, a major period of extensional stresses led to the emplacement of an extensive suite of mafic dikes including the Grenville (Kretz et al., 1985) and the Logan (Jones, 1984) dikes and the development of the Midcontinent Rift System about 1100 Ma. ago. The Midcontinent Rift extends for about 1000 miles (1600 km) from eastern Lake Superior to southern Kansas (see Fig. 2), and may also include an eastern arm that extends southward across Michigan. As the rift developed, tens of thousands of feet of mafic volcanic rocks accumulated in central grabens, and were buried by tens of thousands of feet of clastic sediments when volcanism ceased but subsidence continued. Subsequently, major compressional stresses reversed the throw on many of the graben-bounding faults, producing a feature with a central horst of mafic igneous rocks, flanked by deep, clastic-filled basins. It is not presently clear how this period of failed rifting fits into our present understanding of Wilson Cycles. It may be an unusual part of the Grenville Orogeny (as age constraints would argue), or it may be the taphrohgenic period of another cycle that was not fully developed or currently recognized. All of this activity apparently

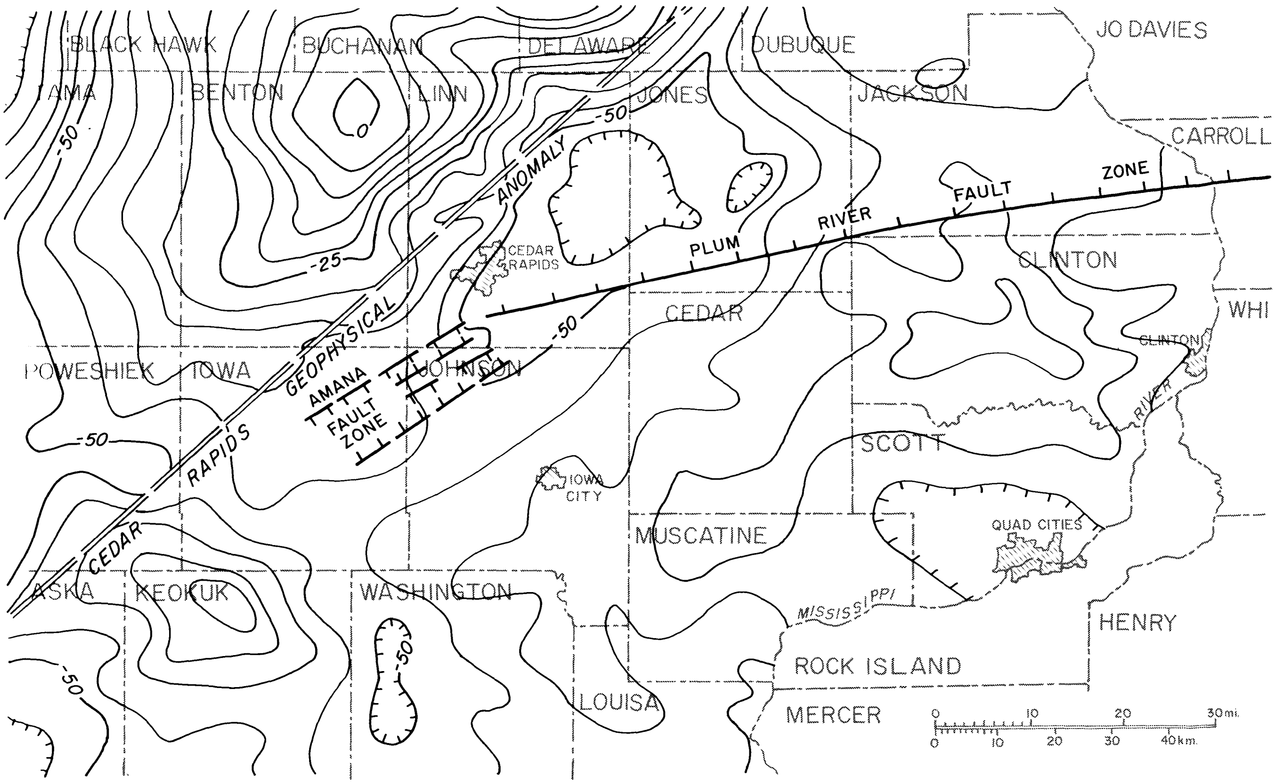

Figure 6. Bouguer gravity anomaly map of the area around the Plum River Fault Zone in Iowa (Anderson, 1981). Contour interval-5 miligals.

occurred from about 1150 Ma to 1000 Ma. In this period of major regional extensional and compressional stresses, it is likely that movements occurred along the Plum River Fault Zone, although the nature and magnitude of these movements cannot be determined with present information.

No rocks associated with the Grenville Cycle are known in the area of the Plum River Fault Zone. However, the region only lies about 450 miles (720 km) to the west of the interpreted Grenville Suture (the western margin of the Grenville Ocean basin), and about the same distance from the Sibley Group of southern Ontario, the unit from which the Sibley Taphrogen is named. So, rocks associated with this part of the cycle may be present but undetected.

Post-Grenville Cycles

Two post-Grenville Cycles have tentatively been identified in the Proterozoic rock record of the North American Craton. Very little is presently known about the Avalonian Cycle. Rocks formed at that time are scattered throughout the Appalachan Mountains, from Tennessee to Canada. The cycle is named for the Avalon Terrane in Newfoundland but also includes the rocks of the Blue Ridge block in the southeast United States and other rocks ranging in age from about 820 Ma to 650 Ma. Rocks of this age are also known from the Atlantic coastal mountains of Europe and Africa.

The Iapetus Cycle begins with the opening of the Iapetus (proto-Atlantic) Ocean about 650 Ma and culminated with the Appalachian (Taconic, Acadian, and Alleghenian) Orogeny during the Early to Late Paleozoic, about 500 to 250 Ma ago. Rocks of the Avalonian Cycle were heavily

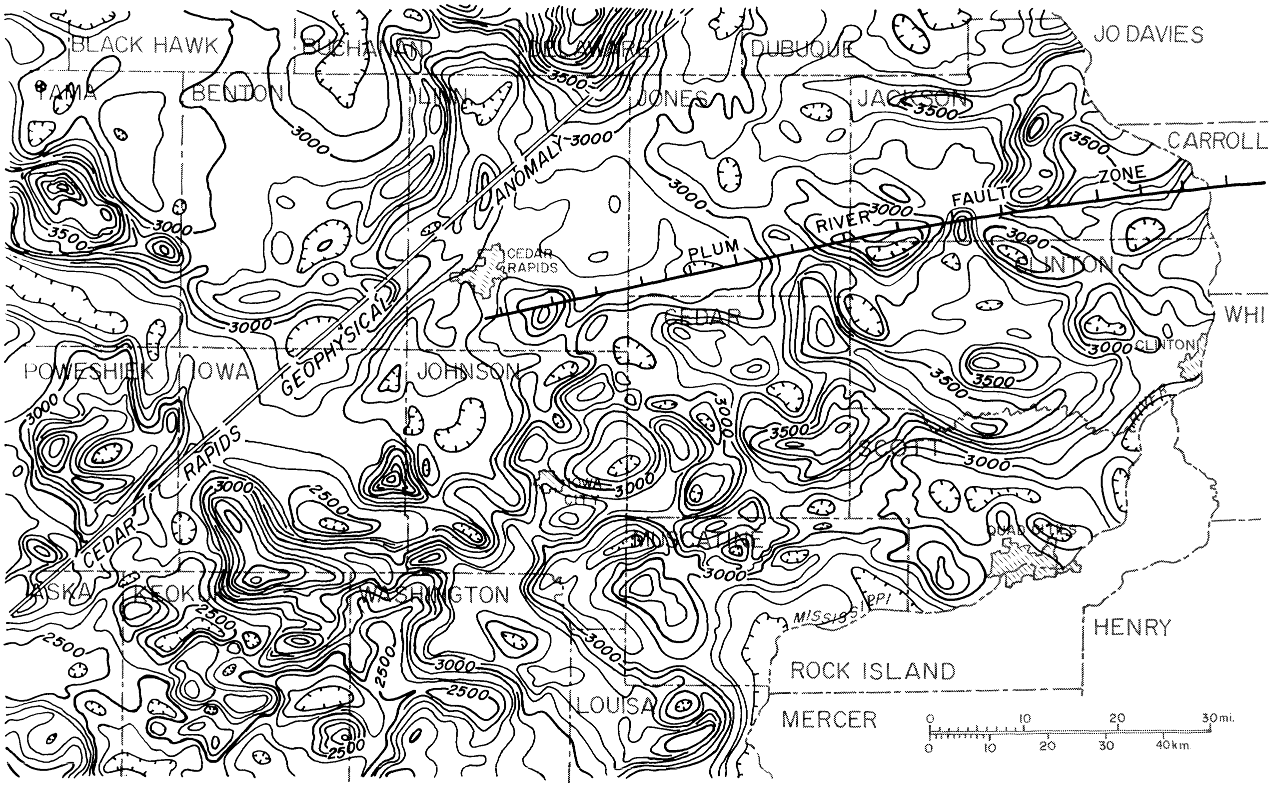

Figure 7. Aeromagnetic anomaly map of the area around the Plum River Fault Zone in Iowa (Zeitz and others, 1976). Contour interval-100 gammas.

overprinted by the Appalachian Orogeny.

No rocks of Avalonian or Iapetus cycles are known west of the Appalachian Mountains. It is unlikely that any post-Grenville Proterozoic rocks exist in the area of the Plum River Fault Zone, however several tectonic reactivations of the zone may have occurred. Paleozoic reactivations along the fault zone were roughly contemporaneous with Paleozoic orogenic events, and were responsible for the displacements that can be seen today.

CONCLUSIONS

Because the Phanerozoic component of the Midcontinent crust represents only about 2% of the total crustal thickness, and the region was at least 500 miles from active continental margins during the Paleozoic, it is likely that any Paleozoic structures in that region represent reactivation of structures that originally formed during the Precambrian. The crust in the area of the Plum River Fault Zone is believed to have formed by accretion of island arc assemblages during the Penokean Orogeny about 1880-1790 Ma ago. The accreted Penokean terrane, the Penokean Volcanic Belt, has a structural grain that appears to trend about N35°E, approximately parallel to its suture with the Superior Craton but quite disparate from the N85°E trend of the Plum River Fault Zone. This discrepancy makes it difficult, but not impossible to reconcile formation of the Plum River Fault Zone ancestral structure with the Penokean Orogeny. The structure does, however, appear to have influenced the emplacement of the Green Island Plutonic Belt about 1485 Ma ago. These considerations bracket the apparent age of origin of the ancestral Plum River Fault Zone structure between about 1880-1485 Ma.

Interpretation and modeling of the Bouguer gravity anomaly (Fig. 6) and aeromagnetic anomaly (Fig. 7) maps of the area suggests that generally only about 1500 feet (450 m) of vertical displacement presently exists at the Precambrian crystalline surface (Fig. 4). This displacement can be explained by net movements during the Phanerozoic. The absence of apparent vertical relief across most of the fault at the end of the Proteozoic does not preclude the possibility of significant Proterozoic movement. Periods of extensive weathering eroded deeply into crustal rocks between about 1760-1650 Ma and 1100-850 Ma and could have leveled considerable structural relief in the area. The exception to this conclusion is the region where the Plum River Fault Zone cuts the Cedar Rapids Quartzite. It is possible that quartzite stood in significant relief, perhaps as much as 1200 feet (360 m), on the Precambrian surface at the time of the initial Phanerozoic marine transgressions.

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STRATIGRAPHY AND LITHOFACIES OF THE DRESBACHIAN (UPPER CAMBRIAN) EAU CLAIRE FORMATION IN THE SUBSURFACE OF EASTERN IOWA

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INTRODUCTION

The Upper Cambrian Eau Claire Formation is named for exposures near the city of Eau Claire, Wisconsin. Its internal stratigraphy and relationship to the overlying Wonewoc Formation and the underlying Mt. Simon Formation have been documented over a five county outcrop region in west-central Wisconsin by several workers (Ostrom, 1966, 1970; Morrison, 1968; Huber, 1975; and Byers, 1978). Subsurface stratigraphic studies of the Eau Claire in the northern Midcontinent include Twenhofel et al., 1935; Workman and Bell, 1948; Buschbach, 1964; Austin, 1970; Lochman-Balk, 1971; Ostrom, 1970, 1978; Howe et al., 1972; Flurkey, 1976; Becker et al., 1978; Palmer, 1982; and Mossler, 1987.

This paper describes the lithostratigraphy and biostratigraphy of the Eau Claire Formation in the subsurface of eastern Iowa. Utilization of Wisconsin outcrop data, published corehole descriptions from Missouri, redescribed cores from Iowa, and unpublished Iowa drillhole descriptions have led to a refined understanding of the Eau Claire's upper and lower boundaries, internal lithofacies distribution, and biostratigraphic age in the subsurface of eastern Iowa (Fig. 1). A north-south transect through the post-Mt. Simon Dresbachian strata from the type area in western Wisconsin to northeastern Missouri illustrates the Eau Claire's: 1) doubling in thickness, 2) facies transition from siliciclastic dominated to mixed siliciclastic/carbonate to carbonate-dominated, and 3) late Dresbachian age (Aphelaspis and Dunderbergia zones) in its upper part as far north as Lousia County, southeastern Iowa. This paper also offers new data concerning the southward extent, magnitude, and timing of the Dresbachian/Franconian (Sauk II/Sauk III, Palmer, 1981) disconformity in eastern Iowa. Additionally, this study notes the array of sedimentary structures as observed in core, and uses these in conjunction with vertical and lateral lithofacies distribution and biostratigraphic age constraints to interpret probable environments of deposition and their lateral and vertical shifting within the inner shelf region of eastern Iowa during Dresbachian time.

STRATIGRAPHY

Data Sources

Wisconsin

Huber's (1975) study of the Eau Claire in west-central Wisconsin was the principal study used to establish a working definition of the formation, which could be extended into the subsurface. Other Wisconsin outcrop studies provided additional information, especially concerning the Eau Claire's upper and lower boundaries in the outcrop region (Fig. 1). Several unpublished well cuttings and geophysical logs from the files of the Wisconsin Geological Survey provided subsurface information in the southwestern part of the state.

Iowa

The highest quality subsurface data in Iowa was derived from two full and continuous cores through the Eau Claire. The Elkader core (1"; Fig. 2b) and the Hutchinson core (3½"; Fig. 3) were described and searched for trilobites and acrotretid brachiopods. Supplemental lithofacies data was derived from a number of drillhole cuttings logs in files of the Iowa Department of Natural Resources - Geological Survey Bureau.

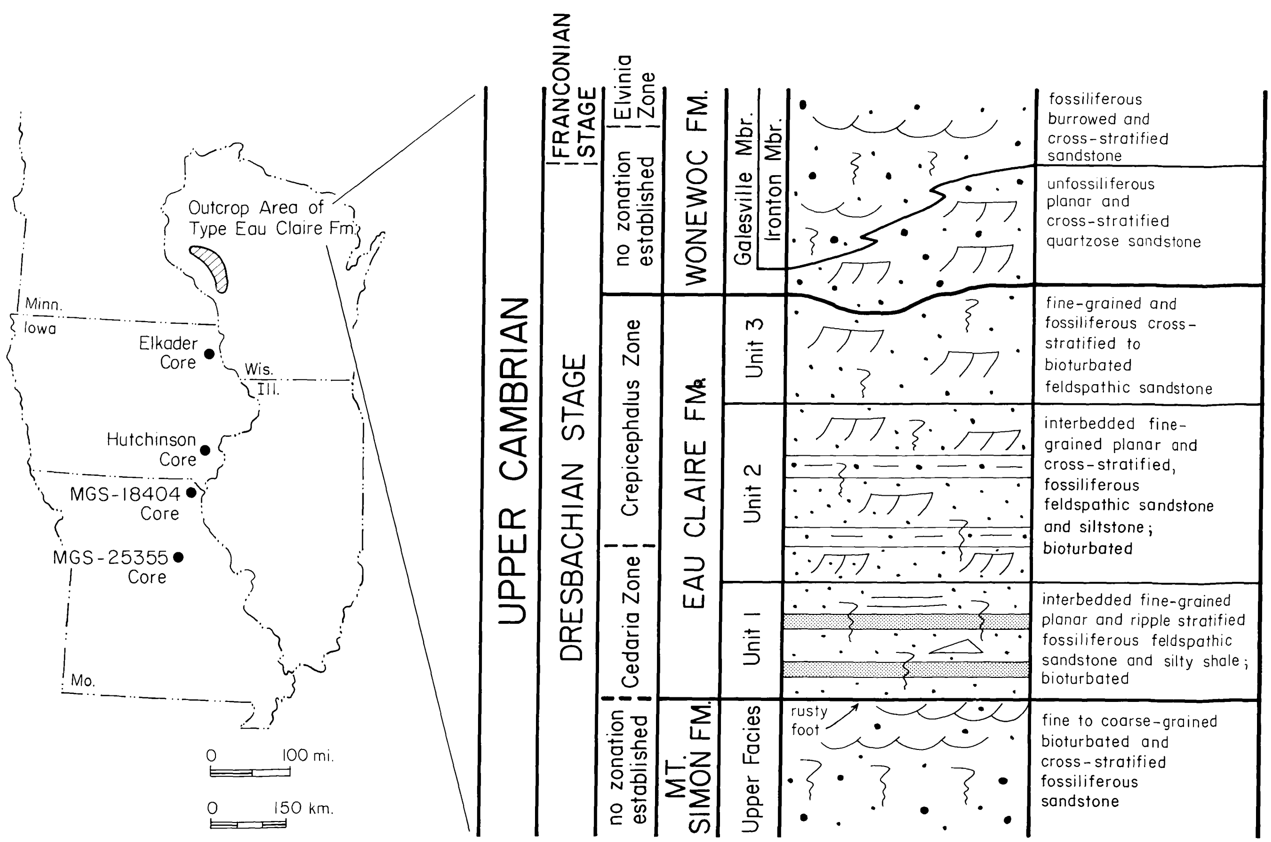

Figure 1. Map showing the location of the Dresbachian Eau Claire Formation outcrop area in west-central Wisconsin, and cores examined or utilized in this study. Generalized stratigraphic column of the post-Mt. Simon Formation Dresbachian sequence in western Wisconsin. (Column compiled and adapted from Ostrom, 1970; Huber, 1975; Driese et al., 1981; and Dott et al., 1986).

Illinois

Buschbach's (1965) published description of the #1 E.A. South stratigraphic test in Henry County was the only Illinois data used.

Missouri

Core descriptions and faunal data contained in Howe et al., 1972, for Clark and Audrain counties completed the data set for the southern part of the study area.

All of these data sources were used in constructing Figure 4, a north-south stratigraphic cross-section of the Eau Claire Formation through eastern Iowa. Figure 4 illustrates the principal lithostratigraphic and biostratigraphic relationships observed and interpreted in this study.

Contact with Mt. Simon Sandstone

Outcrop region

In the outcrop region the contact between the Eau Claire and underlying Mt. Simon occurs where fine- to coarse-grained, inarticulate brachiopod rich, trough cross-stratified quartz arenites of the Mt. Simon (Driese et al., 1981) pass abruptly upward into thin- to thick-bedded fossiliferous fine-grained sandstone and siltstone with interstratified shales of the Eau Claire (Unit 1, Shaly Thin-Bedded Facies of Huber, 1975). At a number of outcrop localities this lithologic change coincides with the occurrence of a red, iron oxide-stained zone 5-30 cm thick within the uppermost Mt. Simon sandstone (Driese et al., 1981). Huber (1975) informally termed this red marker zone the "rusty foot".

Subsurface

In the subsurface, the transition between the two formations is grossly similar to that of the outcrop region, but is often transitional over a 10 to 70 foot interval. In the Elkader core (Fig. 2b) the upper 70 feet of the Mt. Simon consists of thin to medium interbeds of bioturbated, shaly, very fine- to medium-grained sandstones, and fine- to coarse-grained planar laminated to low-angle cross-laminated sandstones. Both lithologies contain whole to abraded valves of the phosphatic inarticulate brachiopod Obolus, and phosphate-cemented clasts. Many, but not all, of the coarser-grained beds contain enough brachiopod valve debris to be considered coquinas. The phosphatic cemented clasts are much more common in the coarser sand layers. Clasts are oval to flat pebble shaped, range from 0.2-1.5 cm in diameter, and display subrounded to rounded edges. Clast framework grains consist of silt to very fine-grained K-feldspar and quartz with common inarticulate brachiopod valves. Clast cement consists primarily of phosphate, scattered pyrite, and minor dolomite. Coarse-grained, phosphatic clast sandstones sharply overlie finer shaly sandstones, and often are normally graded, having a higher clast concentration in the base grading to a greater valve concentration in the upper part.

The finer-grained shale-rich interbeds of the upper Mt. Simon in the Elkader core are similar, both in texture and composition, to the lithologies in the basal Eau Claire (unit EC1, Fig. 2b). Both exhibit horizontal to wavy interlamination of shale and sand on a submillimeter to centimeter scale, and moderate to strong bioturbation has disrupted much of the original fabric. Rare, compressed, v-shaped, sand-filled mudcracks in shale laminae also occur in both units.

The contact between the Eau Claire and Mt. Simon in the Elkader core is placed above the last common occurrence of coarse-grained sand. This coincides with the appearance of two heavily hematitic, inarticulate brachiopod rich, fine- to coarse-grained sandstone beds at 1233.7'-1234.0', and 1235.1'-1236.4'. Hematite dominates as the cement, but also occurs as single to multiple oolitic coatings on numerous quartz grains. This hematite-rich zone occurs at roughly the same stratigraphic position in Allamakee County, where it is underlain by 5 to 10 feet of shale which is placed in the Eau Claire. The iron ooid zone at the Eau Claire-Mt. Simon contact in Clayton and Allamakee counties is here considered a lithologic correlative to the "rusty foot" of the outcrop region.

To the south, in east-central and southeast Iowa, the contact between the two formations is consistently picked at the upward change from fine- to coarse-grained sandstones of the Mt. Simon to shaly and silty, very fine- to medium-grained sandstones of the Eau Claire. In Louisa County this is illustrated by the log of the Hutchinson core and the overlapping Friis core (Fig. 3). In these cores finer and coarser transition zone interbeds are common in the upper 10 feet of the Mt. Simon, but absent are bioturbated fabrics, phosphatic cemented clasts, inarticulate brachiopod material, and any indication of a hematite or iron-ooid zone. Where available, natural gamma log signatures correspond well to the lithologic transition (Fig. 3) between the two formations. Criteria for separating the Lamotte Sandstone from the Eau Claire and basal Bonneterre of northeastern Missouri (Howe et al., 1972) are similar to those used in Iowa.

Division of the Eau Claire

As illustrated in Figure 4, numerous lithofacies have been recognized within the Eau Claire in the subsurface. At this time none of these apparently mappable units are given formalized member status. Instead they are assigned lithic designations which indicate the principal rock types composing each lithofacies. To facilitate description of these lithofacies the Eau Claire will be discussed in terms of the lithofacies which compose it's lower, middle, and upper portions.

Lower Eau Claire

Both in Wisconsin and Iowa the lower Eau Claire is composed of a fine sandstone and shale facies (Fig. 4). In the subsurface, shale content generally varies between 10-20%. Locally, as at Dubuque, shale dominates over sandstone. The sandstone and shale facies of the lower Eau Claire varies in thickness from 45 feet at Eau Claire, Wisconsin, to 135 feet in Lousia County, Iowa. This lithofacies thins appreciably into northern Missouri as it passes laterally into thick shales and finally to carbonate dominated strata of the lower Bonneterre. The facies exhibits a rapid thickening along the line of section in east-central Iowa (Fig. 4).

Faunal Age

Huber (1975) showed that the sandstone and shale facies of the lower Eau Claire in the outcrop region contains a trilobite fauna characteristic of the Cedaria Zone. Rare identifiable trilobite molds from the Elkader and Hutchinson cores consist of Cedaria and Norwoodella (Figs. 2b, 3, & 4); both are Cedaria Zone trilobites. A Crepicephalus at 2177.3' in the upper part of unit A in the Hutchinson core (Fig. 3) indicates that the upper portion of this facies in Louisa County contains a Crepicephalus Zone fauna.

Lithologies

Based on the textures and sedimentary structures observed in the rock cores, a list of bedding types was established to characterize the lithologies present in the lower Eau Claire sandstone and shale facies.

Laminated sandstone - horizontal to low-angle planar stratification in 3-40 cm thick intervals. Laminae 0.1-3 mm thick and often accentuated by finely abraded inarticulate brachiopod shell debris or pelletal glauconite; phosphate-cemented siltstone clasts, oval to flat and 1 mm to 4 cm in diameter are often present.

Small scale ripple laminated sandstone - ripple lamination which has less than 2 cm amplitude and is relatively uncommon.

Megaripple laminated sandstone - cross lamination with 2-8 cm amplitude intervals; not common.

Bioturbated shaly sandstone - intervals 5 cm to 2 m thick, with shale content varying from 5-50%. Strongly bioturbated having 75-90% bioturbation; common.

Interlayered sandstone and shale - finely interlayered fine sandstone and green-grey shale with layers less than 5 mm thick; and coarsely interlayered sand and shale with layers 5 mm to 3 cm thick. Laminae are dominantly parallel, horizontal to wavy and continuous except where altered by bioturbation. Intervals may be weakly to strongly bioturbated with common vertical and horizontal traces. V-shaped, sand-filled

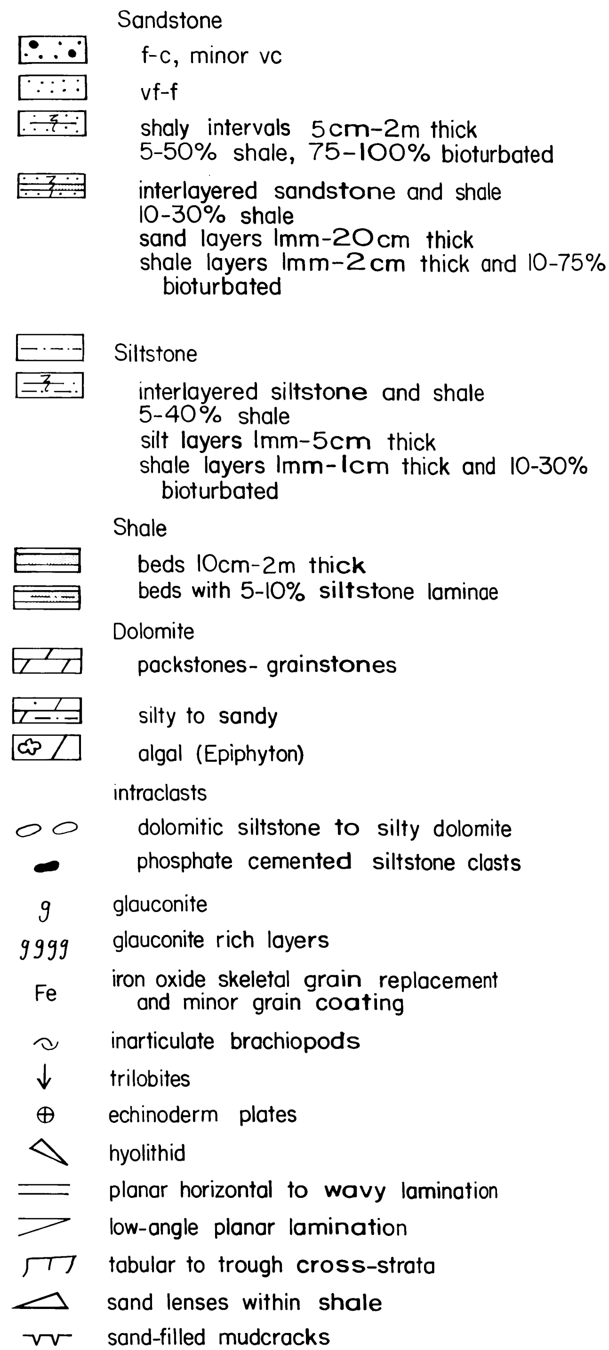

Figure 2a. Lithologic key for Elkader and Hutchinson core logs (Figs. 2b, 3).

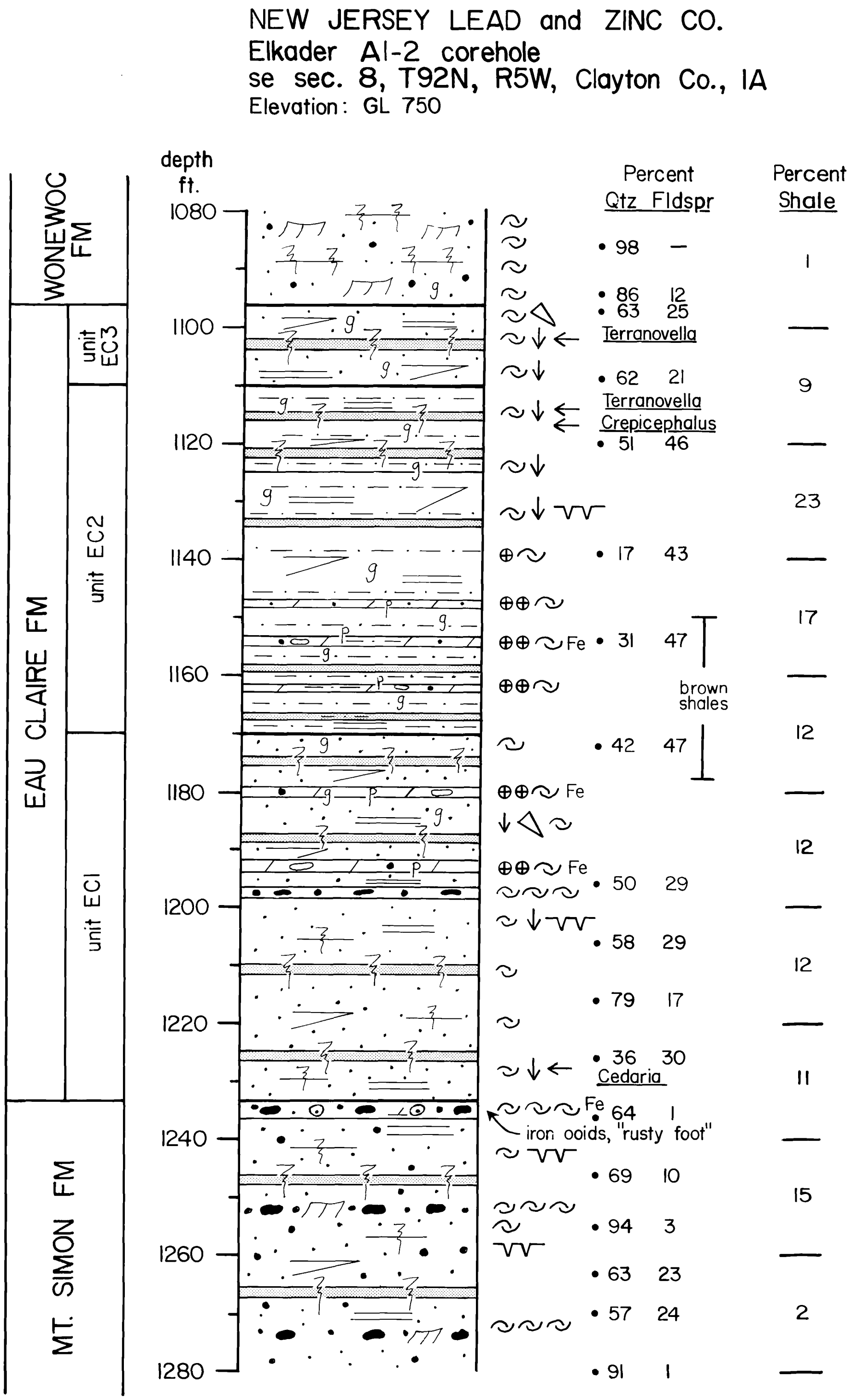

Figure 2b. Graphic core log of the Eau Claire Formation in the Elkader core. Lithologic key in Figure 2a. Percent quartz and feldspar from Flurkey (1976).

mudcracks in shale layers occasionally present. Phosphate-cemented siltstone clasts present in some intervals. Intervals 5 cm to 2 m thick.

Lenticular sandstone - lenses of sand within shale occurring in interlayered sandstone and shale intervals. Lenses 0.5 to 1.5 cm thick.

Shale - rare sandy beds similar to strongly bioturbated shaly sand, but with greater than 50% shale content. Also layers less than 10 cm thick and usually having sand filled mudcracks.

Flat pebble conglomerate - flat pebble clasts of laminated sandstone in a dolomitic sandstone matrix.

Dolomite echinoderm packstones - thin 2-3 cm thick layers containing round medium quartz grains and displaying extensive skeletal grain replacement by hematite.

Discussion

The lithologies which dominate the lower Eau Claire's sandstone and shale facies are, in order of abundance, finely and coarsely interlayered sandstone and shale, laminated sand, strongly bioturbated shaly sand, and lenticular sandstone. Rippled sandstone, thicker shale beds, and flat pebble conglomerates are minor lithologies.

The most common lithology is finely and coarsely interlayered sandstone and shale. The sand layers, where not disturbed by bioturbation, always have a sharp basal contact with the underlying shale layer and often display tool or prod mark impressions on their base. Within some 1-3 cm thick sand layers individual sets of planar laminae may be distinguished where they are truncated and overlain by another set. The overlying set of laminae drape over the truncated set forming a mini-hummock. Truncation relief is typically 0.5-1.0 cm. Although not universal, the sand layers commonly display graded bedding, grading from slightly coarser and shellier sand at the base to shaly and silty sand at the top. The graded sequences, like the ungraded ones, are laminated on a submillimeter to multimillimeter scale. Phosphatic-cemented siltstone clasts occasionally are present in both bioturbated and nonbioturbated interlayered sediments in the lower part of unit A, Hutchinson core. Finely broken and abraded inarticulate brachiopod shell debris is common and trilobite and hyolithid shell molds are much less common. Sand-filled, compressed, v-shaped mudcracks in shale, occurring in the lower two-thirds of unit A, Hutchinson core, usually are present within these interlayered lithologies, but also occur at the base of thicker laminated sandstone intervals.

Laminated sandstones always contain finely broken and abraded inarticulate brachiopod shell debris which tend to accentuate the laminae. Phosphatic-cemented siltstone clasts and whole valve inarticulate brachiopod material are common to abundant in some intervals. These phosphatic clast intervals tend to contain an abundance of medium- and rarely some coarse-grained sand. Laminated sandstones sharply overlie variably bioturbated interlayered sandstone and shale intervals, and strongly bioturbated shaly sand intervals. Trilobite resting traces (Rusophycus), and tool marks are present on some of their soles. Laminated sandstones also overlie and grade upward from small-scale to megarippled sands. Their upper contacts may be gradational, into variably bioturbated intervals, or sharp, into nonbioturbated interlayered to lenticular intervals containing sand-filled mudcracks.

Bioturbation is common in many intervals of the lower Eau Claire. It ranges from absent in the well laminated and rippled sandstones to strongly developed in the bioturbated shaly sandstones. Bioturbation is absent in the mudcracked layers.

Small-scale rippled and megarippled intervals, while comparatively few, usually display bimodal foreset laminae dip directions, and cut and fill structure. Some small-scale ripples display symmetrical cross-laminae with very thin (<1.0 mm) clay laminae. Inarticulate brachiopod valves are often concentrated along megaripple cross-laminae.

Flat pebble conglomerates and echinoderm dolomite packstones occur in the upper portion of the lower Eau Claire in the Elkader and Hutchinson cores (Figs. 2b, 3). The laminated sandstone flat pebble intraclasts range in length from 3 mm - 5 cm and have a maximum thickness of 1 cm. The matrix consists of very fine to medium sand and numerous echinoderm grains. Echinoderm dolomite packstones are present as several thin 2-3 cm thick layers in this portion of the Elkader core. Dolomitized echinoderm grains and glauconite grains are extensively replaced by red iron oxide (hematite). Round, medium quartz grains, some with single to

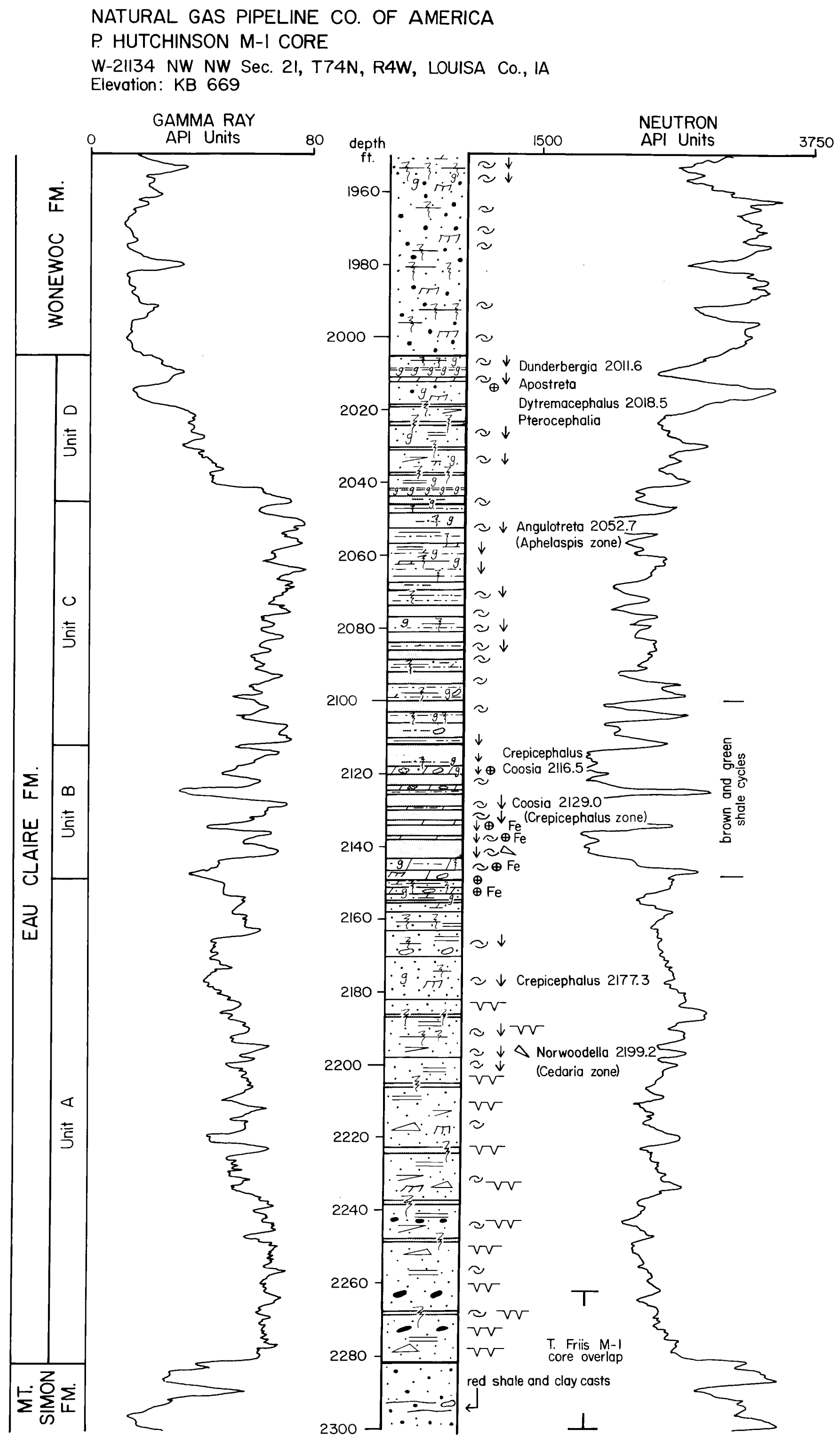

Figure 3. Graphic core log of the Eau Claire Formation in the Hutchinson core. Lithologic key in Figure 2a.

multiple iron oxide coatings, are common in these packstones. Inarticulate brachiopod shell material is common but trilobite grains are rare.

Middle Eau Claire

In the Wisconsin outcrop, the middle portion of the Eau Claire has been informally designated the "lower massive beds" by Huber (1975; Unit 2 in Fig. 1). These consist of thin- to thick-bedded variably cross-stratified sandstones interbedded with horizontally laminated siltstone beds. Ripple marks, mudcracks, and bioturbated fabrics are common. The middle lithofacies of the outcrop region grades laterally southward to a siltstone and shale lithofacies in northeastern and east-central Iowa (Fig. 2b, unit EC2; and Fig. 4). This siltstone and shale lithofacies in turn grades laterally into a shale and dolomite facies in southeastern Iowa, which finally grades into an oolitic limestone and dolomite facies of the Bonneterre Formation in northeastern Missouri (Figs. 3, 4).

Faunal Age

In the outcrop region the lower one-third of the middle sandstone and siltstone lithofacies contains a Cedaria Zone fauna, and the upper two-thirds a Crepicephalus Zone fauna. Rare identifiable trilobite molds from the middle Eau Claire of the Elkader core (unit EC2) and Hutchinson core (unit B) are Crepicephalus and Coosia. Both of these trilobites indicate a Crepicephalus Zone age for the middle lithofacies as well as for the upper part of the lower sandstone and shale facies (unit A - Hutchinson core), and the lower part of the overlying siltstone and shale facies (unit C - Hutchinson core; Fig. 4)

Lithologies

The siltstone and shale facies of the middle Eau Claire in the Elkader core contains similar textures and sedimentary structures to those of the lower Eau Claire's sandstone and shale facies. The main difference is that the grain sizes are in the siltstone range as opposed to the fine sandstone range, and that this interval is consistently glauconitic (Fig. 2b). Laminated siltstones and finely to coarsely interlayered siltstones and shales displaying varying degrees of bioturbation and occasional graded laminations dominate the facies. Thin to medium green-grey shale beds also are present, but small-scale rippled and megarippled intervals are absent. Sand-filled mudcracks have been noted at only one interval. Laminated glauconite wackestones to arenites occur sporadically with laminated reddish-brown shale near the base of unit EC2, Elkader core. Dolomite echinoderm packstones having sparse iron oxide skeletal grain replacement occur in several thin beds. Bioturbation, while occasionally present, is much less common, especially in the central portion of unit EC2.

The middle portion of the Eau Claire in the Hutchinson core displays significant lithologic differences from the biostratigraphically equivalent siltstone and shale facies of northeastern Iowa. The principal lithologies are outlined below.

Shale - alternating "cycles" of laminated reddish-brown and green shales. Brown shale intervals, 9 cm to 1.75 m thick, dominate; green shale intervals 3 cm to 60 cm thick. Sparse trilobite mold and inarticulate brachiopod shell material is present.

Skeletal dolomite packstones to grainstones - variably cross-stratified and silty beds, 6 cm to 75 cm thick, occurring with sharp contacts, within brown and green shale intervals; and occurring at the base and associated with algal stromatolite or thrombolite intervals. Numerous beds exhibit extensive hematite replacement of skeletal and glauconite grains. Dominant skeletal grains are echinoderms and trilobites; subordinate grains include inarticulate brachiopods, hyolithids, sponge, indeterminates, and intraclasts.

Bioturbated dolomites - fine crystalline, silty to very fine sandy, glauconitic, bioturbated dolostones in beds 3 to 15 cm thick.

Dense crystalline dolomite - homogenous, fine crystalline, nonsilty, nonglauconitic dolomite occurring as two 6 and 15 cm thick beds immediately beneath algal stromatolite and thrombolite intervals.

Dolomite algal thrombolites and stromatolites - unlaminated (thrombolites) and laminated (stromatolites) algal structures (Aitken, 1967; Pratt & James, 1982) occurring as 12 to 24 cm thick beds developed on either dense crystalline dolostone, skeletal packstone to grainstone, or green shale.

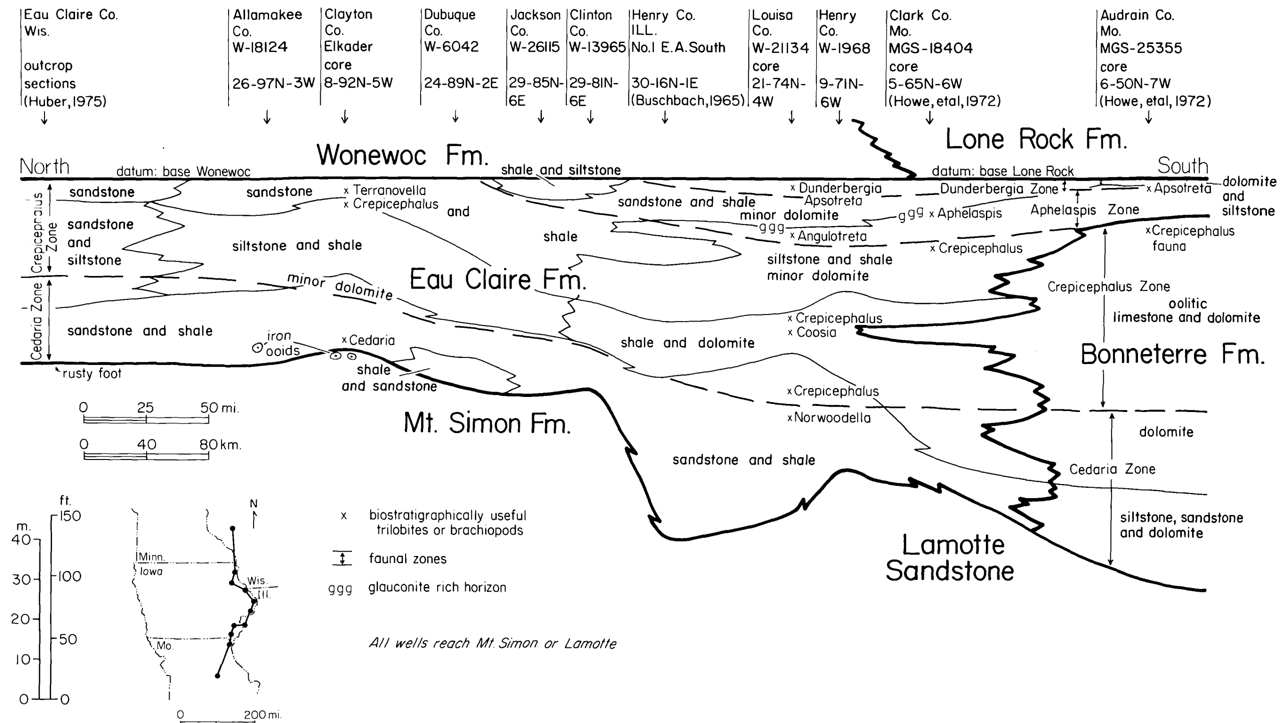

Figure 4. Stratigraphic cross-section of the Eau Claire Formation extending from the western Wisconsin outcrop region through eastern Iowa into northern Missouri.

Discussion

The middle portion of the Eau Claire exhibits the greatest amount of lithofacies variation from north to south, and also contains the highest concentration of shale and carbonate in the formation. Shale dominates over dolomite by about seven to three in the Hutchinson core, and this same interval grades to an oolitic limestone and dolostone dominated sequence in northeastern Missouri (Fig. 4). The occurrence of Crepicephalus and Coosia documents this interval as being the same age as the "lower massive beds" of the outcrop region.

The shale and dolomite facies of the Hutchinson core can be characterized as consisting of alternating green to brown shale "cycles" containing subordinate intervening carbonate beds. Reddish-brown shale comprises approximately 80% of the shale. Both green and brown shales are sparsely fossiliferous, containing limited numbers of trilobite molds and inarticulate brachiopod valves. Intervening dolostone beds usually occur with sharp upper and lower contacts, but a few intervals exist where skeletal grains (primarily echinoderm) occur as discrete laminae within shale. This always occurs as a gradation into an overlying echinoderm packstone-grainstone bed.

Partial hematite replacement of glauconite and skeletal grains is common within this facies, but is restricted to the packstones and grainstones of the lower two-thirds. Echinoderm and glauconite grains are most commonly replaced, while trilobites and other skeletal grains, including sponge, hyolithid, and indeterminates, are less affected. Cross-stratification is discernable in only a few grainstone beds as probable low-angle megaripple laminations.

Silty to very fine sandy and glauconitic dolostones, which comprise less than 10% of the lithologies, are the only beds which exhibit bioturbation. Prebioturbation remnant laminae suggest that these thin beds were orignally laminated. These beds occur within both green and brown shale intervals throughout the facies, but glauconite or skeletal grains always lack hematite replacement.

Dolomite algal thrombolites and stromatolites comprise another distinctive lithotype within the shale and dolomite lithofacies. These are known from the Hutchinson core where they form four discrete beds within two intervals. The lower interval, from 2123.1' to 2125.6', contains three algal layers, and the upper interval, from 2118.0' to 2118.8' contain one algal layer. Both laminated and nonlaminated clotted structure is apparent in the algal intervals, and following the use of Aitken (1967), and Pratt and James (1982), these are referred to respectively as stromatolitic and thrombolitic. Epiphyton appears to be the dominant blue-green algal growth form in the thrombolites. Two of the intervals can be classified as thrombolites as they display a distinctly massive and clotted appearance due to the growth of Epiphyton. The other two algal layers are somewhat composite since they exhibit internal transitions from clotted to laminated growth form in the same layer. As noted earlier, algal layers are developed on green shale, skeletal packstones-grainstones and dense crystalline dolostone.

The upper Eau Claire and its contact with the Wonewoc Formation

The upper Eau Claire of the outcrop region, referred to as the "upper massive beds" by Huber (1975; Unit 3 in Fig. 1), consists of five to twenty feet of fine-grained, thick-bedded, cross-stratified sandstones. Mudcracks, vertically oriented biogenic structures and skeletal material are common (Huber, 1975). The contact with the overlying Wonewoc is disconformable (Ostrom, 1966, 1970) and sharp, displaying up to twenty feet of local relief. The sandstone facies of the outcrop region of the upper Eau Claire passes laterally in the subsurface into a sandstone and shale facies as observed in the Elkader core (unit EC3, Figs. 2, 4). The sandstone and shale facies is interpreted to grade southward into a siltstone and shale facies in southeastern Iowa, which in turn grades to oolitic carbonates of the upper Bonneterre in northeastern Missouri (Fig. 4). Sparse faunal control in the Elkader core, the Hutchinson core, and the northeast Missouri cores indicate that these lithofacies contain a

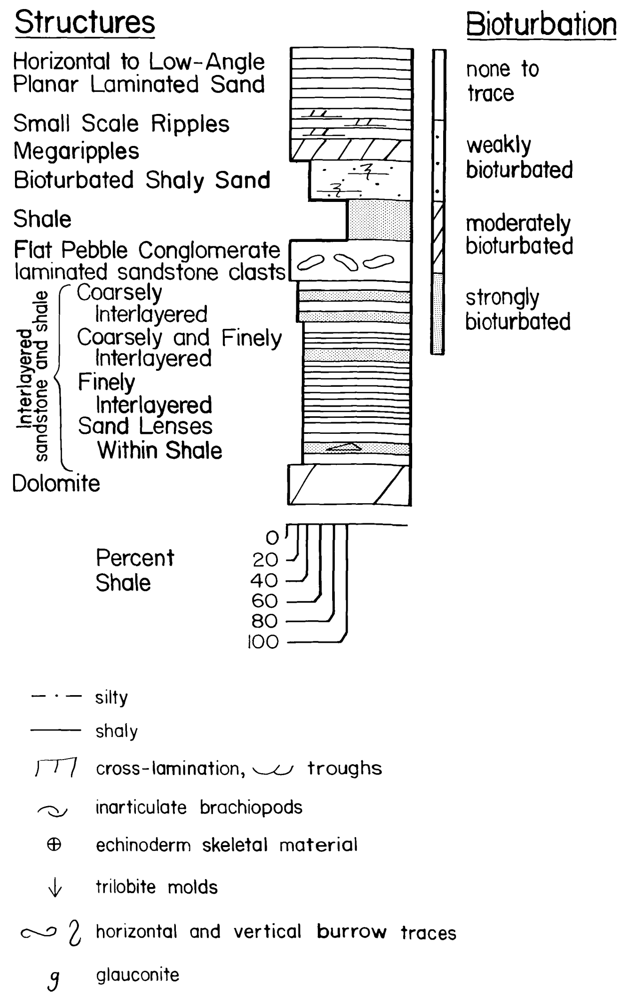

Figure 5a. Lithologic key for Figure 5b, Wonewoc/Eau Claire contact.

Crepicephalus Zone fauna, and are thus similar in biostratigraphic age.

Figures 3 and 4 illustrate that additional and biostratigraphically younger lithofacies comprise the uppermost Eau Claire in the subsurface of southeastern Iowa and northeastern Missouri. The upper portion of the siltstone and shale facies in the Hutchinson core (unit C, Fig. 3) contains the phosphatic acrotretid brachiopod Angulotreta; and the trilobite Aphelaspis occurs in this same facies in the Clark County, Missouri core. Above this, in both of these cores, occurs a sandstone and shale facies (unit D, Hutchinson core) which contains Dunderbergia and Apsotreta. The presence of the inarticulate

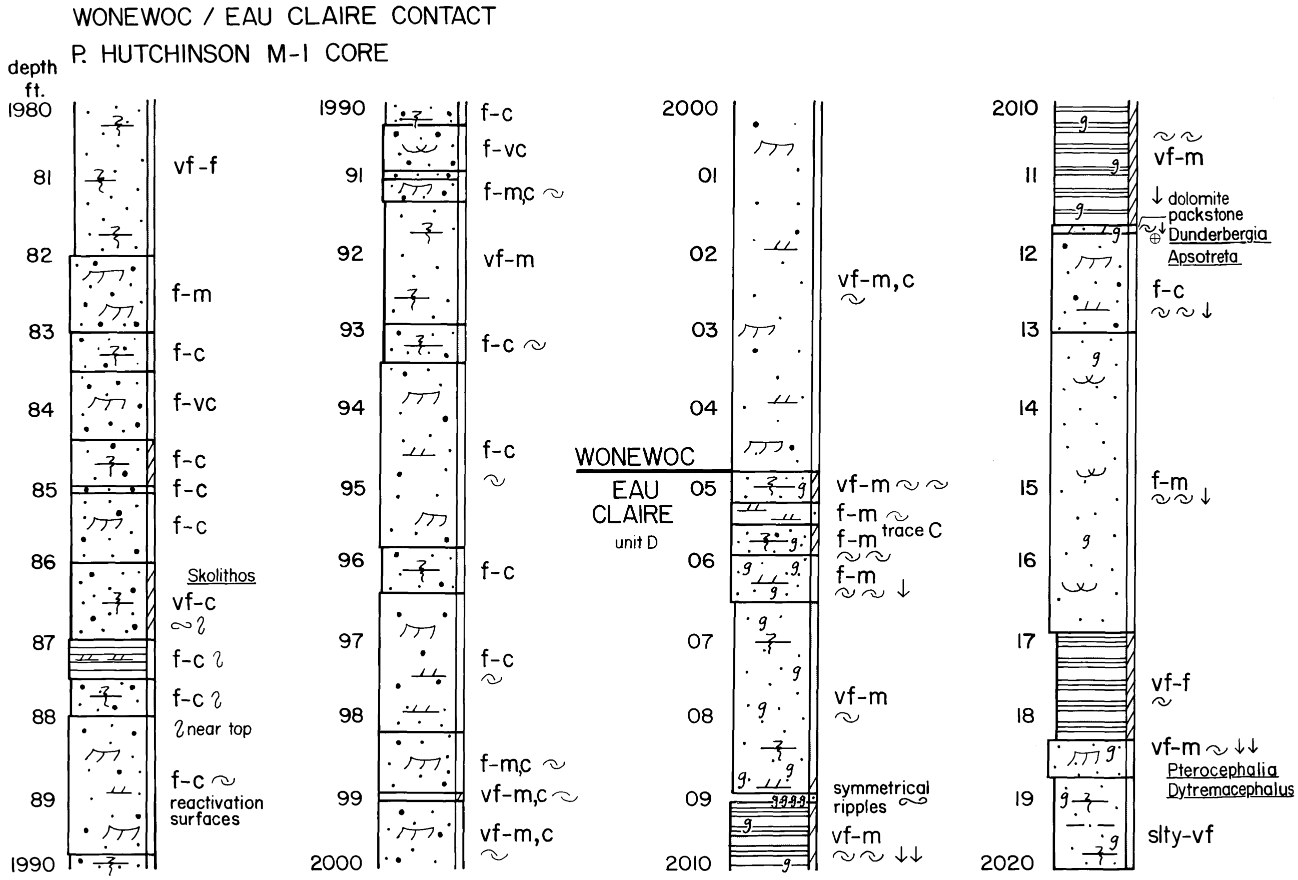

Figure 5b. Detailed graphic core log of the Wonewoc Formation/Eau Claire Formation contact in the Hutchinson core. Lithologic key in Figure 5a.

brachiopods Angulotreta and Apsotreta, and the trilobites Aphelaspis and Dunderbergia indicate a late Dresbachian age (Kurtz, 1971; Palmer, 1955) for a significant portion of the upper Eau Claire in southeastern Iowa and northeastern Missouri. These faunas also suggest that the upper Eau Claire of this region is in part a biostratigraphically equivalent lateral lithofacies to the upper Sullivan Siltstone Member and the lower Whetstone Creek Member of the Bonneterre and Davis formations of Missouri (Howe et al., 1972; Kurtz, 1986).

Lithologies and Discussion

The upper sandstone and shale facies of the Elkader core (unit EC3) consists of finely and coarsely interlayered fine sandstone and shale, and laminated sandstone (Fig. 2b). The sandstones are feldspathic and glauconitic, and the shales are green-grey in color. Skeletal grains include broken and abraded inarticulate brachiopod shells, and trilobite and hyolithid molds. Bioturbation is represented by horizontal traces (Planolites). The contact with the overlying Wonewoc Formation is placed at the change to megarippled and bioturbated, fine to medium sandstones. This lithologic change coincides with a distinctly higher concentration of quartz in the Wonewoc (Flurkey, 1976).

The siltstone and shale facies of the Hutchinson core (unit C, Fig. 3) is dominated by variably glauconitic and dolomitic, feldspathic, finely to coarsely interlayered siltstone and shale. Grey-green variably silty shales, up to 1 m thick and containing sparse inarticulate brachiopod shell valves and fragments of trilobite molds are the second major lithotype. Normally graded silt to shale layers, 1 to 2 cm thick, are common, and truncation and scour and fill of underlying layers are noted in several intervals. Biogenic structures consist of horizontal traces of Planolites, which are usually present in shales at the top of graded beds. Tool and prod mark casts are common on the soles of some coarse siltstone graded layers.

Several thin 3 to 5 cm thick dolomite beds of trilobite packstone to grainstone are present near the middle of unit C, Hutchinson core. These dolostones are glauconitic, silty to very fine sandy, and cross-laminated. Subordinate grains include inarticulate brachiopods and echinoderms. The upper few feet of unit C, Hutchinson core, contains several glauconite rich horizons, which could be characterized as greensilts.

The youngest and uppermost lithofacies of the upper Eau Claire in the study region is the upper sandstone and shale facies of the Hutchinson core (unit D, Fig. 3). The lower twenty-five feet of this unit consists of fine-grained, glauconitic and feldspathic interlayered sandstones and shales, laminated sandstones, and strongly bioturbated shaly sandstones. Overall shale content varies between 10 and 15%. The upper fifteen feet of this facies contains several 9 cm to 1.2 m thick intervals of variably cross-stratified very fine- to medium-grained, glauconitic sandstone (Fig. 5b). These intervals are interbedded with interlayered sandstones and shales, and strongly bioturbated shaly sandstones. One cross-stratified interval (Fig. 5b, 2011.7'-2013.0') contains coarse quartz sand and immediately above this is a thin 3 cm thick dolomitized trilobite packstone containing Dunderbergia and Apsotreta. Three feet above this trilobite packstone is a thin 3 cm thick very glauconitic shale layer. The contact with the overlying Wonewoc is placed at 2004.8' at the abrupt transition to thick bedded, very fine- to coarse-grained, nonglauconitic, cross-stratified sandstones. These sandstones, like others in the lower Wonewoc of the Hutchinson core are quartz arenites and contain highly broken and abraded inarticulate brachiopod shell debris. The remainder of the Wonewoc as illustrated in Figure 5b consists of interbeds of bioturbated and variably shaly sandstone, and cross-stratified sandstones. This facies of the lower Wonewoc is similar to the Wonewoc's burrowed and trough cross-stratified facies of the outcrop region as described by Dott et al. (1986).

Southward from the Lousia County Hutchinson core the Wonewoc Formation thins to zero in extreme southeastern Iowa (Fig. 4). In this area and northeastern Missouri Howe et al. (1972) have shown the Birkmose Member of the Lone Rock Formation to rest upon sandstones, siltstones, and shales of the upper Bonneterre. Kurtz (in Howe et al., 1972) has listed an abundant Linnarsonella fauna from the Birkmose and this is considered to be a partial faunal equivalent to the Elvinia Zone (Kurtz,1971), a portion of which is known to occur in the upper Wonewoc of the outcrop region.

DEPOSITIONAL ENVIRONMENTS

Previous Interpretations

The Eau Claire Formation has long been recognized as a fine-grained, siliciclastic, and fossiliferous marine interval in the Upper Cambrian (Croixan Series) sequence. Early interpretations of the environment of deposition of the Eau Claire established the idea that it's sediments were deposited in a shallow inland sea, and that at times, portions of the sea bottom were exposed to the atmosphere (Twenhofel et al., 1935). Later work by University of Minnesota graduate students, under the direction of W.C. Bell, placed the Eau Claire within an independent lithostratigraphic and biostratigraphic framework erected for the entire Croixan sequence (Berg et al., 1956). Within this framework the majority of the Eau Claire was classified as a fine clastic lithotope of thin-bedded, fine-grained to shaly sandstone and siltstone. It was interpreted as an offshore phase of a relatively simple Dresbachian transgressive-regressive cycle of marine sedimentation across an area of moderately low relief (ibid.). M.E. Ostrom's work on the Upper Cambrian of the Upper Mississippi Valley in the 60's and 70's modified and greatly expanded the lithofacies concept approach to studying the sequence (Ostrom, 1964, 1966, 1970, 1978). He recognized assymetric unconformity-bounded lithologic cycles consisting of recurrent lithostromes, and using a modern analogue, interpreted each unique lithostrome as the product of a distinct shelf depositional zone. In this model Ostrom (ibid.) characterized the Eau Claire as an argillaceous sandstone or shale lithostrome, which on a regional scale, is laterally transitional to carbonates of the Dresbachian age Bonneterre Formation. He interpreted the Eau Claire as representing deposits of an offshore shelf depositional zone, analogous to present day middle shelf regions in the northwestern Gulf of Mexico. Most recently, detailed sedimentological investigations of the Eau Claire and other fine-and coarse-grained Croixan lithofacies by Huber (1975), Byers (1978), Driese et al. (1981), and Byers and Dott (1986) have recognized a number of extremely shallow water to emergent-related sedimentary structures within the Eau Claire. In light of these and other sedimentological observations, these workers have interpreted the Eau Claire, in Wisconsin, to represent deposits of an intertidal and shallow subtidal tidal-process-influenced shelf regime.

Although all these excellent studies have advanced the understanding of numerous aspects of ancient cratonic shelf sedimentation within the Croixan Series, they cannot provide a comprehensive picture of this ancient shelf, simply because they have, for the most part, been limited to a relatively small outcrop area of the Cambrian on the craton (see figure 25 in Lochman-Balk, 1971). Most of the Dresbachian sequence of the Midcontinent is present in the subsurface, and while outcrop studies are the backbone to any understanding of this geology, subsurface studies must be conducted to fill in some rather large regional blanks.

In this author's attempt to add another piece to the Croixan puzzle, it appeared that it might be useful to step back and try to visualize the craton on a larger perspective. In doing so it became apparent that Midcontinent Upper Cambrian lithofacies distributions on the eastern side of the Transcontinental Arch are crudely similar to those on the western side of the arch. This fact is well illustrated in some of Lochman-Balk's (1971) paleogeographic maps, but Palmer's (1960) work on contemporaneous Cambrian lithofacies in western North America first recognized their distribution into three shoreline-parallel interfingering facies belts. These have been termed the inner detrital belt, the middle carbonate belt, and the outer detrital belt, and have been utilized as a lithostratigraphic and paleogeographic framework for many Cambrian studies in western North America. Fitting the Upper Cambrian of the northern Midcontinent into a similar framework appears plausible, and may prove to be useful in larger stratigraphic and depositional syntheses.

In the application of a similar framework to the Cambrian of the northern Midcontinent, the Croixan Series would clearly lie within the inner detrital belt, and lateral time-equivalent strata such as portions of the Bonneterre, Davis-Doe Run, and Potosi formations of Missouri and southern Illinois would lie within the middle carbonate belt. Expansion or contraction of either of these belts would shift shelf facies tracts laterally, resulting in variations of the vertical and lateral arrangement of lithofacies. The rather rudimentary lithostratigraphic and biostratigraphic relations documented and interpreted for the Eau Claire in this study (Fig. 4) clearly show that such facies shifting occurred in this region during the Dresbachian, and that the Eau Claire, when viewed on a more regional scale, is more than a relatively uniform fine-grained cratonic interval.

Lower Eau Claire

The lower portion of the Eau Claire in the subsurface of eastern Iowa bears a close similarity to the lower Eau Claire of the Wisconsin outcrop area. Sparse trilobite collections from core material show it to be the same age as the lower Eau Claire of the outcrop belt and sedimentary structures and lithologies are also somewhat similar. Many of the lithologies and sedimentary structures, when considered individually, are ambiguous as to the specific environment of deposition. Finely and coarsely interlayered sandstones and shales, lenticular sands in shale, and horizontal to low-angle laminated sandstones, as well as small-scale rippled, megarippled sands, and bioturbated shaly sandstones have been documented to form in both intertidal and subtidal settings in modern siliciclastic marine shelf environments (Evans, 1965; Van Stratten, 1954; Reineck, 1967; Reineck & Singh, 1967, 1972; Howard & Reineck, 1972; Klein, 1977; and Aigner & Reineck, 1982). It is apparent, however, that the suite of lithologies and sedimentary structures are indicative of deposition in repetitive, alternating, and fluctuating hydraulic flow conditions. Variable flow conditions dominated the lower Eau Claire inner shelf as evidenced by the very abundant stratification.

Finely and coarsely interlayered sandstone and shale intervals of variable thickness (5 cm - 2 m) are the most common form of bedding. Horizontal or planar to low-angle laminations dominate the very fine-grained sand layers; smaller scale ripple cross-lamination is rare. Both convex-up and convex-down whole to finely broken abraded inarticulate brachiopod valves are common. Some layers are graded, and are transitional to interlaminated shale and sand, and finally to shale, but many display no such gradation. It is difficult to decipher the origin of these ungraded layers. For very fine sands planar laminations can form from traction fallout at high flow velocity (>60 cm/sec; Harms et al., 1982), but can also form from suspension cloud fallout at low flow velocities (<20 cm/sec; Reineck & Singh, 1972). Additionally, there is speculation that planar lamination may also form by fallout from oscillatory flow with high orbital velocities in association with the formation of hummocky and swalely cross-stratification (Harms et al., 1982). Shallow (1 mm or less), faintly oriented tool and prod mark casts on the soles of some sand layers are supportive of the notion that laterally moving flows were operative prior to sand deposition; this would tend to rule-out deposition from suspension cloud fallout at low flow velocities. Other supportive evidence for higher flow velocities occur in some sand layers: 1) truncation of planar sets within a sand layer and deposition of another planar set over it; 2) presence of phosphatic-cemented siltstone clasts (up to 1.2 cm in length and 3 mm thickness) with rounded edges; and 3) graded layers having fine to medium sand, rare phosphatic clasts, and slightly higher inarticulate brachiopod valve concentration at the base grading upward to interlaminated sand-silt and shale, and finally to shale. The graded layers, especially those with phosphatic clasts, suggest deposition from waning high to low velocity flow conditions. Deposits from high velocity storm-generated flows, termed tempestites, are known from the modern North Sea and other ancient shelves (Aigner, 1982; Sepkoski, 1982; and Aigner & Reineck, 1982) and appear to be a plausible mechanism for generating graded beds such as these. However, very few of the sand layers contain oscillatory ripple-laminations in the upper portions, as recognized in the ideal or standard tempestite sequence.

The shale layers in the interlayered intervals vary from less than 1 mm to 3 cm thick. These obviously were deposited under very low flow velocities. Shale layers tend to exhibit more bioturbation traces than sand layers; the dominant form is sand-filled horizontal Planolites. Some degree of bioturbation, varying from a trace to strong bioturbation, is present in all interlayered intervals except the ones which contain v-shaped sand-filled cracks in shale. Crack-fillings range up to 2.5 cm in length and 0.7 cm in top width, and are usually connected to an overlying sandstone layer. Most fillings are oriented oblique to bedding; this is presumably due to compaction effects. Some cracks extend across the width of the core, while others do not; polygonal paterns are moderately well developed. Many fillings contain brachiopod valves and fragments similar to the connected overlying sand layer. These cracks appear at multiple positions within the lower Eau Claire of the Hutchinson core and at fewer positions within the upper Mt. Simon and lower Eau Claire of the Elkader core. The cracks are interpreted as representing periods of subaerial exposure and desiccation of mud, followed by filling with high-velocity flow-transported sand.

Planar laminated sandstone intervals up to 40 cm thick and bioturbated shaly sandstones are the other two lithologies which dominate the lower Eau Claire. The thicker laminated sandstones display many of the same features as the interlayered sandstones. Brachiopod valve and minor trilobite and hyolithid concentrations along laminae, occasional phosphatic clasts, and some set truncations, indicate deposition from high velocity flow conditions. The association of laminated sandstones with mudcracked intervals necessitates the interpretation that they probably were deposited in shallow water and perhaps on a surface periodically exposed to subaerial processes.

The bioturbated shaly sandstones represent strong disturbance of interlayered sand and shales by burrowing fauna. Remnant stratification in some intervals supports the case for an originally stratified sediment.

Megarippled and small-scale rippled intervals are not common, but can easily be accommodated into a shallow water setting with periodic high velocity flows. Brachiopod valves commonly line cross-laminae, and phosphatic clasts are common.

Iron Ooids and Phosphatic Clasts

The occurrence of iron ooids at the Eau Claire/Mt. Simon contact, and the presence of common phosphatic-cemented clasts in the upper Mt. Simon of the Elkader core and the lower Eau Claire of the Hutchinson core merit consideration. Van Houten and Bhattacharyya (1982) have summarized occurrences of Phanerozoic oolitic ironstones and offered a facies model to account for these occurrences. They concluded that ferric oxide oolitic ironstones most commonly develop in nearshore shallow marine siliciclastic-dominated environments. They maintain that most oolitic ironstones are condensed deposits which developed either at the end of a regressive shoaling upward sequence or the beginning of a renewed transgression; this situation could be considered as a stillstand. Driese et al. (1981) interpreted the bulk of the Mt. Simon in Wisconsin as a shoaling upward sequence terminating as an intertidal flat deposit transitional to the Eau Claire. The "rusty foot" occurs at the contact between the Mt. Simon and the Eau Claire in Wisconsin. This red iron oxide stained zone appears to be correlative with the occurrence of iron ooids in the subsurface of Allamakee and Clayton counties, Iowa (Fig. 4). It is logical to assume that the "rusty foot" and the iron ooids formed in similar facies positions and under similar environmental conditions. The co-occurrence of the previously discussed shallow-water indicators, the iron ooids and the position of the "rusty foot" in Driese et al.'s (1981) and Huber's (1975) facies models suggests that the iron ooids fit well into Van Houten and Bhattacharyya's (1982) shallow-water shoaling upward model.

Phosphatic cemented siltstone clasts are common in coarse-grained beds of the upper Mt. Simon in the Elkader core and finer-grained beds of the lower Eau Claire in the Hutchinson core. Considering the previous discussion, it must be concluded that these clasts also formed in close association with a shallow-water shoaling upward regime. Wanless (1981) noted the presence of similar phosphatic clasts in association with similar sediments from shoaling upward cycles in the Cambrian Bright Angel Shale of the Grand Canyon. He considered these clasts to be present in cycle caps which record phases of emergence.

Emergent conditions might well be necessary for the formation of these clasts. Clasts might form as follows: 1) high intensity storms deposit the laminated coarse silts and very fine sands along with phosphatic inarticulate brachiopod shell debris in a supratidal position; 2) dissolution of phosphate from shells in the surrounding environment mobilizes phosphate, which is precipitated as a cement in favorable layers; and 3) subsequent erosion and reworking of these layers forms the flat-pebble clasts with rounded edges.

A supratidal position for the formation of the phosphatic cemented layers appears warranted only because intact cemented layers from which the clasts were formed have not been observed in the cores. This leads to the suggestion that most (maybe all) of these original layers were eroded and reworked into clasts. Comprehensive destruction of original layers in a supratidal position experiencing periodic high intensity storms seems reasonable. Formation of cemented layers in a subtidal position would enhance the likelihood of preservation.

Lateral Relationships

Inspection of the stratigraphic cross-section in Figure 4 leads to speculation that the upper Mt. Simon of the Elkader core might be a lateral time-equivalent facies to a portion of the lower Eau Claire of the Hutchinson core. Biostratigraphy cannot presently confirm or refute this contention, since both intervals lack biostratigraphically useful fauna. However, the complementary thickness patterns, the gentle regional slope of the better documented faunal zones higher in the Eau Claire, and co-occurrence of similar lithofacies supports such an interpretation.

Paleoenvironment

The spectrum of lithologies and sedimentary structures within the cores, and the apparent stratigraphic relationships between the upper Mt. Simon and lower Eau Claire suggest that both were deposited in very shallow marine waters that experienced repetitive fluctuating flow conditions. The dominant process responsible for these flow conditions is uncertain, but storm-induced current flow may be a reasonable candidate. It is uncertain whether tidally-induced current flow played much of a role in the paleoenvironment. Periodic subaerial exposure is also indicated by the mudcracks.

It is diffcult to construct a facies model for this paleoenvironment based on the study of a few cores over a 300 mile-long transect. Instead, it might be useful to borrow and adapt another model. Pratt and James (1986) recently set forth a novel new model for carbonate sedimentation in shallow epeiric seas. This model, called the tidal-flat island model, envisions early Phanerozoic epeiric seas as dotted by low-relief tidal flat islands surrounded by open water. Modifications of this model to silicilastic sedimentation in epeiric seas might prove fruitful in understanding the lower Eau Claire of eastern Iowa. Perhaps the environment of the upper Mt. Simon and lower Eau Claire can be envisaged similarly as a shallow sea dotted with siliciclastic banks and shoals which were periodically built above sea level by Cambrian storms and/or tidal activity.

Middle Eau Claire

The middle portion of the Eau Claire in the subsurface of eastern Iowa is interpreted to have been deposited in deeper water during a transgression. The lithologies of the middle Eau Claire are quite different from the lower Eau Claire and indicate somewhat reduced detrial input and slightly clearer-water sedimentation.

Accumulation of shales within the central Eau Claire attests to periods of quiet sedimentation. The skeletal dolomite packstones to grainstones, which are interbedded with shales, indicate periodic energetic flow conditions. Echinoderm plates are common in these packstones. Sprinkle (1976) observed that in the western U.S., echinoderms are most commonly found in shales of the outer detrital belt, but that a few echinoderm occurrences are located in shales of the inner detrital belt and the middle carbonate belt, especially during the Late Cambrian. He noted that shales are the most favorable matrix for preserving complete Cambrian echinoderms. No articulated echinoderms have been found in the middle Eau Claire shales of the cores, and echinoderm debris has not been observed to be disseminated in the shales. However, paleoecological studies of Cambrian echinoderms indicate that they inhabited and even thrived in shaly environments (Sprinkle, 1976). This probably indicates that echinoderms were inhabitants of the shale and dolomite facies of the middle Eau Claire and that the echinoderm plates in the packstones had not undergone extensive (tens of kilometers) lateral transport from a different facies. The thin echinoderm packstones in the Elkader core could be interpreted similarly.

The relatively thin thrombolite and stromatolite intervals of the middle Eau Claire in the Hutchinson core also indicate a subtidal environment (Aitken, 1967; and Pratt & James, 1982, 1986). Intraclasts associated with the algal layers are probably indicative of submarine cementation of favorable layers and later reworking and deposition by storms (Sepkoski, 1982). Bioturbated, silty glauconitic dolomites indicate periodic oxygenated bottom sediments. Extensive predepositional replacement of skeletal grains and glauconite grains by iron oxides also indicates periodic well-oxygenated conditions, and the green to red-brown shale cycles may be indicative of cyclic oxidizing and reducing conditions on the sea floor or in the seawater.

Glauconite is consistently present in low percentages in siltstones of the middle Eau Claire. This is indicative of reduced sedimentation rates and possibly slightly cooler bottom waters (Porrenga, 1967; and Van Houton & Purucker, 1984). The common occurrence of graded beds within interlayered siltstones and shales may indicate grain fallout from storm generated sediment suspension clouds, but rare truncated sets of planar laminae also indicate occasional bottom-scouring flow conditions.

The lateral north to south grain size trend of fining southward is good evidence for a shallow nearshore to deeper offshore profile (Figs. 4, 6). The actual and inferred thin tongues of the Bonneterre Formation, which extend north towards the Iowa/Missouri line, imply a shifting of the middle carbonate belt to the north. Expansion of the carbonate belt in this manner is most easily explained as a consequence of a

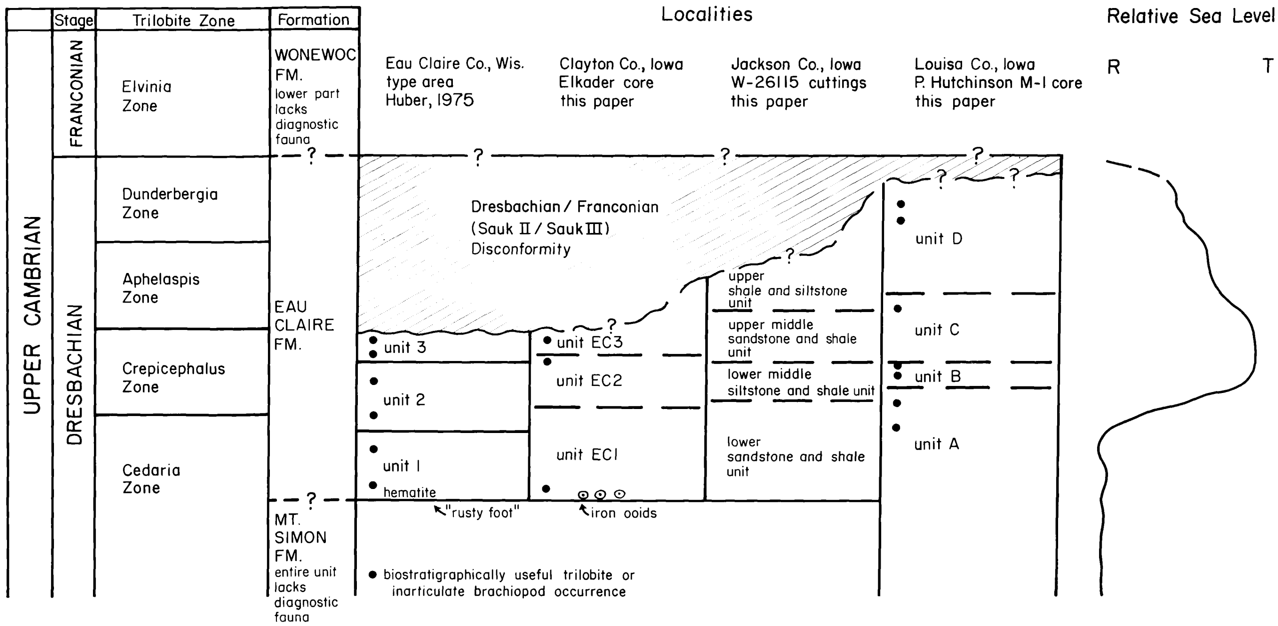

Figure 6. Chart illustrating biostratigraphic ages of Eau Claire lithofacies, magnitude of the Dresbachian/Franconian disconformity, and inferred sea level curve for the Dresbachian of eastern Iowa.

marine transgression which reduced the terrigenous influx and led to clearer-water sedimentation. In summary, it appears that the middle portion of the Eau Claire, which is more or less coincident with the Crepicephalus Zone, was deposited during a transgression in deeper and slightly clearer-water conditions (Fig. 6).

The upper Eau Claire and the Dresbachian/Franconian Disconformity

Figures 4 and 6 illustrate that the upper Eau Claire contains younger lithofacies in the subsurface to the south of the Wisconsin outcrop belt. Sequentially older facies and faunal zones are truncated beneath the Wonewoc along a south to north transect. The lithofacies containing the younger Aphelaspis and Dunderbergia zones are included within the Eau Claire, despite the fact that they are not present in the type area, by virtue of their overall lithic similarity to the rest of the formation. These or other time-equivalent facies may have extended northward, but truncation beneath the Dresbachian/Franconian (Sauk II/Sauk III disconformity of Palmer, 1981) disconformity has destroyed the rock record in that northern area.

Consideration of Figures 4 and 6 leads to the posing of several questions: 1) How much erosion and truncation, if any, is represented in the upper Eau Claire of the Hutchinson core?; and 2) How far southward does the Sauk II/Sauk III disconformity extend? Part of the answer to these questions lies in the interpretation of the depositional environment of the younger Eau Claire lithofacies.

The upper sequence of the Eau Claire in the Hutchinson core is interpreted to represent a shallowing upward trend, probably related to a regression punctuated by stillstands. The upward increase in grain size, decrease in shale content and appearance of megaripple cross-stratification in the upper Eau Claire is suggestive of a shallowing upward trend (Figs. 3, 5b). Rare grains of coarse quartz sand present in two intervals of unit D, Hutchinson core, indicate encroachment to the south of the coarser-grained facies, which presumably were deposited in shallower and more energetic flow conditions. The presence of disbursed glauconite in the sandstones indicates slow sedimentation, and the occurrence of two glauconite wackestones near the base and the top of unit D suggest near-cessation of detrital input for short periods of time. These may be the products of short stillstands.

The transition from the Eau Claire to the Wonewoc appears to be rather abrupt in the Hutchinson core. Interbedded glauconitic ripple cross-stratified sands and bioturbated shaly sands of the Eau Claire are sharply overlain by coarser, megarippled, nonglauconitic sandstone of the Wonewoc. The Wonewoc sandstones contain finely abraded inarticulate brachiopod valve debris, but biostratigraphically useful trilobites or brachiopods have not been recovered. Is this contact a disconformity? Due to the present lack of faunal control it is unkown whether part of the Dunderbergia Zone or roughly equivalent Apsotreta Zone is missing. Poor lateral stratigraphic control in the immediate vicinity of the Hutchinson core precludes the discovery of a physical disconformity, such as in the outcrop belt at Galesville, Wisconsin (Ostrom, 1966). That leaves only the core for interpretation and my interpretation is that the contact is disconformable. Although there is the slight suggestion of a transition between the two units because of the coarse grains in two upper Eau Claire intervals, the contact lacks the overall transitional character exhibited in other Croixan formational contacts.

To the south of the Hutchinson core, the Wonewoc pinches out and the Lone Rock Formation overlies the Eau Claire. Farther south, lithologies becomes more carbonate rich and stratigraphic nomenclature changes to the Davis and Bonneterre formations. In this area Kurtz (1986) interpreted the Dresbachian/Franconian boundary to represent a shallowing of the epicontinental sea. When viewed on a regional scale, the southward decrease in magnitude of the Sauk II/Sauk III disconformity is apparent, but deciphering the nature of its origin is another matter. Palmer (1981) urged stratigraphers to check their sections again for evidence of this disconformity which may represent a continent-wide but short-lived eustatic lowering of sea level. This study has attempted to answer that urging, but obviously it will take more studies of other subsurface materials to complete the story.

SUMMARY

The lower Eau Claire and upper Mt. Simon are in part lateral lithofacies, which were deposited during a shoaling upward regressive phase during latter Cedaria Zone time. Abundant shallow water to exposure indicators suggest that periodically exposed islands or shoals dotted the epeiric sea at this time. Transgression peaked during the middle Eau Claire as evidenced by the development of a variety of carbonate lithologies, and the lateral arrangement of lithofacies on the shelf during Crepicephalus Zone time. These interpretations are in accord with outcrop belt studies of the Mt. Simon and Eau Claire by Ostrom (1970, 1978), Driese et al. (1981) and Huber (1975). Regressive, shoaling upward conditions dominated the upper Eau Claire during late Dresbachian time (Aphelaspis and Dunderbergia zones). The upper Eau Claire in the subsurface of southeastern Iowa is of late Dresbachian age, and apparently is truncated by the Sauk II/Sauk III disconformity. The nature of this disconformity in the subsurface needs further study.

ACKNOWLEDGMENTS

A.R. (Pete) Palmer has been of special help to me in this and other ongoing subsurface Cambrian studies. He has identified all of the trilobites and acrotretid brachiopods used in this study, and provided council and enthusiastic encouragement. Bill Bunker has patiently listened to my ideas on the Cambrian and provided valuable feedback. Bill and Greg Ludvigson carefully edited this manuscript and made many helpful suggestions. Orrin Plocher helped with rock preparation during the study, and Patricia Lohmann and Mary McMurrer of the Iowa Department of Natural Resources drafted the illustrations.

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CHANGING STRUCTURAL AND DEPOSITIONAL PATTERNS, ORDOVICIAN CHAMPLAINIAN AND CINCINNATIAN SERIES OF IOWA-ILLINOIS

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INTRODUCTION

Thickness and lithofacies patterns and general stratigraphic relations displayed in the Middle and Upper Ordovician (Champlainian and Cincinnatian Series) sequence of Iowa and Illinois are used to interpret structural movements, depositional events, and sea-level changes for this interval in the Midcontinent area. Isopachous patterns for various stratigraphic units help delineate areas of relative uplift or subsidence if analyzed in conjunction with lithofacies distributions. It should be emphasized that a region of thick isopachs does not necessarily reflect that the area was a bathymetric basin during deposition, merely that it was an area of increased subsidence or sedimentation. Therefore, it is important to contrast structural-stratigraphic basins from bathymetric-depositional basins; these two basinal concepts may or may not be coincident. Relative rates of sedimentation vs subsidence determine whether or not a region was bathymetrically depressed with respect to adjacent regions during deposition. Thus, depositional interpretations of lithofacies patterns become vital in understanding relative bathymetry in the epeiric seas.

Depositional patterns were affected not only by local structural settings and relative crustal movements, but also by regional and global changes in sea level. Global eustatic changes in sea level should exert profound influence on shallow-water epeiric sedimentation patterns and are considered to be the major external control on deposition in many settings. Relative changes in sea level can be interpreted for any given stratigraphic sequence at specific localities, but the problem remains as to whether these changes are due to local or regional structural influences or to global eustacy. Only through broad-scale regional, inter-regional, and inter-continental correlations and comparisons can the eustatic overprint be interpreted with confidence. The Ordovician sequence of the Upper Mississippi Valley area displays sedimentary patterns and stratigraphic relations that can be interpreted in the context of relative changes in sea level. Apparent changes in sea level interpreted for the Iowa-Illinois area are used to delineate deepening and shallowing trends, which serve to bracket depositional cycles of varying magnitude. Those cycles which display consistent regional and inter-regional patterns are the cycles most likely to be of eustatic origin.

Middle and Upper Ordovician rocks are well exposed in Iowa, Illinois, and surrounding states, an area which serves as a standard of reference for the central craton of North America. Superb exposures, abundant fossils, and fascinating sedimentary features make these strata well suited for study. However, much work remains to be done, not only in the outcrop belt, but particularly in the vast area of subsurface Ordovician rocks. This report merely provides a broad-brush early-stage attempt at some sort of regional synthesis. We encourage additional workers to join in the fun. The attendance and enthusiasm at the 1987 Ordovician symposium at the North-Central GSA in St. Paul certainly was encouraging, providing truly significant contributions to our understanding of these rocks (see Sloan, ed., 1987).

STRUCTURAL AND DEPOSITIONAL PATTERNS

The New Madrid Rift Complex (Braile et al., 1982), which underlies the present Mississippi Embayment, had a major influence on depositional patterns in and adjacent to the proto-Illinois Basin throughout the Paleozoic (Kolata and Nelson, in press). The rift complex developed as a failed rift coincident with the breakup of a supercontinent during late Precambrian to early Cambrian time (Braile et al., 1986). Major displacement along the rift's bounding faults ceased by late Cambrian time. The tectonic setting changed from a rift basin to a broad, slowly subsiding cratonic embayment open to the continental margin and Iapetus Ocean to the South. Regional downwarping of both the rift complex and surrounding cratonic areas continued episodically throughout the Paleozoic. As a result, a broad southward-plunging trough developed in the proto-Illinois Basin. Open-marine circulation dominated, and depositional facies aligned with tectonically defined bathymetric zones. Configuration of the proto-Illinois Basin varied throughout the Paleozoic in response to periodic uplift of the confining arches and domes (Quinlan and Beaumont, 1984).

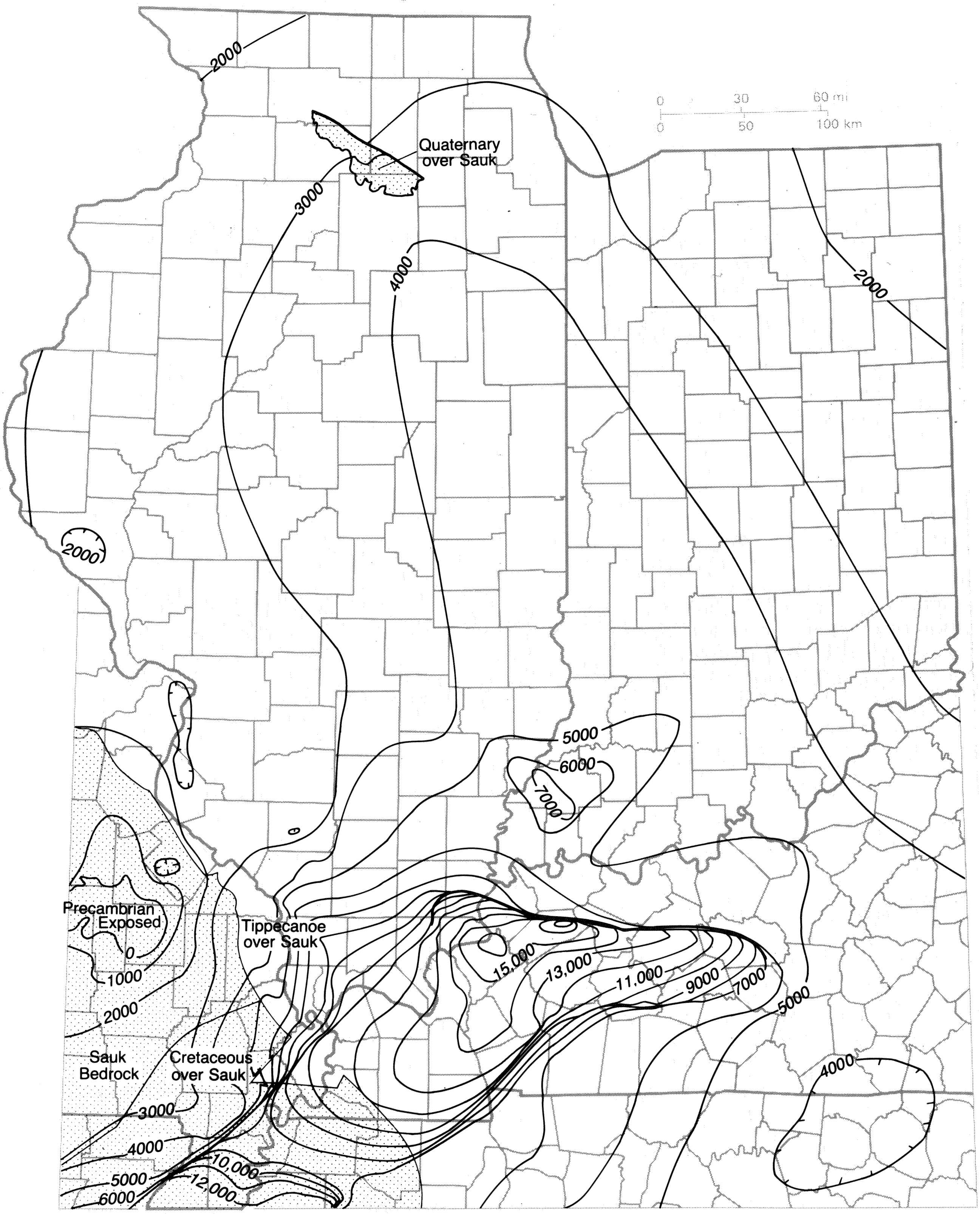
Sauk Sequence (Cambrian-Lower Ordovician) isopach trends reflect relatively rapid subsidence rates within the southern part of the proto-Illinois Basin. The Sauk is in excess of 15,000 feet thick in southern Illinois and western Kentucky and thins northward to approximately 2000 feet in northwestern Illinois and eastern Iowa (Fig. 1). Similar isopach trends developed during initial deposition of the Tippecanoe Sequence (through Blackriveran time), and again during deposition of the Kaskaskia and Absaroka sequences. Changes in subsidence rates within the proto-Illinois Basin during the Early Paleozoic influenced sedimentation and thickness patterns over large areas of the Midcontinent region.

St.Peter-Glenwood

Middle Ordovician (Champlainian) sedimentation was initiated in the Upper Mississippi Valley area with deposition of the St. Peter Sandstone, a diachronous sheet sand that spread northward as seas transgressed into the interior of the continent during the Chazyan and early Blackriveran. Earlier pre-St. Peter Middle Ordovician (Whiterockian) deposits are found to the south in Oklahoma, southern Illinois and adjacent states in basins marginal to the continent (inboard from the Ouachita continental margin). St. Peter sands buried the regional unconformity surface developed on Sauk Sequence rocks (the sub-Tippecanoe surface) and infilled valleys and sinkholes developed on that surface, reaching maximum thicknesses of up to 520 feet in northern Illinois (Kolata et al., 1978). Because the St. Peter infilled a highly irregular unconformity surface, its thickness patterns do not necessarily reflect structural patterns in a straight-forward manner.

Coincident with the northward spread of St. Peter sand facies, Dutchtown carbonate facies were deposited in southern Illinois and adjacent parts of Indiana, Kentucky and Missouri. The Dutchtown depocenter is situated in southern Illinois and western Kentucky, over the rift complex. Subsequent deposition of an east-west trending sand body (the Starved Rock Sandstone)

Figure 1. Isopach map of the Sauk Sequence in the Illinois area (prepared by Michael L. Sargent, Illinois Geological Survey). The New Madrid Rift Complex of southern Illinois and western Kentucky was the site of rapid subsidence and sedimentation during Early and Middle Cambrian time. The Sauk Sequence thins in all directions away from the northern part of the rift complex.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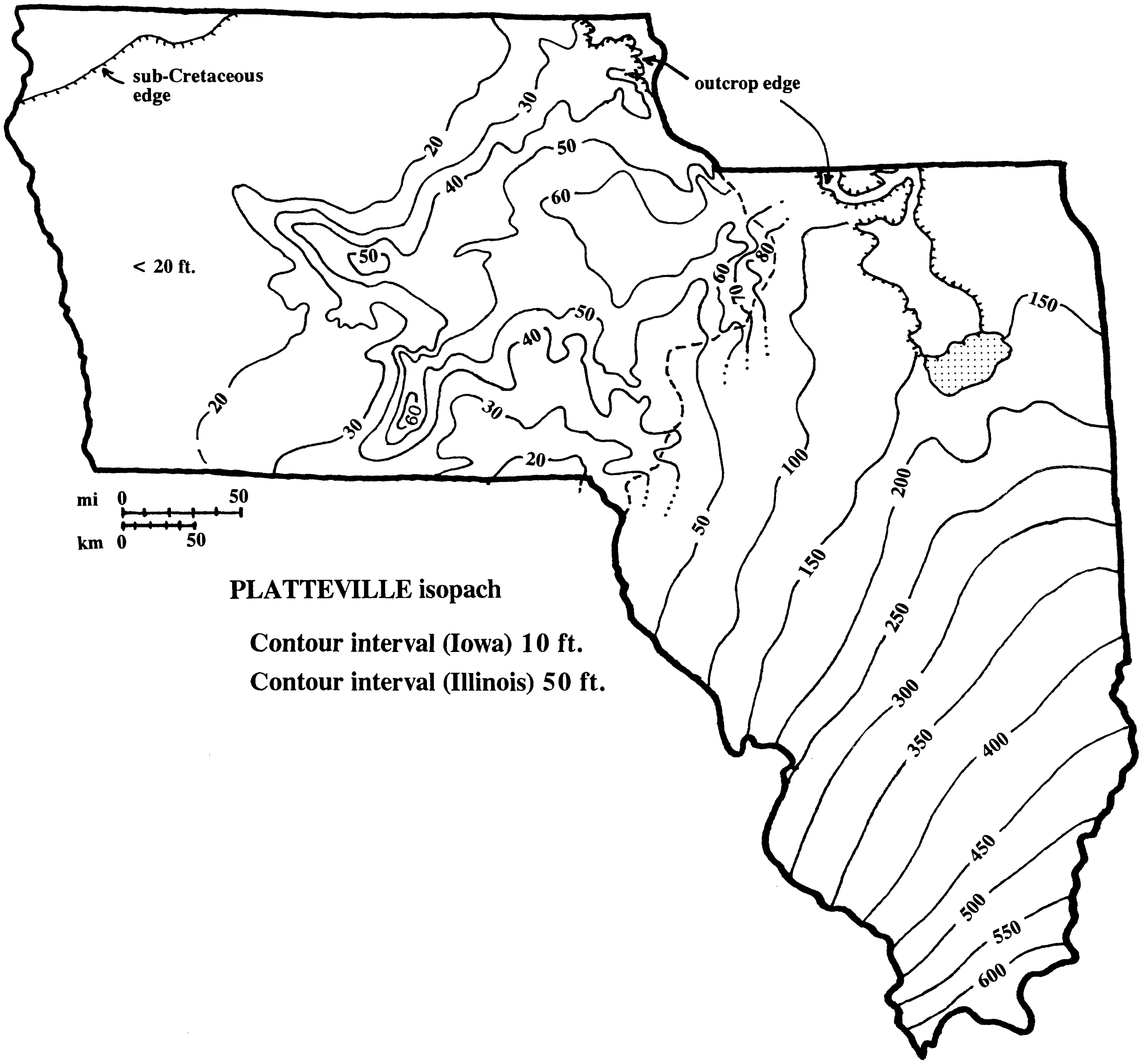


Figure 2. Generalized Platteville isopach map for Iowa and Illinois. Shaded where covered by post-Ordovician rocks.

across northern Illinois, southeast Iowa, and northern Missouri served to separate two regions of deposition: 1) the Glenwood Shale to the north, and 2) the Joachim Dolomite to the south. The combined Glenwood-Joachim package thickens dramatically southward in Illinois (see Willman and Buschbach, 1975, p. 63), basically paralleling isopachous trends seen in underlying Sauk Sequence (Upper Cambrian-Lower Ordovician) strata. Glenwood thickness patterns in Iowa generally thicken southeastward; the Glenwood is thin to absent in southwest Iowa and adjacent Nebraska. These relations suggest that general structural patterns remained relatively unchanged in the region during Sauk and early Tippecanoe (Chazyan-Blackriveran) deposition. The eastward thickening of strata in Iowa and the southward thickening across Illinois is referred to here as a Sauk-type pattern.

Platteville

Carbonate strata of the Platteville Formation in Iowa, the Platteville Group in Illinois, and the Plattin Formation in Missouri overlie Glenwood, Starved Rock, and Joachim strata in the region. The lower Platteville shares facies relations with the Glenwood Shale in Minnesota (Mossler, 1985). This extensive body of fossiliferous open-marine Platteville carbonate rocks thickens eastward in Iowa and southward in Illinois and displays a Sauk-type isopachous pattern (Fig. 2). The uppermost Quimbys Mill Formation and underlying Nachusa Formation of the Platteville Group are approximately 15 and 20 feet thick, respectively, in north-central Illinois and south-central Wisconsin (Willman and Kolata, 1978). Both formations thin westerly and terminate near the Mississippi River. Within the same area, the overlying Decorah units (discussed below) show an inverse relationship and overstep the Platteville from west to east and gradually pinch out from the base upwards. Carbonate facies indicate that relatively shallow depositional environments existed in northern Illinois, where upper Platteville strata (Quimbys Mill) locally display mudflat and coralline facies in the Rochelle area. The end of Platteville deposition marked the end of Sauk-type structural patterns in the area; a major structural reorganization would dramatically alter later thickness patterns.

Smaller-scale structural movements are also evident during Platteville deposition. Basal Platteville strata (Pecatonica Formation) are absent in the area of the Forreston Dome south of the Plum River Fault Zone in northwestern Illinois where younger Platteville strata (Mifflin Formation) overlap the St. Peter Sandstone (Willman and Kolata, 1978). This suggests that anticlinal folding probably was coincident with early Platteville deposition in that area.

Galena Structural Reorganization

In general, Galena Group strata thin to the south in the Iowa-Illinois area, and reach maximum thicknesses in northern Iowa and adjacent Minnesota (Fig. 3). These patterns are virtually reversed from the Sauk-like patterns seen in underlying Platteville strata (Fig. 2) and indicate a major structural reorganization. The lower Galena Group (Decorah Formation) of northern Iowa and adjacent Minnesota is dominated by shales derived from northern Transcontinental Arch terrains. These shales grade to carbonate-dominated facies in southeastern Iowa and throughout most of Illinois (Witzke, 1980). This northern shale package, however, does not by itself explain the northward thickening seen in the Galena which is primarily related to thickening of carbonate strata above the Decorah. The Galena Group is dominated by open-marine carbonate facies throughout its extent. Regional changes in carbonate facies are coincident with isopachous trends; deeper-water skeletal wackestone facies dominate in the north whereas shallower-water higher-energy bottom environments are indicated southward in Illinois and Missouri where packstone and grainstone lithologies dominate (Kimmswick facies). Southwestward thinning of Galena strata in the subsurface of Iowa is more difficult to evaluate since it occurs entirely within a pervasively dolomitized sequence lacking adequate sampling (Witzke, 1983).

The Wisconsin Arch in north-central Illinois and south-central Wisconsin was a positive element during the initial phases of Galena deposition, where Guttenberg strata overstep basal Galena beds (Spechts Ferry) and onlap the Platteville (Kolata et al., 1986). Dunleith strata overstep the Guttenberg in parts of Ogle County, Illinois to disconformably overlie the Platteville. Thickness variations seen in the Galena of central Iowa (Fig. 3) probably relate to small-scale movements along the trend of the Keewenawan Midcontinent Rift feature (Witzke, 1983). Southeastward deflection of the 250-foot contour in east-central Iowa and northeastward deflection in northwestern Illinois outlines the general position of the East-Central Iowa Basin, a basinal feature that strongly influenced later Ordovician and Silurian deposition (Bunker et al., 1985). This basin was bisected by the Plum River Fault Zone sometime after the Devonian. It seems probable that development of the East-Central Iowa Basin was initiated during Galena deposition.

The southward thinning of Galena strata in Illinois and adjacent Indiana and Missouri probably is due to structural changes associated with a slow-down of subsidence rates in the

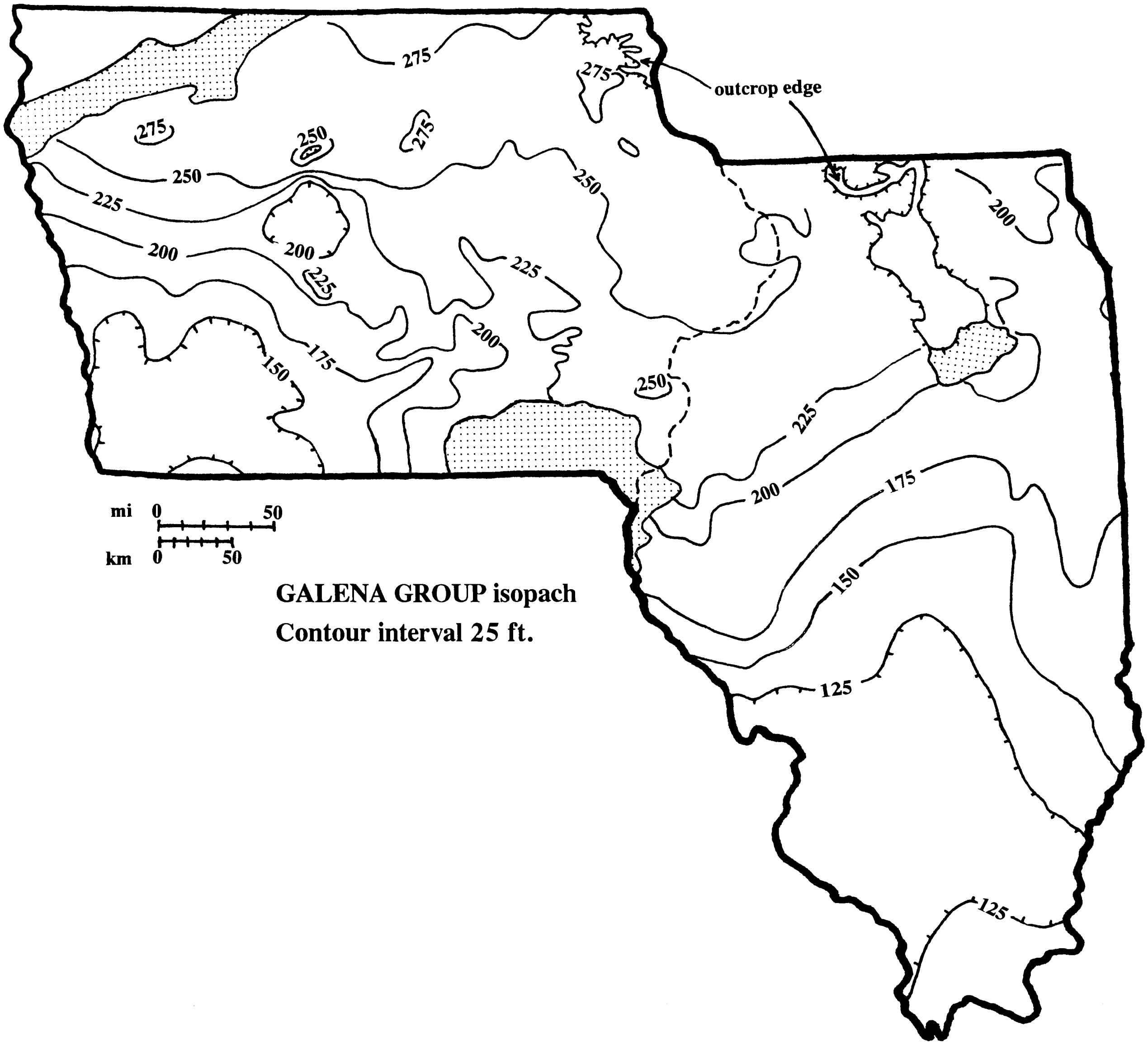


Figure 3. Generalized isopach map of the Galena Group for Iowa and Illinois. Shaded where beveled by post-Ordovician rocks.

southern basin area. Some of the southern thinning of Galena strata probably is due to uplift around the Ozark Dome and adjacent Sangamon Arch, as well. Galena depositional facies likewise shallow southward, indicating that the thinner Galena sequence in the south was not a function of reduced sedimentation in a starved basin setting. The structural changes that accompanied the initiation of Galena deposition completely disrupted Sauk-like structural patterns and marked a major episode of structural reorganization in the region. Deposition of the Galena and overlying Maquoketa strata was uninterrupted in the northern area, whereas Galena (Kimmswick) strata may be erosionally bevelled beneath the Cape Limestone in the southern area (Templeton and Willman, 1963). These interpreted stratigraphic relationships likewise support the idea that subsidence rates in the southern area significantly slowed during

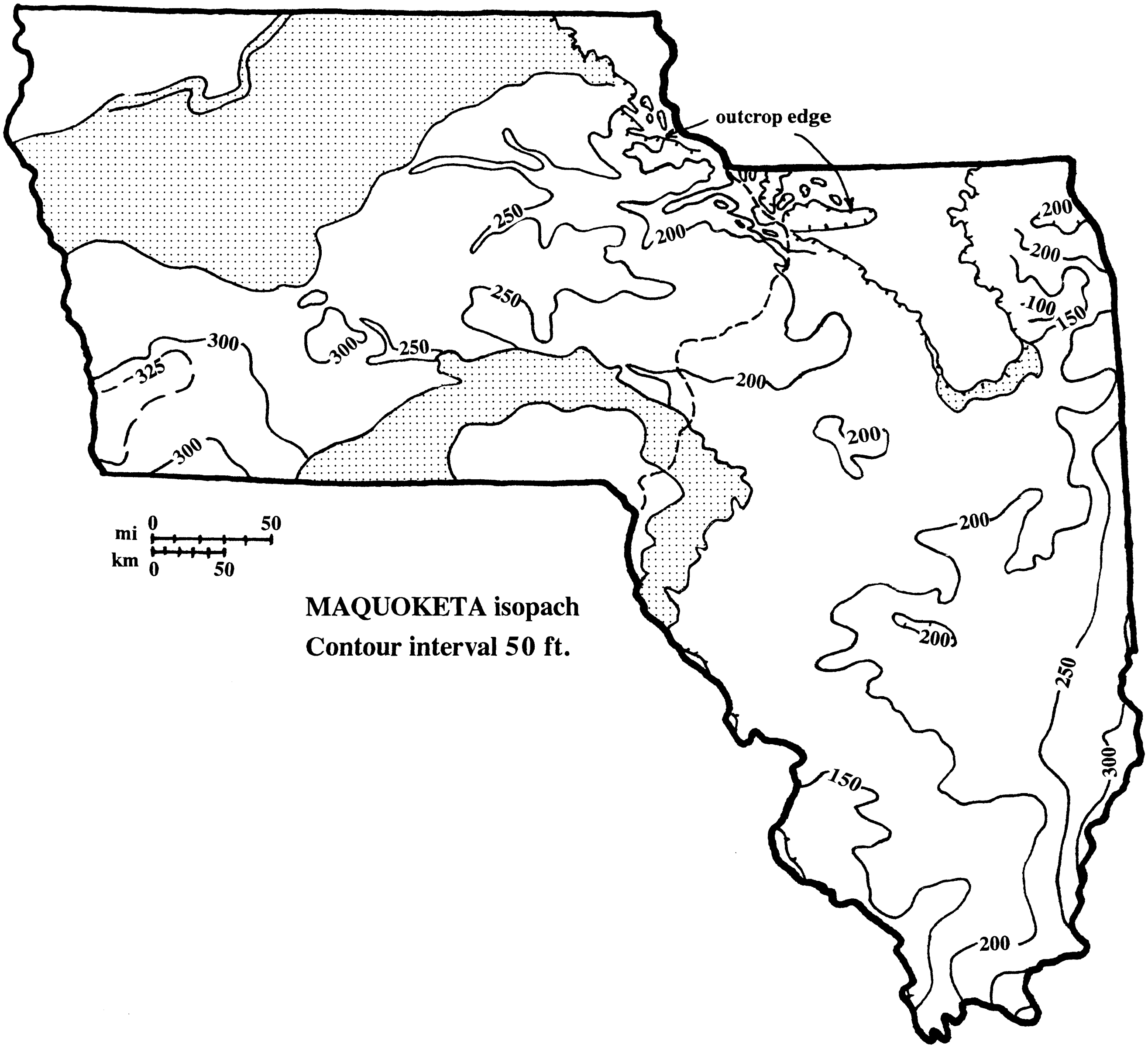
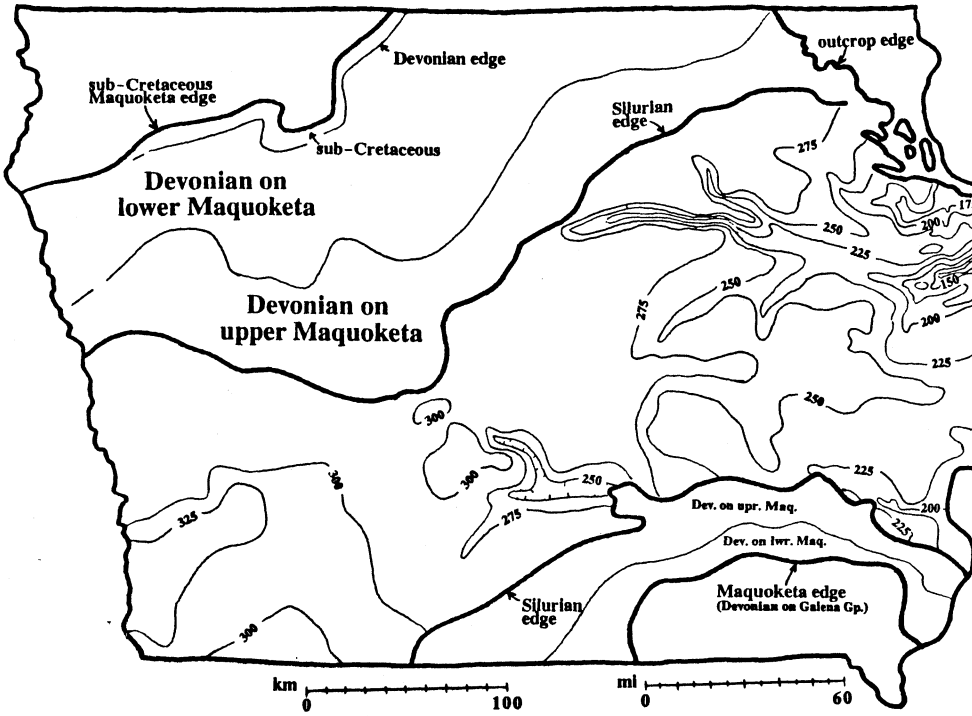


Figure 4. Generalized isopach map of Maquoketa strata for Iowa and Illinois. Isopached only where covered by Silurian rocks; shaded where beveled by post-Silurian stata.

Galena deposition.

Maquoketa

Thickness patterns of the Maquoketa Formation in Iowa and the Maquoketa Group in Illinois (Fig. 4) were strongly influenced by sub-Silurian erosional patterns developed on the upper Maquoketa surface. An eastward-sloping dendritic paleodrainage network is apparent in eastern Iowa (Fig. 5), which roughly parallels eastward-sloping structural trends seen in the East-Central Iowa Basin area. Additional sub-Silurian erosional effects are also seen in areas of northern Illinois; probable eastward-draining incision is identified in west-central Illinois (Brown County) and deep sub-Silurian incision of Maquoketa strata is seen in northeastern Illinois (Fig. 4). Aside from the local erosional channeling, regional thickness



contour interval 25 ft.

Figure 5. Isopach map of the Maquoketa Formation in Iowa; isopached only where covered by Silurian rocks. Note interpreted eastward-draining paleovalleys.

patterns of Maquoketa strata (Fig. 4) show general westward and southwestward thickening across northern Illinois and Iowa, slight thinning in central Illinois, and eastward thickening in southern Illinois. The Scales Shale (lower Maquoketa) is unusually thin along the axis of the LaSalle Anticlinal Belt in Livingston and LaSalle counties, Illinois, suggesting minor uplift of that structure during deposition of the Scales (Kolata and Graese, 1983, p. 14). Maquoketa thickness patterns are dissimilar to those noted in underlying Galena strata (Fig. 2), particularly with respect to southwest Iowa. Unfortunately, regional depositional facies of Maquoketa rocks are not coincident with the isopachous patterns, and the complex interplay of rates of subsidence vs sedimentation rates needs to be evaluated.

In general, Maquoketa strata are carbonate-dominated in western Iowa but become progressively more shale-rich eastward, in the direction of the distant Taconic source terrains (Witzke, 1983). Maquoketa rocks incorporate shallower depositional facies westward in Iowa, suggesting that east-central Iowa was bathymetrically depressed with respect to adjacent areas during Maquoketa deposition. This is best reflected by lateral facies relations in lower Maquoketa strata (Elgin Member in Iowa, lower Scales Shale in Illinois and eastern Iowa). Mixed skeletal carbonates (wackestones, packstones) characterize this interval to the west and northwest in Iowa and brown graptolitic shales dominate in east-central and southeast Iowa and adjacent northwestern Illinois (Fig. 6); an intermediate facies belt in central and northeastern Iowa is dominated by micritic carbonates, commonly with graptolites and trilobite-dominated faunas (Witzke, 1987a). The brown shale facies is interpreted to have been deposited in deeper stratified water than the coeval carbonate shelf environments to the west. An intracratonic shelf margin apparently was

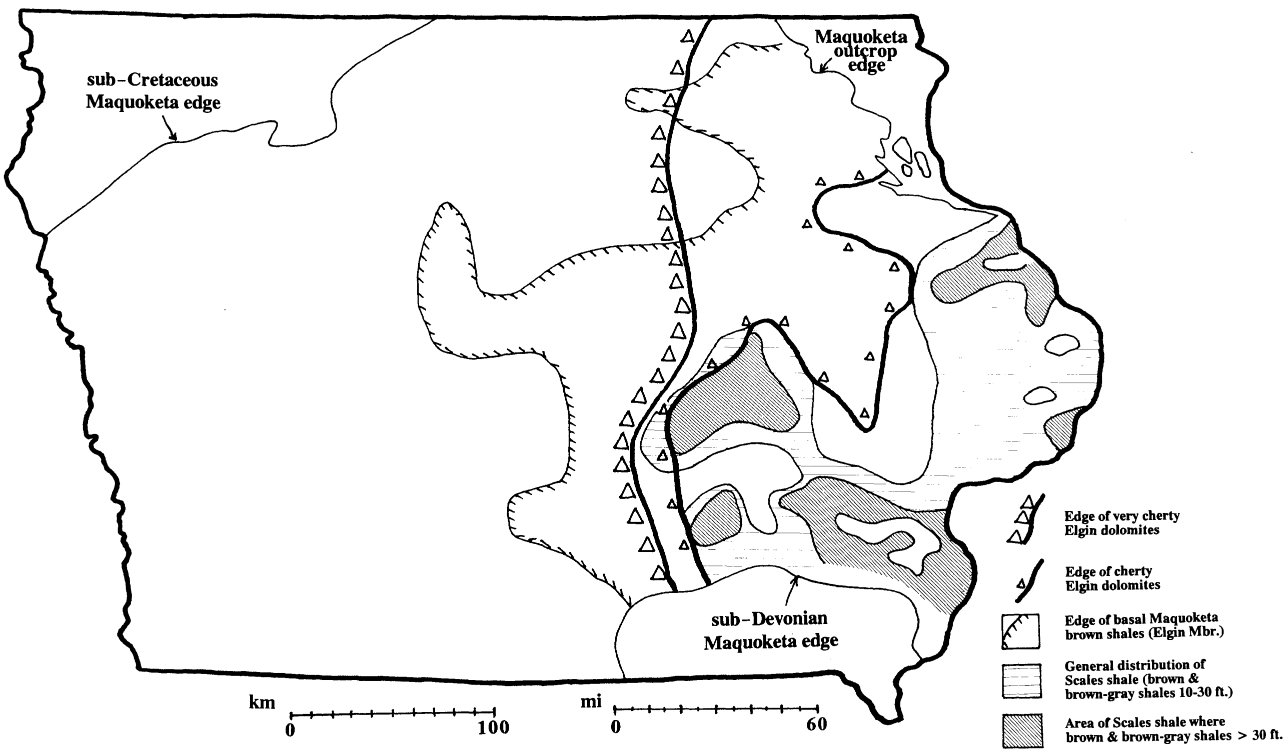


Figure 6. General distribution of lower Maquoketa brown shales in Iowa. Eastern edge of Elgin cherty carbonate facies noted. Maximum thicknesses of brown shales occur within the East-Central Iowa Basin area.

developed in eastern Iowa (see edge of cherty Elgin carbonates, Fig. 6), with brown shales deposited downslope from the shelf margin. The Scales Shale of northwestern Illinois grades eastward in north-central Illinois to dolomitized skeletal packstones-grainstones (Kolata and Graese, 1983. p. 12-13). As such, deeper-water graptolitic shale facies are rimmed on all sides by shallower-water shelf carbonates; the deeper facies occupy the area of the East-Central Iowa Basin and indicate that this basin area was a bathymetric depression during Maquoketa deposition. The significance of this basin is not reflected by the isopach patterns (Fig. 4).

In southwestern Illinois the influence of possible structural upwarping is seen. Silt and sand (Thebes Sandstone) was shed off the Ozark Dome during Maquoketa deposition, indicating that the dome was structurally elevated (Witzke, 1980). Maquoketa strata also thin to the west in that area (Fig. 4). Trending northeast across central Illinois, Maquoketa rocks are also observed to thin (Fig. 4), possibly indicating that the Sangamon Arch was also structurally elevated at that time. Brown organic-rich graptolitic shale facies, which characterize portions of the lower Maquoketa sequence in the East-Central Basin area, are less common in the southern area. If the deposition of such facies is depth-related (in stratified settings), then this observation may suggest that southern Illinois was an area of slightly shallower water during Maquoketa deposition than was the northern basin area.

Maquoketa rocks are observed to thicken southwestward in Iowa, reaching thicknesses in excess of 325 ft near the southwest corner of the state (Figs. 4, 5). This thickening, however, is not reflected by any interpreted deepening in depositional facies; instead, relatively shallow mixed carbonate facies dominate the Maquoketa across western Iowa. The thickening of Maquoketa rocks to the southwest probably reflects increasing subsidence in that area during Maquoketa deposition, but sedimentation apparently kept pace with subsidence preventing formation of a bathymetric basin. Nevertheless, southwestward thickening of Maquoketa strata in Iowa indicates that a locus of subsidence was initiated in that area during Maquoketa deposition, which contrasts markedly with the thin sequence of Galena rocks in the same area. This region of subsidence is interpreted to herald the initial phases of development of the North Kansas Basin (Witzke, 1983), a prominent stratigraphic and bathymetric basin area during Silurian deposition.

Evidence for the final phases of Ordovician deposition (Hirnantian) in the area has not been found in Iowa, but Hirnantian limestones are noted in southern Illinois and adjacent Missouri as well and Arkansas and Oklahoma (Amsden and Barrick, 1986). These Hirnantian strata were deposited in shelf environments inboard from the Ouachita continental margin. The absence of such strata in the northern area is due to either nondeposition or erosion; in either case it appears that differential subsidence permitted preservation of latest Ordovician strata only in the southern area. The northern area was subjected to erosion during the latest Ordovician and/or earliest Silurian with incision locally to 150 ft (Kolata and Graese, 1983).

DEPOSITIONAL CYCLES IN THE MIDDLE-UPPER ORDOVICIAN SEQUENCE

Although changing structural patterns during the Middle and Late Ordovician certainly affected sedimentation in the cratonic seas, broad-scale sedimentary rhythms seem to be superimposed on the epeirogenic effects. Apparent synchronous changes in depth-related sedimentation factors can be seen across broad areas of the craton, suggesting that global eustatic events may be involved. Such an idea certainly is not unreasonable--changes in global sea level through time are the rule, not the exception. Such changes should leave a profound impact on sedimentation patterns, especially in the epeiric seas. Therefore, the recognition of regional shallowing and deepening sedimentary episodes may ultimately prove to have application not only regionally, but also in our understanding of global events. An attempt is made here to outline some apparent trends in relative sea level in the Ordovician sequence of the Iowa-Illinois and Midcontinent areas.

Sloan (1987), Witzke (1987b), and Schutter (1987) recently proposed interpretations of relative changes of sea level for the Ordovician of the Midcontinent, and the latter author related these changes to a proposed global sea-level curve. In this report three large-scale depositional cycles are identified in the Middle and Upper Ordovician referred to in ascending order as, the St. Peter-Platteville cycle, the Galena cycle, and the Maquoketa cycle (Fig. 7). Each of these is marked by a basal transgressive episode and displays shallowing-upward sedimentary features in the upper part. Locally, unconformities separate these major cycles. The larger-scale cycles are, in turn, subdivided into smaller-scale transgressive-regressive depositional cycles which may correspond to the "fourth-order" T-R cycles of Busch et al. (1988). To eliminate excess verbiage, these smaller-scale cycles will be termed subcycles.

St. Peter-Platteville Depositional Cycle

The northward spread of St. Peter sand facies occurred during the transgressive phase of the first Middle Ordovician depositional cycle represented in the central Midcontinent area. The resumption of sedimentation ended an erosional hiatus of some 25 or 30 million year duration in the Iowa area. St. Peter deposition marked the initial phases of a depositional cycle here termed the St. Peter-Platteville cycle. Maximum sea level stand during this cycle is probably represented by deposition of the Mifflin and Grand Detour formations of the Platteville Group in Illinois and the lower to middle parts of the McGregor Member of the Platteville Formation in Iowa. These portions of the Platteville sequence contain an abundant and diverse stenohaline marine fauna in shaly skeletal wackestone (and packstone) lithologies. Carbonate facies became widespread during Platteville deposition; the expansion of carbonate facies implies that siliciclastic influx was reduced. This reduction was probably related to the foundering of source terrains as sea level rose to its cyclic maximum. The shallowing-upward phases of Platteville deposition will be discussed subsequently.

St. Peter-Glenwood (Ancell) Subcycle

The transition from St. Peter sand to Glenwood shale deposition is interpreted to represent a deepening-upward sequence associated with initial transgression of the Middle Ordovician seas into the Iowa area (Schutter, 1978). In general, this fining-upward sequence marks the change from nearshore and offshore sand deposition to more distal mud deposition as shorelines transgressed onto the Transcontinental Arch. The Glenwood contains a moderately diverse although poorly preserved benthic invertebrate fauna in northeast Iowa and southeast Minnesota (ibid.). The Glenwood includes organic-rich shale facies, locally with up to 25% organic carbon, in southeastern Iowa (Hatch et al., 1987). Oxygen-deficient stratified bottom waters probably developed in the deepening seas of that area. Glenwood shale deposition is interpreted to record maximum transgression of the first transgressive-regressive episode of deposition, here termed the St. Peter-Glenwood or Ancell subcycle (the Ancell Group includes the St. Peter-Glenwood interval in the Upper Mississippi Valley).

A minor regressive phase of deposition is interpreted for the upper Glenwood in Iowa. The Glenwood shales of southeastern Iowa are overlain by the Starved Rock Sandstone, which records a coarsening-upward sedimentary sequence. The progradation of the expansive Starved Rock Sandstone body is interpreted within the context of a shallowing-upward sedimentary sequence. Migration of sand facies above shale implies that a shift from lower-energy to higher-energy bottom currents had occurred, probably a response to shallowing depositional conditions. North of the main body of the Starved Rock Sandstone, the upper Glenwood shales in northern and central Iowa also display a coarsening-upward sequence in many sections. Thin sandstones or siltstones commonly cap the Glenwood Shale in those areas.

South of the Starved Rock Sandstone body in southern Illinois and adjacent Missouri, the transgressive St. Peter sandstone is overlain by fossiliferous carbonate strata of the Dutchtown Formation. The change to marine carbonate deposition is interpreted to record upward deepening. The fossiliferous Dutchtown is overlain by more restricted carbonate and carbonate-evaporite facies of the Joachim Dolomite. Occurrences of dessication features and evaporites in the Joachim record a general shallowing phase of deposition. The Joachim was probably deposited in restricted subtidal to supratidal settings. The transgressive-regressive sedimentary episode represented by the St. Peter-Dutchtown-Joachim sequence is equated with the St. Peter-Glenwood subcycle to the north. The St. Peter in northern Arkansas is unconformably overlain by Plattin carbonates (Craig, 1988), possibly suggesting that the regressive phase of the Ancell subcycle led to subaerial exposure in that area.

Platteville Subcycle

The spread of Platteville carbonate deposition in the Midcontinent records a major transgressive phase of sedimentation. The initial phases of this transgression are recorded by deposition of the Pecatonica Formation in Illinois and the Pecatonica Member in Iowa. The Pecatonica incorporates sand and reworked phosphatic grains at many localities. Basal Platteville strata in Minnesota apparently onlap and interfinger with Glenwood shales northward (Mossler, 1985). As noted, the Pecatonica is absent locally above anticlines in northwest Illinois, where the St. Peter is overlain by younger Platteville strata of the Mifflin Formation. These relations, although undoubtedly influenced by epeirogenic movements, are consistent with the development of a local unconformity developed during offlap of the Ancell subcycle followed by renewed transgression during the Platteville subcycle. Overstepping of the Pecatonica by the Mifflin suggests that the Mifflin represents a higher stand of sea level than the Pecatonica. A widespread hardground surface at the top of the Pecatonica in the Upper Mississippi Valley area may represent a transgressive discontinuity. Continued deepening coincident with Mifflin deposition was marked by an increase in diversity of benthic invertebrates.

Although the Platteville records a general transgressive-regressive sedimentary episode, smaller-scale sea-level changes are seen within the Platteville (Plattin) sequence of eastern Missouri, southern Illinois and Indiana as well as western Kentucky. Carbonate oolite in the Brickeys Member (lower Plattin) records a shoaling phase in that area; Templeton and Willman (1963) assigned the Brickeys to the lower part of the Mifflin. The Victory Member (about mid Plattin, upper Grand Detour) in eastern Missouri includes mudflat facies with evidence of subaerial exposure locally (Thompson and Spreng, 1988; and personal communication).

The upper Platteville interval in northern Illinois displays features suggestive of a general shallowing-upward sedimentary sequence. Calcarenitic lenses are common in the upper Grand Detour Formation (Willman and Kolata, 1978), which contrasts with the dominantly shaly wackestone lithologies in the underlying Grand Detour-Mifflin interval. A general upward shift from diverse brachiopod-mollusc dominated faunas in the Mifflin-Grand Detour to less diverse faunas containing tabulate and rugose corals in the overlying Nachusa-Quimbys Mill interval may also be depth-related. This may be analogous to paleocommunity successions seen in the Iowa Silurian (see M. Johnson's summary in this guidebook), where shallower-water coral-rich faunas contrast with deeper-water and more diverse brachiopod-echinoderm-mollusc faunas. Upper Platteville Quimbys Mill strata over the Wisconsin Arch in Rochelle (quarry east side of Hwy 51, NW NW SE sec. 13, T40N, R1E, Ogle Co.), north-central Illinois, include unfossiliferous laminated mudflat facies, clearly indicating significant depositional shallowing. Large coral heads up to 60 cm in diameter are also seen locally in the Quimbys Mill about 20 km northwest of the mudflat facies locality (near Stillman Valley, SW NE SW sec. 11, T24N, R11E). The top of the Platteville is marked by a hardground or ferruginized surface (diastem?) over much of the Upper Mississippi Valley area, and locally erosional incision is seen at the top of the Platteville (e.g., near South Wayne, Lafayette County, Wisc.).

Can the upper Platteville shallowing phase be recognized at other localities outside the Iowa-Illinois area? Probable Platteville equivalents in the upper Bromide of Oklahoma (Corbin Ranch beds) include birdseye carbonate fabrics (mudflat deposition) with an unconformity surface at the top (see Amsden and Barrick, 1988; Sweet, 1984). The Platteville subcycle in northern Arkansas (Plattin Fm) is also marked at the top by an unconformity (Craig, 1988). Extensive supratidal and intertidal environments are represented in probable Platteville equivalents (latest Blackriveran) in Kentucky (Cressman, 1973), southwestern Ohio (Stith, 1979), southeastern Indiana (Droste and Shaver, 1983), and New York (Walker, 1973). Within these areas Blackriveran strata show unconformable or disconformable relations with overlying strata. Shallowing deposition with local subaerial exposure, therefore, seems to be of interregional scope, and the regressive phase of deposition in the upper Platteville subcycle is most reasonably interpreted as eustatic in origin.

Galena Depositional Cycle

The Galena Group in the Upper Mississippi Valley includes two general lithologies: 1) shale and carbonate in the lower part, and 2) pure carbonates in the upper part. In general, shale content in the lower Galena (Decorah Shale) generally increases to the northwest toward source areas on the Transcontinental Arch (Witzke, 1980). Continued transgression of these source areas during upper Galena deposition reduced siliciclastic influx and permitted the spread of widespread subtidal carbonate facies throughout the Midcontinent. Depositional shallowing in the upper Galena Group is reflected by a relative increase in grainstone lithologies. This sequence records a broad-scale transgressive-regressive (T-R) depositional episode here termed the Galena cycle. Smaller-scale T-R depositional patterns are present, and are included as subcycles within the Galena cycle.

While the Galena was being deposited, the east coast of North America (Laurentia) underwent continental collision (Stillman, 1984). This was the setting for the development of explosive subduction-related arc volcanics which extended from Alabama to maritime Canada. Prevailing winds carried ash from island-arc volcanoes out over the Midcontinent where it fell into widespread epeiric seas. At least 13

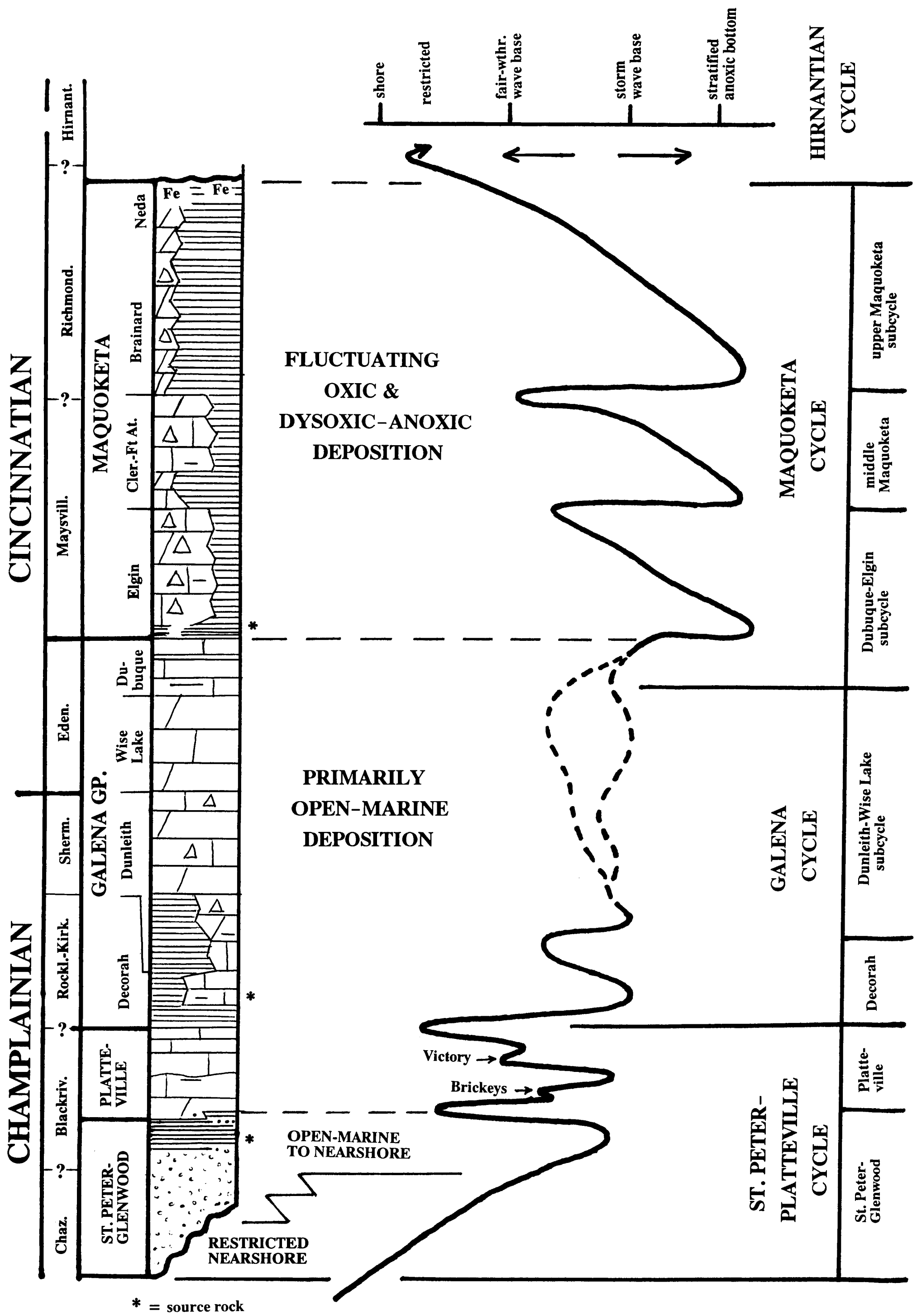


Figure 7. Qualitative sea-level curve for Champlainian-Cincinnatian deposition in the Iowa - northern Illinois area. Dashed where relative depth relations are unclear.

K-bentonite beds (altered volcanic ash) are preserved in Galena rocks of the Mississippi Valley. Some beds have been correlated from southeastern Minnesota through the Illinois Basin to the southern Appalachians, a distance of 800 miles (1300 km) (Huff et al., 1986; Kolata et al., 1986).

Decorah Subcycle

Galena deposition was initiated in most of the Upper Mississippi Valley area with widespread shale and carbonate deposition of the Spechts Ferry Member (formation in Illinois) and its equivalents in the undifferentiated Decorah Shale of northern Iowa and Minnesota. However, this initial phase of Decorah deposition is not present over the trend of the Wisconsin Arch, where overlying Decorah carbonate strata of the Guttenberg Formation (member in Iowa) overstep the Spechts Ferry edge to lie directly on the Platteville (Kolata et al., 1986). The Guttenberg, therefore, is interpreted to represent a further deepening in the seaway. The regional expansion of Guttenberg subtidal carbonates is also interpreted to correspond to continuing foundering of siliciclastic source terrains coincident with transgressive deepening. Maximum transgression during this deepening phase is interpreted to coincide with deposition of the lower Guttenberg, an interval containing diverse and well-preserved benthic faunas. Guttenberg skeletal wackestones are interbedded with organic-rich shales (containing up to 43% organic carbon) over a broad area of the Upper Mississippi Valley area, indicating that periodic bottom anoxia affected deposition. These organic-rich shales record a major excursion in stable carbon isotopes (Hatch et al., 1987) that apparently coincides with the transgressive phase of Decorah sedimentation. Major global positive $\delta^{13}C$ excursions are believed to reflect an increase in the storage of isotopically light organic carbon in the sedimentary record, thus depleting the oceanic reservoir of ^{12}C. This is consistent with decreased oceanic ventilation and general eustatic deepening of the world's oceans.

The Guttenberg and lower Ion members of the Decorah Formation in the type area of northeast Iowa display a general shallowing-upward depositional sequence. Whole-shell skeletal wackstones dominate in the lower Guttenberg, but broken grains become abundant in the middle Guttenberg. The upper Guttenberg and lower Ion contains interbeds of broken and abraded grain packstones. The Guttenberg also displays a general upward increase in intergrain cements. This sequence reflects increasing bottom energy developed during the regressive phase of Decorah sedimentation. Progradation of siliciclastic sediments during regression is seen by the southward spread of Ion shales in Iowa and by the presence of sandy Buckhorn (= Ion) strata in northern Illinois. Where the Guttenberg oversteps the Platteville along the trend of the Wisconsin Arch, it is characterized by a thinner sequence with numerous coquinoid horizons. At Rockton, Illinois (SE NW NE sec 25, T46N, R1E), it is only 50 cm thick and is dominated by brachiopod coquinas displaying mega-ripple bedforms.

Southward in eastern Missouri and adjacent Illinois the Guttenberg is erosionally truncated beneath the Dunleith (Kimmswick) Formation (Kolata et al., 1986, p. 24). This erosional hiatus corresponds to the regressive phase of the Decorah subcycle, and suggests that the area around the Ozark Dome was subaerially exposed at that time. Shallowing-upward depositional sequences equivalent to the Decorah in other geographic areas have not been confirmed, although the post-Roughlock unconformity in Wyoming and the post-Tyrone unconformity in the Cincinnati region (see Sweet, 1984) may equate to the post-Decorah unconformity in Missouri. If so, the Decorah subcycle may be of interregional scope.

Dunleith-Wise Lake Subcycle

Expansion of subtidal carbonate facies (Dunleith and Wise Lake formations) above the Decorah suggests that siliciclastic source terrains on the Transcontinental Arch foundered coincident with renewed transgression. This is reflected by the the upward decrease in shale within the Cummingsville Formation (a lower Dunleith equivalent) of Minnesota. The transgressive phase of this second subcycle in the Galena Group, the Dunleith-Wise Lake subcycle, is also marked by overstepping of underlying Ordovician strata; the lower Kimmswick (equivalent to the Dunleith) unconformably overlies the Decorah in parts of eastern Missouri and adjacent Illinois, and the Dunleith oversteps the Decorah edge to lie directly on the Platteville along the trend of the Wisconsin Arch in eastern Ogle County in northern Illinois.

Maximum transgression of the Dunleith-Wise Lake subcycle may have occurred roughly coincident with deposition of the Rivoli-Sherwood interval in the middle Dunleith Formation (also see Sloan, 1987, p. 19). It was at that position that influx of Decorah-style shales (seen in underlying upper Decorah and Cummingsville formations in northern Iowa and Minnesota) was significantly reduced and purer carbonate facies (although still argillaceous) became more widespread (see Stone, 1983). However, one of the authors (DRK) suggests that maximum transgression was coincident with Wise Lake deposition; the purest carbonate rocks in the Galena Group are noted in the Wise Lake, suggesting maximum inundation of siliciclastic source terrains during deposition of that interval. Differences in interpretation are shown by dashed lines in Figure 7. Carbonates of the Dunleith-Wise Lake in Iowa are dominated by bioturbated skeletal mudstones and wackestones deposited in subtidal settings, but thin skeletal packstones-grainstones punctuate the sequence which record episodic storm events in the epeiric seas (Delgado, 1983). Stratigraphic equivalents to the south in Illinois and eastern Missouri (Kimmswick Limestone) are dominated by calcarenitic packstone and grainstone (in part crossbedded) lithologies which were deposited in higher-energy and shallower environments than those in the northern area.

Minor sea-level fluctuations during deposition of the Dunleith-Wise Lake sequence may be represented by contrasting sedimentation patterns of receptaculitid-bearing (calcareous green algae) and non-receptaculitid-bearing intervals. In general, the lower, middle, and upper receptaculitid "zones" (for stratigraphy see Levorson and Gerk, 1983, and Levorson et al., 1987) in the Dunleith-Wise Lake contain prominent receptaculitid algae, and dasyclad algal grains are represented in thin-section (Bakush, 1985). The presence of calcareous algae in these intervals indicates deposition within the photic zone. The absence of calcareous algae in portions of the lower (St. James-Beecher) and middle Dunleith (lower Rivoli) suggests sedimentary variations during deposition of the sequence, possibly related to sea-level fluctuations, depth of the photic zone, and turbidity factors. Further studies are needed to contrast the algal and non-algal intervals, both depositionally and paleontologically.

The regressive phase of the Dunleith-Wise Lake cycle can best be seen upward through the Wise Lake and basal Dubuque sequence. The Wise Lake Formation in northeast Iowa displays a general upward increase in receptaculitids and dasyclad grains (Bakush, 1985), and grainstone units ("SCB"-sparry calcarenite beds of Levorson and Gerk, 1972) are more numerous in the Wise Lake than in the underlying Dunleith. However, in northern Illinois and southern Wisconsin tempestites appear to be more common in the Dunleith than in the Wise Lake. Faunal changes are also observed upward in the Wise Lake; an interesting fauna containing large gastropods (including Maclurites) is conspicuous in upper Wise Lake strata. Vertical burrows (especially Paleosynapta) are present in the uppermost Wise Lake and basal Dubuque. The sedimentary and biotic changes observed upward in the Wise Lake are interpreted as a shallowing-upward depositional sequence. Sloan (1987) likewise interpreted the Wise Lake to be a shallowing-upward sequence.

The Dunleith-Wise Lake subcycle is capped by an erosional unconformity in southern Illinois and adjacent Missouri. Kimmswick calcarenitic strata were interpreted to be truncated southward in that area by Templeton and Willman (1963), and are overlain unconformably (locally channeled) by the Cape Limestone in the southern area. Kimmswick strata are also unconformably overlain by Cape-equivalent beds ("Fernvale") in northern Arkansas (Craig, 1988). The intra-Viola Springs unconformity of Sweet (1983) in Oklahoma probably corresponds to the post-Kimmswick unconformity in Missouri. Therefore, it seems likely that the regressive phase of the Dunleith-Wise Lake subcycle (and the Galena cycle as well) is represented by a shallowing-upward depositional sequence in the northern area and by a widespread erosional surface in the southern area. The interregional character of these temporal changes is interpreted to record a eustatic lowering of sea level during the latter phases of the Galena cycle.

Maquoketa Depositional Cycle

A vast influx of siliciclastic sediments into the Midcontinent area occurred following deposition of the Galena Group. Unlike the shales of the lower Galena Group (Decorah) which were derived from Transcontinental Arch source terrains, the Maquoketa shales were derived primarily from distant eastern sources associated with uplift of the Taconic orogen. Some Maquoketa siliciclastic influx was also derived from the area of the Ozark Dome (Thebes Sandstone) and perhaps minor amounts from the Transcontinental Arch. The change from carbonate-dominated deposition of the Galena Group to shale-dominated deposition of the Maquoketa Group in eastern Iowa and Illinois is interpreted to mark the beginning of a new depositional cycle here termed the Maquoketa cycle. Evidence for upward-deepening deposition is first seen in the upper part of the Galena Group (Dubuque Formation), but maximum transgression probably occurred during deposition of portions of the overlying Maquoketa Group. Three transgressive-regressive depositional episodes are interpreted within the Dubuque-Maquoketa sequence, here described as subcycles within the Maquoketa cycle (see Fig. 7).

Dubuque-Elgin (lower Maquoketa) Subcycle

The shallowing phase of Wise Lake-basal Dubuque deposition, marked by abundant dasyclad grains and multiple grainstone beds, is succeeded by an interpreted upward-deepening depositional sequence in the Dubuque and lower Elgin (lower Scales Shale in Illinois) interval. Dubuque carbonates are characterized by regularly-bedded crinoidal wackestone-packstone beds interbedded with thin shales. The Dubuque, above its basal portion, displays a notable absence of calcareous algae (perhaps due to influx of siliciclastics), contrasting with underlying Wise Lake strata, and also shows an increase in trilobite grain frequency and matrix mud over that seen in the Wise Lake (Bakush, 1985). These observations suggest that the Dubuque was deposited in deeper and quieter-water depositional conditions than the underlying Wise Lake. Ludvigson (1987) suggested that the depositional and diagenetic features of the Dubuque "may record a gradual deepening of Galena Group carbonate deposition, marking a transition to depositional realms characteristic of overlying Maquoketa strata." The Dubuque commonly displays an abundance of organic-walled microfossils, and their preservation implies that lower bottom oxygenation contrasted Dubuque from Wise Lake deposition. Anoxic and dysoxic bottom conditions characterized much of succeeding Maquoketa deposition, and the appearance of such conditions in the Dubuque seemingly heralds the advance of anoxic bottom waters in the area. A widespread brown graptolite- and trilobite-bearing shale in the upper Dubuque of northern Iowa is dissimilar to any known shale in the Galena Group but is largely indistinguishable from many shales in the Maquoketa sequence.

Continued and significant depositional deepening is interpreted above the Dubuque in the Upper Mississippi Valley area, where organic-rich graptolite-bearing shales and phosphorites of the lower Maquoketa (Scales Shale and Elgin Member) abruptly overlie the top of the Dubuque south of the Minnesota line. These basal Maquoketa facies are interpreted to have been deposited under dysoxic to anoxic conditions in a stratified seaway (Witzke, 1987a) and record a major transgressive episode. Although some previous workers have interpreted the Dubuque-Maquoketa contact to be unconformable, there is no evidence of subaerial exposure or erosional truncation at the contact. Instead, the occurrence of multiple hardgrounds and phosphorites in the basal Maquoketa suggests transgressive sediment starvation marked by phosphatic sediment enrichment and reworking. The widespread deposition of abundant organic matter, apatite, and pyrite in the lower Maquoketa suggests that deepening was sufficient to establish a stratified water mass with oxygen-poor phosphate-rich bottom waters. Maximum transgression is interpreted above the basal Maquoketa within an interval of laminated graptolitic brown shale with sparse to absent benthic faunas (Argo Fay bed of Illinois) seen in the area of the East-Central Iowa Basin. The basal phosphorites and brown shales of the lower Maquoketa are replaced laterally by skeletal carbonates deposited in shallower depositional settings, although the basal brown shales have considerable geographic extent in eastern Iowa (see Fig. 6) and northwestern Illinois.

The Elgin carbonate sequence in the lower Maquoketa of Iowa records a general shallowing-upward trend that marks the regressive phase of the Dubuque-Elgin subcycle. In northern and central Iowa the Elgin displays an upward shift from 1) sparsely burrowed graptolite- and trilobite-bearing mudstones, to 2) mixed skeletal wackestones, to 3) mixed skeletal wackestones and packstones (the shallowest phase)(Fig. 8). In the East-Central Iowa Basin area the shift from brown shales to greenish-gray shales likewise is interpreted to record a shift from deeper anoxic-dysoxic deposition to slightly shallower dysoxic deposition (still with stratified bottom waters). The Elgin surface in central Iowa is marked by a hardground (on skeletal packstone lithologies) that is abruptly overlain by laminated unfossiliferous shaly carbonates of the succeeding middle Maquoketa subcycle.

The transgressive phase of the Dubuque-Elgin subcycle is marked by deposition of the Cape Limestone (and "Fernvale") above the upper Kimmswick unconformity in the southern area. The Cape probably correlates biostratigraphically with the Dubuque-lower Maquoketa sequence to the north (Sweet et al., 1975). It is suggested here that the Dubuque-Elgin subcycle corresponds to deposition of the Cape or "Fernvale" interval in Missouri, Illinois, and Arkansas and the upper Viola Group (including the Welling Formation) in the Oklahoma area. The initiation of extensive carbonate deposition in the western United States (Fremont, lower Bighorn) may also correlate with transgression of the Dubuque-Elgin subcycle (see correlations of Sweet, 1984). An accelerated transgressive episode marked by deposition of the lower Maquoketa in the Iowa area is probably of eustatic origin. Schutter (1987) recognized a major transgressive event at or near the Caradoc-Ashgill boundary that is marked by drowning of carbonate shelf sedimentation, followed by organic-rich black or brown shale deposition, on four continents. This apparent deepening event is equated with the transgressive phases of the lower Maquoketa subcycle.

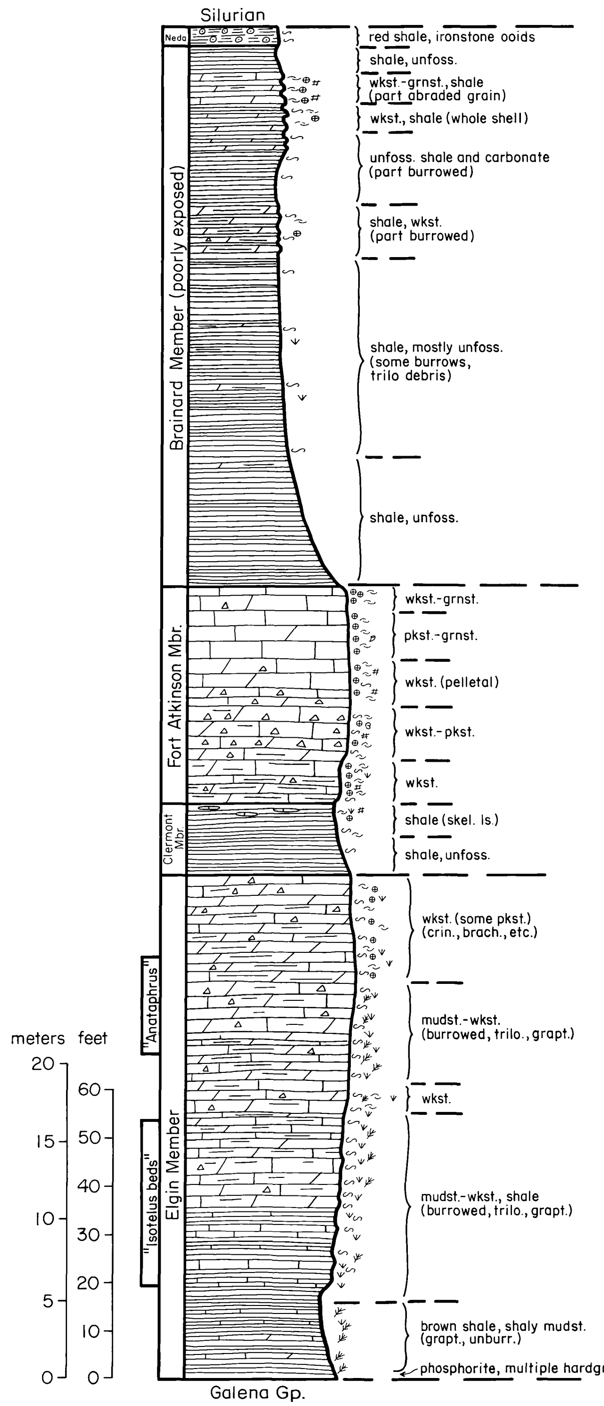


Figure 8. General composite section of the Maquoketa Formation as exposed in northeast Iowa (primarily Fayette County).

Middle Maquoketa Subcycle

A second Maquoketa T-R subcycle is interpreted for the Clermont-Fort Atkinson sequence in the middle part of the Maquoketa Formation of Iowa and is termed the middle Maquoketa subcycle (Witzke, 1987b). It is marked by an apparent deepening trend in the Clermont Shale of eastern Iowa, which is characterized by unfossiliferous shale in the lower part. It was probably deposited as dysoxic bottom waters spread across the area. In central and western Iowa the transgressive phase is marked by deposition of non-burrowed, faintly-laminated, graptolite-bearing argillaceous carbonates; it occurs above a pyrite and phosphate-encrusted hardground at the top of the Elgin.

The regressive phase of the middle Maquoketa subcycle is marked by a shallowing-upward sequence in the Fort Atkinson Member in Iowa. The Fort Atkinson is represented by discontinuous carbonate "banks" within the shale-dominated sequence of eastern Iowa, northern Illinois and northwestern Indiana and by continuous carbonate shelf deposits to the west in central and western Iowa. The Fort Atkinson carbonates display a shallowing-upward sequence characterized, in ascending order, by 1) whole-shell skeletal wackestones (burrowed graptolitic carbonate mudstones to the west), 2) mixed skeletal wackestones and packstones, and 3) skeletal packstones and grainstones (locally at the top)(see Fig. 8 and Torney, 1983). Within shale-dominated intervals in eastern Iowa that lack the prominent Fort Atkinson carbonate banks, the general sequence in ascending order includes 1) unburrowed shale, 2) burrowed shale, and 3) shale interbedded with skeletal wackestones/packstones (locally grainstones).

Correlation of Clermont-Fort Atkinson strata with other geographic areas in cratonic North America is not yet clear, although biostratigraphic studies of the diverse fauna should clarify correlations. Therefore, it cannot yet be demonstrated whether the middle Maquoketa cycle is of interregional extent. Nevertheless, comparisons between western Iowa and Kansas seem to indicate that the Viola-Sylvan contact to the southwest occurs at roughly the same position as the Elgin-Clermont contact in Iowa. If so, the lower Sylvan sequence in Kansas may equate with the middle Maquoketa subcycle; brown and green-gray shales occur in that area. Sweet (1984) correlated a widespread sub-Herald/Priest Canyon unconformity in the western United States to a position near the middle part of the Maquoketa. This unconformity may have formed during the regressive phase of the middle Maquoketa or preceding subcycle. Likewise, the extensive development of evaporite units in the upper Red River sequence of the Williston Basin (Fuller, 1961) may relate to a major regressive phase of regional sedimentation, possibly equivalent to the middle Maquoketa subcycle. T-R event stratigraphy offers potential means for independent correlation in cratonic settings, but biostratigraphic studies also remain essential.

Upper Maquoketa Subcycle

The Brainard-Neda interval is interpreted as a third transgressive-regressive subcycle within the upper Maquoketa sequence. This interval is shale-dominated in eastern Iowa and much of Illinois and generally overlies fossiliferous Fort Atkinson carbonates over much of its extent. To the west in Iowa it incorporates progressively more carbonates, and it becomes carbonate-dominated in southwest Iowa and portions of adjacent Missouri-Nebraska (Witzke, 1987b). The lower Brainard shale interval is interpreted to record a transgressive phase of deposition analogous to the previous two Maquoketa subcycles. The lower Brainard is dominated by generally unfossiliferous burrowed to non-burrowed shale over most of its extent, and two phosphatic horizons are noted in northeastern Illinois (Kolata and Graese, 1983). Graptolitic brown shales are locally present in the lower Brainard of central Iowa. Deposition of the lower Brainard is interpreted to have occurred under generally dysoxic bottom conditions within a stratified seaway. The contrast between lower Brainard unfossiliferous shale and fossiliferous packstones-grainstones of the underlying Fort Atkinson is striking and suggests a significant deepening event during lower Brainard deposition.

The upper Brainard interval is interpreted to record a shallowing-upward depositional sequence, as displayed at STOP 3 (Wacker, Illinois) for this field trip and at other localities in eastern Iowa. In general, the upper Brainard shales show an upward increase in burrowing fabrics; interbedded carbonates display the shallowing trend best, in ascending order: 1) largely unfossiliferous argillaceous carbonates with scattered burrows, trilobites, or other fossils in some beds, 2) fossiliferous beds with whole-shell wackestone fabrics, and 3) skeletal packstone and grainstone beds (Fig. 8). The upward increase in packstone-grainstone beds within the shale sequence is interpreted to reflect the increasing frequency with which storm wave base affected the bottom sediments as the seas shallowed. Packstone-grainstone beds are most abundant near the top of the Brainard and may indicate bottom shallowing above general fair-weather wave base.

The Brainard is overlain by red shales of the Neda Formation (or member) in areas of Iowa, Illinois, and Wisconsin where sub-Silurian erosion was minimal. The Neda locally contains ferruginous and phosphatic ooids and clasts. The Neda is apparently conformable with the underlying Brainard and contains a sparse fauna of Brainard aspect. It does not seem to represent a separate T-R unit, and is interpreted as the final phase of nearshore deposition of the upper Maquoketa subcycle. The seaways probably withdrew from the northern area following Brainard-Neda deposition, and a significant erosional hiatus separates these strata from overlying Silurian units.

The Brainard Shale is Richmondian in age, but its exact correlation with other areas in North America is not yet clear. Therefore, it is difficult to evaluate whether or not the upper Maquoketa subcycle is of interregional extent. The transgression of marine shales and carbonates of the Stony Mountain Formation above Herald evaporites in the Williston Basin (Kendall, 1976) may equate with the transgressive phase of this subcycle, and the deposition of upper Stony Mountain and lower Stonewall evaporites in that area may correspond to the regressive phase of the upper Maquoketa subcycle. Additional biostratigraphic studies are needed to further evaluate these suggestions.

HIRNANTIAN DEPOSITS

Although Hirnantian (uppermost Ordovician) deposits are apparently absent across Iowa and northwestern Illinois, Hirnantian skeletal and oolitic carbonate strata are present in the southern area (Amsden and Barrick, 1986). These strata may represent the final phase of deposition of the upper Maquoketa subcycle (they grade laterally into Sylvan shales in Oklahoma) or alternatively may record a separate T-R cycle in the latest Ordovician. The latter interpretation is favored in this report, as Hirnantian strata overlie an unconformity surface in at least several areas of the North American craton. Hirnantian rocks unconformably overlie the "Fernvale" in northern Arkansas (ibid.), and are locally disconformable on the Maquoketa Shale in eastern Missouri (Thompson and Satterfield, 1975). The Late Ordovician onlap of Precambrian rocks along the western margin of the Hudson Bay Basin reported by Johnson and Witzke (1987) is now known to be of probable Hirnantian (Gamachian) age, and is interpreted to represent the transgressive phase of a latest Ordovician (Hirnantian) T-R cycle.

CONCLUSIONS

A highly interpretive summary of structural and depositional patterns for the Middle and Upper Ordovician of the Iowa-Illinois area is presented to stimulate further studies, and many interpretations outlined in this report will undoubtedly need significant revision. Nevertheless, gross changes in structural patterns, as reflected by variations in regional thickness and lithofacies patterns, show that significant structural reorganization occurred during the Middle and Late Ordovician. Sauk-type structural patterns were still present during the initial phases of Tippecanoe deposition in the area, and are typified by thickness trends seen in the Platteville interval. These patterns were significantly altered during subsequent Galena deposition, primarily as a result of declining subsidence rates within the New Madrid Rift Complex of the proto-Illinois Basin. Uplift of the Ozark Dome also affected thickness and lithofacies during deposition of the Galena and Maquoketa. Downwarping was initiated in the East-Central Iowa Basin during Galena and Maquoketa deposition. The dramatic shift in structural patterns seen in the Middle Ordovician of the Midcontinent area is probably related to large-scale tectonic changes that affected broad areas of the North American continent. Perhaps orogenic changes along the eastern margin of the continent influenced structural development in the cratonic interior.

Shallowing and deepening depositional trends seen in the Iowa-Illinois area record bathymetric changes in the Midcontinent seaways that may be of eustatic origin. Three large-scale transgressive-regressive (T-R) episodes are recognized and named, in ascending order, the St. Peter-Platteville cycle, the Galena cycle, and the Maquoketa cycle. Smaller-scale transgressive-regressive subcycles are recognized within these larger cycles. T-R units in the Middle and Upper Ordovician of the Iowa-Illinois area apparently correlate across large areas of the continent and may ultimately prove useful for defining an event stratigraphy of continental or global scale.

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EARLY SILURIAN CARBONATES FROM THE UPPER MISSISSIPPI VALLEY AREA AS A KEY TO PLATFORM DEVELOPMENT ON A CRATONIC SCALE

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ABSTRACT

Cyclic carbonates in the Lower Silurian of the Upper Mississippi Valley area provide a key to the reconstruction of craton-wide platform seas which inundated at least 65% of the ancestral North American continent. The pattern typical of the East Iowa Basin is one of recurrent coral-stromatoporoid, pentamerid, and stricklandiid communities. Through association with fossil algae and other factors, this pattern is attributed to fluctuating sea levels. The deepest-water communities were the stricklandiid communities and the evolution of their key brachiopod members enables peak transgressions to be confidently dated. Comparison of Iowa cycles with carbonate cycles from other regions, such as the Michigan and Williston basins, reveals a set of overlapping but progressively shallower patterns. A series of four transgressive-regressive events may be correlated to many widespread areas. Taken as a whole, the distribution of staggered cycles provides a template for the interpretation of paleobathymetry on a cratonic scale. Nothing approaching the vast epicontinental seas of the past exists today but basic principles of stratigraphy and paleoecology help bring them alive for us.

INTRODUCTION

Why study fossils when we have so much left to learn about the living? Why study ancient sea floors when so much remains to be discovered beneath the waves of our present oceans? These are tough questions typical not only of doubtful students beginning a course in historical geology, but also of doubtful authorities controlling ever dwindling funds for potential research projects in basic paleontology and stratigraphy. I am surely a romantic, but I believe our understanding of living ecosystems is enhanced immensely by even an incomplete impression of their development through geologic time. Paleoceanography offers not only the opportunity to study marine environments and their biotas in the context of time, but also the chance to investigate features such as large epicontinental seas poorly represented today.

This review deals both with carbonate-platform seas which occupied the present area of the Upper Mississippi Valley during Early Silurian times, as well as their bathymetric relationship in a setting of cratonic dimensions. As a beginning paleontology student at the University of Iowa during the late 1960s, I was introduced to the work by Ziegler (1965) on the Early Silurian paleogeography and distribution of marine communities in the Welsh Basin. This was for me highly stimulating fare, which brought to life the Lower Silurian of my native eastern Iowa. Brachiopods such as Pentamerus and Stricklandia, which defined some of the Welsh communities, also are abundant in Iowa outcrops. Silurian faunas were highly cosmopolitan, a fact appreciated long before the Silurian System was even named. Buckland (1822), for example, realized the affinities of English fossils with those collected as far afield as Estonia, Gotland (Sweden), northern Michigan, and central Saskatchewan.

With modifications to the community approach advocated by Ziegler, my first serious research was on the Silurian of Iowa (Johnson, 1975; 1977; 1979; 1980). This work emphasized sea-level fluctuations as the main control of recurrent community patterns and I was inevitably led to search for similar patterns in other places such as the Michigan Basin (Johnson and Campbell, 1980; Johnson, 1981) and the Williston Basin (Johnson and Lescinsky, 1986). I believe far too much is made of basin boundaries, which often are tectonically overprinted. It also seems to me that there is a basin mind-set, in which some workers band together and confine themselves to a basin as a career. The Silurian faunas which moved from basin to basin paid no attention to these boundaries. Consequently, I have been more interested in trying to find out what different depositional regions shared in common rather than what made them unique from one another. The same may be said for Silurian basins outside North America. It has been my good fortune to study regions with well developed Lower Silurian strata in Norway, Estonia, China, and Australia. Despite sometimes profound scenic and cultural differences, I have always felt "at home" in the field or museum with familiar fossil assemblages which may as well come from Zwingle, Iowa.

This contribution and my participation in the conference on the Plum River Fault Zone across the Upper Mississippi Valley, represent both an intellectual and physical homecoming for me. I have always felt that the Iowa-Illinois-Wisconsin area not only launched me on what has been an exciting research project covering several independent cratons, but that it also is an important key to unlocking the structure of carbonate-platform seas on a scale much larger than any existing today. The task I set for myself here is rather unconventional. Very little new data beyond those already reported in my local and inter-regional summaries (Johnson, 1983; 1987) will be added in this report. I am certain that had I started my research in Winnipeg, Oslo, or Sydney, however, the key would have been much more difficult to find and I would not have come nearly so far. In a way more personal than normal for journal articles, I hope to explain why the Lower Silurian of the Upper Mississippi Valley is so very special.

EVENT STRATIGRAPHY

The greatest challenge in stratigraphy is to identify events of an inter-regional to global character. Bentonites are precise tools but they are uncommon on a truly inter-regional scale. Geomagnetic reversals and rare-earth anomalies are global in scale but they have not been convincingly demonstrated as useful tools for the Paleozoic. Events marked by changes in sea level, however, have attracted increasing attention of stratigraphers working on all systems of the Paleozoic. The usual form of expression comparing one locality or region with another is the sea-level curve. In order to adequately document events of at least inter-regional significance, three important features are required: 1) cyclicity must be evident, 2) repetitive faunas or other depositional traits should be clearly associated with changes in water depth, and 3) the correlation of events from one locality to another must be confirmed through use of a standard time scale with independent tie points. Data from the Lower Silurian of the Iowa-Illinois-Wisconsin area fill each of these criteria especially well.

Cyclicity

The sea-level curves I have drawn from the Lower Silurian apparently arouse suspicion. I am sometimes asked: "Where do those curves come from?" It may seem almost as if a rabbit were plucked from a magician's hat. The first step in producing such a curve is simply the recognition of repetitive events in an objective stratigraphic sequence. While I have discovered some previously unrecognized events, by far most of the stratigraphic patterns that I utilized in my curves were already described by other stratigraphers long ago. I have learned, in fact, to pick out and follow many of the significant observations made by my predecessors. In Iowa, for example, Wilson (1895) was struck by the clear recurrence of coral beds. He distinguished between the "Lower Coralline" and "Upper Coralline" beds, corresponding in today's nomenclature to the Sweeney and Picture Rock members of the Hopkinton Dolomite (Johnson, 1983). The succession of fossil communities recorded in the Hopkinton and Scotch Grove formations provide some of the sharpest and most predictable patterns I have encountered anywhere. This is probably due in large part to the frequent preservation of brachiopod populations in their growth position (Johnson, 1977).

Whittlesey (1851, p. 188–189) was early to grasp the meaning of recurrent coral beds in the Silurian of northern Michigan and he even proposed that an alternately rising and falling sea bed caused the cyclic destruction and reintroduction of corals limited to a specific depth range. In a summary of over 60 years of field work in northern Michigan, Ehlers (1973, p. 132) made the following observations: "The lower and upper Pentamerus dolomite and the thin, even-bedded gray dolomites between them are remarkable for their continuity from St. Martin's Island, Big Bay de Noc, to Drummond Island." For those unfamiliar with the Great Lakes, the distance described is over 200 km. Ehlers goes on to state his opinion that the paired Pentamerus beds probably extend in opposite directions to Wisconsin and Ontario. This is a highly valuable observation based on the patient accumulation of data year after year in a terrain more difficult than Iowa's. When it is realized that two additional beds rich in other pentamerid genera are widely distributed from northern Michigan to Manitoulin Island (Ontario) and that coral beds frequently occur stratigraphically amongst these four, one begins to be impressed with the notion of cyclicity.

Equally distinctive events have been noted by stratigraphers in the Williston Basin. In particular, early subsurface studies of the Upper Ordovician and Lower Silurian revealed widespread marker beds distinguished by their red coloration and content of frosted quartz grains (Porter and Fuller, 1959). The subsequent recognition of these thin units in the Manitoba outcrop belt (King, 1964), made it possible to better assess the cyclic nature of intervening strata. In Iowa, no single part of a cycle is more distinctive than any other. In Michigan and Manitoba, it was necessary to tease out the details of rock intervals between prominent marker beds. By itself, the recognition of cyclicity bears no overt revelation of cause. Cyclic patterns may be due to local, regional, or global influences. Events are easily placed in relative order, but may have been diachronous or more synchronous in timing.

Water depth

Ziegler (1965) described marine communities from the Lower Silurian of Wales and the Welsh Borderland, which he was able to map in zones parallel to a migrating shoreline. He reasoned that community distribution was regulated by factors associated with variation in water depth and distance from shore. Because some of the same communities are well represented in the Upper Mississippi Valley, it is logical to invoke changes in water depth as the source of local cyclicity. There are, however, some important differences between the Welsh and East Iowa basins. The Welsh Basin was the site of a genuine topographic depression with a paleoslope demarked by a coastline. It was filled with clastic sediments. The East Iowa Basin developed only modest topography during deposition of Early Silurian carbonates, with no link to a shoreline. It was much more a simple platform than a topographic basin. While sea-level fluctuation was the likely agent of cyclicity in Iowa, it was still important to demonstrate independent evidence for this relationship.

One line of evidence for depth control is the orderly sequence of paleocommunities in stratigraphic succession. A drop in sea level is indicated by beds containing a diverse fauna including stricklandiid brachiopods, followed by beds dominated by pentamerid brachiopods, followed in turn by beds filled with tabulate corals and stromatoporoids. This pattern (Fig. 1) is typical of strata from the base of the Farmers Creek Member to the top of the Picture Rock Member (Hopkinton Dolomite). A rise in relative sea level is indicated by exactly the reverse sequence, as shown by the strata of the Picture Rock Member (Hopkinton Dolomite) through the overlying Johns Creek Quarry (inter-reef facies) and Weldon members of the Scotch Grove Formation. Another line of evidence is the development of coral bioherms (reefal facies of the Johns Creek Quarry Member) growing directly over biostromal coral beds of the Picture Rock Member. These bioherms (Johnson, 1980, fig. 6) are coeval with but rise above inter-reef facies consisting largely of Pentameroides populations. Both facies appear to have responded to a steady rise in sea level. The biostromal corals survived this change only locally by forming biohermal structures.

The most convincing evidence, however, comes from data on size variation in Cyclocrinites

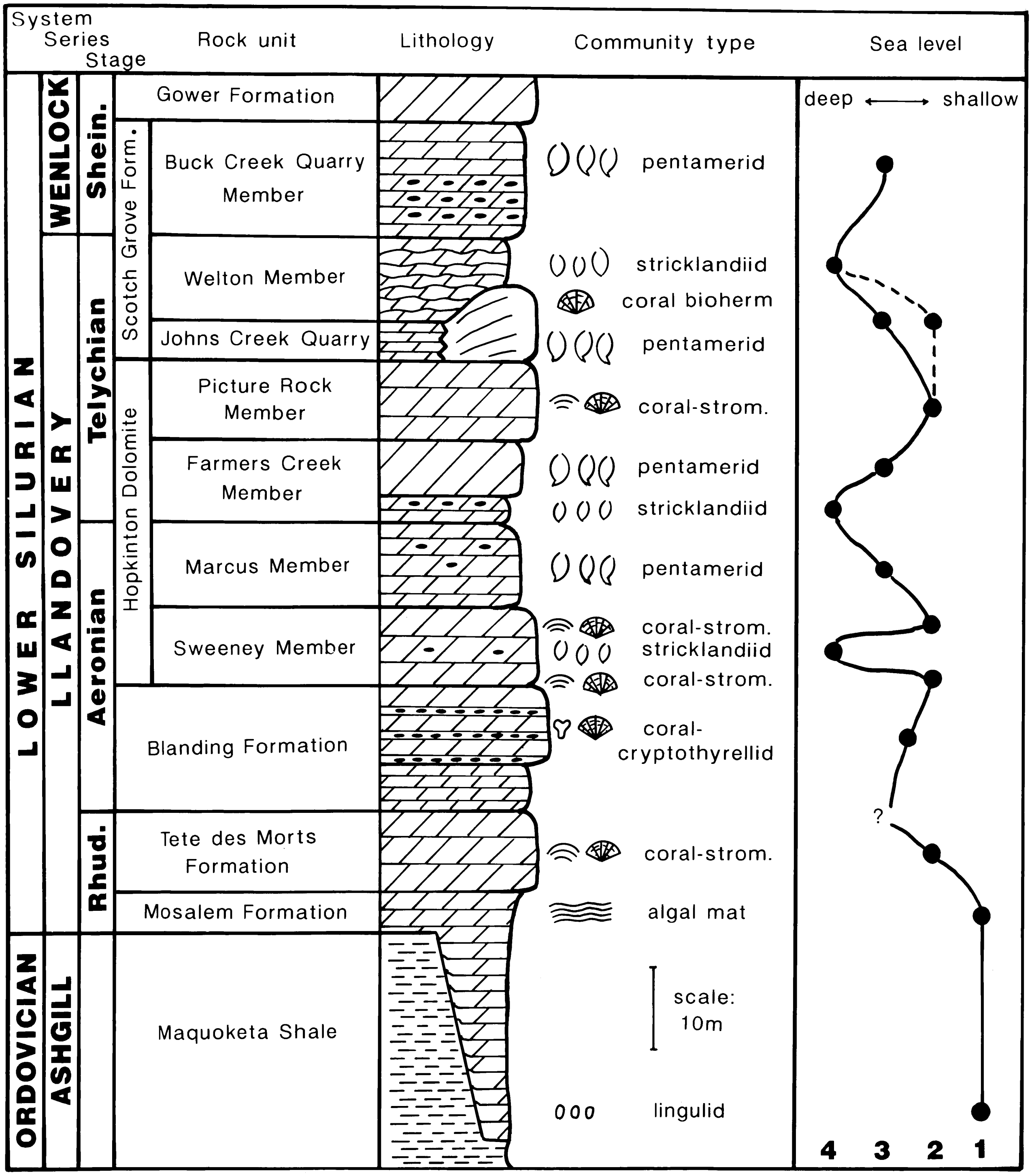

Figure 1. Sea-level curve for the Iowa-Illinois-Wisconsin area of the Upper Mississippi Valley, based on the succession of Lower Silurian strata and fossil communities (modified from Johnson, 1983). Numbers to the lower right refer to benthic assemblage zones.

BATHYMETRY OF EARLY SILURIAN MARINE COMMUNITIES								
Welsh Basin ZIEGLER (1965)		Williston Basin JOHNSON & LESCINSKY (1986)		Michigan Basin JOHNSON & CAMPBELL (1980)		East Iowa Basin JOHNSON (1980)		depth (m)
0	Rocky shore	0	Red beds & frosted sands					0±
1	Lingulid	1	Stromatolite	1	Stromatolite			0-10
2	Eocoeliid	2	Coral-Stromatoporid	2	Coral-Stromatoporid	2	Coral-Stromatoporid	10-30
3	Pentamerid	large	size variation	3	Pentamerid	3	Pentamerid	30-60
4	Stricklandiid	small	in Cyclocrinitid algae / lower photic zone limits			4	Stricklandiid	60-90
5	Clorindid		BEADLE & JOHNSON (1986)					90-120
6	Graptolite		Sedimentary Environments					120-?
← clastics →		← platform carbonates →						

Figure 2. Bathymetry of Early Silurian fossil communities preserved in the Welsh, Williston, Michigan, and East Iowa basins (after Johnson, 1987). Numbers in the left column refer to benthic assemblage zones.

dactioloides. This is a globose dasycladacean alga found in a wide range of paleocommunities. Average thallus size of specimens associated with the stricklandiid community in the lower part of the Farmers Creek Member is that of a large marble (about 2 cm in diameter). Specimens associated with the pentamerid community in the upper part of the Farmers Creek Member are commonly golf-ball size (about 3 cm in diameter). Although very rare, even larger specimens have been found associated with the coral-stromatoporoid fauna of the overlying Picture Rock Member. This unusual case was reviewed in detail by Beadle and Johnson (1986), including a treatment of variation in facet size. Filtration of sunlight with increasing water depth is concluded to have caused size variation. Those algae living in shallower waters were able to sustain more growth than those living at a greater depth at the lower limits of photosynthesis. Presumably, the shallowest ecotypes were rarely preserved due to predominantly rough-water conditions in the coral-stromatoporoid communities. Similar patterns were found in cyclocrinitids from the Lower Silurian of Anticosti Island (Quebec) and southern Norway but the Iowa data is superior.

Physical evidence for extreme rough water conditions or intermittent subaerial exposure is unknown in the Iowa sequence, but occurs in the Michigan Basin and is common in the Williston Basin. Features most characteristic of such conditions include ripple marks, brecciated stromatolite crusts, desiccation cracks, and the thin accumulations of frosted (wind-blown) quartz grains. Depth relationships of Early Silurian communities in different sectors of the North American platform are suggested in Figure 2, as well as their relationship to the classic Welsh communities of Ziegler (1965). The step-wise arrangement of the North American communities is meant to reflect inter-regional variation in bathymetry. Estimates of absolute water depth are conjectural but the lower limits of fair-weather wave base are reasonably thought to fall at the boundary between benthic-assemblage zones 2 and 3, and the lower limit of functional photosynthesis is placed between benthic-assemblage zones 4 and 5.

Time scale

A composite sea-level curve for the full Llandovery Series (Lower Silurian) of the Upper Mississippi Valley area is given in Figure 1. Three or possibly four transgressive-regressive cycles are represented. The first peak (at or about the Rhuddanian to Aeronian transition) is poorly fossiliferous and it is surmised only on the basis of lithology. The other peaks correspond to three well defined stricklandiid communities, which occurred in Mid-Aeronian, Early Telychian, and Late Telychian times. Dating is on the basis of brachiopod biostratigraphy, which confirms that individual members or even beds were virtually isochronous for considerable distances. In addition to being the name bearers of the Stricklandia or Costistricklandia communities (generally on the basis of macro-fossil dominance), these brachiopods also belong to an evolutionary lineage. Changes in the size and shape of internal skeletal features and muscle scars are gradualistic. Most of the evolving features were first described by Williams (1951) from collections in the type Llandovery district of Wales. Comparison of the Iowa stricklandiids with Welsh samples was undertaken by Johnson (1979). A more recent biometric review of the Stricklandia part of the lineage was completed by Baarli (1986), using materials from Wales, Norway, and Estonia. Back-up dating in Iowa is provided by the Pentamerus-Pentameroides lineage (Johnson, 1979). The transition between the two genera is widely observed in North America, the British Isles, and Norway and it is regarded as indicative of an Early Telychian age.

Granted arguments for local patterns of cyclicity due to sea-level events peaking at specific times, the question still remains whether or not these events may be traced to other regions. Such events were due either to recurrent uplift and subsidence of a local nature (as first advocated by Whittlesey, 1851, for patterns in Michigan) or else they were due to more widespread causes possibly of eustatic nature. The only way to decide is through biostratigraphic correlation of peak events from cyclic patterns in many other regions. Adoption of a standard time scale is essential to the test. The standard graptolite, conodont, and brachiopod zones for the Euro-American Llandovery were reviewed by Johnson (1987, fig. 5). Graptolite zones are practically useless for dating sequences in platform carbonates. Ten zones constructed from the overlapping ranges of lineage members including Stricklandia, Pentamerus, and Eocoelia, however, provide temporal resolution nearly as refined as the much valued graptolite zones (Johnson, 1987).

Figure 3A shows a set of ideal carbonate cycles from the East Iowa, Michigan, and Williston basins. Four such cycles are typical for the Llandovery Series of North America. The overlapping ranges of brachiopod lineage zones may be applied to demonstrate the equivalence of specific sea-level events in many different regions. Figure 3B shows an inter-basin correlation of the third, or Early Telychian event. As we have just seen, elements of two key brachiopod lineages occur in the East Iowa Basin. The co-occurrence of Stricklandia laevis and Pentamerus oblongus is diagnostic of the earliest Telychian. Stricklandiids are comparatively rare in the Michigan Basin, occurring only as secondary elements of pentamerid communities, as opposed to the dominant member of their own community. In Michigan, however, occurrence of Pentamerus oblongus and Eocoelia curtisi is also diagnostic of the earliest Telychian. This means that event correlation between the two basins is very good. Unfortunately, none of the lineages are represented at this interval in the Williston Basin. An earlier cycle there may be dated by the occurrence of the pentamerid Virgiana. Thus, the extra-basin correlation of subsequent Williston cycles depends largely on their number and order. In this case, the third Llandovery cycle in the Williston Basin is matched with the better correlated number 3 cycles in the East Iowa and Michigan basins.

CRATONIC DATA BASE

To date, parts or all of the four transgressive-regressive cycles belonging to the Llandovery Series have been identified and correlated from 22 different areas scattered across the ancestral North American craton. The individual sea-level curves for these areas and the specific basis for their correlation are summarized by Johnson (1987, fig. 6). For the most part, each of these areas fits one of the models represented by the Williston, Michigan, or East Iowa basins (Fig. 3A). The Hudson Bay and Tobosa basins, for example, are most similar to the Williston Basin. The Great Basin and western New York bear a closer relation to the pattern of the Michigan Basin. Parts of the North Kansas, Anadarco, and Illinois basins contain strata rich in stricklandiid brachiopods. They are, thus,

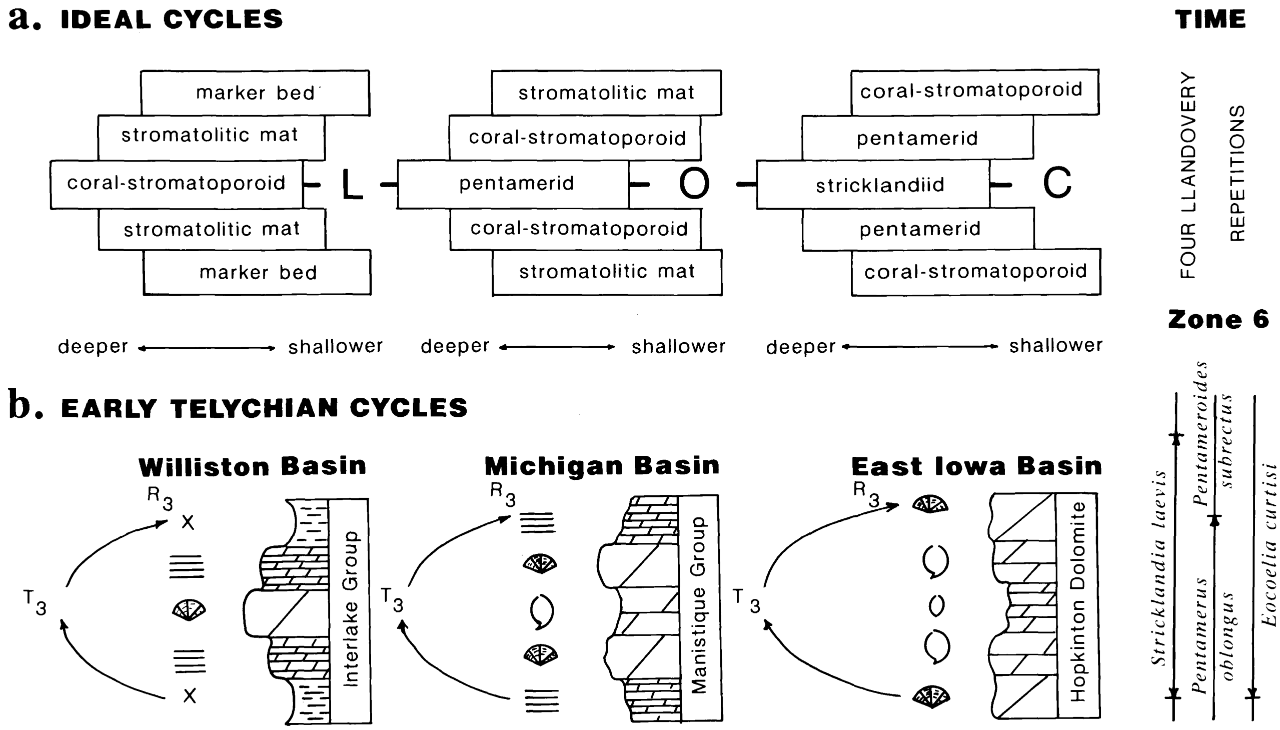

Figure 3. Correlation of typical carbonate cycles from the Lower Silurian of the Williston, Michigan, and East Iowa basins. A) Ideal cycles (L-O-C denotes line of correlation linking maximum trangressive peaks in each region). B) Example of the third transgressive-regressive cycle of Early Telychian age. Figures modified from Johnson and Lescinsky (1986).

more similar to the situation in the Upper Mississippi Valley.

A practical application of Llandovery sea-level curves representative of widespread districts, is the reconstruction of sea-floor bathymetry on a cratonic scale. In other words, the correlated curves provide a means to visualize the submarine topography of vast platform seas in a temporal series. Using the data base, it is possible to reconstruct time specific maps for the four prominent high stands in Llandovery sea level and the four succeeding low stands in sea level. Four paleoceanographic maps were assembled by Johnson (1987) and two of them are reproduced here.

The Early Telychian maximum

Figure 4 shows the paleobathymetry of the ancestral North American continent during co-existence of the index fossils Stricklandia laevis, Pentamerus oblongus, and Eocoelia curtisi (brachiopod zone 6). More than 65% of the paleocontinent was inundated during this Early Telychian maximum stand in sea-level. Stricklandiid communities were developed in three basic regions. Western Newfoundland and British Columbia qualify as shelf-margin regions. A major incursion toward the heart of the platform followed the Reelfoot Rift along the axis of the Mississippi Valley. This feature may have been related to part of an extensive Cambrian aulacogen (Lowe, 1985) and it explains the occurrence of deeper-water faunas far from the shelf margins. In any case, Lower Telychian stricklandiids are found from the Red Mountain Formation of Alabama to the Sexton Creek Formation of southern Illinois, to the Hopkinton Dolomite of eastern Iowa and its equivalent strata in eastern Nebraska.

Somewhat shallower banks with simple

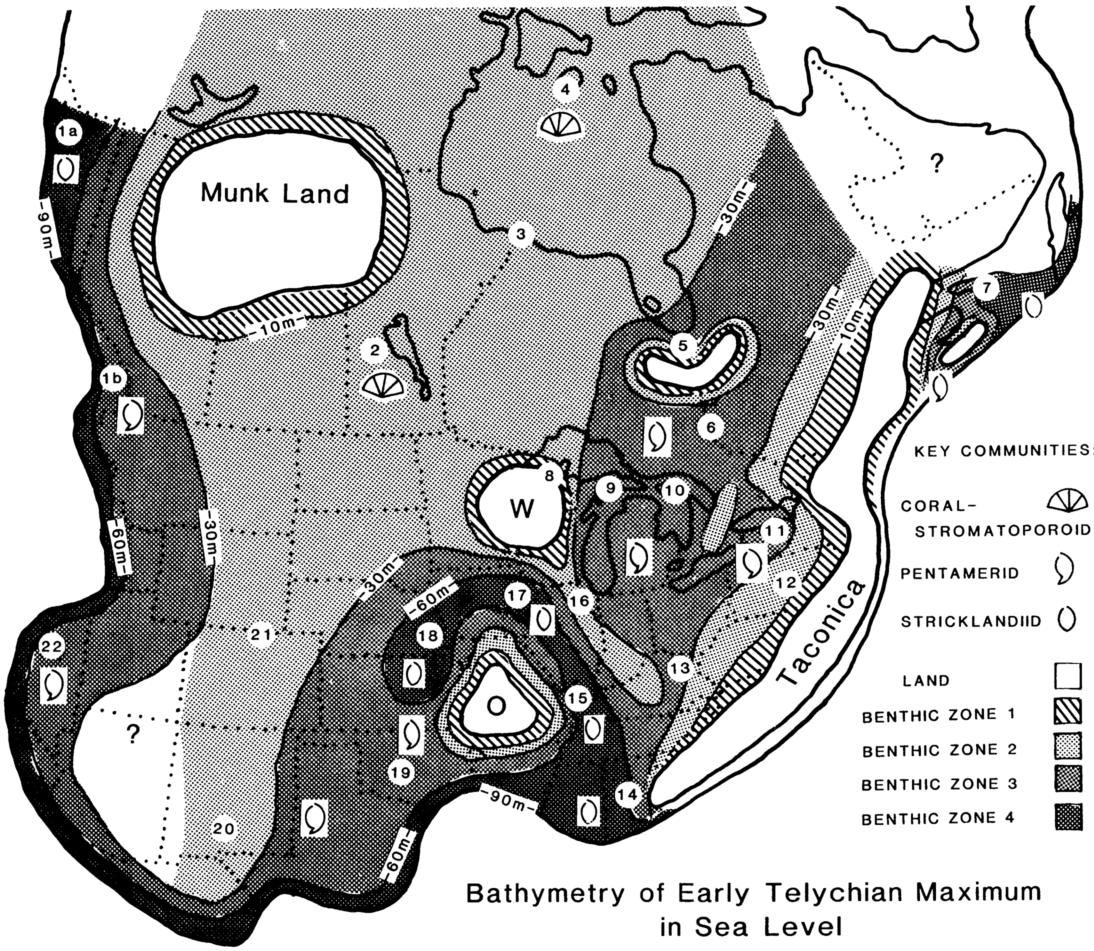

Figure 4. Bathymetric map reconstruction of the ancestral North American platform during the Early Telychian maximum in sea-level. The O and W refer to the Ozark and Wisconsin dome areas, respectively. Figure from Johnson (1987).

pentamerid communities were broadly developed in three separate regions: 1) from Nevada to Alberta, 2) across the Texas and Oklahoma panhandles, and 3) extending from western New York and the Michigan Basin to the Lake Timiskaming District of Ontario. Coral-stromatoporoid communities thrived under the most shallow-water, carbonate conditions across the broad back of the Transcontinental Arch stretching from New Mexico to Manitoba and beyond to Southampton Island in Hudson Bay. The range of bathymetries between these key regions and the few known land areas is mainly extrapolated. Compatible clastic cycles along the inner side of Taconica, however, have been described recently in the central Appalachians (Cotter, 1988), as well as the southern Appalachians (Easthouse and Driese, 1988).

The Mid-Telychian minimum

Figure 5 shows the contrasting paleobathymetry for a slightly younger time, at the minimum stand in sea level during the Middle Telychian. Good index fossils exist in theory but they are exceedingly rare in the very widespread, shallower carbonate facies. In practice, these facies are correlated by their intermediate position bracketed between the well dated Early

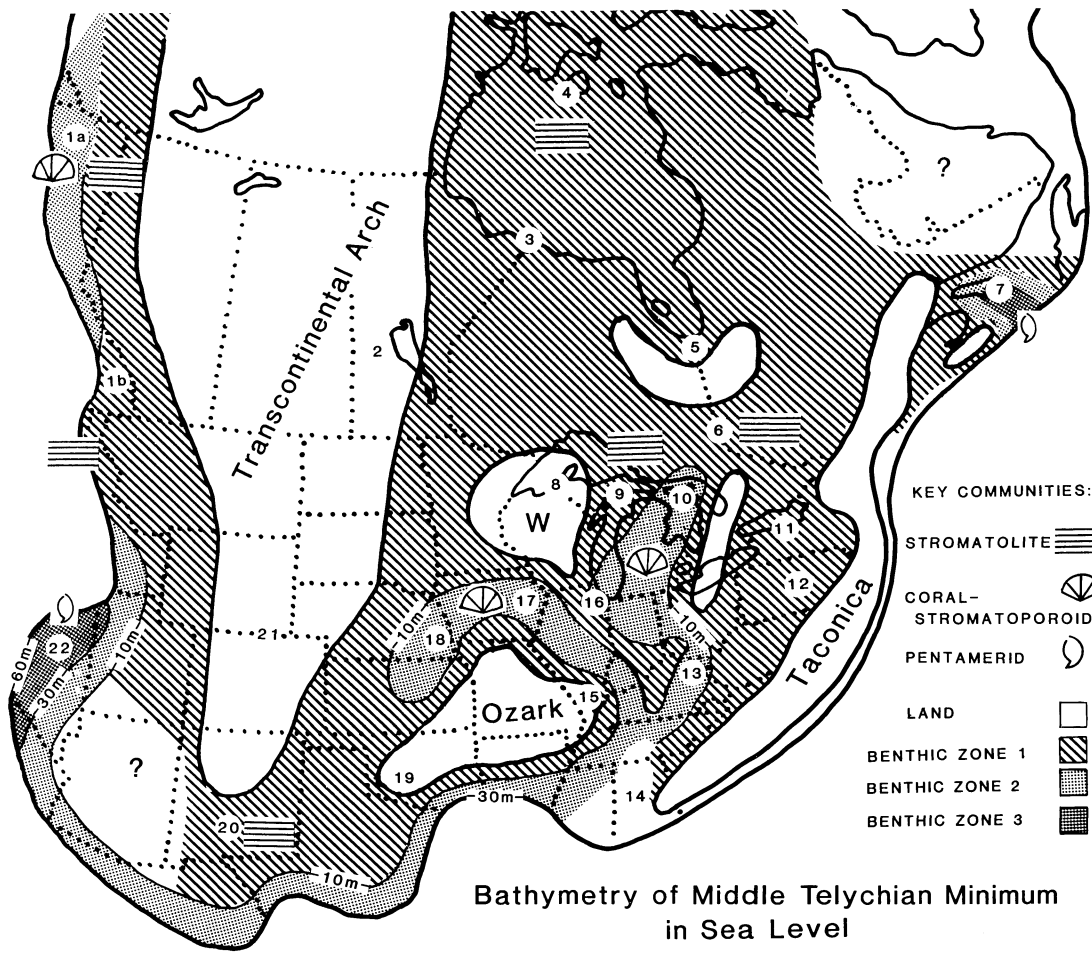

Figure 5. Bathymetric map reconstruction of the ancestral North American platform during the Middle Telychian minimum in sea level. The O and W refer to the Ozark and Wisconsin dome areas, respectively. Figure from Johnson (1987).

and Late Telychian high stands in sea level. The most striking differences are the presence of coral-stromatoporoid communities in the central platform region formerly occupied by stricklandiid communities and the dramatic expansion of low lands. Exposure of the Transcontinental Arch expanded to include those regions formerly occupied by the coral-stromatoporoid communities. In the Williston Basin region, such emergence is probably represented by the widespread u_2 marker bed of King (1964). Stratigraphic disconformities on the present southern and eastern flanks of the Ozark Dome (Amsden and Barrick, 1988) also imply land expansion. The coeval Prices Falls Member of the Clarita Formation in south-central Oklahoma, the "Button" Shale of north-central Arkansas, and the Seventy-six Shale of northeast Missouri and southwest Illinois seem to represent shallow-water clastics deposited at maximum exposure of the Ozark lowlands in Middle Telychian time. The most widespread carbonates of this interval are laminated deposits typically generated by algal mats. These often bear a restricted fauna of ostracods and gastropods.

MODEL LIMITATIONS AND REALITY

The making of models as a tool to understand the past is a fundamental part of geological research. We are inescapably part of the present world, however, and we feel nervous if our models stray to far from its reality. The model I apply to the large-scale development of carbonate platforms has definite limitations and it exhibits some features markedly foreign to our present experience. Estimation of absolute changes in water depth is the most important limitation on my model. Stricklandiid communities including fossil algae are reasonably interpreted as having lived in some of the deepest water, carbonate environments on the North American platform during Early Silurian time. The critical question is, however: how deep? The diminution of light with increasing water depth varies with water clarity. I suggest that stricklandiid communities may have been restricted to a range of 60-90 m. It is certainly possible to argue that their bathymetric range was broader or narrower. Whatever the actual range, some aspects of the maps in Figures 4 and 5 are rather startling. The Silurian craton was extraordinarily flat, with very little variation in sea-floor bathymetry. If 90 m is taken as the lower limits of the stricklandiid communities, then maximum variation in platform bathymetry was between 0-90 m. Whether this estimate is halved or doubled, relief was very slight. Extraordinary as this may seem to eyes accustomed to the present world, I believe it is close to reality. Two lines of supporting evidence may be drawn.

Peneplanation of the Canadian Shield

Lowering of continents to base level is a concept less fashionable among geomorphologists than it has been, but peneplanation of extensive areas such as the Canadian Shield is readily apparent. Ambrose (1964) studied the geomorphology of Precambrian plains in Canada, which he argued were only slightly modified by the wear of recent continental glaciations. Lower Paleozoic outliers widely scattered across the Canadian Shield indicate that the topography of the presently exhumed surface is much the same as it was in Late Precambrian time. With the exception mainly of Labrador, Ambrose (1964, p. 850) concluded that topographic relief varied by only about "200-300 feet" (60-90 m). Move the Canadian Shield to a tropical setting, raise sea-level 90 m to allow transgression of carbonate sediments, and the final stratigraphic-bathymetric configuration would be close to that depicted in Figures 4 and 5, but with Labradoran Islands.

Analog of the Bahamian Banks

With shallow sub-tropical banks less than 20 m deep covering an area of 200,000 square km, the Bahamas are one of the few large contemporary examples of a carbonate platform. Carbonate sediments on the banks occur today in a complex mosaic of facies (Enos, 1974) controlled by seasonal salinity gradients related to poor circulation. During Pliocene times, on the other hand, a laterally uniform biostrome was deposited across an extensive part of the banks under conditions of much better circulation presumably related to higher sea level (Beach and Ginsburg, 1980). This cover was not unlike some of the Silurian biostromes of the Upper Mississippi Valley area. A realistic analog for a carbonate platform of cratonic dimensions might be imagined by stitching together a patch-work of dozens of Bahamian Banks grouped randomly at slightly different bathymetric levels. Such a configuration would produce a range of biofacies under various conditions of circulation and other physical factors related to water depth. The shallowest facies would be expected to form a mosaic (Williston-type basin), while the deeper facies would be more laterally homogeneous (Michigan or Iowa-type basins). Add fluctuating sea levels to this scenario and coordinated cyclic carbonates must be the result.

OTHER SILURIAN CRATONS

From all available evidence, the Early Silurian (Llandovery) craton of North America was mostly flat, extensively flooded, and widely affected by a series of at least four fluctuations in sea level. Periodicity was perhaps on the order of 2.5 million years. The calculation depends on whose radiometric time scale is used, since the length of the Silurian Period has been calibrated with various results. The interesting question remains whether or not the North American pattern is global. Some workers believe that eustasy is a clouded issue due, for example, to the irregularity of the geoid (Mörner, 1976).

Nonetheless, it is possible for a group of different continents with roughly the same hypsometry to record similar patterns of sea-level change. It is worthwhile to search for inter-continental patterns and McKerrow (1979) made the first serious attempt to check for Silurian eustasy on more than one paleocontinent. So far, good results have been obtained from the South China Plate (Johnson et al., 1985) with three out of four events matched. The Balto-Scandinavian Plate includes superb shelly sequences in Norway and Estonia. Baarli (1985) found two cycles in the central Oslo region which correlate with the first two events in North America. Two subsequent deepenings also are represented in the central Oslo region and correlation with Estonian cycles is being attempted (work in progress). Nestor (1972) has already drawn attention to the Late Llandovery "eustatic" transgression on this plate. Exposures of shelly Llandovery strata are limited to the area near Orange, in New South Wales, Australia (Jenkins, 1977). At least two Australian cycles correspond to the first and fourth cycles in North America.

It is interesting to note that only the Estonian geologists have attempted independently to trace Lower Silurian cycles outside their own boundaries. In tectonically complex regions such as Norway and Australia, it certainly would have been more logical to first attribute such events to local causes. The carbonate platforms of the Baltic and Upper Mississippi Valley areas share in common, however, a simple layer-cake structure in settings of relative tectonic stability. These are the very factors which promote a search for something other than local causes. A foreign visitor to the Upper Mississippi Valley once confided to me that midwestern geology is "child's play" and nothing to occupy a "real" stratigrapher. I see it another way. Potentially important global signals are easy to overlook in structurally messy regions where it is often impossible to see the forest for the trees.

LOCAL VALLEY TECTONICS

The Upper Mississippi Valley area was far from structurally or tectonically dormant. The theme of this field conference is the Plum River Fault Zone and I would be remiss not to add a few words on the relationship of the Lower Silurian to this important feature. Early investigators described what was called the "Savanna Anticline" along the trend of the presently recognized fault zone. This term has become unpopular. The lineament which crosses the Mississippi River near Savanna was first reassessed by Kolata and Buschbach (1976). They found clear evidence for faulting along a corridor which was named the Plum River Fault Zone. Work at this stage was focused on the Illinois side of the Mississippi River and some smaller structures were located adjacent to the fault zone. These include the Forreston Dome, the Brookville Dome, and the Leaf River Anticline. Later work by Bunker et al. (1985) extended the fault zone well into eastern Iowa. Graben style faulting seems to have taken place in post-Silurian time, but was coincidental with a line of pre-existing structures.

Measured sections of Lower Silurian strata indicate that along strike of the lineament, the Picture Rock Member of the Hopkinton Dolomite is at least twice as thick as it is to the north (Fig. 6) or south. This is clearly shown by natural outcrops along the south fork of the Maquoketa River near Baldwin (on the lineament and 11 m thick), as compared to quarry sites like the Johns Creek Quarry near Farley (50 km north of the lineament and 5.25 m thick). A similar pattern of thinning strata is found on the south side of the lineament. In contrast, underlying units such as the Farmers Creek, Marcus, or Sweeney members of the Hopkinton Dolomite seem to be far more laterally consistent in their thickness. On the lineament near Sabula, for example, the Sweeney Member is about 11 m thick, while near Dubuque it is still 9.5 m thick. Even more striking is the complete absence of the Johns Creek Quarry Member (inter-reef facies) along the lineament (Fig. 6).

One possible interpretation is that a loose string of small domes or anticlines began to push up during later Llandovery time, coincidental with an episode of rising sea level (the Early Telychian

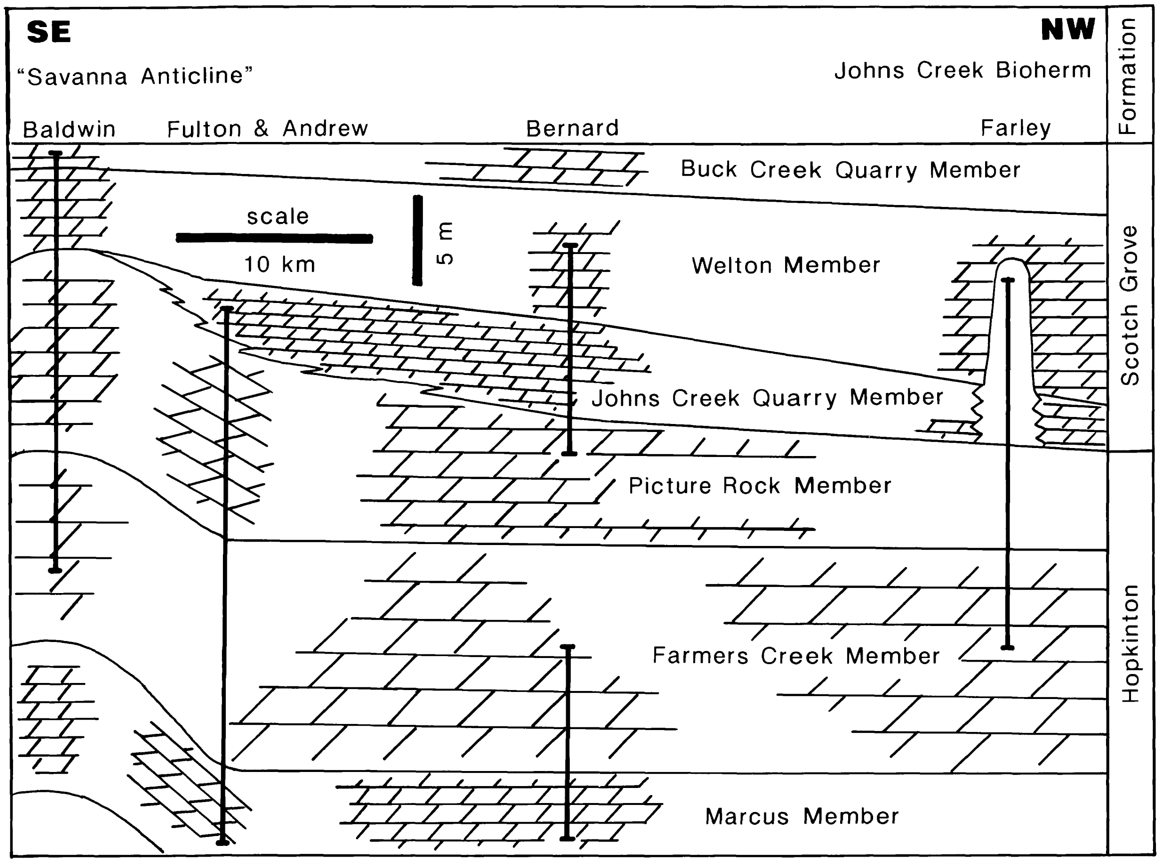

Figure 6. Stratigraphic cross-section across the lineament of the "Savanna Anticline" extending from northwestern Illinois to eastern Iowa. The Picture Rock Member (Hopkinton Dolomite) thickens above this structure, while the inter-reef facies of the Johns Creek Quarry Member (Scotch Grove Formation) pinches out. Modified from Johnson (1983).

event). North and south of the lineament, bioherms grew up from the coral-stromatoporoid biostromes forming the Picture Rock Member of the Hopkinton Dolomite. Good examples are found to the north near Farley (Dubuque County) and Monticello (Jones County) and to the south near Elwood (Clinton County). Among the bioherms, biostromal beds with Pentameroides were emplaced. In contrast, the biostromal character of the Picture Rock Member is maintained throughout its extra thickness along its lineament sites and the unit is directly draped by the deeper-water Weldon Member of the Scotch Grove Formation (omitting the inter-reef facies of the Johns Creek Member). My interpretation is that small, local uplifts of the sea bed along the lineament temporarily offset the effect of rising sea level. Something like a set of carbonate ramps was created by this process, but their slopes were slight (1 vertical meter/10 km on the north side). Good surficial evidence is limited to a few natural or quarry localities where entire rock units are exposed. Collection of pertinent subsurface data on both sides of the Mississippi River may pull the picture into better focus.

The line of small domes or anticlines stretching from Illinois to Iowa is not the only feature to cut across the trend of the former Reelfoot Rift. A short 150 km long structure called the Pascola Arch crosses the southeastern toe of Missouri and it apparently formed sometime after the Carboniferous but before the end of the Cretaceous (Schwalb, 1982). This arch forms one flank of the Illinois Basin and overprints a previous trough-like connection to the continental shelf margin. Possibly the lineament of the Plum River Fault Zone also marks the site of an aborted arch which started but never reached the dimensions of the Pascola Arch.

CONCLUSIONS

During the Early Silurian, epicontinental seas covered over 65% of the ancestral North American continent. Much of the sea bed was formed by a vast carbonate platform. The present Upper Mississippi Valley area was one of the deeper parts of this platform and was connected to the margin of the paleocontinent along the axis of a former Cambrian aulacogen called the Reelfoot Rift. During times of peak transgression, diverse stricklandiid communities lined this depression from Alabama to Iowa and even extended across to Nebraska. Much of the plateau beyond a gentle rise in the direction of the Michigan Basin was blanketed by enormous populations of pentamerid brachiopods. A similar rise stretched beyond Nebraska toward the Williston Basin, where pentamerid populations gave way to shallower coral-stromatoporoid communities covering the broad back of the Transcontinental Arch. Bathymetric relief may have been as little as 90 m. During episodes of maximum regression, similar coral-stromatoporoid communities migrated to the "low ground" once occupied by the stricklandiid communities. At these times, pentamerid and stricklandiid communities were excluded mostly to the continental margins.

This is a panoramic picture of the bathymetry and ecology of an ancient carbonate platform having cratonic dimensions. The interpretation is based on the widespread occurrence of carbonate cycles attributed to sea-level fluctuations, most of which may be well dated biostratigraphically. The key to this vision lies in the Upper Mississippi Valley area, pointed toward the heart of the paleocontinent. Silurian strata in the East Iowa Basin are well exposed. The fossil communities are easily defined and their stratigraphic order is unambiguous. The relatively stable setting, with regard to structural tectonics, requires an extra-basin search for the cause and continuity of sea-level events. In contrast, exposures are far more limited in the neighboring Michigan and Williston basins. Cyclicity is evident, but previous stratigraphic work focused more on distinctive marker beds. Furthermore, the history of subsidence in these larger basins promotes satisfaction with internal solutions for problems of cause.

I am fatally attracted to the "big picture" and I will be delighted if the prevalent pattern of Lower Silurian cyclicity in North American may help decode heretofore hidden patterns on other cratons where internal solutions are traditionally sought. The advantage could ultimately be a more detailed portrait of a Silurian planet.

ACKNOWLEDGMENTS

The topic and scale of this contribution owe an unmistakable debt of inspiration to Alfred M. Ziegler (University of Chicago). My path to Chicago and beyond, however, first passed through Iowa City. This review is gratefully dedicated to my former Iowa professors: William M. Furnish, who taught me sensible stratigraphy, and Gilbert Klapper, who understood and encouraged my passion for paleoecology. Above all, their frequent travel to far-off places for bits and pieces of a greater puzzle made a lasting impression.

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LATE MIDDLE THROUGH LOWER UPPER DEVONIAN STRATIGRAPHY ACROSS EASTERN IOWA

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INTRODUCTION

New stratigraphic nomenclature has recently been introduced (Witzke et al., 1988) to delineate major lithic units within Middle and lower Upper Devonian rocks across eastern Iowa. A conodont zonation scheme for cratonic biofacies in eastern Iowa area has also been proposed (Bunker and Klapper, 1984; Witzke et al., 1985; Witzke et al., 1988), providing a basis for correlations in the Midcontinent Shelf area.

Devonian strata in Iowa and surrounding areas unconformably overlie an eroded surface of Ordovician and Silurian rocks (Fig. 1). Upper Devonian strata overstep onto the Precambrian surface adjacent to the Sioux Ridge in northwestern Iowa (Fig. 1). Devonian strata in the central Midcontinent region are bounded to the north by the Transcontinental Arch (including the Sioux Ridge), to the west by the Cambridge Arch-Central Kansas Uplift, to the south by the Chautauqua Arch-Ozark Uplift-Sangamon Arch, and to the east by the Devonian outcrop belt and sub-Pennsylvanian Devonian edge (Figs. 1, 2). Devonian strata trend across the Transcontinental Arch in the area of the Nebraska Sag (Fig. 1). Devonian seaways may also have breached the arch to the east of the resistant highlands of the Sioux Ridge in central Minnesota, but pre-Cretaceous erosion apparently removed Devonian strata from that area. The East-Central Iowa and North Kansas basins (Fig. 1) are primarily Silurian features, but persisted as structural depressions during the initial stages of Middle Devonian (late Eifelian-early Givetian) deposition in the region. Eifelian deposits are restricted to these two basin areas and are absent from intervening areas in central and western Iowa. Late Givetian and Frasnian strata thicken markedly toward central and northern Iowa, where the thickest sequences of Devonian strata in the region are preserved (to 230 m). Total Devonian isopachs delineate this area as a stratigraphic basin (Fig. 2), here referred to as the Iowa Basin (Central Iowa Basin of Collinson et al., 1967).

STRATIGRAPHY

Wapsipinicon Group

The Wapsipinicon Formation was named by Norton (1895) for exposures along the Wapsipinicon River in northeastern Linn County, Iowa. It was elevated to group status by Witzke et al. (1988) to include, in ascending order, the Bertram, Otis, and Pinicon Ridge formations in eastern Iowa, and the Spillville and Pinicon Ridge formations in northern Iowa and southeastern Minnesota (Fig. 3).

The Wapsipinicon Group overlies an erosional surface on Ordovician and Silurian strata (Fig. 1) and is overlain disconformably by the Cedar Valley Group over its regional extent in Iowa, western Illinois, northern Missouri, and southeastern Minnesota. Its edge is overlapped by Cedar Valley strata to the west and south. The Wapsipinicon Group reaches thicknesses of up to 60 m south of the type area adjacent to the Plum River Fault Zone (Fig. 4), and is dominated by carbonate rock, but gypsum and anhydrite are significant components in southern and central Iowa (Fig. 4). The group encompasses two major transgressive-regressive depositional cycles.

Bertram Formation

The Bertram Formation (Fig. 3) is the most geographically restricted unit of the Wapsipinicon Group (Church, 1967; Bunker et al., 1985, p. 92)

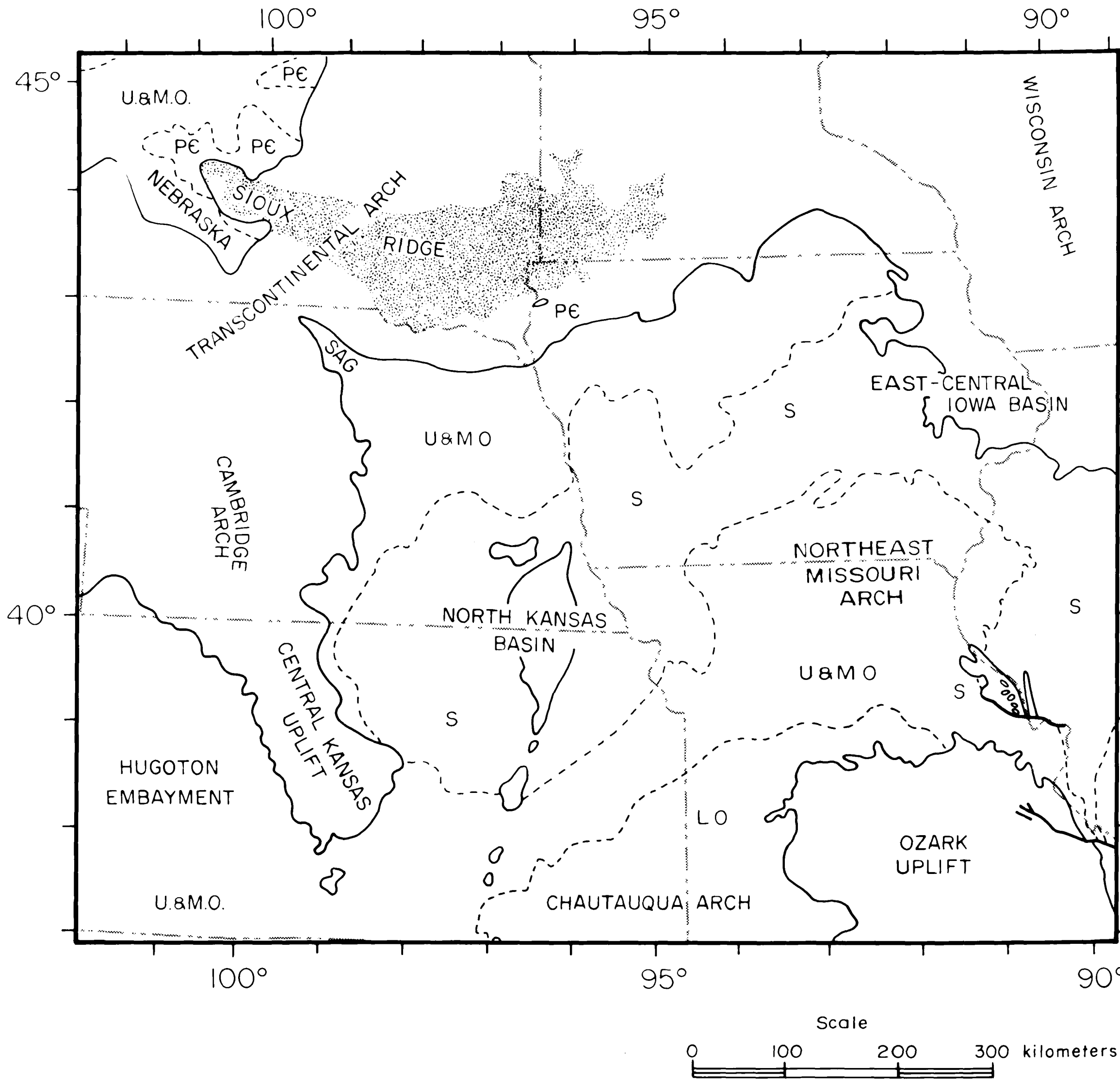

Figure 1. Pre-Kaskaskia paleogeologic map of the central midcontinent region, U.S.A. (from Bunker et al. 1988). P -Precambrian; LO-Lower Ordovician; U&MO-Upper & Middle Ordovician; S-Silurian.

and is limited to the area 35 to 45 km north of the Plum River Fault Zone near its western terminus (Fig. 5). The Bertram occupies a small asymmetrical basin with its greatest thicknesses (50-96+ ft; 15-30 m) adjacent to the fault zone. The Bertram overlies an eroded Silurian surface, and the fractured upper beds are overlain sharply, perhaps unconformably, by the Otis Formation.

The Bertram Formation is dominated by unfossiliferous vuggy dolomite, in part laminated to sandy, and in places extensively calcitized. Sandy shale occurs commonly in the basal Bertram. Dolomites are locally fractured, brecciated, or conglomeratic, and accretionary caliche fabrics and possible gypsum pseudomorphs are noteworthy (Sammis, 1978). Pelletal fabrics and "birdseye" are present, and possible ostracodes have also been noted (Church, 1967).

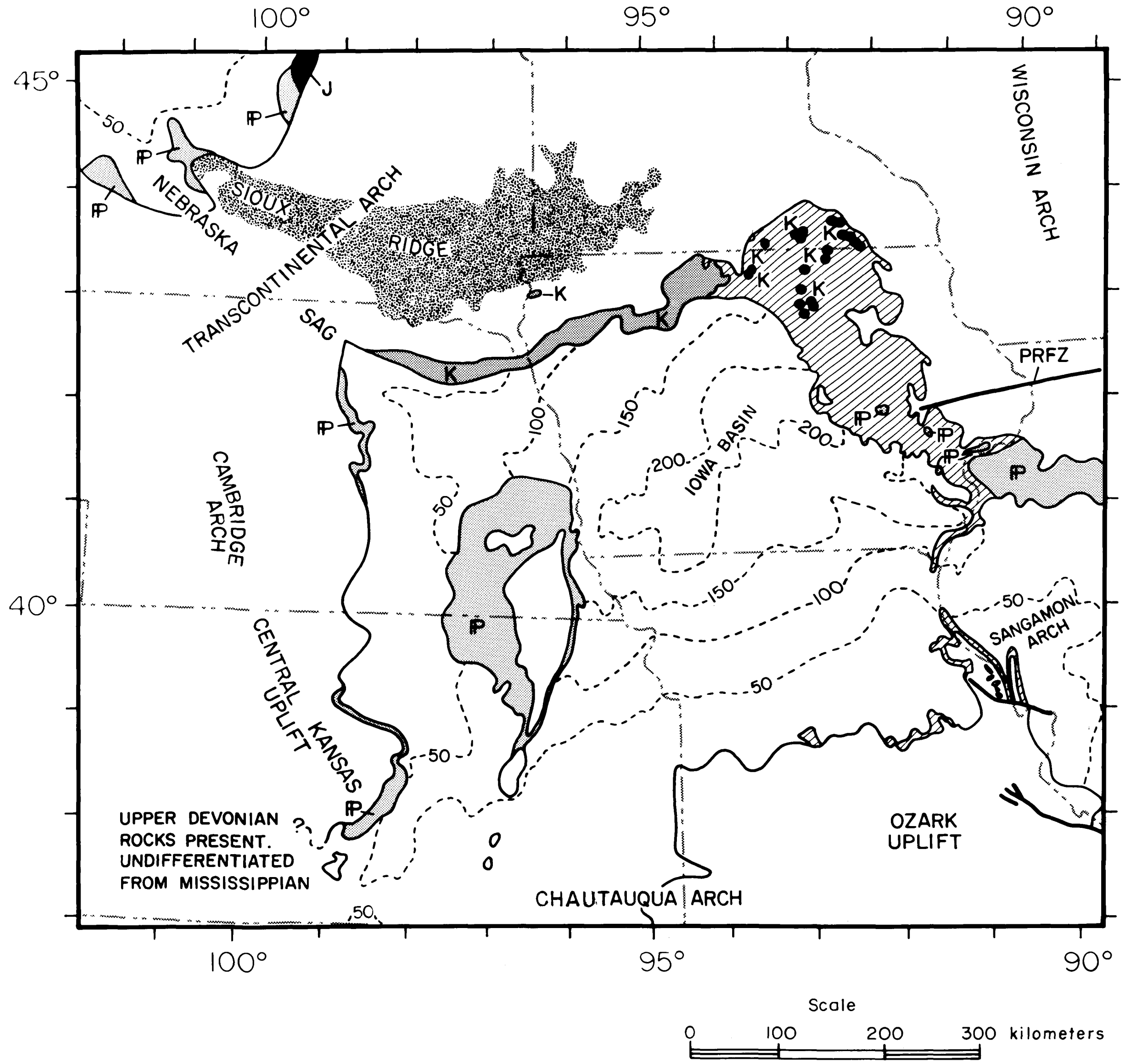

Figure 2. Total Devonian isopach map of the central midcontinent region, U.S.A. (from Bunker et al., 1988). Contour interval is 50 m. Patterned areas denote where Devonian strata have been eroded and overstepped by younger strata. P-Pennsylvanian; J-Jurassic; K- Cretaceous; diagonally-lined areas denote present day outcrop. PRFZ-Plum River Fault Zone.

Otis Formation

The Otis Formation (Fig. 3), named for the old Otis railroad station in southeastern Linn County, Iowa, overlies the eroded Silurian surface over most of its extent in east-central Iowa and northwestern Illinois (Fig. 6). It is thickest (with a maximum of up to 53 ft; 16 m) where it overlies the Bertram Formation adjacent to the Plum River Fault Zone. The Otis is consistently overlain by the Kenwood Member of the Pinicon Ridge Formation. The Otis Formation typically consists of medium- to thick-bedded dolomite in the lower part (Coggon Member), and thin-to medium-bedded limestone and dolomite in the upper part (Cedar Rapids Member). The contact between the two members is transitional, and the members are not differentiated at some localities. Although characteristically dominated by dolomite and calcitized dolomite at most localities, the undifferentiated Otis is entirely limestone in the Davenport, Iowa, area where oolitic grainstones have been noted (Witzke et al., 1985). The Coggon Member contains scattered molds of the brachiopod Emanuella along with rare specimens of gastropods, rostroconchs, and trilobites (Dechenella). The Coggon Member has produced the conodont Ozarkodina raaschi at a single locality (Klapper and Barrick, 1983). The Cedar Rapids Member contains varied lithologies including: 1) limestone, dolomite, and chert with scattered Emanuella and rare gastropods and bryozoans; 2) pelletal limestones, locally with spirorbids; 3) laminated carbonates, in part with "birdseye" and mudcracks; 4) unfossiliferous carbonate, in part intraclastic, fractured, or brecciated; and 5) silicified beds and chert nodules (Church, 1967; Sammis, 1978).

Spillville Formation

The basal Devonian dolomite sequence in northeastern Iowa and southeastern Minnesota was named the Spillville Formation by Klapper and Barrick (1983) for the quarry at Spillville, Winneshiek County, Iowa and is now included within the Wapsipinicon Group (Witzke et al., 1988). The Spillville Formation (Fig. 3) unconformably overlies an eroded surface of Upper Ordovician rocks and is overlain abruptly by the Kenwood Member of the Pinicon Ridge Formation. The Spillville is 19 m thick at the type locality and reaches maximum thicknesses of 25 to 30 m (82-98 ft) in parts of northeastern Iowa (Fig. 6).

The Spillville Formation is dominated by medium- to thick-bedded dolomite, in part stylolitic, with scattered to abundant fossil molds and vugs; dolomitic limestones are present locally. By contrast, the basal 0.5 to 4.2 m of the formation is characterized by argillaceous to shaly dolomite and shale, in part silty, sandy, or conglomeratic with reworked Ordovician chert clasts; this basal clastic-rich unit has been informally termed the "Lake Meyer member" (Bunker et al., 1983; Witzke and Bunker, 1985). The remainder of the formation is informally and crudely subdivided into a fossiliferous lower unit, the "Productella beds" (Calvin, 1903), and an upper vuggy and calcitic coral-rich unit, the "calcite zone" (Dorheim and Koch, 1966). Late Eifelian conodonts (approximately kockelianus and lower ensensis zones) were described by Klapper and Barrick (1983) from the Spillville Formation. The macrofauna of the Spillville has not been studied in detail, but the moldic fauna is diverse and allied with the Lake Church Formation of eastern Wisconsin (Raasch, 1935; Griesemer, 1965). Brachiopods include "Productella" (=Spinulicosta), Strophodonta, chonetids, Schizophoria, Gypidula, Athyris, Emanuella, spiriferids, Cyrtina, atrypids, and others (Stauffer, 1922). Gastropods, bivalves, rostroconchs, nautiloids, crinoid debris, bryozoans, tentaculites, and trilobites (Phacops and Dechenella) also occur. Corals are scattered throughout the formation but become especially noteworthy in the upper "calcite zone," where favositids, and solitary and colonial rugosans (Hexagonaria) are common; massive stromatoporoids are present locally.

Pinicon Ridge Formation

The Pinicon Ridge Formation (Fig. 3) is the unfossiliferous Devonian rock sequence which lies above the Otis or Spillville formations and disconformably below the fossiliferous Cedar Valley Group. The type locality is at Pinicon Ridge Park in northwestern Linn County, Iowa, in the historic type area of the Wapsipinicon Group. The formation consists of three members, in ascending order, the Kenwood, Spring Grove, and Davenport. The Pinicon Ridge Formation overlaps the Spillville and Otis edges to overstep an eroded surface of Ordovician and Silurian rocks in parts of central and southern Iowa, northern Missouri, and western Illinois (Fig. 1). It partially buries paleotopographic ridges or paleoescarpments of resistant Silurian carbonate rock (Hardin-Bremer, Washington-Louisa, and Pike-Schuyler highs, Fig. 4). The Pinicon Ridge ranges from about 20 to 40 m in thickness across most of Iowa (with a maximum thickness of up to 57 m in southeastern Iowa), but varies from 8 to 13 m in northeastern Iowa. The formation is dominated by unfossiliferous carbonate (variably shaly, laminated, or brecciated), but gypsum and anhydrite, in part of economic importance, are widespread in central and southern Iowa (Fig. 4). The stratigraphic position of the formation suggests an early Givetian age (Upper ensensis-Lower varcus subzones), although the

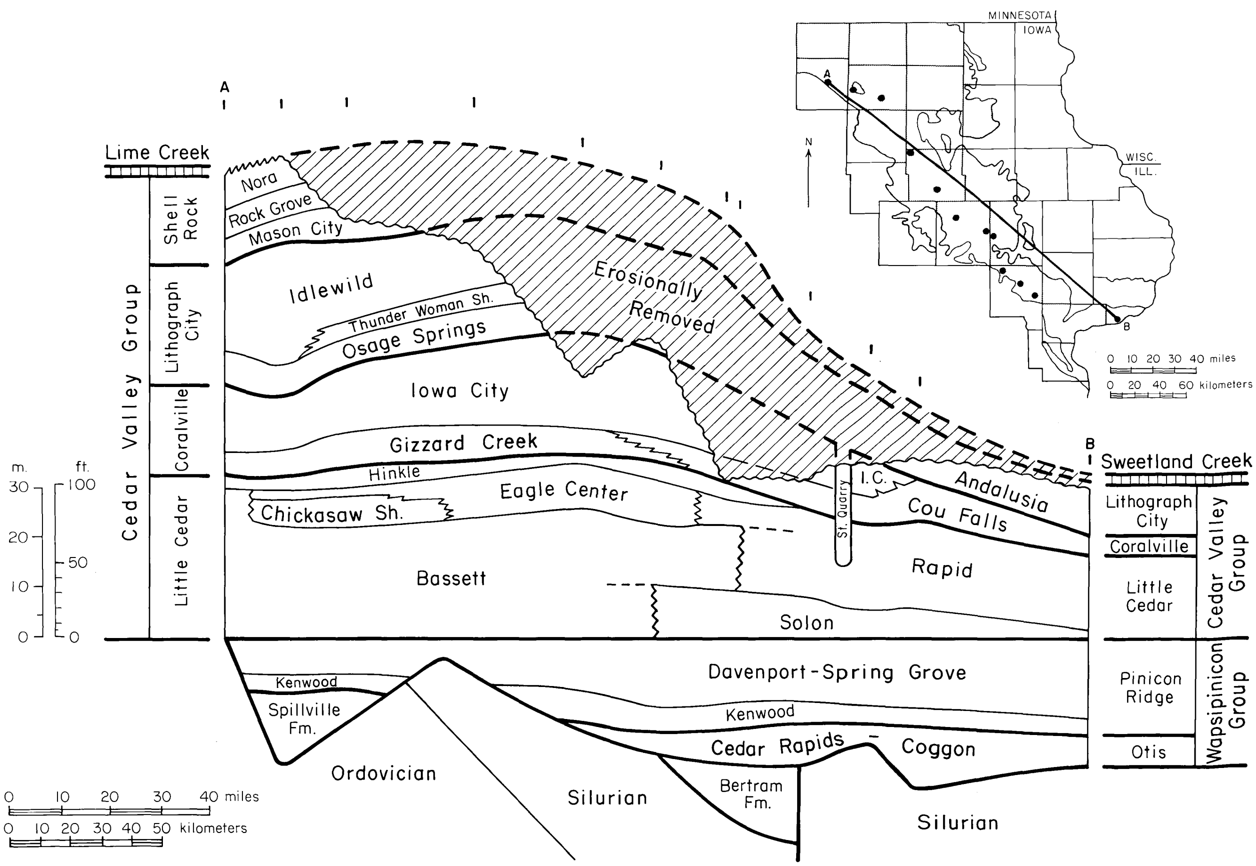

Figure 3. Generalized stratigraphic cross-section from north-central to extreme east-central Iowa, showing interpreted stratigraphic relationships of the various units of the Wapsipinicon and Cedar Valley groups (from Witzke et al., 1988).

magnitude of the disconformity at the top of the sequence is unknown. Erosional relief at the top of the formation is difficult to evaluate because of extensive brecciation, but is not presently known to exceed 40 cm. Constituent members of the Pinicon Ridge Formation are each lithologically distinct, as discussed below.

The Kenwood Member (Fig. 3) is dominated by thin-bedded, unfossiliferous, argillaceous to shaly dolomite and dolomitic limestone with scattered to abundant quartz/chert silt and sand, locally forming sandstone. It is laminated in part, and is locally brecciated to intraclastic. Shale interbeds are noteworthy, and nodular, mosaic, and bedded gypsum/anhydrite are present in Kenwood evaporite sequences of southern Iowa (Giraud, 1986). Nodular and irregular masses of megaquartz are present locally. The Kenwood ranges from about 4.5 to 8 m in thickness across most of Iowa and northern Illinois; it is generally thinner in northeastern Iowa (0.5-4.5 m) and reaches a maximum thickness of 13 m in the evaporite-bearing region of southeastern Iowa. Where the Kenwood overlies the Spillville and Otis formations, no evidence for an erosional contact at the base of the Kenwood has been identified. However, the Kenwood unconformably overlies Ordovician or Silurian strata over most of its extent.

The Spring Grove Member is dominated by thin- to medium-bedded dolomite, in part calcitic to calcitized. The member is typically laminated and petroliferous; laminations characteristically are continuous laterally. Breccias commonly are present at the base and occur locally throughout the interval in northeastern Iowa. Nodular to

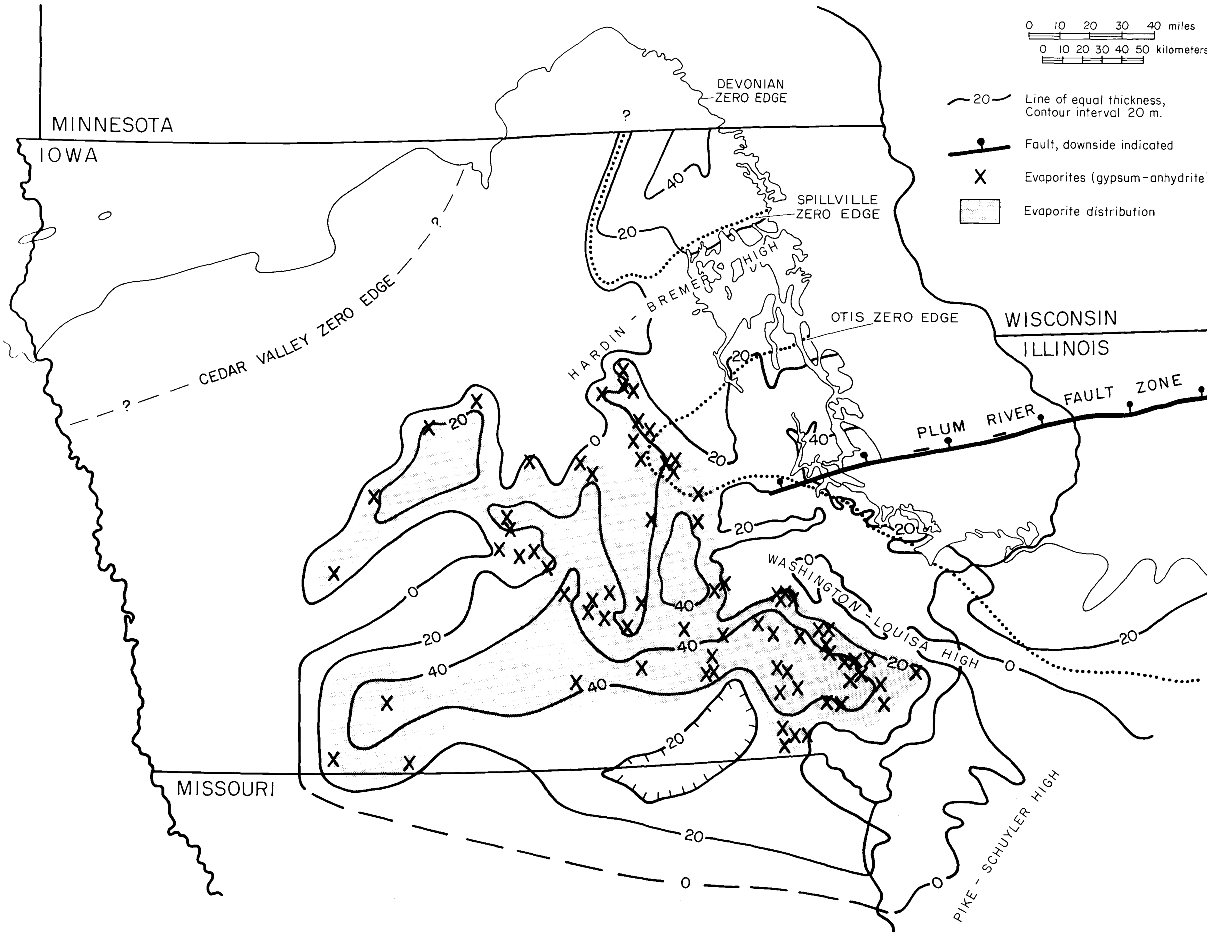

Figure 4. Isopach and evaporite distribution map of the Wapsipinicon Group (from Witzke et al., 1988).

mosaic anhydrite is present in the lower Spring Grove in evaporite-bearing Wapsipinicon sequences of southern Iowa (Giraud, 1986). The member is generally unfossiliferous, but burrows, stromatolites, ostracodes, and articulated fish skeletons with soft tissue (William Hickerson, Augustana College, pers. comm., 1987) have been noted. The Spring Grove ranges from 4 to 7 m in thickness across most of Iowa but is thinner in northeastern Iowa (1-5 m). It overlies the Kenwood Member, apparently conformably, but locally lies on eroded Silurian strata along the Hardin-Bremer High (Bunker et al., 1983).

The Davenport Member consists primarily of dense limestone (commonly "sublithographic" and stylolitic), dolomitic limestone, and dolomite. The limestones are laminated in part, and pelletal, oolitic, intraclastic, and "birdseye" fabrics are present locally. The Davenport is generally unfossiliferous, although ostracodes, burrows, and stromatolites have been noted. Shale lenses and partings are scattered, and sandy to argillaceous carbonate is present in some beds and fracture fills, especially in the upper part. The member is extensively brecciated across most of its extent, but becomes less brecciated eastward into Illinois. The breccias are composed of angular clasts and blocks of carbonate in a matrix of limestone or dolomite. Breccia clasts are composed predominantly of dense limestone, in part laminated, which are indistinguishable from many nonbrecciated limestone lithologies of the Davenport. However, clasts derived from the overlying Solon Member of the Cedar Valley

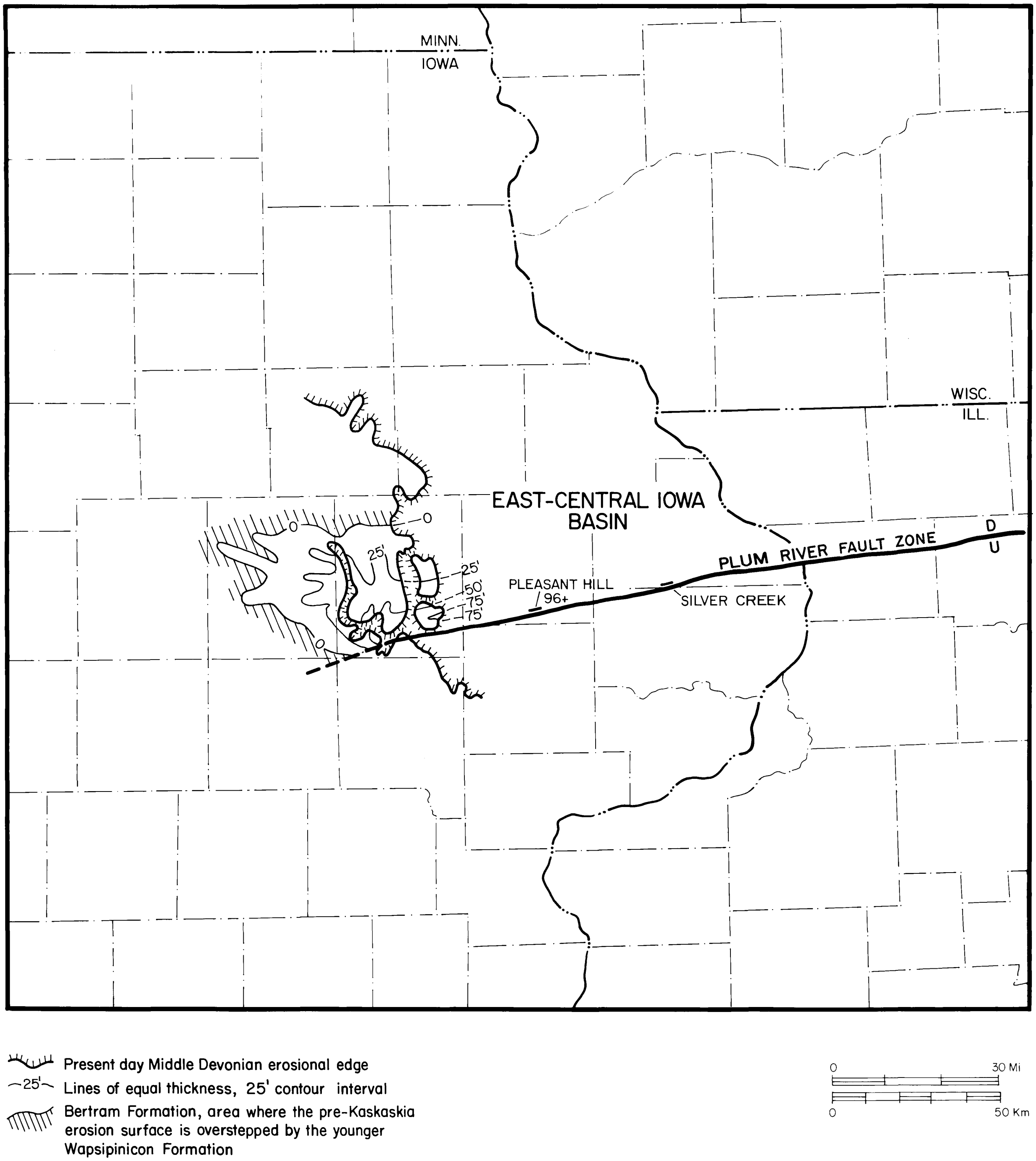

Figure 5. Isopach of the Bertram Fromation in east-central Iowa (modified from Bunker et al., 1985).

Group become increasingly abundant upward in areas where the Davenport Member is extensively brecciated (Sammis, 1978). The Davenport is an evaporite-dominated unit in parts of southern Iowa, where nodular, mosaic, and massive gypsum and anhydrite interbed with limestone and dolomite (Giraud, 1986). The extensive Davenport breccias are interpreted to have formed by evaporite-solution collapse processes, which were operating, at least in part, during open-marine deposition of the lower Cedar Valley Group. The Davenport Member overlies the Spring Grove with apparent conformity, and the contact is gradational or undefineable at some localities. The Davenport Member is about 6 to 10 m thick across most of its extent (3-6 m in northeastern Iowa). Thicknesses of up to 31 m are known in the evaporite-bearing region of southern Iowa.

Cedar Valley Group

Owen (1852) termed the Middle Devonian carbonate sequence of eastern Iowa the "limestones of Cedar Valley," and McGee (1891) formally designated this interval the "Cedar Valley limestone." Subsequent definition of the Wapsipinicon Formation restricted the Cedar Valley Limestone to the interval above the Wapsipinicon and below the Upper Devonian shales of the Sweetland Creek and Lime Creek formations. The Cedar Valley was elevated to group status (Witzke et al., 1988) to include four formations, each corresponding to a major transgressive-regressive cycle of deposition, and each separated from adjacent formations by an erosional unconformity or discontinuity surface. The constituent formations are, in ascending order, the Little Cedar, Coralville, Lithograph City, and Shell Rock (Fig. 3).

No type locality for the Cedar Valley Limestone was ever designated, but a primary reference section at the Conklin Quarry near Iowa City has been proposed (Bunker et al., 1985). Where overlain by the Sweetland Creek or Lime Creek formations, the Cedar Valley Group varies considerably in thickness, ranging from 23 to 40 m in southeastern Iowa and reaching maximum thicknesses of 80 to 120 m in northern and central Iowa. The Cedar Valley Group disconformably overlies the Wapsipinicon Group over much of its extent, but Cedar Valley strata overlap the Wapsipinicon edge to the south and west to overstep Ordovician or Silurian rocks in parts of northern Missouri, northern Illinois (Collinson and Atherton, 1975), and western Iowa (Figs. 1, 7). The Cedar Valley Group is dominated by fossiliferous limestone in southeastern Iowa and northern Illinois, by dolomite and limestone in northern Iowa, and by dolomite and anhydrite in central Iowa.

Little Cedar Formation

The Little Cedar Formation (Fig. 3) includes lower Cedar Valley strata which are bounded below by the Wapsipinicon Group (or Ordovician-Silurian rocks where the Wapsipinicon is absent) and above by the Coralville Formation. The type locality is at the Chickasaw Park Quarry (Witzke et al., 1988) adjacent to the Little Cedar River in southwestern Chickasaw County, Iowa; this locality exposes one of the most complete sections (17 m) of the formation in northern Iowa. The Little Cedar Formation ranges from 15 to 37 m in thickness; it is thinnest in southeastern Iowa and thickest in northern and central Iowa. The Coralville overlies a disconformity or discontinuity surface at the top of the Little Cedar Formation at most localities. However, the Lithograph City Formation is locally incised into the Little Cedar Formation in parts of Johnson County and the Sweetland Creek Shale overlies the formation at a few localities in southeastern Iowa. The Little Cedar Formation is interpreted to have been deposited during a large-scale transgressive-regressive cycle (part of cycle IIa of Johnson et al., 1985), the Taghanic onlap.

The formation is subdivided into three to four members in northern and central Iowa (in ascending order, Bassett, Chickasaw Shale, Eagle Center, and Hinkle) and two members in southeastern Iowa (Solon and Rapid). Subsequent discussion of the constituent members defines lithologic variations within the formation.

The Solon Member (Fig. 3) is dominated by fine skeletal muddy calcarenite (biomicrite and some biosparite) with scattered shaly or carbonaceous partings (Kettenbrink, 1973). Argillaceous calcilutite is present, especially near the northern limits of the member. Hardgrounds

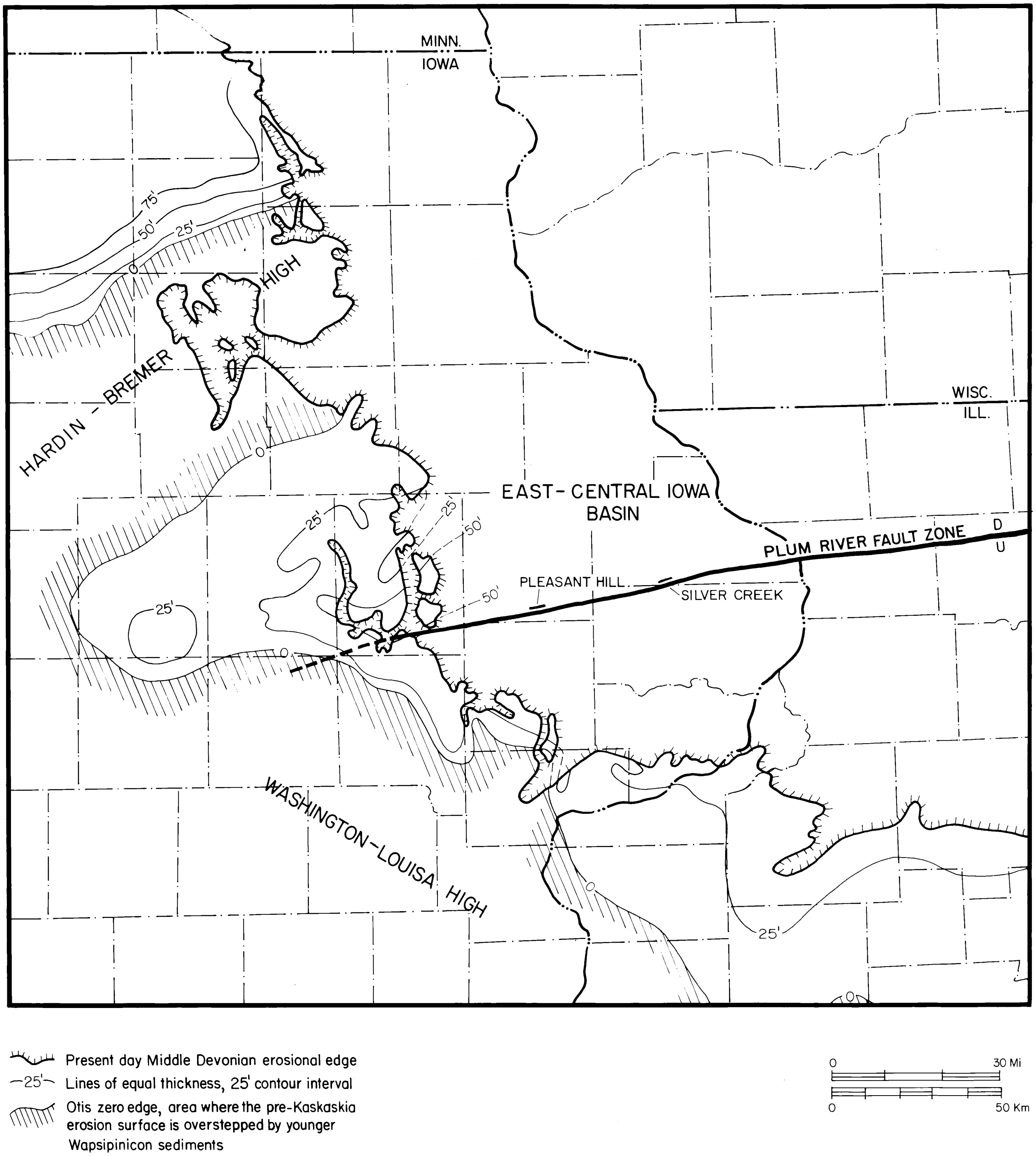

Figure 6. Isopach of the Otis and Spillville formations in east-central and northern Iowa (modified from Bunker et al., 1985).

are developed locally. A thin sandy limestone is commonly present at the base; this sandy interval locally includes sandstone facies (Hoing Sandstone) in parts of northern Missouri and western Illinois (Collinson and Atherton, 1975; Tissue, 1977). The basal contact is disconformable, but the upper contact with the Rapid Member is variably gradational or sharp (locally a burrowed discontinuity surface or hardground). The Solon Member is limited geographically to areas of east-central and southeastern Iowa and adjacent areas of northern Missouri and Illinois. The member varies from 1.5 to 12 m in thickness and is thinnest to the southeast. It generally thickens to the north, and skeletal calcarenites of the Solon are replaced by argillaceous dolomites of the lower Bassett Member in that direction.

Conodont faunas of the Solon span the upper Middle varcus through Lower hermanni-cristatus subzones (Bunker and Klapper, 1984; Witzke et al., 1985). The macrofauna of the Solon is abundant and diverse. The "independensis zone" of Stainbrook (1941a) spans approximately the lower half of the Solon and is named for the characteristic atrypid Independatrypa independensis. The "profunda beds" of Stainbrook (1941a) span the upper half of the Solon Member, and are named for the characteristic colonial rugosan Hexagonaria profunda. This interval is commonly coral and/or stromatoporoid rich and is locally biostromal (Mitchell, 1977). Solitary and colonial rugosans (including Asterobillingsa) and tabulates are generally common, but the biostromes are locally dominated by laminar stromatoporoids. The northern portion of the Solon outcrop belt, near its transition into the lower Bassett Member, locally contains packed accumulations (shell banks) of the terebratulid Rensselandia in the upper beds.

The Rapid Member (Fig. 3) is dominated by argillaceous calcilutite (ranging from micrite to sparse biomicrite), but shaly partings and lenses of calcarenite interbed with the sequence. The Rapid is divided into three widely recognizable descriptive lithologic units: 1) a lower fossiliferous calcilutite interval (some calcarenite) with common shaly partings ("bellula zone" of Stainbrook, 1941a); 2) a middle unit ("Pentamerella beds" of Stainbrook, 1941a) composed of interbedded fossiliferous calcilutite (some calcarenite) and unfossiliferous to sparsely fossiliferous burrowed calcilutite; this unit is capped by two widespread coralline biostromes (Zawistowski, 1971) and locally includes concentrations of glauconite and apatite pellets in some beds; and 3) an upper unit ("waterlooensis zone" of Stainbrook, 1941a) of fossiliferous calcilutite interbedded with lenses of echinoderm-rich calcarenite (packstones and grainstones to the north). The upper unit is glauconitic (in the lower half) in the type Rapid area, and is commonly cherty with prominent hardgrounds. The upper Rapid displays a greater degree of thickness and facies variations than is noted in the lower and middle strata; it varies in thickness from about 2 m in the south to 5 m in the north.

The Rapid Member conformably overlies the Solon, but its upper surface is marked by a widespread burrowed discontinuity surface over most of its geographic extent. It is sharply overlain by calcarenites of the Coralville Formation except near its northern limits where it is conformably overlain by the Hinkle Member (Fig. 3). The State Quarry Member of the Lithograph City Formation is locally incised into the Rapid in the type area, and the Sweetland Creek Shale overlies the Rapid at a few localities in the subsurface of extreme southeastern Iowa. The Rapid Member is recognized across southeastern and east-central Iowa and adjacent areas of western Illinois and northeastern Missouri. It is relatively uniform in lithology and thickness over its geographic extent, ranging in thickness from 13.5 to 18 m. The Rapid Member is replaced north of Palo by strata of the middle and upper Bassett, Chickasaw Shale, Eagle Center, and Hinkle members. The Solon and Rapid members are replaced southward in western Illinois and northern Missouri by skeletal calcarenites equivalent to part of the Callaway Formation.

Conodonts from the "bellula zone" span parts of the Lower and Upper hermanni-cristatus subzones (Witzke et al., 1985). The lower Rapid ("bellula zone") contains a diverse brachiopod-dominated fauna, which includes Spinatrypa bellula, Pseudoatrypa, Orthospirifer, Eosyringothyris, Tylothyris, Cyrtina, Schizophoria, Strophodonta, Protoleptostrophia, Productella, Striatochonetes, and others (Stainbrook, 1938a-1943b; Witzke et al., 1985).

Calcilutites of the middle Rapid contain faunas of slightly lower diversity than in underlying beds. Brachiopods (Schizophoria, Pseudoatrypa, Orthospirifer, Cyrtina, Tylothyris, and others), bryozoans, and echinoderm debris are characteristic. The upper part of the middle Rapid, by contrast, is an abundantly fossiliferous calcilutite to calcarenite containing conspicuous coralline biostromes over much of the geographic extent of the member. Conodonts of the subterminus Fauna, a probable equivalent of the disparilis Zone, first occur within this coralline interval (Witzke et al., 1985).

Conodonts from the upper Rapid ("waterlooensis zone") are assigned to the Lower subterminus Fauna (Witzke et al., 1985). The upper Rapid is characterized by crinoidal and bryozoan-rich calcilutites and calcarenites. A relatively diverse brachiopod fauna is present and includes Desquamatia waterlooensis, Orthospirifer, Tylothyris, Cyrtina, Eosyringothyris, Schizophoria, Strophodonta, Protoleptostrophia, Floweria, Productella, Striatochonetes, and others. Articulated specimens of camerate, inadunate, and flexible crinoids, blastoids, rhombiferans (Strobilocystites), starfish, edrioasteroids, and echinoids are known (Calhoun, 1983; Strimple, 1970).

The Bassett Member (Fig. 3) is dominated by slightly argillaceous to argillaceous dolomite, commonly vuggy, and containing scattered to abundant fossil molds. Limestones and dolomitic limestones generally increase in abundance throughout the member southward in the Iowa outcrop belt. Chert nodules occur locally, generally in the upper half. Stylolites and hardgrounds are present, especially in the lower half of the member. The basal Bassett Member is silty, sandy, and/or conglomeratic in areas where it overlies Ordovician strata.

The Bassett Member disconformably overlies the Wapsipinicon Group throughout most of the outcrop belt, but overlaps the Wapsipinicon edge to the west in the subsurface to overstep the eroded surface developed on Upper Ordovician strata. The member locally overlies beveled Silurian rocks along the trend of the Hardin-Bremer High (Fig. 4), where its basal unit is coralline (Dorheim and Koch, 1962). The Bassett is overlain conformably by the Chickasaw Shale in the northern outcrop belt, where it ranges in thickness from 19 to 25 m. The Bassett is overlain conformably by the Eagle Center Member to the south where it ranges in thickness from 15 to 25 m. The Chickasaw Shale and Eagle Center members are not recognized west of the outcrop belt in the subsurface of north-central and central Iowa; the Bassett averages 30 m in thickness and is conformably overlain by the Hinkle Member in those areas. The Bassett Member is dominated by dense, sparsely fossiliferous calcilutite and coralline to brachiopod-rich calcarenites near its southern limit. It interfingers with characteristic lithologies of the Solon and lower to middle Rapid members in that area, where it locally overlies the Solon Member and is overlain by upper Rapid strata.

Conodonts from the lower part of the Bassett include Icriodus brevis, I. latericrescens latericrescens, Polygnathus linguiformis linguiformis (gamma and epsilon morphotypes), P. ovatinodosus, P. alveoliposticus, P. ansatus, P. varcus, P. xylus xylus, and others (Klug, 1982b; Klapper and Barrick, 1983; Witzke et al., 1988). This fauna indicates assignment to the Middle varcus Subzone, and suggests correlation with most or all of the Solon Member to the south. The lower unit contains an abundant brachiopod fauna, typically atrypid-dominated (Independatrypa and Spinatrypa).

Conodonts of the middle and upper intervals of the Bassett Member are not zonally significant but include I. brevis, I. latericrescens latericrescens, P. xylus xylus, and P. ovatinodosus (middle unit); stratigraphic position suggests correlation with the lower and middle parts of the Rapid Member. The middle part of the Bassett is typified by sparsely fossiliferous burrowed calcilutite fabrics, but fossiliferous beds occur within the unit. This unit also includes local packstone beds of Rensselandia and sparse corals (pachyporids and favositids) near the southern limits of the member. The upper part contains biostromal beds rich in corals and/or stromatoporoids, variably dominated by favositids, solitary rugosans, Hexagonaria, or domal or laminar stromatoporoids, and including Asterobillingsa near its southern limit. Glauconitic and phosphatic strata below the

Eagle Center Member have produced an interesting fish fauna, as well as conodonts of the basal subterminus Fauna (Denison, 1985).

The Chickasaw Shale Member (Fig. 3) is composed of medium-gray dolomitic shale and argillaceous to shaly dolomite, in part silty. Nonskeletal, sparse to abundant burrow-mottled fabrics dominate, but skeletal material is noted in the lower 1 to 1.8 m at most localities (bryozoans, Desquamatia, and other brachiopods). The Chickasaw Shale ranges from 5.4 to 6.5 m in thickness. It is replaced to the south by strata of the Eagle Center Member and to the west by argillaceous and silty beds in the upper Bassett Member (Witzke and Bunker, 1984).

The Eagle Center Member (Fig. 3) consists of an interval of argillaceous and generally cherty and laminated dolomite below the Hinkle Member and above the Chickasaw Shale or Bassett Member. The member is dominated by sparsely fossiliferous to unfossiliferous burrowed argillaceous dolomite and contains prominent chert nodules and bands in the lower one-half to seven-eighths. Faint to prominent laminations, in part disrupted by scattered burrow mottles, characterize much of the member at most sections; some laminations are pyritic (Anderson and Garvin, 1984). Thin dolomitized or silicified fossiliferous calcilutite and calcarenite beds are interspersed locally within the generally unfossiliferous sequence. The Eagle Center is not dolomitized to the southeast of the type area, where it is dominated by sparsely fossiliferous to unfossiliferous, burrowed, cherty, argillaceous calcilutite, in part laminated, and contains thin skeletal calcarenite beds. Upper Eagle Center strata, primarily in areas where the member overlies the Chickasaw Shale, contain stromatoporoids or corals and are locally biostromal. The Eagle Center Member ranges in thickness from 8 to 11 m where it overlies the Bassett Member, and is 1.4 to 4.2 m thick where it overlies the Chickasaw Shale.

Conodonts from the Eagle Center Member (Icriodus subterminus and Polygnathus xylus xylus) are assigned to the Lower subterminus Fauna. Macrofauna is sparse in the member, but scattered fish debris (placoderm and shark) is noted in the laminated dolomites. Thin fossiliferous beds within the laminated sequence have yielded brachiopods (Desquamatia waterlooensis, Orthospirifer, Cranaena, and rhynchonellids), bryozoans, and crinoid debris (Anderson and Garvin, 1984). Upper strata are locally biostromal, primarily in the northern sections, and have yielded corals (Hexagonaria, solitary rugosans, and favositids), domal stromatoporoids, brachiopods, crinoid debris, and rostroconchs. The fauna and stratigraphic position indicates correlation of the Eagle Center with upper Rapid strata.

The Hinkle Member (Fig. 3) is the uppermost member of the Little Cedar Formation in northern and central Iowa, where it conformably overlies the Eagle Center or Bassett Member and is disconformably overlain by the Coralville Formation. It conformably overlies the upper Rapid Member along its southernmost extent. The Hinkle Member is characterized by dense unfossiliferous "sublithographic" limestone and dolomitic limestone, in part with laminated, pelletal, intraclastic, and "birdseye" fabrics. Similar fabrics are noted at all known sections, but the member is partially to completely dolomitized over most of north-central and central Iowa. Hinkle strata are generally unfossiliferous, but burrows, ostracodes, and sparse brachiopods have been noted locally. The member is commonly fractured to brecciated, and argillaceous beds and minor shale (locally carbonaceous) are present at many sections. Laminated carbonates are petroliferous in part, and desiccation cracks and minor erosional disconformities occur within some Hinkle sequences. Gypsum molds are present locally (e.g., Klug, 1982b, p. 47), and the member includes extensive evaporites (gypsum and anhydrite) in central Iowa (Fig. 7). The Hinkle changes character near its eastern margin where faintly laminated limestones are interbedded locally with thin fossiliferous limestone beds carrying brachiopods, echinoderm debris, favositids, and domal stromatoporoids. The Hinkle Member averages about 2.5 m in thickness and is known to vary between 0.4 and 4.1 m. Erosional relief, locally to 1 m, is evident below the Coralville Formation at some localities.

Coralville Formation

Keyes (1912) proposed the Coralville as a stratigraphic unit within the Cedar Valley Limestone, and Stainbrook (1941a) designated the type section at Conklin Quarry adjacent to the city of Coralville, Johnson County, Iowa. The Coralville Formation includes a lower fossiliferous carbonate member with an abundant marine fauna (Cou Falls or Gizzard Creek members) and an upper carbonate-dominated unit with laminated, brecciated, or evaporitic textures and some restricted-marine faunas (Iowa City Member). The Coralville Formation was deposited during a single transgressive-regressive depositional cycle and is bounded above and below by disconformities or discontinuity surfaces. The formation overlies the Little Cedar Formation at all known localities, and where capped by younger Devonian strata is variably overlain by the Lithograph City, Sweetland Creek, or Lime Creek formations. The Coralville formation varies greatly in thickness across Iowa, reaching a maximum thickness of 20 to 25 m in areas of central and northern Iowa. It is as thin as 3.9 m in parts of southeastern Iowa.

The Cou Falls Member (Fig. 3) is characterized by fossiliferous fine-grained calcarenite (primarily an abraded-grain packstone) with coral and stromatoporoid biostromes through much of the sequence (Kettenbrink, 1973). Thin shaly and dark carbonaceous partings occur in the lower half. The Cou Falls Member sharply overlies a prominent discontinuity surface at the top of the Rapid Member; calcarenites of the Cou Falls infill vertical burrows along this surface which locally penetrate up to 30 cm into upper Rapid strata. The Cou Falls Member is conformably overlain by the Iowa City Member in the type area. The Cou Falls Member encompasses the entire Coralville Formation east of the Iowa City Member edge (Fig. 3), where it contains calcarenites (generally coralline) in the lower part and argillaceous calcilutite to calcarenite in the upper part. The Andalusia Member of the Lithograph City Formation overlies a discontinuity surface at the top of the Cou Falls Member in parts of southeastern Iowa. The Cou Falls Member is replaced to the north and west by the Gizzard Creek Member and locally overlies Gizzard Creek strata in a transitional belt near its northern limits. The Cou Falls disconformably overlies the Hinkle Member of the Little Cedar Formation along the southern margin of that unit. The Cou Falls Member ranges from 5 to 7 m in thickness in the type area, and varies between about 3.5 and 11 m in thickness over its geographic extent.

Conodonts of the Cou Falls Member are sparse but include Icriodus subterminus, Mehlina gradata, and undescribed species of Polygnathus; these indicate assignment to the Upper subterminus Fauna (Witzke et al., 1985). Stainbrook (1941a) and Kettenbrink (1973) subdivided the lower Coralville sequence in Johnson County (Cou Falls Member) into two faunal intervals, the lower "Cranaena zone" and the upper "Idiostroma beds." The "Cranaena zone" contains prominent coralline biostromes dominated by colonial (Hexagonaria) and solitary rugosans (Pitrat, 1962), favositids, and massive stromatoporoids. Brachiopods are common in some beds (Day, 1988); Pseudoatrypa, Cranaena, Pholidostrophia, and Pentamerella generally dominate. The "Idiostroma beds" are characterized by biostromal strata containing branching ("Idiostroma") and massive stromatoporoids, colonial (Hexagonaria) and solitary rugosans, and favositids.

The Cou Falls Member east of the Iowa City edge resembles "Cranaena zone" strata in the lower part, but includes argillaceous calcilutites and calcarenites in the upper part with brachiopods and crinoid debris, locally with corals, stromatoporoids, or abundant bryozoans (Klug, 1982a; Witzke et al., 1985; Day, 1988).

The Gizzard Creek Member (Fig. 3) is dominated by dolomite, generally medium- to thick-bedded in the lower part and medium- to thin-bedded in the upper part, but dolomitic limestones and calcite-cemented (poikilotopic sparites) dolomites are present. The Gizzard Creek Member is slightly argillaceous in part, and calcite-filled vugs are common. Intraclasts are present locally in some beds. The member contains scattered to abundant fossil molds and displays wackestone (calcilutite) to rare packstone fabrics, in part burrow mottled. The Gizzard Creek Member disconformably overlies the Hinkle Member at all localities, and is conformably overlain by the Iowa City Member at most localities. The Gizzard Creek Member ranges from 3.7 to 7 m in thickness.

Conodonts of the Gizzard Creek Member include Icriodus subterminus, Mehlina gradata, and Polygnathus angustidiscus (Witzke et al., 1988) which are assigned to the Upper

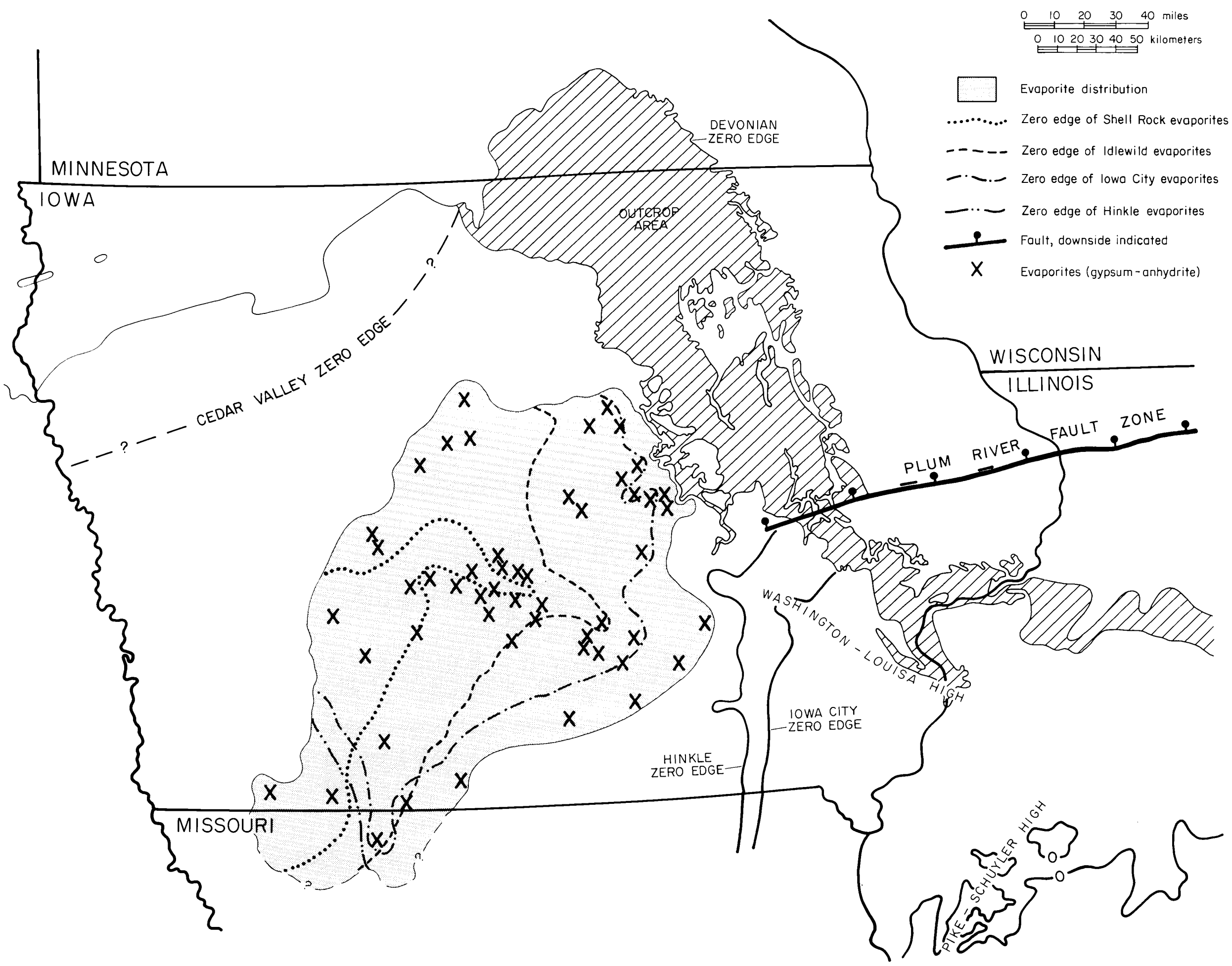

Figure 7. Evaporite distribution map of the Cedar Valley Group and the zero edges of the Hinkle and Iowa City members (from Witzke et al., 1988).

subterminus Fauna. Faunas of the Gizzard Creek are generally of low diversity and are characterized by sparse to abundant crinoid debris and brachiopods (Independatrypa, Athyris, and rare Tecnocyrtina; Day, 1988). Rare gastropods and bryozoans have been noted, and branching stromatoporoids and favositids are present locally near the southern limits of the member in the outcrop belt.

The Iowa City Member (Fig. 3) is characterized by a diverse assemblage of lithologies that commonly share significant lateral facies variations over short distances. The member in the type area of central Johnson County includes the following lithologies: 1) laminated and pelleted calcilutites, commonly "sublithographic" with "birdseye" voids and stylolites; 2) pelleted calcilutites with scattered to abundant corals and/or stromatoporoids; 3) intraclastic, brecciated, or oncolitic limestones; and 4) some thin shales, in part carbonaceous (Kettenbrink, 1973; Witzke, 1984). Mudcracks and vadose pisoliths are noted in some beds, and erosional surfaces occur locally within the sequence (Witzke, 1984).

The Iowa City Member in the northern outcrop belt and in the subsurface of central Iowa is characterized by sedimentary fabrics similar to those of the type area, but includes dolomites and dolomitic limestones. There is a general increase in the relative abundance of shale, with shaly intervals locally up to 2 m thick, breccia, and intraclastic strata in this area, and some beds are locally sandy. Evaporite molds have been identified locally. The thickest development of evaporites (gypsum and anhydrite) in the Cedar Valley Group occurs within the Iowa City Member of central Iowa. The Iowa City Member in the type area is disconformably overlain by the State Quarry Member of the Lithograph City Formation or by the Lime Creek Formation. The member is disconformably overlain by the Osage Springs Member of the Lithograph City Formation across northern and central Iowa. The Iowa City Member ranges from 0 to 8 m in thickness in the type area, and from 8 to 17 m across northern and central Iowa. The Iowa City Member is absent 12 km to the southeast of the type locality, where the entire Coralville Formation is represented by fossiliferous calcarenites of the Cou Falls Member. The edge of the Iowa City Member trends south-southwest from the type area (Fig. 7), and the member is absent in southeastern Iowa and adjacent parts of northeastern Missouri and western Illinois.

Conodonts have not been recovered from the Iowa City Member. Laminated and "birdseye"-bearing strata are sparsely fossiliferous in part (stromatolites, calcareous algae, foraminifers, ostracodes and gastropods), and some calcilutites are burrow mottled. Fossiliferous calcilutites and some calcarenites interbed with the sequence and contain low-diversity macrofaunas generally dominated by favositid corals and/or branching stromatoporoids (locally biostromal). A biostromal interval in the middle to upper part of the Iowa City Member ("Amphipora bed" of Kettenbrink, 1973) contains abundant branching stromatoporoids in the Johnson County type area, and this interval probably correlates with stromatoporoid-rich strata to the north.

Lithograph City Formation

The Lithograph City Formation was proposed for the interval lying disconformably between the Coralville Formation below and the Shell Rock Formation or Sweetland Creek Shale above. The type locality of the formation was designated in the old quarry area adjacent to the former town of Lithograph City, Floyd County, Iowa, where high quality stone for lithographic engraving was quarried in the early 1900s (see discussion and map in Bunker et al., 1986). The Lithograph City Formation in northern Iowa includes limestone, shale, and dolomite, variably fossiliferous, laminated, or brecciated; evaporites are present in central Iowa. The formation is dominated by fossiliferous limestone, dolomite, and shale in southeastern Iowa. Three members of the formation are recognized in northern Iowa (Osage Springs, Thunder Woman Shale, and Idlewild; Fig. 3). Two distinctive facies south of the northern outcrop belt are assigned member status within the Lithograph City Formation (Fig. 3; State Quarry Member in eastern Iowa and the Andalusia Member in southeastern Iowa and adjacent areas of northeastern Missouri and western Illinois). Where capped by younger Devonian strata, the formation ranges from about 20 to 36 m in thickness in northern and central Iowa. It is thinner to the southeast where it ranges from 0 to 12 m in thickness.

The Osage Springs Member (Fig. 3) is characterized by fossiliferous dolomite and dolomitic limestone, in part slightly argillaceous, in the type area. Calcite-filled vugs and stylolites are common, and poikilotopic calcite cements are present locally in the upper part of the member. Thin intervals containing faintly laminated to intraclastic fabrics have been noted at some localities. The Osage Springs Member in its type area of north-central Iowa is similar both in thickness and lithology to the Gizzard Creek Member of the Coralville Formation, but is distinguishable by its higher stratigraphic position and differing fauna. The Osage Springs Member becomes limestone-dominated (skeletal calcilutite and calcarenite) southward in the northern Iowa outcrop belt, and stromatoporoids (locally biostromal) become increasingly common in that direction. Fossiliferous and locally oolitic limestones and dolomites have been noted in central Iowa (Klug, 1982b). The member is conformably overlain by laminated carbonates of the Idlewild Member in the northern outcrop belt, and is conformably overlain by the Thunder Woman Shale in the southern outcrop belt and in the subsurface of central Iowa. The Osage Springs Member varies from 3.4 to 7.5 m in thickness.

The conodont Pandorinellina insita first occurs in northern Iowa in the basal Osage Springs Member; the insita Fauna correlates with the Lowermost asymmetricus Zone of the standard conodont zonation (Witzke et al., 1985). The macrofauna is dominated by brachiopods in northern outcrops; Allanella, Athyris, Independatrypa, and Strophodonta are characteristic (Day, 1988). Stromatoporoids become abundant to the south and include both massive and branching forms. Echinoderm debris is present in all sections, and bryozoans, gastropods, corals, and burrows have been noted locally.

The Thunder Woman Shale Member (Fig. 3) is characterized by light to medium gray, slightly dolomitic and silty shale; argillaceous dolomite is present locally, in part laminated and with gypsum molds. Shelly fossils are absent in the member, but horizontal and subhorizontal burrow mottles are common in the upper half. Conodont fragments and fish debris have been noted in the subsurface of north-central Iowa (Klug, 1982b). The Thunder Woman Shale is present in the southern part of the northern outcrop belt of the Lithograph City Formation, and extends into the subsurface of central Iowa (Bunker et al., 1986). It is erosionally truncated to the south within the Devonian outcrop of eastern Iowa. The member is replaced northward in the outcrop belt of northernmost Iowa and adjacent Minnesota by carbonate-dominated strata of the lower Idlewild Member (Fig. 3). The Thunder Woman Shale ranges from 3 to 6 m in thickness.

The Idlewild Member (Fig. 3) is characterized by an interbedded sequence of contrasting lithologic groupings: 1) laminated and pelleted lithographic and "sublithographic" limestones and their dolomitized equivalents, in part with mudcracks, "birdseye," or evaporite molds; 2) non-laminated dolomite and limestone, in part "sublithographic," pelleted, oncolitic, intraclastic, brecciated, and/or sandy, and locally containing mudcracks and "birdseye"; 3) calcareous shale, in part brecciated to intraclastic; and 4) fossiliferous dolomite and limestone (calcilutite and minor calcarenite), with scattered to abundant brachiopods and/or stromatoporoids (locally biostromal). Lithologic groupings 1 and 2 dominate the sequence at most localities, but group 4 lithologies are subequal in importance at some sections. Fossiliferous carbonates that interbed with the sequence cannot generally be correlated from section to section, although an interval of fossiliferous strata in the middle part of the member occurs at a similar stratigraphic position in most sections (lower "Unit D" of Witzke and Bunker, 1984) and probably correlates regionally. The Idlewild Member contains gypsum and anhydrite in the the subsurface of central Iowa (Fig. 7), primarily in the lower part of the member. The member is replaced by fossiliferous carbonates of the middle and upper Andalusia Member in southeastern Iowa. Where capped by the Shell Rock Formation, the Idlewild Member ranges from 16 to 24 m in thickness.

Conodonts from fossiliferous beds in the Idlewild Member include Pandorinellina insita and Polygnathus angustidiscus; these are assigned to the insita Fauna, however, regional relations suggest that the member spans a portion of the range of the Lowermost and Lower asymmetricus zones. Lithologic groupings 1 and 2 commonly contain ostracodes and are burrowed in part; stromatolites and gastropods have been noted locally. Fossiliferous beds in the member contain brachiopods (Day, 1988); Allanella and Athyris typically dominate. Echinoderm debris is common in some beds, and bryozoans, gastropods, and ostracodes also occur. Stromatoporoids are abundant in some beds, and locally form biostromes (domal or branching forms variably dominate). Favositids are present locally.

The State Quarry Member (Fig. 3) is restricted to Johnson County, Iowa, where it occupies broad channels (1 to 1.5 km wide) incised into the Coralville and Little Cedar formations (to as low as the middle Rapid Member). It is covered by Quaternary sediments at most localities, but it is overlain locally by the Lime Creek Formation. The State Quarry Member is characterized by fossiliferous calcarenites and calcilutites (Watson, 1974). Skeletal calcarenites (packstones and abraded grainstones) predominate at most localities, and are crossbedded in part. These are dominated by echinoderm, brachiopod, and/or stromatoporoid grains. Intraclastic and pelletal calcarenites also occur. Skeletal calcilutites are present near the channel margins. Fish bone lags are noted locally at or near the base of the member. The State Quarry Member reaches thicknesses of up to 12 m.

The conodont fauna of the State Quarry

Member includes Pandorinellina insita, Polygnathus angustidiscus, P. norrisi, and Icriodus subterminus (Watson, 1974; Witzke et al., 1985); it is assigned to the insita Fauna, and correlated with the Lowermost asymmetricus Zone. Other conodonts are present, many apparently reworked from Rapid and lower Coralville strata. The State Quarry Member contains a macrofauna characterized by abundant echinoderm debris, brachiopods, and stromatoporoids. A variety of brachiopods occur (Day, 1988), including Allanella, Independatrypa, Radiatrypa, and Ladogioides. Branching and massive stromatoporoids, solitary rugosans, favositids, auloporids, gastropods, nautiloids, spirorbids, ostracodes, trilobites, calcareous algae and foraminifera, and fish debris (placoderms and dipnoans) are noted (Watson, 1974).

The Andalusia Member (Fig. 3) is characterized by argillaceous and fossiliferous dolomitic limestone, limestone, and dolomite with fossiliferous calcareous shales in the lower part. Dolomite content generally increases upward in the section. Coral and stromatoporoid biostromes are present in the upper one-third of the member in its type area. Hardground and discontinuity surfaces, in part auloporid encrusted, occur within the Andalusia sequence in the lower and upper parts. The member overlies a discontinuity surface at the top of the Coralville Formation, and where capped by younger Devonian strata, is disconformably overlain by the Sweetland Creek Shale. The Andalusia Member is replaced by strata of the Osage Springs and Idlewild members to the northwest along the outcrop belt, and in subsurface sections it locally interfingers up depositional slope with the State Quarry Member in the basal part. The Andalusia Member, where capped by the Sweetland Creek Shale, ranges from about 6 to 12 m in thickness.

Conodonts of the insita Fauna range through most of the Andalusia Member and include Pandorinellina insita, Mehlina gradata, Icriodus subterminus, and Polygnathus sp. (Witzke et al., 1985). Uppermost strata of the member have yielded Ancyrodella rugosa, A. africana, A. alata (late form), Polygnathus asymmetricus, P. dubius, I. subterminus, and M. gradata; these forms indicate assignment to the upper part of the Lower asymmetricus Zone (ibid.). Brachiopods of the Andalusia Member include Strophodonta, Independatrypa, Schizophoria, Allanella, and Tecnocyrtina; Orthospirifer and Cranaena occur in the upper beds (faunal list in Day, 1988). Echinoderm debris is common to abundant, and bryozoans, bivalves, gastropods, rostroconchs, nautiloids, and fish debris are noted in some beds. Biostromal units in the upper Andalusia Member are variably dominated by solitary rugosans (but with some favositids) or massive stromatoporoids.

Shell Rock Formation

Belanski (1927) named the "Shellrock stage" (formation) for a limestone-dominated interval exposed along the Shell Rock River in northern Iowa, and subdivided it into three "substages" (members), in ascending order, the Mason City, Rock Grove, and Nora. The Shell Rock Formation is now included in the upper Cedar Valley Group (Witzke et al., 1988; Fig. 3). A comprehensive summary of the stratigraphy of the formation in the type area is given by Koch (1970). The Shell Rock Formation is characterized by fossiliferous carbonates with some shale in the type area, but incorporates laminated, "birdseye"-bearing, brecciated, and intraclastic facies in the western outcrop and subsurface. Where capped by younger Devonian strata, the Shell Rock Formation ranges from about 17 to 24 m in thickness over its known geographic extent in northern and central Iowa. It disconformably overlies the Idlewild Member, and erosional relief has been noted locally. The eroded upper surface of the formation is buried by the Lime Creek Formation. Varied lithofacies characterize the different members as discussed below.

The Mason City Member (Fig. 3) is dominated by skeletal limestone and dolomitic limestone (calcilutite to calcarenite), in part argillaceous, in the type area; it includes prominent stromatoporoid biostromes at all known localities. Hardground and discontinuity surfaces are noted in the member, and are locally encrusted with edrioasteroids and auloporids (Koch and Strimple, 1968). The Mason City Member becomes more dolomitic west of the type area, where it includes interbeds of "birdseye"-bearing (Anderson, 1984, p. 15, 40), intraclastic, and laminated carbonates. The member ranges from 6 to 11 m in thickness over most of its extent, but may thin to 1.5 m in the southernmost outcrop area (Belanski, 1927; Koch, 1970).

The Rock Grove Member (Fig. 3) is dominated by pelleted to skeletal limestone and dolomite, in part vuggy, and is poorly fossiliferous through much of the lower interval. Crossbedded calcarenites occur locally in the upper beds. The member includes shales and burrowed shaly limestones in the type area. Rock Grove strata incorporate laminated, intraclastic, brecciated, mudcracked, and "birdseye"-bearing lithologies to the west and southwest, especially in the lower half. Minor evaporites in the middle Shell Rock Formation of central Iowa (Fig. 7) probably correlate with the Rock Grove Member. The member varies from 2.5 to 8 m in thickness.

The Nora Member (Fig. 3) in the type area is characterized by prominent stromatoporoid biostromes separated by an intervening shale or argillaceous skeletal calcilutite. The member becomes more dolomitic, less shaly, and contains fewer stromatoporoids to the west and southwest. It locally includes "birdseye"-bearing and intraclastic beds. Where capped by the Lime Creek Formation, the Nora Member ranges from 3 to 7 m in thickness.

Conodonts of the Shell Rock Formation, which include Ancyrodella gigas, Polygnathus asymmetricus, and others (Anderson, 1964, 1966; Witzke et al., 1988), indicate correlation with the Middle and/or Upper asymmetricus zones. Brachiopod faunas of the Shell Rock are correlated with Faunal Interval 30 (= Middle asymmetricus Zone) of the western United States by Day (1988). Brachiopods and echinoderm debris are present in all members, and articulated specimens of crinoids, rhombiferans, edrioasteroids, and disarticulated echinoids are known from the Mason City Member (Belanski, 1928; Koch and Strimple, 1968; Strimple, 1970). Molluscs are common locally and include bivalves, gastropods, nautiloids, and scaphopods. Biostromal beds in the Mason City and Nora members are dominated by stromatoporoids, and massive (tabular to subspherical) and branching forms are present (see taxonomic studies by Stock, 1982, 1984a, b). Corals (solitary and colonial rugosans, and tabulates) occur in some beds. Additional fossils include ostracodes, spirorbids, conularids, calcispheres, calcareous algae, charophytes, and fish debris (Koch, 1970).

DEPOSITIONAL CYCLES

The Wapsipinicon and Cedar Valley groups display stratigraphic and biogeographic relations that are critical for understanding paleogeography and depositional systems in the Devonian seaways of the central midcontinent region of North America. The first marine transgression into the area was marked by deposition of Otis and Spillville strata during the Late Eifelian. This transgression also apparently breached the Transcontinental Arch, establishing faunal communication between eastern and western regions of North America across shallow cratonic facies in the Iowa area. Subsequent deposition of the Pinicon Ridge Formation marked a regional expansion of the seaway, but the expanded seaway apparently displayed restricted circulation patterns that excluded normal-marine benthos across the region. Antiestuarine circulation (Witzke, 1987) with circulatory restrictions to the east and northwest may have promoted the development of hypersalinity in the region, and extensive shallow-water and/or supratidal evaporites were deposited.

Subsequent deposition of the Cedar Valley Group was marked by significant expansion of the seaway, and open-marine facies spread across most of Iowa. Stratigraphic relationships within the Cedar Valley Group show a marked thinning of all formations into southeastern Iowa (Fig. 3). Although stratigraphic thinning is normally associated with shallowing depositional trends in many basins, facies in southeastern Iowa are consistently deeper-water and more open-marine than those to the north and west. In fact, the shallower-water facies, including evaporites (Fig. 7), occupy the central region of the Iowa Basin (Fig. 2). Therefore, the Iowa Basin did not develop as a bathymetric basin, but represents an intershelf basin in which shallow-water and mudflat sedimentation kept pace with increased subsidence during deposition of the Cedar Valley Group (and Lime Creek Formation as well, Witzke, 1987). Tidal-flat facies did not prograde out of the intershelf basin area during regressive episodes, but terminated at an intracratonic shelf margin, which is preserved in southeastern Iowa and adjacent northeastern Missouri (Fig. 7, Hinkle and Iowa City edges). This shelf margin

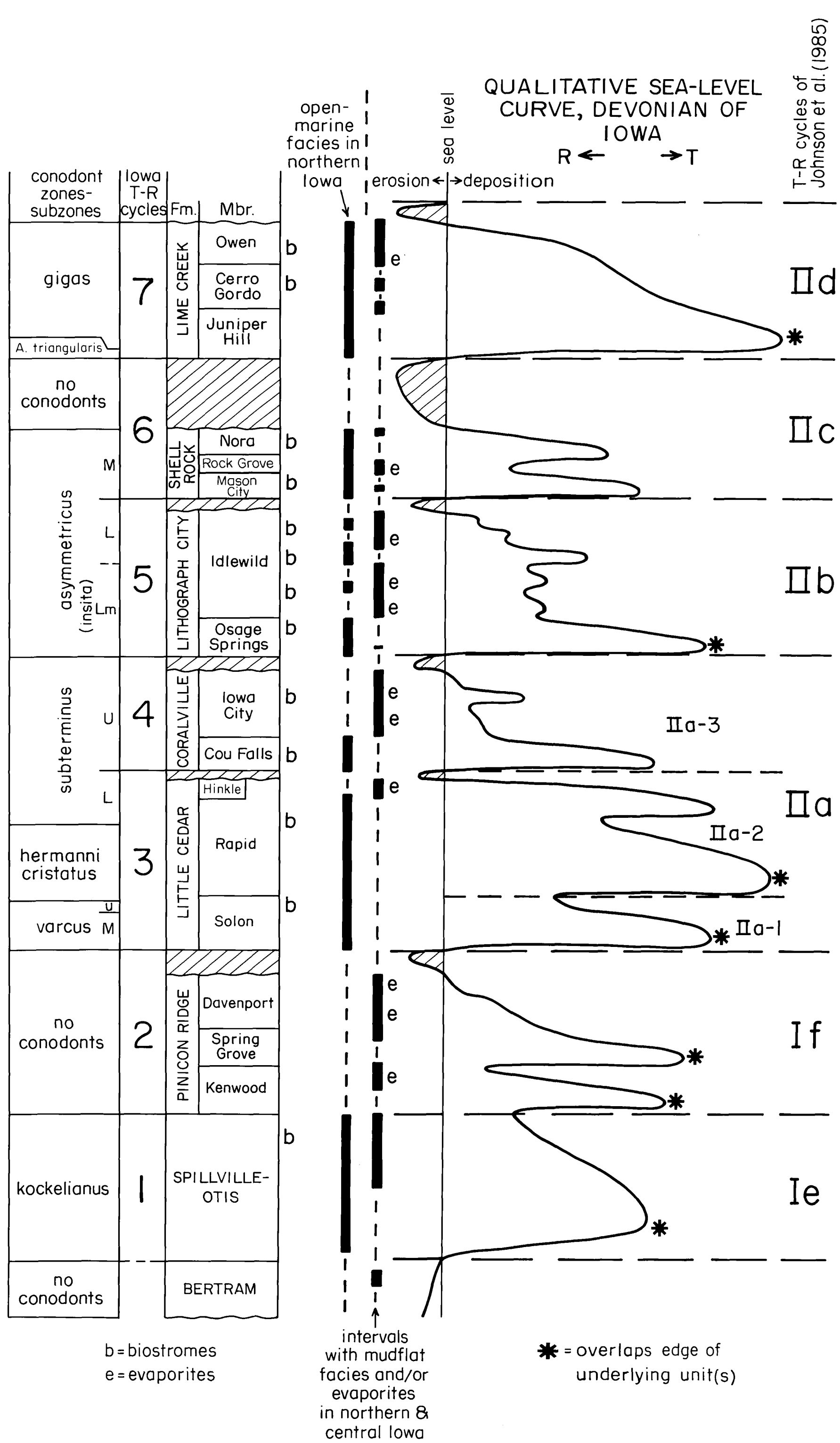

Figure 8. Qualitative sea-level curve for the late Middle and early Upper Devonian rocks of Iowa and their relationship to the T-R cycles of Johnson et al. (1985). (from Witzke et al., 1988).

bounded an area to the west termed the "Midcontinent Shelf" by Slingerland (1986), who numerically modelled tidal effects in the Late Devonian epicontinental seaway. Tidal influence is evident by extensive intertidal and supratidal mudflat facies in the Midcontinent Shelf area (i.e., the Iowa Basin area), and by tidal-channel facies (e.g. the State Quarry Member) along the intracratonic shelf margin.

The progressive deepening of depositional facies of the Cedar Valley Limestone to the southeast may relate to subsidence in the Illinois Basin, in a manner similar to that described for succeeding deposition of the Lime Creek Formation (Witzke, 1987). Sedimentation patterns in the Illinois Basin throughout the Devonian indicate consistently deeper-water depositional conditions than that interpreted for coeval strata in northern and central Iowa. However, a linear deepening trend between the Midcontinent Shelf area and the deep Illinois Basin is not as evident during Cedar Valley deposition, primarily because of an intervening structural high in central Illinois, the Sangamon Arch (Whiting and Stevenson, 1965).

Deposition of the Wapsipinicon and Cedar Valley groups in Iowa was marked by a series of six major transgressive-regressive depositional cycles (Figs. 3, 8). Each cycle of the Cedar Valley Group is bounded regionally by disconformities and each was terminated by the progradation of mudflat facies. Evaporite deposition generally occurred during the regressive portions of each cycle (Fig. 8). These cycles correspond closely to T-R cycles Ie through IIc of Johnson et al. (1985), and additional subcycles are recognized. In particular, Iowa T-R cycle 4 (Coralville Formation) apparently was not identified by Johnson et al. (1985) at other Euramerican localities. This suggests that Iowa T-R cycle 4 may not be entirely of eustatic origin, but may possibly relate to local subsidence rates in the Iowa Basin area. T-R cycle IIa of Johnson et al. (1985) is provisionally subdivided into three subcycles in the Iowa area as shown on Figure 8. Additional minor transgressive-regressive subcycles are interpreted for the following intervals: Kenwood, Spring Grove-Davenport, lower-middle Rapid, upper Rapid, upper Iowa City, middle-upper Idlewild, Mason City-lower Rock Grove, and upper Rock Grove-Nora.

A significant erosional hiatus (Upper *asymmetricus* and most or all of the *A. triangularis* zones) separates the Shell Rock and Lime Creek formations (Figs. 3, 8), indicating complete withdrawal of Devonian seas from the Iowa area following Shell Rock deposition. If general southeastward thinning of stratigraphic units and depositional trends observed through most of the Cedar Valley sequence also hold for the Shell Rock Formation, a thin Shell Rock section would be expected to have been deposited in southeastern Iowa (Fig. 3). The apparent absence of Shell Rock strata in this area would be anomalous were it not for the development of a significant regional unconformity following Shell Rock deposition. It is suggested that Shell Rock strata in southeastern Iowa were removed by erosion rather than nondeposition. Lime Creek sediments ("Independence shale") locally infill karstic openings within Middle Devonian carbonates of east-central Iowa, and an episode of pre-Lime Creek erosion and karstification has been interpreted (Bunker et al., 1985).

FIELD RELATIONSHIPS BETWEEN DEVONIAN ROCKS AND THE PLUM RIVER FAULT ZONE

The Plum River Fault Zone was named for exposures along Plum River near Savanna, Illinois (Kolata & Buschbach, 1976). Detailed geologic investigations in this area led to the recognition of a narrow belt of high angle faults trending roughly east-west for approximately 60 miles (97 km) through northwestern Illinois and east-central Iowa. The fault zone forms the principal element in an east-west trending belt of structural deformation formerly termed the Savanna-Sabula Anticline (Cady, 1920).

As originally defined (Kolata & Buschbach, 1976) the western terminus of the fault zone was in the area to the south of Maquoketa, Jackson County, Iowa. A Devonian outlier (Dorheim, 1953; Silver Creek Outlier of Bunker et al, 1985) delineates the location of the fault zone in this area. Further detailed field and subsurface studies (Bunker & Ludvigson, 1977; Ludvigson et al., 1978; Bunker et al., 1985) in Iowa, extended the western terminus of the fault zone approximately 60 miles (80 km) to an area south of Cedar Rapids, Linn County, Iowa. The field observation of other structurally preserved Devonian outliers (Skvor-Hartl Outlier, Dows & Mettler, 1962; Pleasant Hill Outlier, Ludvigson et al., 1978; Bunker et al., 1985) along the trend of the fault zone have helped to constrain it's geographic extent and orientation.

A number of geologic field mapping projects have been completed along the fault zone in Iowa, where bedrock is well exposed. These studies have been of critical importance in recognizing the internal structure of the fault (summarized in Bunker et al., 1985). The structural geometry of the fault zone is best displayed with the least ambiquity where rocks of the Devonian and Silurian systems are exposed in close proximity. The Devonian stratigraphy associated with three structurally preserved outliers along the Plum River Fault Zone is briefly discussed below:

1) Silver Creek Outlier (STOP 6, NE SE SW 33-T84N-R3E, Jackson County, Iowa) - This Devonian outlier was first observed and described by Dorheim (1953; Fig. 9). Baik (1980, p. 129-151) and Bunker et al. (1985, p. 58-61) described the structural relationships of this outlier to the Plum River Fault Zone. The Silver Creek Outlier is enclosed in an east-west trending graben, and its tectonic setting strongly suggests that various Middle Devonian and Silurian units are juxtaposed at this locality.

The oldest Middle Devonian rocks exposed within the graben belong to the Davenport Member (exposure "a" of Dorheim, 1953) of the Pinicon Ridge Formation. Dolomitic limestone breccias characterize the Davenport at this locality. Although older stratigraphic units of the Wapsipinicon Group are not exposed at Silver Creek, the inference by regional stratigraphic and structural relationships suggests their presence (as implied by Bunker et al., 1985, fig. 15).

Fossiliferous limestones of the Little Cedar Formation occur within one of at least two rotated fault blocks that collectively define the Silver Creek Graben (Ludvigson, 1988). Dorheim (1953) reported numerous brachiopods, corals, and bryozoans which are associated with the Rapid and Solon members from their type area in Johnson County, Iowa. Of particular note is the occurrence of Spinatrypa bellula from the northernmost Little Cedar exposures (exposure "b" of Dorheim, 1953), which is indicative of the

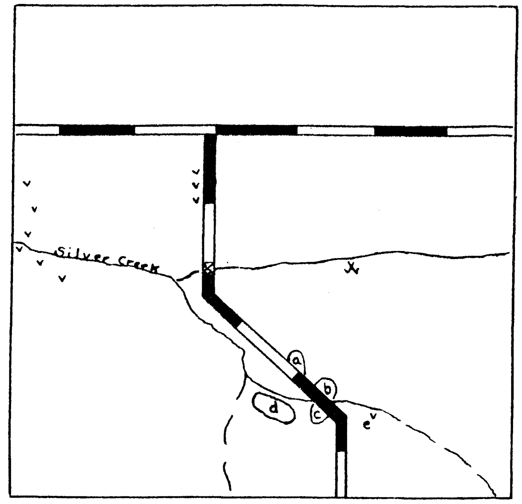

Figure 9. Silver Creek Outlier, map from Dorheim (1953). See STOP 6 for most recent geologic map of outlier.

"bellula zone" of the lower Rapid Member. Hexagonaria profunda typically characteristic of the upper Solon Member is noted at the southernmost exposures (base of exposure "b" and all of "c" of Dorheim, 1953). A single conodont sample from exposure "b" yielded common to abundant Icriodus brevis and Polygnathus xylus xylus, as well as two specimens of Schmidtognathus wittekindti. The occurrence of S. wittekindti is associated with the hermanni-cristatus Zone, which is consistent with the stratigraphic assignment to the lower Rapid Member ("bellula zone") at exposure "b" as defined by Dorheim (1953). Open marine strata of the Little Cedar Formation at the Silver Creek Graben are cut by abundant fissure-fills of unfossiliferous lime mudstones with calcite spars. Ludvigson (1988) has presented arguments that the internal sediments in these fissures are remnants of nonmarine strata from the upper Iowa City Member of the Coralville Formation.

2) Pleasant Hill Outlier ($s^{1}/_{2}$ 20-T83N-R2W, Jones County, Iowa) - The Pleasant Hill Outlier is a 150-350 m wide block of Middle Devonian carbonates that have been faulted into juxtaposition with Silurian dolostones. The structure and field relationships of this graben have been described in Ludvigson et al. (1978, p. 30-34), Cumerlato (1983), Bunker et al. (1985, p.

60-72), and Ludvigson (1988, p. 204-236). The estimated vertical displacement is approximately 300 feet (91 m).

The oldest Middle Devonian unit exposed is the Bertram Formation. These beds have a pronounced northward dip, which allow for a calculated thickness of 96 feet (29 m) making this the thickest known section of the Bertram in the region. The Otis Formation is poorly exposed at the west end of the graben. At the east end of the outlier the Spring Grove and Davenport members of the Pinicon Ridge Formation can be observed in continuous exposure with the Solon Member of the Little Cedar Formation. The youngest Middle Devonian rocks are exposed along the northern margins of the graben and belong to the "Idiostroma beds" of the Cou Falls Member, Coralville Formation.

3) Skvor-Hartl Outlier - The geology of the Skvor-Hartl area was first described by Dow and Mettler (1962), who recognized a pair of west-southwest plunging folds in sections 9 and 16, T82N, R6W, Linn County, Iowa. Ludvigson et al. (1978, p. 27) suggested that the Plum River Fault Zone continues through the Skvor-Hartl area, bisecting the paired west-southwest folds. Bunker et al. (1985) described the general structural framework of this area. The Skvor-Hartl area is the westernmost area where the general structural characteristics of the Plum River Fault Zone can be observed from surface features.

Strata of the Otis and Pinicon Ridge formations are exposed in a series of roadcuts along the north-facing bluffs to the south of the Cedar River in the central part of the Skvor-Hartl area. Approximately 1 mile north of the Skvor-Hartl area are exposures of the Bertram Formation. Outcropping along the axis of the west-southwest plunging syncline of Dow and Mettler (1962) are strata of the "Idiostroma beds", Cou Falls Member, Coralville Formation.

4) Westward extension of the Plum River Fault Zone into the Amana Colonies area. Although most previous interpretations of the western terminus of the Plum River Fault Zone have shown an questionable northeast-southwest trending fault with an opposite sense of throw at it's western end, the maps in this report have been modified by removing this questionable fault and extending the Plum River Fault Zone westward (by a dashed line) into northern Iowa County. Structural discordance, however, is noted in Iowa County near the Amana Colonies to the south of the Plum River Fault Zone, where exposures of Early Mississippian rocks (Late Kinderhookian, Wassonville Formation, Stookey, 1909, pp. 167, 184; Van Tyl, 1922, p. 78; Laudon, 1931, p. 377-378) are noted in near proximity to Upper Devonian (Late Frasnian) exposures of the Lime Creek Formation (Müller & Müller, 1957). Statigraphic displacements of 250-300 feet are estimated within the Amana Colonies area, with the downthrown side to the south. Unfortunately surface exposures and subsurface information are rather limited in this area and the trend of the "Amana fault" is not known with certainty, although it appears to subparallel the Plum River Fault Zone near it's western terminus (Wahl & Bunker, 1986).

SUMMARY

The hierachal order of stratigraphic nomenclature within the Middle Devonian rocks of the Midcontinent Shelf region is best developed within the framework of cyclic patterns of deposition. Transgressive-regressive sequences are common throughout the geologic column as noted in the numerous articles contained in this guidebook and generally express themselves within shelf sequences with greater clarity than in the deep basinal areas. The Wapsipinicon and Cedar Valley groups of the Midcontinent Shelf region record repetitive patterns of deposition, which can observed over large areas of the continental interior and correlated with eustatic sea-level events elsewhere across Euramerica. Extrapolation of these larger scale patterns of cyclic sedimentation across the Midcontinent region can be better tied together by careful observation and integration of remote outliers, such as those noted along the Plum River Fault Zone. These outliers can help provide critical detail for integration into paleogeographic and paleostructural reconstructions of the continental interior.

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A CASEYVILLE (MORROWAN) MIOSPORE ASSEMBLAGE FROM JACKSON COUNTY, IOWA

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Previous studies of miospores from Pennsylvanian deposits in Iowa by Ravn (1979,1986) and Ravn and Fitzgerald (1982) have shown that the Morrowan Caseyville Fm., the Atokan-Desmoinesian Kilbourn, Kalo, Floris, and Swede Hollow formations of the Cherokee Group, as well as post-Cherokee deposits all have diagnostic spore assemblages. The Cherokee Group deposits studied by Ravn (1979,1986) are located in numerous counties in the south-central part of the state (Forest City or Western Interior Basin), whereas the Caseyville assemblages are from Muscatine and Scott counties in the eastern part of the state (Illinois or Eastern Interior Basin; Ravn, 1986, Ravn and Fitzgerald, 1982).

Ludvigson's (1985) petrologic studies of sandstones in the Pennsylvanian rocks of Jackson County suggested that some are probably remnants of Caseyville strata. The carbonaceous mudstone from a measured creek bed section at DBF-2 (see Stop 5, this guidebook) was sampled and analyzed for plant microfossil content. The mudstone contains numerous lycopsid spores, with Lycospora being the most abundant genus in the assemblage. Three species that previously have been found only in Caseyville strata, Pustulatisporites papillosus, Densosporites variabilis, and Lycospora noctuina, are present. (Plate 1, Figs. g,n,d). The abundance of lycopsid spores (Lycospora, Densosporites, etc.) also corresponds to Ravn's (1986) Caseyville assemblage in eastern Iowa. Miospore assemblages in the younger Atokan-Desmoinesian Cherokee Group deposits from south-central Iowa do not contain characteristic Caseyville spores, and are marked by the first appearance of certain spores that are not seen in the DBF-2 assemblage.

All of the described Caseyville coals have been found to be dominated by spores with lycopodiaceous affinities. Many of these genera are related to small herbaceous lycopods (Densosporites spp., Cirratridites saturni), but arborescent lycopods (Lycospora spp.) are present as well. The assemblage found in the Caseyville coals to the south is thought to represent a partly open, partly forested fresh water swamp (Ravn and Fitzgerald, 1982). The miospore flora at DBF-2 comes from a mudstone in a coarse-grained fluvial sequence (see STOP 5, Fig. 10). Pennsylvanian sandstones at the DBF locality fill a channel complex that is incised into lower Paleozoic rocks, and probably include both Morrowan and Atokan-Desmoinesian units (see discussion of Stop 5 in the field trip guide). It's association with coarse-grained fluvial deposits suggests that the deposit originated as a cutoff channel or floodbasin swamp. The Jackson County assemblage probably represents a different paleoecology (floodplain rather than coal swamp), and is less diverse than assemblages studied in the Quad Cities area. Although many of the species are identical to those found in previous work on the Caseyville in Iowa, 5 forms could not be identified. This site is located to the north of what was previously recognized as the northernmost limit of the Illinois (Eastern Interior) Basin, and more information is needed to better understand the paleogeography and paleoecology of this region during the Morrowan.

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PLATE 1

a. Lycospora granulata Kosanke 1950, 29 μm

b. Lycospora pellucida (Wicher) Schopf, Wilson and Bentall 1944, 38.7 μm

c. Lycospora pellucida (Wicher) Schopf, Wilson and Bentall 1944, 35.5 μm

d. Lycospora noctuina Butterworth and Williams 1958, 29 μm

e. Cirratriradites saturnii (Ibrahim) Schopf, Wilson and Bentall 1944, 79 μm

f. ?Waltzipora sagittata Playford 1962, 22.5 μm

g. Pustulatisporites papillosus (Knox) Potonie and Kremp 1955, 29 μm

h. unknown sp. 1, 32 μm

i. unknown sp. 2, 22.5 μm

j. unknown sp. 3, 24.5 μm

k. unknown sp. 4, 25.8 μm

l. Ahrensisporites ornatus (Neves) Ravn 1986, 49 μm

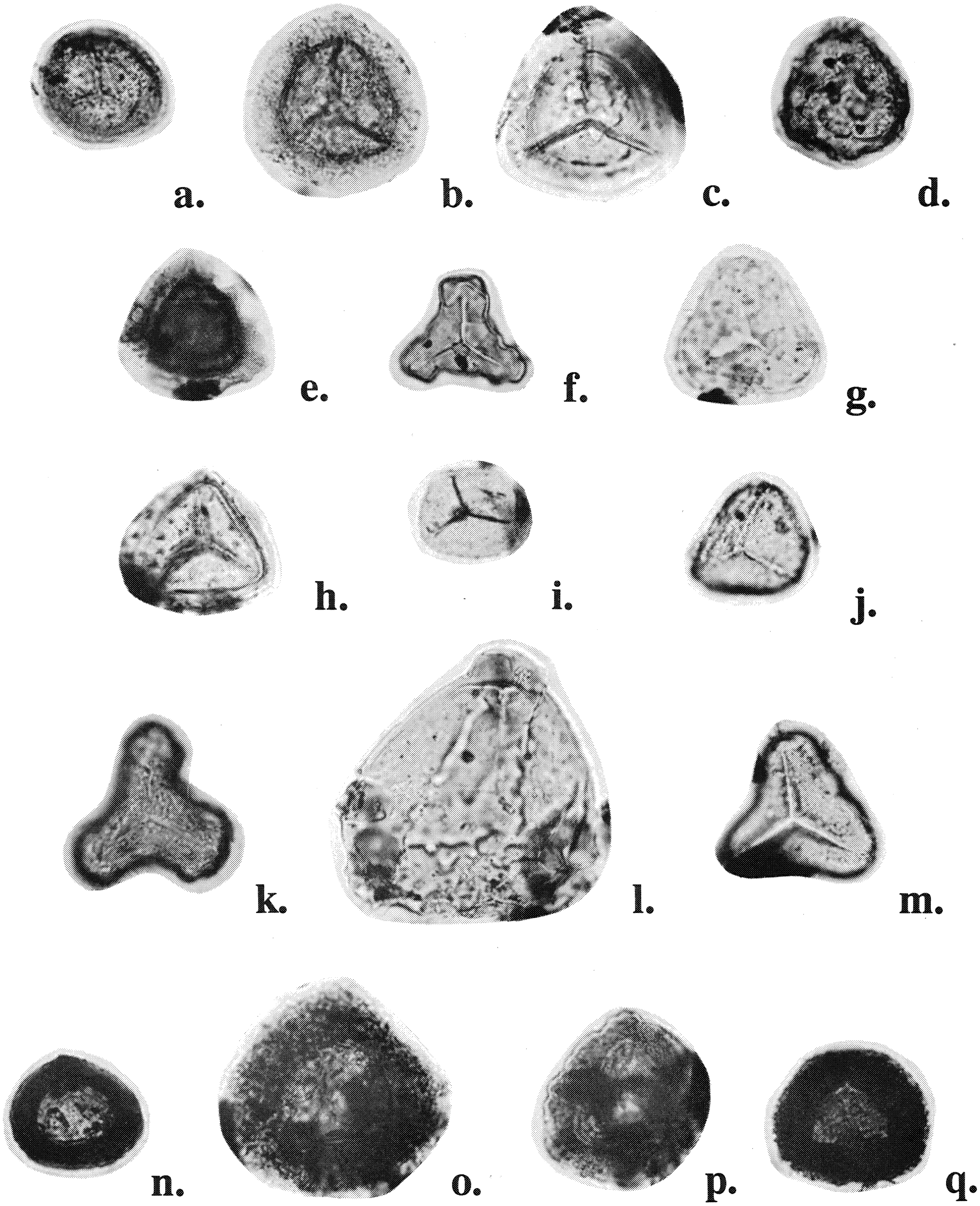
m. unknown sp. 5, 30 μm

n. Densosporites variabilis (Waltz) Potonie and Kremp 1956, 26 μm

o. Densosporites sphaerotriangularis Kosanke 1950, 43 μm

p. Densosporites sp., 37 μm

q. Densosporites sp., 29 μm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PART 2

Perspectives on Late Diagenetic Features

INTRODUCTION

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The presence of epigenetic lead sulfide minerals in the Paleozoic rocks of the Upper Mississippi Valley (UMV) stimulated the early exploration and settlement of the region. David Dale Owen's (1844) pioneering geological description of the UMV was originally chartered as an evaluation of new mineral lands. The zinc-lead ores of the UMV region are still regarded as the "type" example of the Mississippi Valley-Type (MVT) ore deposit, an important class of base metal ores found in cratonic sedimentary rocks. Although the UMV mining district has been inactive since 1979, the genesis of these deposits has remained a focus of serious scientific study (eg. Bethke, 1986; Sverjensky, 1986; Gerla, 1987). Epigenetic sulfide mineral deposits bearing general MVT affinities occur in the region surrounding the UMV mineral district (Fig. 1), and have been interpreted as products of the same hydrothermal fluid flow system that produced the commercial zinc-lead ores (Heyl and West, 1982). Other studies suggest, however, that the fringe area deposits may be the products of multiple processes and events (Garvin et al., 1987).

Perhaps the most (and maybe the only) endearing aspect of Paleozoic rocks to many frustrated midwestern "hard rock" geologists is their proclivity to host epigenetic mineral deposits. In addition to yielding a large number of museum-quality specimens to Geology Department collections, these deposits also have the potential to yield new insights into the physical and chemical processes that have acted on their host rocks since their early lithification. A dilemma is apparent, however, in deciding just who these "problem children" belong to. Mineralogists, economic geologists, analytical and theoretical geochemists, sedimentary petrologists, hydrogeologists, structural geologists, and stratigraphers all have important contributions to make in developing a holistic understanding of epigenetic mineral deposits. Undoubtedly, the cross-disciplinary nature of this endeavor has discouraged many "well-behaved" specialists from wading into the interpretive quagmire.

The advent of "basin analysis" as a new soft rock discipline may become something more than a merely restyled name for "regional stratigraphy" when practicing sedimentary specialists show that they are ready and willing to focus their attention on <u>all</u> of the fabric elements and mineral phases present in the rocks that they are studying. A survey of current geoscience journals suggests to me that this broadened perspective, recognizing the importance and utility of late diagenetic minerals, is gaining a foothold in mainstream research on sedimentary rocks. There appears to be a growing recognition that these features subtly preserve a record of both the kinematic and hydrologic phenomena associated with the evolution of sedimentary basins.

Calcite is probably the most abundant and widely distributed of the secondary minerals found in sedimentary rocks, and forms in a wide range of environmental settings. This fact, and the voluminous literature on it's geochemistry makes calcite and allied carbonate minerals especially useful for environmental interpretations. The papers by Ludvigson, Garvin and Ludvigson, and Spry and Kutz in this guidebook exemplify this utility.

Sulfide minerals are also widely distributed diagenetic minerals in sedimentary rocks, and can be emplaced by a variety of processes. Isotopic studies of sulfide sulfur, and chemically-bonded cations in some instances, can be used to draw environmental inferences about the diagenetic phases in which they reside. Papers by Spry and Kutz, Ludvigson and Millen, and Garvin and Ludvigson from this guidebook are examples.

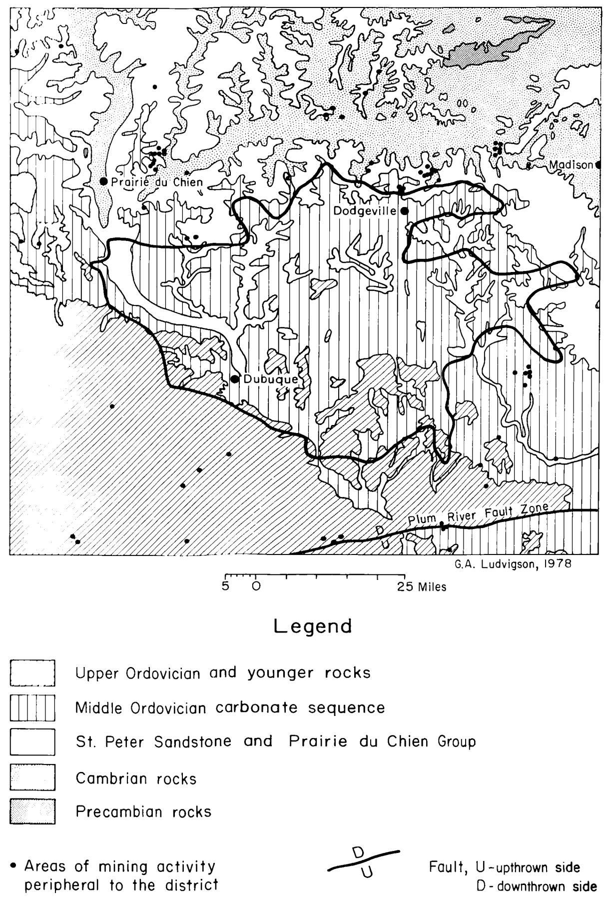

Figure 1. Regional bedrock geology of the Upper Mississippi Valley zinc-lead district, with areas of past mining activity in fringe area deposits. Boundaries of the mining district are shown with the heavy black line. Note the concentration of fringe area deposits near the Plum River Fault Zone. From Ludvigson et al. (1983).

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GEOLOGY AND ISOTOPE GEOCHEMISTRY OF MISSISSIPPI VALLEY-TYPE LEAD-ZINC MINERALIZATION AT MOUNT CARROLL, ILLINOIS

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INTRODUCTION

The Plum River Fault Zone (PRFZ) of eastern Iowa and northwest Illinois has been proposed to be a locus for possible concentrations of undiscovered Mississippi Valley-type (MVT) lead-zinc deposits (Heyl and West, 1982; Ludvigson et al., 1983). It has been further suggested that the deposits found along the fault are cogenetic with the formerly commercial zinc-lead ore deposits of the Upper Mississippi Valley (UMV) district (ibid.), a hypothesis that is evaluated here for the lead mining area at Mount Carroll, Illinois, the largest known group of deposits along the PRFZ, and the only group that has had a past history of economic exploitation.

The Mount Carroll lead mining area occurs in a locality where Silurian dolostones are in fault contact with Ordovician dolostones of the Galena Group along the PRFZ. Figure 1 shows that the Silurian Hopkinton Formation is exposed to the north of the PRFZ, whereas the Ordovician Dubuque, Wise Lake, and Dunleith formations are exposed to the south. Exposures of the Ordovician Maquoketa Group and the Silurian Scotch Grove and Gower formations occur within the PRFZ. While the precise brittle infrastructure of the near-surface rocks in the fault zone cannot be uniquely interpreted from the outcrop distribution, it is clear that a pattern of intersecting, or braided faults is present within the PRFZ in this area.

Figure 2 is a cross-section through the Mount Carroll lead mining area showing a structural interpretation of the PRFZ that assumes that all faults are vertical. The fault zone consists of three fault blocks, the northernmost containing exposures of the Silurian Gower Formation. The presence of Gower strata in the northern fault block suggests stratigraphic throws on the order of 600 feet (183 m) with respect to rocks exposed immediately to the south of the fault zone.

GEOLOGIC CONTROLS ON SULFIDE MINERAL OCCURRENCES AT MOUNT CARROLL AND RELATIONSHIPS TO UMV ORE DEPOSITS

All of the mining sites at Mount Carroll known to have been productive are contained within the Wise Lake Formation in the upper part of the Ordovician Galena Group, and occur south of the PRFZ (see paper by Witzke and Kolata in this guidebook for a discussion of the stratigraphy of the Galena Group). Sites for which structural data are available on the geometry and orientation of the deposits indicate that sulfide mineralization occurs in closely-spaced (decimeter to meter), centimeter-thick high angle veinlets striking west-northwest to east-northeast

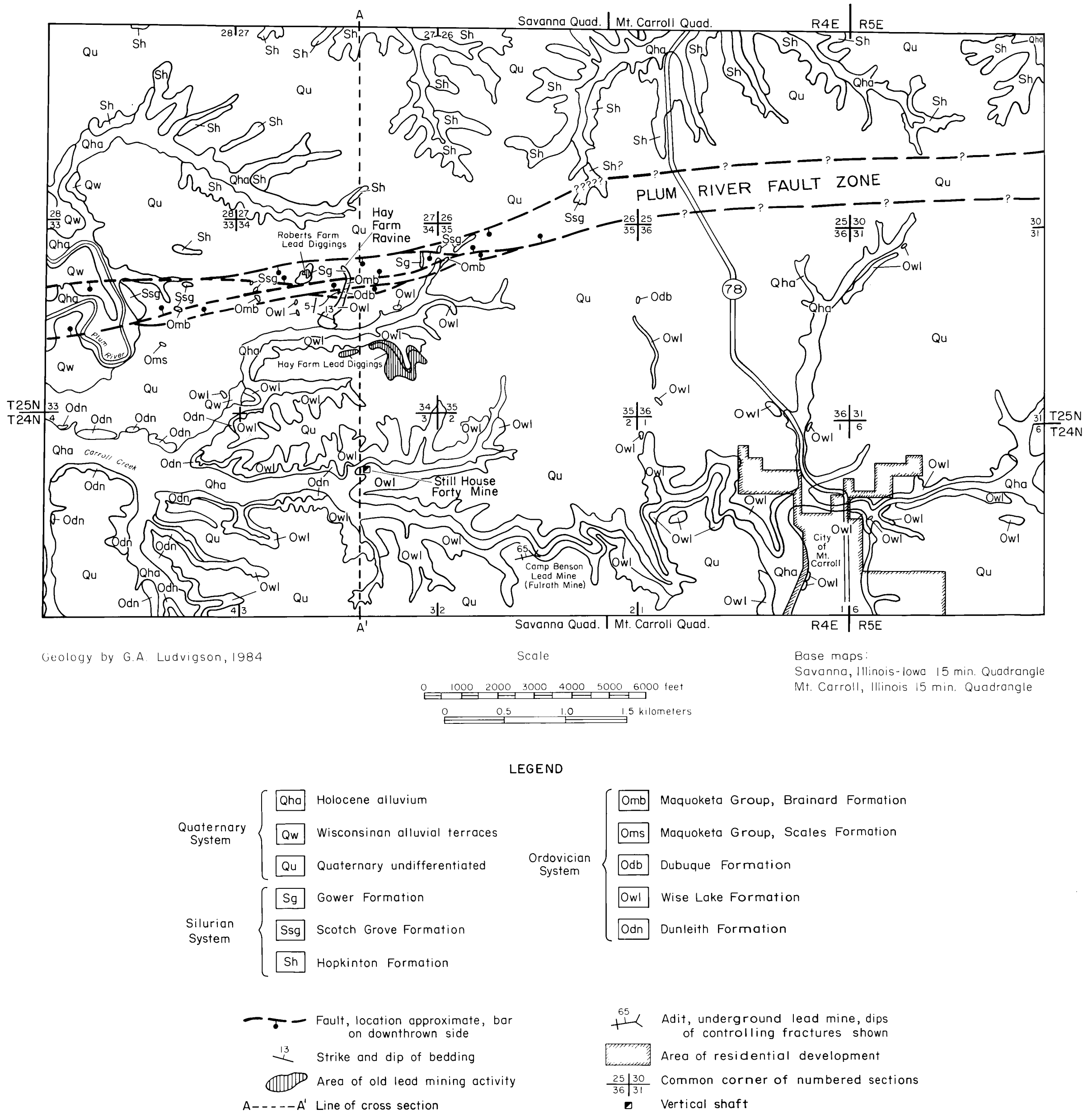


Figure 1. Geologic map of the Mount Carroll Lead Mining Area. From Ludvigson (1988).

(Ludvigson, 1988). These veinlets occur along the axis of a broad anticlinal uplift parallel to the PRFZ (Fig. 2), and are themselves sub-parallel to the PRFZ. MVT sulfide mineralization has not been observed in cataclastic rocks within the PRFZ at Mount Carroll, and appears to be confined to dilational structures on the upthrown side of the fault zone.

The stratigraphic position and structural geometry of the deposits at Mount Carroll are closely similar to one of two major types of deposits that were mined in the UMV zinc-lead district. The major UMV ore body types are discussed at length in Heyl et al. (1959), and are the gash vein and pitch and flat deposits. The gash vein deposits occur within the Wise Lake and

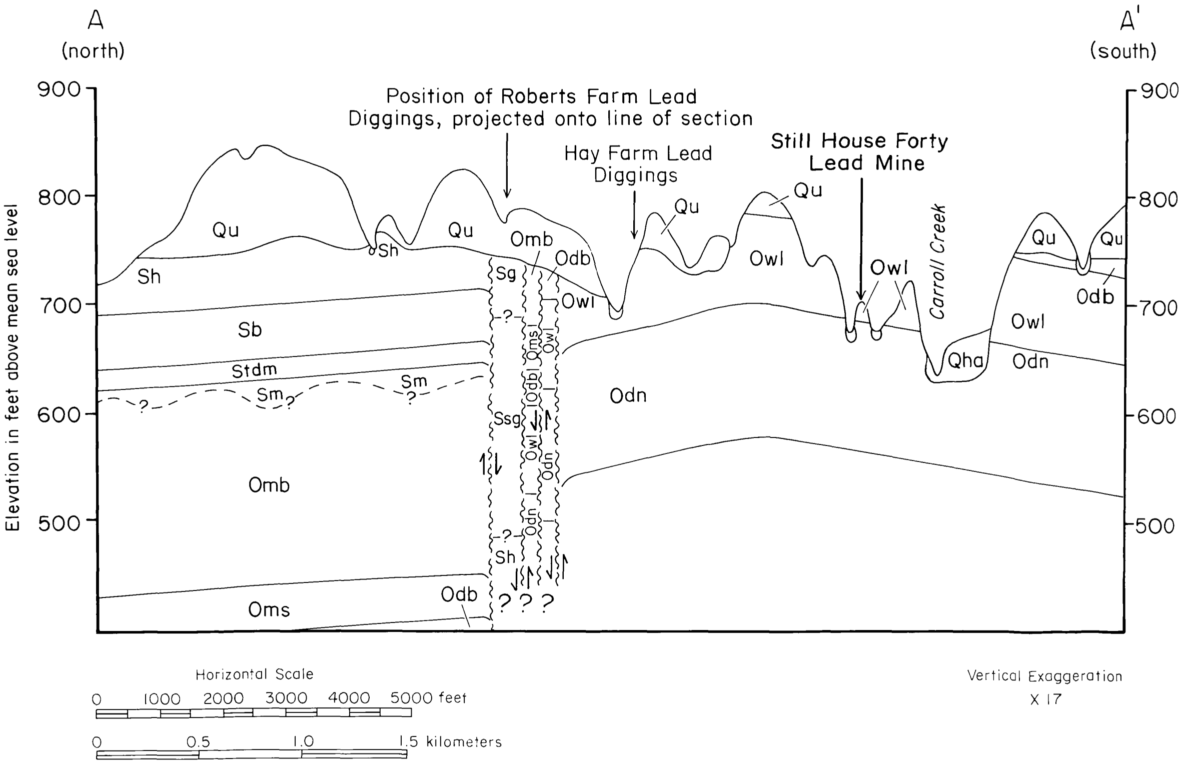


Figure 2. Geologic cross section along line A-A' in Figure 1. From Ludvigson (1988).

Dunleith formations, and occur as high-angle fracture fills (ibid., p. 81-82). They comprise the so-called "upper run" or "lead range" deposits that were mined in the early history of the district (Willman et al., 1946). The Mount Carroll lead deposits very closely resemble the gash vein deposits of the UMV district.

The more recently-exploited pitch and flat deposits principally occur in the Decorah Formation, but extend upward into the lower portions of the Dunleith Formation and downward into uppermost strata of the Platteville Formation. They are located along synclinal axes, commonly with arcuate terminations (Heyl et al., 1959, p. 81). Gerla (1987) has presented evidence that pitch and flat ore bodies originated from the development of fracture permeability as a consequence of the bending of rock strata during the regional development of low-amplitude folds. Precise three-dimensional and genetic relationships between the pitch and flat deposits and the overlying gash vein deposits, however, are not well known. The most widely-cited set of sulfur isotopic data on UMV minerals (McLimans, 1977) was assembled from collections from 14 different mines developed in pitch and flat ore bodies.

Earlier sulfur isotopic studies of minerals from the Mount Carroll area (Garvin et al. 1985, 1987) showed a gross dissimilarity to the deposits studied by McLimans (1977). This fact, and the geometric similarity between the Mount Carroll and the UMV gash vein deposits suggested the need to study the isotopic systematics of authentic gash vein deposits in the UMV district. The underground Fessler Mine No. 2 at the state-owned Mines of Spain property at Dubuque, Iowa provided such an opportunity. In 1985, Curtis Wright of the University of Dubuque sampled in-place galenas from the Fessler Mine for isotopic analysis. The results were reported by Ludvigson et al. (1986, 1987), and helped provide the basis for the current expanded view of the isotopic geochemistry of UMV zinc-lead ore deposits.

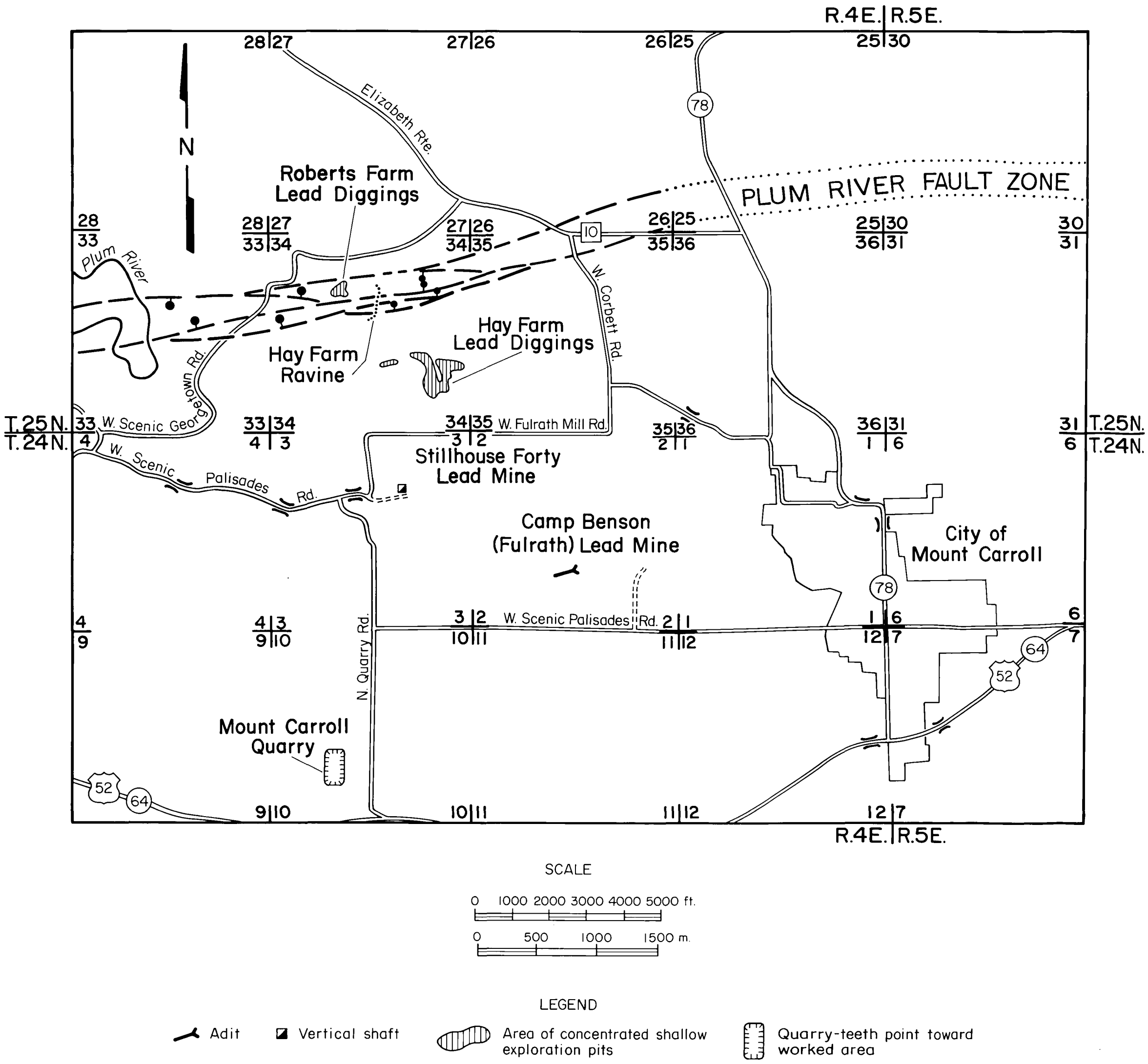


Figure 3. Location map of sampling areas in the Mount Carroll Lead Mining Area. From Ludvigson (1988).

HISTORY OF MINING ACTIVITY AT MOUNT CARROLL

The chronology of mining activity at Mount Carroll is poorly documented, but probably began shortly after the settlement of the area. Shaw (1873) referred to two failed commercial underground mining ventures, but provided no further information. General descriptions of 19th and 20th century lead mining operations in the Mount Carroll area are contained in Heyl et al. (1959, p. 293) and Heyl and West (1982, p. 1807, 1817). During geologic mapping in the area (see Figs. 1, 2) an effort was made to systematically collect samples of galena and any coexisting sulfide minerals from a representative group of deposits shown in Figure 3. Descriptive information on these deposits is included in Ludvigson (1988). During our work in this area, two former underground mining operations were examined, and they are described below.

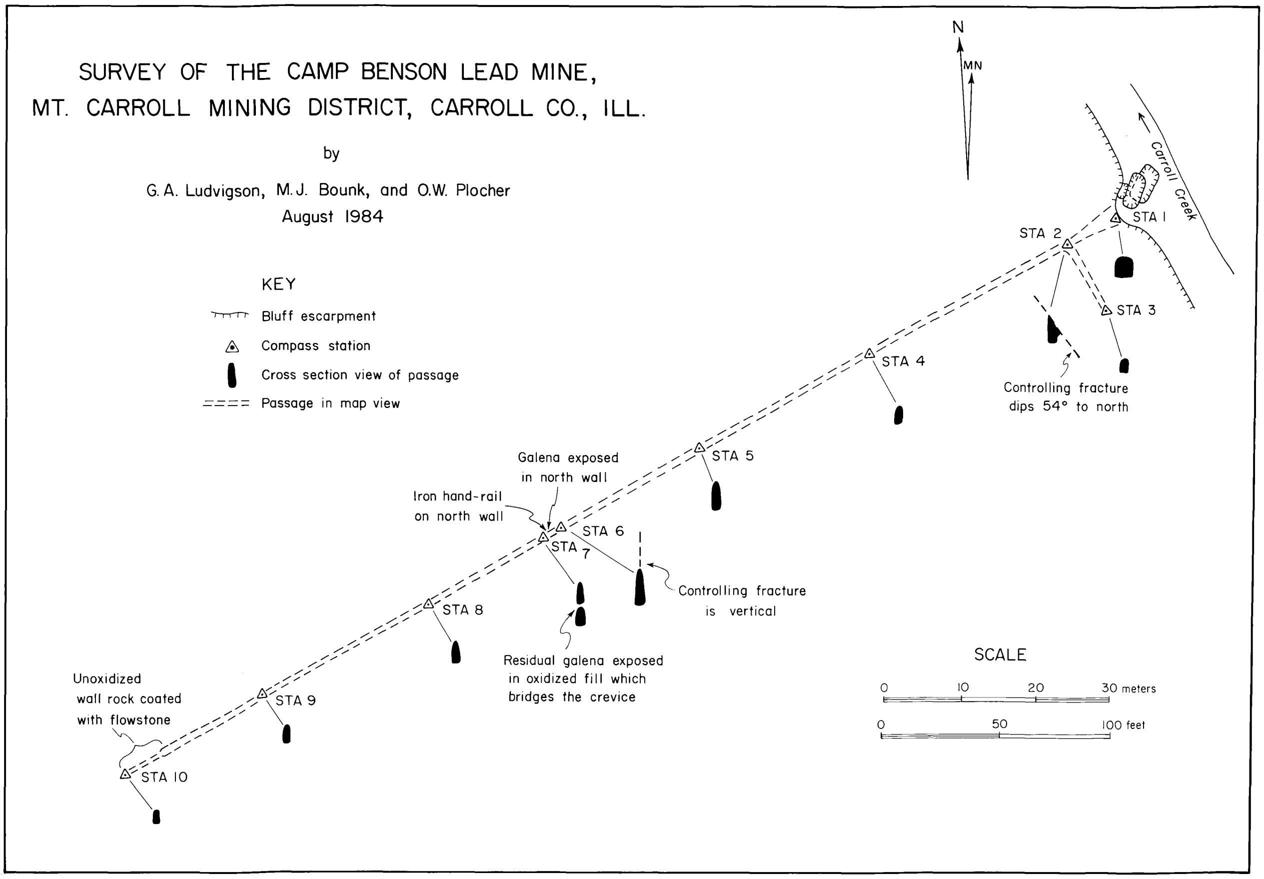


Figure 4. Map of the Camp Benson Lead Mine.

The Camp Benson Mine

This 600 foot-long (183 m) adit (Fig. 4) is located in Camp Benson, operated by the Sterling, Illinois, YMCA. The underground workings are currently being used by the camp, and access is provided by a cable cart spanning Carroll Creek. The opening is developed in a rock escarpment of the Wise Lake Formation, forming the creek bank along the outside circumference of an incised meander of the valley of Carroll Creek. The adit apparently was driven into an oxidized fissure for the recovery of residual galena. Tool marks in the workings indicate that most of the excavation was driven into soft, weathered rock above the water table. The last 20 feet were driven into wet, fresh rock that was mined with considerably more difficulty. Mine walls in this section are coated by flowstone, obscuring geologic relationships.

Galena occurs as abundant fragments in oxidized fill, on the floor of the mine, and as crystals coating the walls of the vein system along which the mine was developed. The size of the adit and the mining methods recorded by the workings indicate that this was a very small operation, employing only a few individuals.

According to Fay Christian, a long-time resident in the area, the Camp Benson Mine was operated by Adam Fulrath, whose family operated a mill along Carroll Creek. He further reported that mining operations were active at the turn of the 20th century. Electric lighting equipment of uncertain vintage was once strung throughout the length of the adit, but we do not know whether the installation was related to the period of mining activity.

The Still House Forty Mine

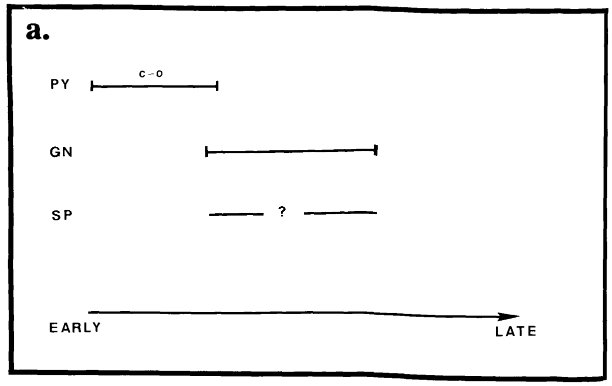
Hostetter (1913, p. 620) identified the area of the Still House Forty Mine as the area of the most extensive surface digging activity in the Mount Carroll area. Heyl and West (1982) also referred to a mine dump in the area of this mine, the most recent of the operations that we have encountered in the area. The mine, located on the Fecke farm property, consists of a trench driven into the wall of a small dimension stone quarry developed in the lower portion of the Wise Lake Formation. The trench leads eastward to the opening of an adit supported by timbers, now collapsed. The mine dump is at the north end of the quarry, and tailings evidently were hauled out through the trench. Further to the east is an abandoned, collapsed vertical shaft equipped with an internal combustion engine-powered steel hoist mounted on a concrete pad. Electric lighting was employed in the mine, and some of that equipment still remains.

Galena has been collected from the mine dump, as all underground workings are currently inaccessible. Alignment of the trench and shaft indicate that the deposit was developed in an east-west striking high-angle fracture system.

PETROLOGY OF SULFIDE MINERALIZATION

A suite of pristine pyrite, sphalerite, and galena samples was collected from the operating Mount Carroll Quarry during earlier visits, while marcasite and hydrothermal(?) dolomite were recogized and collected during later visits. Pyrite veinlets in dolowackestones of the Wise Lake Formation occur as smooth-walled fracture fillings 2-3 mm in width, with no appreciable impregnation of wallrock. Pyrite veinlets in argillaceous dolowackestones of the Dubuque Formation, however, are enveloped within zones of pervasive pyritic impregnation up to several centimeters thick.

The mineral paragenesis at Mount Carroll, reported by Garvin et al. (1987, see Fig. 5a) shows the first sulfide mineral phase as cubo-octahedral pyrite, followed by galena and sphalerite. Relationships between galena and sphalerite were uncertain because the two sulfide



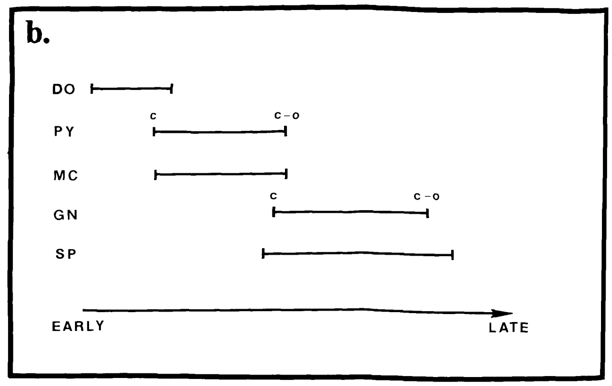


Figure 5. Mineral paragenesis of MVT mineralization at Mount Carroll, Illinois. a. As reported by Garvin et al. (1987); b. this study.

phases had not been observed together. Recently collected veinlet samples contain both minerals and permit a more complete assessment of mineral paragenesis.

The revised paragenetic sequence (Fig. 5b) begins with deposition of dolomite spars lining vugs. These spars partially envelope the earliest stages of cubic and cubo-octahedral pyrite and bladed marcasite mineralization, indicating inclusion of dolomite in the mineralizing sequence. Dolomite has not been observed in vein-filling sequences. Sphalerite precipitation followed in the vug-filling mineralization sequence, but was preceded by dolomite dissolution, as indicated by corroded contacts along vug walls (Ludvigson, 1988). The presence of darker- and lighter-colored sphalerites in the vug-filling sequences may be related to changes in

Fe contents (e.g., McLimans et al., 1980).

Pyrite and/or marcasite occur with galena and sphalerite in veinlets, and were the first sulfide minerals to precipitate. Galena and sphalerite both grow on the iron sulfide phases. Sphalerite deposition appears to span the entire length of galena deposition. New samples will permit more accurate determination of sulfur isotopic temperatures by enabling us to microsample co-precipitated sphalerite and galena. Early galenas precipitated as cubic crystals, while later growths appear as cubo-octahedrons and octahedrons. The paragenetic sequence shown in Figure 5b is similar to that described by Heyl (1968) for the UMV district. This suggests that the deposits at Mount Carroll and those in the UMV may have been produced during the same mineralizing event.

ISOTOPIC DATA

Chips of mineral samples from several mines and the operating quarry at Mount Carroll were extracted by chisel and analyzed for $\delta^{34}S$ at the laboratory of Dr. E. M. Ripley at Indiana University. Chip samples from galenas at Mount Carroll were also analyzed by Millen for $^{206}Pb/^{204}Pb$, $^{207}Pb/^{204}Pb$, and $^{208}Pb/^{204}Pb$ at the laboratory of Dr. C. W. Montgomery at Northern Illinois University. Analytical results reported in Ludvigson (1988) have been combined here with new unpublished lead and sulfur data generated from T. M. Millen's dissertation research at Northern Illinois University. The additional sulfur analyses were generated in the laboratory of Dr. E. C. Perry at Northern Illinois University.

Sulfur Isotopic Data

Histograms of the sulfur isotopic ratios of sulfide minerals at Mount Carroll and the Fessler Mine (Ludvigson et al., 1987; Millen and Ludvigson, 1987) are compared with McLiman's (1977) data from pitch and flat ore deposits in the UMV zinc-lead district in Figure 6. The data illustrate the depletion of the Mount Carroll and Fessler Mine deposits with respect to UMV pitch and flat deposits.

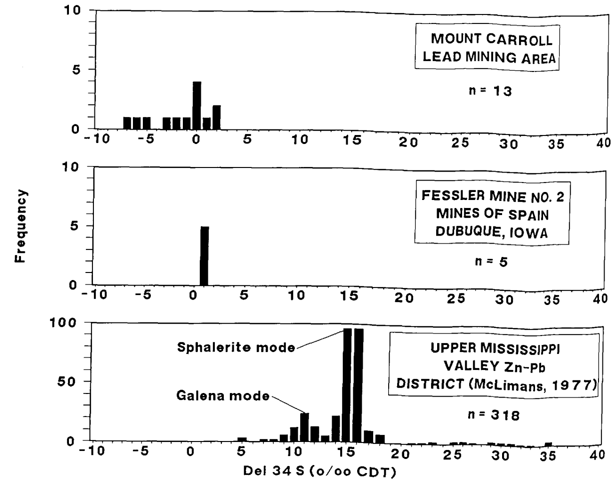


Figure 6. Histograms comparing the sulfur isotopic ratios from sulfide minerals from the Mount Carroll Lead Mining Area, the Fessler Mine, and the Upper Mississippi Valley Zinc-Lead District.

Lead Isotopic Data

Lead isotopic ratios from galenas in the Mount Carroll deposits are shown in Figure 7, and are compared with published data from Heyl et al. (1966) on ore leads from the UMV zinc-lead deposits. The plots show overlap in the $^{206}Pb/^{204}Pb$, $^{207}Pb/^{204}Pb$ and $^{208}Pb/^{204}Pb$ ratios between the two populations, and the ratios from the Mount Carroll leads appear to roughly conform to the linear trends documented for the UMV leads. Figure 8 shows that the $^{206}Pb/^{204}Pb$ ratios of galenas at Mount Carroll fit a regional trend of increasingly radiogenic leads in galenas from the southwest to the northeast across the UMV district. Cross-plots of $\delta^{34}S$ and $^{206}Pb/^{204}Pb$ measured from the Fessler Mine and all galenas at Mount Carroll except those from the Still House Forty Mine (Fig. 9) indicate a negative correlation between the two isotopic systems. This relationship has been previously documented from MVT ore deposits in southeast Missouri by Sverjensky et al. (1979) and by Millen and Ludvigson (1987), and Ludvigson et al. (1987) from the Upper Mississippi Valley. The galenas sampled from the dump at the Still House Forty Mine are heavier with respect to lead and sulfur than other galenas from Mount Carroll. In

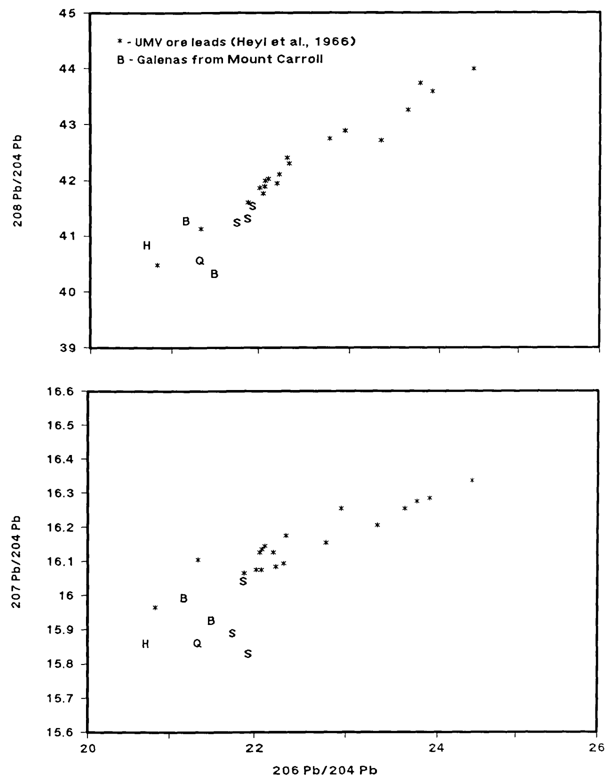


Figure 7. Comparison of lead isotopic ratios in galenas from the Mount Carroll lead mining area and the Upper Mississippi Valley zinc-lead district. B - Camp Benson Mine; H - Hay Farm Diggings; Q - Mount Carroll Quarry; S - Still House Forty Mine.

addition, the positive correlation between the two isotopic systems suggests a different evolutionary history for this deposit than for the other deposits at Mount Carroll.

Application of the Sulfur Isotope Geothermometer

The sulfur isotope geothermometer evaluates the temperature-dependent fractionation between two coprecipitated sulfide mineral species from a fluid with constant isotopic composition. Pyrite, sphalerite, and galena from separate vein- and vug-filling sequences at the Mount Carroll Quarry have closely similar $\delta^{34}S$ values, and appear to have formed in isotopic equilibrium, based on theoretical considerations discussed in Sakai (1968; where $\delta^{34}S_{pyrite} > \delta^{34}S_{sphalerite} > \delta^{34}S_{galena}$). Fractionations between these minerals from Mount Carroll have been used to calculate temperatures of sulfide mineral precipitation using the equations and constants specified by Ohmoto and Rye (1979, p. 518). Calculations for three different mineral pairs are shown in Table 1, with temperatures ranging from 107° to 295°C. The pyrite-galena pair probably provides the most reasonable estimate (200° to 239°C) since their coexistence in veinlets has been frequently observed. Although the sulfur isotopic ratios of sulfide minerals at Mount Carroll are markedly different from those in the UMV main district ores, the calculated temperatures of precipitation may establish some similarity between the two. McLimans (1977) reported isotopic temperatures ranging from 52° to 227°C for galena-sphalerite mineral pairs from UMV main district mines; temperatures based on fluid inclusion studies of UMV ores range from 75° to 220°C (ibid.).

DISCUSSION

Spatial relationships between the PRFZ and the sulfide mineral deposits at Mount Carroll indicate that while brittle deformation associated with cataclasis in major faults may develop extensive networks of effective fracture porosity marginal to the fault, zones of cataclastic deformation apparently have little effective porosity. The sulfide mineral occurrences of interest are associated with dilational mesofaults that are near, but not within zones of penetrative cataclasis. Mineralization at the Mount Carroll Quarry, and the Camp Benson and Still House Forty mines is controlled by dilational fractures that are sub-parallel to the PRFZ, but no MVT mineralization has been found to occur in the cataclastic rocks of the major fault. Finely-divided sulfides dispersed in cataclastic rocks from the Dubuque Formation (HFT-3C in Ludvigson, 1988) were analyzed for sulfur isotopes by whole rock methods, and were found to have a $\delta^{34}S$ value of -20 o/oo (reported in Ludvigson, 1987; Garvin et al., 1987; and Ludvigson, 1988). These sulfides are isotopically much lighter than and dissimilar to the sulfide minerals associated with MVT lead mineralization in the area (Fig. 6), and

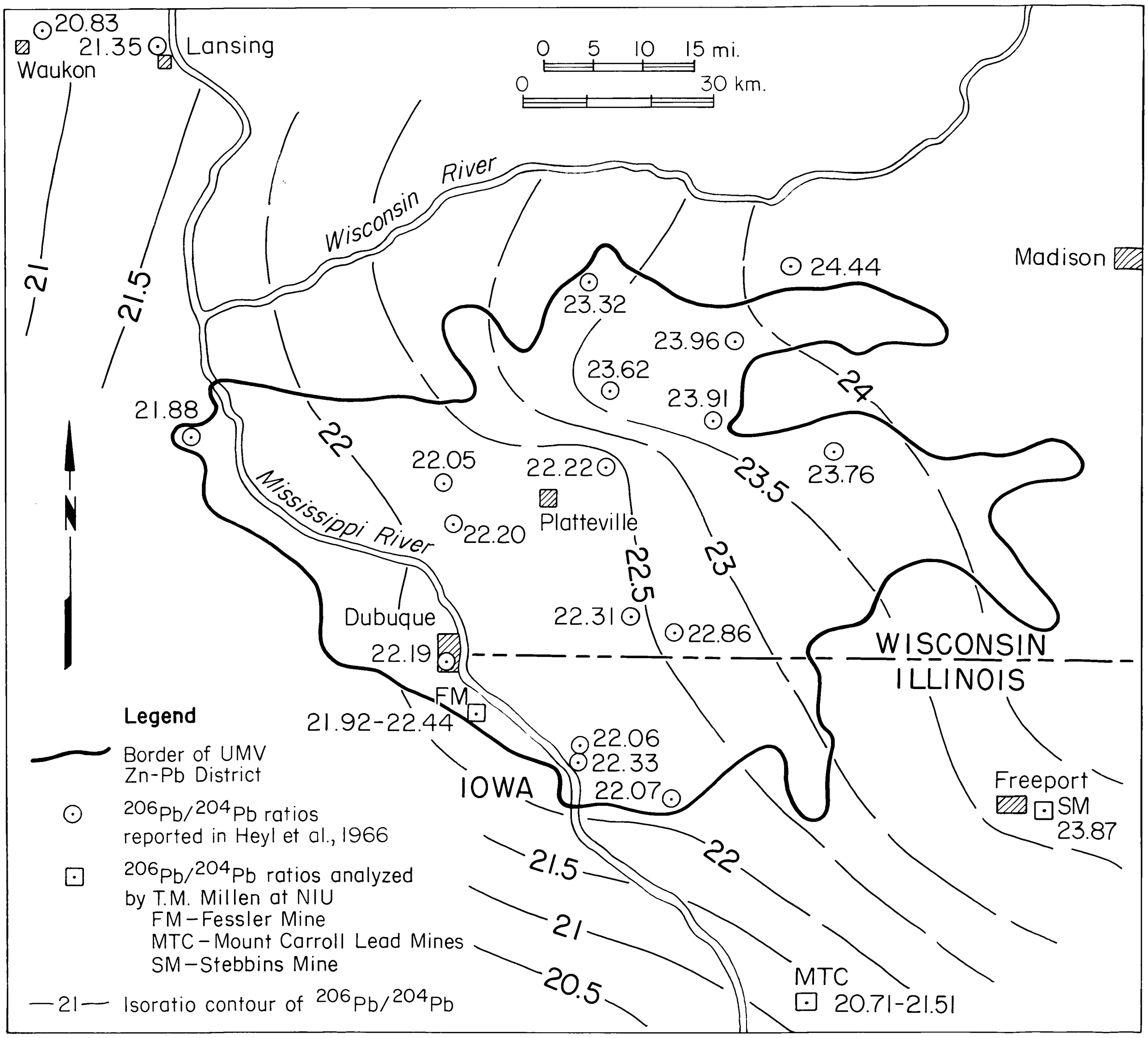


Figure 8. Regional lead isotopic zonation in galenas from the Upper Mississippi River Valley.

are interpreted to be early diagenetic sedimentary sulfides that are known to be widely dispersed in the Dubuque Formation on a regional basis (Levorson et al., 1979; Ludvigson, 1987). Modern hydrologic studies of fault rocks have also shown that while cataclasites generally are zones of low permeability, adjacent rock masses typically are hosts to enhanced fracture permeability (Snipes et al., 1986; Read, 1987). Sibson (1985, 1986) has noted that sites of extensional slip transfer along dilational fault jogs and zones of fluid pressure-induced rock dilation bordering seismogenic faults, rather than fault rocks, are the specific localities that host hydrothermal mineralization along brittle faults in the shallow crust.

Data presented in Figures 7 and 9 indicate that the $^{206}Pb/^{204}Pb$, $^{207}Pb/^{204}Pb$, and $^{208}Pb/^{204}Pb$ ratios in galenas from the Mount Carroll lead mining area conform to and overlap those from the UMV district, and fit into a regional zonation documented across the Upper Mississippi Valley. This information would seemingly support cogenesis between the two populations, or at least support arguments for a similar source of ore leads.

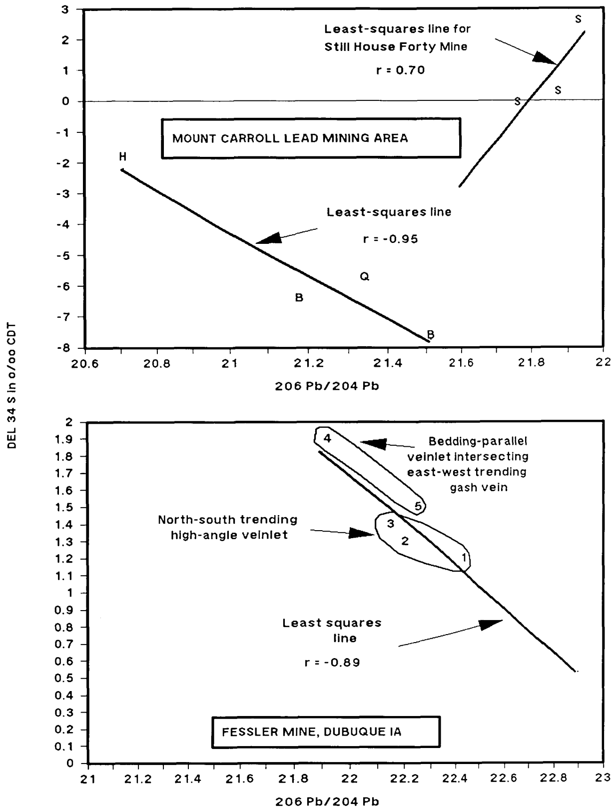


Figure 9. Covariation between lead and sulfur isotopic ratios in galenas from the Mount Carroll and Fessler Mines. B-Camp Benson Mine; H-Hay Farm Diggings; Q-Mount Carroll Quarry; S-Still House Forty Mine. Samples 1 through 5 in the Fessler Mine are described in Ludvigson et al. (1986).

The leads of the UMV district are enriched in ^{206}Pb, ^{207}Pb, and ^{208}Pb relative to many other terrestrial and extraterrestrial leads, and are part of a population of leads whose isotopic ratios yield negative (future) ages using the Holmes-Houtermans single-stage model for Pb-Pb dating (Faure, 1986, p. 317-320). The isotopes ^{206}Pb, ^{207}Pb, ^{208}Pb are the end-products of the decay series of ^{238}U, ^{235}U, and ^{232}Th, respectively, and thus suggest that the highly radiogenic leads in some MVT ores may have been leached from uranium- and thorium-rich Precambrian basement rocks (Doe and Zartman, 1979, p. 46). Doe et al. (1983) studied the U, Th, and Pb isotopic systems of the uraniferous 1.44 Ga UPH granite of northern Illinois (see paper by R. R. Anderson in this guidebook) and concluded that these rocks may have been affected by a Pb leaching event sometime between 400 and 260 Ma (Early Devonian-Permian). The UPH granite is part of a broad belt of so-called "anorogenic granites" that were emplaced in older North American continental crust during Proterozoic time (Anderson, 1983). The UPH granite apparently was enriched in uranium and thorium by the crustal remelting of older continental rocks that were formed during the 1.8 to 1.9 Ga Penokean Orogeny (Nelson and DePaolo, 1985; see paper by R. R. Anderson in this guidebook). The uppermost 70 m (230 ft.) of the UPH granite appears to have been hydrothermally altered, as indicated by disturbed oxygen isotopic fractionations between coexisting quartz and feldspar, an event believed to have coincided with the remobilization of lead by hydrothermal formation waters from overlying Paleozoic sandstone aquifers (Shieh, 1983). Doe et al. (1983, p. 7341-7344) calculated the possible range of lead isotopic ratios of the UPH granite during the postulated mid- to late-Paleozoic hydrothermal event(s), and concluded that these rocks could have served as a credible source of radiogenic leads for MVT deposits.

The linear arrays of "anomalous" radiogenic leads that characterize many MVT ores (Fig. 7) have been attributed to the mixing of "ordinary" single-stage leads with varying proportions of radiogenic Pb contributed from U- and Th-rich crustal rocks (Faure, 1986, p. 316-317). Doe et al. (1983, p.7344) suggested that the ordinary single-stage end-member of these leads may have been contributed from minerals in the sandstone aquifer units through which the mineralizing fluids passed. Thus, the lead isotopic data from galenas at Mount Carroll and the UMV district are characteristic of uranogenic and thorogenic Pb that has been complexly cycled through the upper continental crust. Colinear trends on Pb-evolution diagrams (Fig. 7) and matching patterns of regional isotopic zonation (Fig. 9) indicate that the leads at Mount Carroll were derived from the same reservoir(s) as those from the UMV main district deposits.

The $\delta^{34}S$ values of sulfide minerals at Mount Carroll and the Fessler Mine are lighter than and

Table 1. Sulfur isotopic temperatures from minerals at the Mount Carroll Quarry. From Ludvigson (1988).

Mineral Pair	Δ^{34}S o/oo	Temperature ^{o}C
pyrite-galena	4.2	200-239
sphalerite-galena	2.4	256-295
pyrite-sphalerite	1.8	107-167

do not overlap those from the UMV main district deposits studied by McLimans (1977), indicating a different evolutionary history for the respective mineralizing fluids. This information might suggest that the Mount Carroll deposits originated from a different mineralizing episode than the main district UMV ores, as suggested by Garvin et al. (1987). Studies at the Fessler Mine (Ludvigson et al., 1987; Millen and Ludvigson, 1987) have shown, however, that at least some UMV gash vein deposits developed in the upper part of the Galena Group also have lighter δ^{34}S values than the stratigraphically lower pitch and flat deposits studied by McLimans (1977). δ^{34}S data from the Fessler Mine have demonstrated that a wide range of values can occur within a single galena crystal (Ludvigson et al. 1986), similar to the range of values reported by Hart et al. (1981), Sverjensky et al. (1979), Sverjensky (1981), and Deloule et al. (1986) from southeast Missouri leads. The new data from the Fessler Mine contradict the widely-cited generalization that UMV ores are characterized by uniform strongly-positive δ^{34}S values (Sverjensky, 1986, p. 187-188), and may suggest that either multiple episodes of mineralization (each with distinctive δ^{34}S values) occurred in the UMV district, or that the $\delta^{34}S_{hydrogen\ sulfide}$ of the ore fluid(s) responsible for UMV mineralization varied more widely than was previously believed.

The negative linear correlation between ^{206}Pb/^{204}Pb (abcissa) and δ^{34}S observed in most of the galenas from Mount Carroll (Fig. 9) parallels similar observations made from MVT deposits in southeast Missouri (Sverjensky et al., 1979; Sverjensky, 1981) and at the Fessler Mine in the UMV main district (Millen and Ludvigson, 1987; Ludvigson et al., 1987). These relationships, previously unreported in the UMV district, have been used to argue for variations in the lead and sulfur isotopic compositions of mineralizing fluids, multiple sources of both lead and sulfur, and compatibility with reduced sulfur transport models (Sverjensky et al., 1979; Sverjensky, 1986, p. 188). The significance of the positive correlation between lead and sulfur isotopes at the Still House Forty Mine is not known at this time.

The strongly positive δ^{34}S values observed in sulfide minerals from the UMV main district ores suggest that the contained sulfur was aquired by the reduction of evaporitic sulfate by abiologic processes, possibly reactions with organic compounds (Sverjensky, 1986, p. 181). The lighter δ^{34}S values observed at Mount Carroll and Fessler Mine could possibly be explained by the dilution of isotopically heavy dissolved sulfide by the aquisition of isotopically light sulfur from the leaching of early diagenetic sedimentary sulfides. This hypothesis is not very satisfying, however, because the Wise Lake Formation does not appear to contain any higher concentrations of sedimentary sulfide than the underlying units hosting the pitch and flat deposits. Alternatively, the clustering of the Mount Carroll and Fessler Mine δ^{34}S values near the Canyon Diablo troilite standard raises the possibility that the sulfur in these fracture-controlled deposits was aquired by the leaching of igneous sulfides (0 o/oo CDT; Ohmoto and Rye, 1979), perhaps from underlying Precambrian basement rocks, as suggested by Garvin et al. (1987). This consideration further suggests the possibility that the solutes contained in the sulfide mineral deposits at Mount Carroll may have been leached from deformed, pulverized granites in the basement rocks of the PRFZ, or from dilational fracture networks that were developed in the Proterozoic igneous rocks buried beneath the axis of the Paleozoic Savanna-Sabula Anticlinal System of Bunker et al. (1985, p. 98-99).

GENESIS OF UPPER MISSISSIPPI VALLEY ZINC-LEAD ORES

The origins of the UMV zinc-lead ores and all MVT sulfide deposits in cratonic sedimentary sequences have represented a challenging enigma to generations of geologists. Conceptions of MVT ore genesis historically have been subject to fundamental shifts in majority opinion (summarized in Heyl et al., 1959), a phenomenon that continues today. While there is general agreement that large-scale commercial concentrations of MVT base-metal deposits must originate from the regional migration of large volumes of metalliferous hydrothermal brines from sedimentary basin centers, interpretations of mechanisms of fluid flow, of migration pathways, of timing, and of solute transport and precipitation mechanisms are still subjects of major dispute. At this time, the prevailing opinion is that the UMV zinc-lead ores were precipitated from brines that flowed northward from the center of the Illinois Basin in late Paleozoic time, although compaction-driven (Cathles and Smith, 1983) and gravity-driven (Bethke, 1986a) fluid-flow systems have been defended in the recent literature.

One of the most recent developments in the study of MVT ore fields is the growing recognition that the regionally pervasive brine flow systems required for the formation of these deposits may have originated from large-scale orogenic processes operating along convergent plate boundaries at continental margins. Oliver (1986) has suggested that the orogenic emplacement of thrust sheets over sedimentary wedges at continental margins may lead to the inboard migration of overpressured, heated brines. He also noted that the tectonic elevation of collisional orogens can lead to the establishment of inboard-draining, gravity-driven fluid-flow systems in foreland basins, a paradigm applied by Garven (1985) to the Pine Point deposits in western Canada.

Leach and Rowan (1986) proposed that the hydrothermal brines responsible for the formation of MVT deposits in the Midcontinent United States may have flowed from the Ouachita collisional margin in Permian time. Their hypothesis is supported by a large and steadily-growing body of radiometric data from metamorphic and authigenic minerals indicating a continental-scale heating event that affected Pennsylvanian sedimentary rocks in the Ouachita region (Desborough et al., 1985), and lower Paleozoic sedimentary rocks of the Ouachita, Ozark, and Alleghanian regions between 250 to 300 Ma (Shelton et al., 1986; Hearn et al., 1986; Hearn and Sutter, 1985). Bethke (1986a) suggested that the UMV ores were precipitated by a northward-draining gravity-driven hydrothermal flow system that was recharged into the Illinois Basin (the site of conductive heating) from the Pascola Arch, a Permian uplift that developed in the Ouachita foreland (Viele, 1983). Further evidence for this fossil basinal or "oil-field brine" flow system is provided by the discovery of traces of aliphatic oil in the pore networks of the upper 516 m of the UPH granite of northern Illinois (Couture and Seitz, 1986), the same Proterozoic rocks from which radiogenic ore leads were leached between 400 and 260 Ma (Doe et al., 1983). A preponderance of the recent literature thus seems to support a genetic link between the formation of the UMV zinc-lead ore deposits and a complex interplay of tectonic and hydrologic processes operating along the Ouachita collisional margin of the North American craton during the Permian.

GENESIS OF MVT DEPOSITS AT MOUNT CARROLL, ILLINOIS

Isotopic data from the Mount Carroll deposits indicate that the lead in these deposits was derived from the same reservoir(s) as the radiogenic leads in UMV ores, but that the sulfur in galena, sphalerite, and pyrite at Mount Carroll are all significantly lighter than that documented from well-studied UMV pitch and flat type ore deposits. This difference needs to be considered in light of new sulfur isotopic data from a gash vein type deposit in the UMV district (Millen and Ludvigson, 1987), where similar light sulfurs have been encountered. It is probable that the apparent uniqueness of the Mount Carroll deposits is an artifact of incomplete sampling in the UMV main district, and that high-angle vein systems with isotopically light sulfurs are common, but as yet poorly studied.

Paleotemperatures at Mount Carroll determined by the sulfur isotope geothermometer (107^{o}C to 295^{o}C) are slightly higher than those determined by McLimans (1977) for UMV ore minerals (52^{o}C to 227^{o}C). The data must be further refined, however, and analyses of confirmed coprecipitated phases need to be acquired. The elevated temperatures suggested from the current data imply that the hydrothermal fluids responsible for the precipitation of sulfide minerals at Mount Carroll, if derived from a sedimentary basin, must have been part of a flow system driven by gravitational head as opposed to a system driven by compactive overpressuring, as greater conductive heat losses (and lower temperatures of mineralization) would result from the lower fluid velocities associated with the latter mechanism (Bethke, 1986b). Given the slightly higher temperatures of sulfide mineral precipitation calculated for the deposits at Mount Carroll, they could be interpreted as resulting from: 1) a separate episode of mineralization resulting from a hydrothermal flow system like that proposed by Bethke (1986a), with a unique population of isotopically light sulfurs, possibly leached from igneous sulfides in underlying fractured basement rocks, or 2) the same episode of mineralization as that proposed by Bethke (1986a), with the higher temperatures resulting from Mount Carroll's more southward position (than the UMV main district), closer to the source of heating in the Illinois Basin, or 3) an episode of coseismic deformation during which conductively-heated pore fluids in the underlying 1.44 Ga anorogenic granites were injected into dilational fracture systems in the Paleozoic rocks bounding the PRFZ, similar to mechanisms suggested by Sibson (1986).

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A FLUID INCLUSION AND STABLE ISOTOPE STUDY OF MINOR UPPER MISSISSIPPI VALLEY-TYPE SULFIDE MINERALIZATION IN IOWA, WISCONSIN AND ILLINOIS

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INTRODUCTION

The Upper Mississippi Valley (UMV) district is one of the largest base metal districts in the United States; it covers more than 7,800 square km in northeast Iowa, southwest Wisconsin, and northwest Illinois (Fig. 1). The largest deposits are close to the southern margin of the district in Iowa and Illinois. Heyl and West (1982) suggested that potentially productive extensions of the district occur up to 100 km to the south and southwest in these states. Zinc, lead, copper, iron sulfides, barite, and less commonly nickel and cobalt minerals occur in the district, but they are known from over 60 localities in an area which covers 100,000 square km surrounding the main district. The deposits in the UMV district are primarily located in Ordovician sedimentary rocks, whereas those in the outlying districts occur in rocks ranging from Cambrian to Mississippian in age (Fig. 1). Heyl et al. (1959) and Heyl and West (1982) suggested that the minor occurrences may be remnants of fluid pathways from main UMV mineralization. Although considerable attention has been directed toward understanding the genesis of the lead-zinc mineralization in the main UMV district, an increasing number of investigations have centered on the origin of the minor base metal occurrences which surround the main UMV district. The principal focus of these investigations relates to the hypothesis that many of the minor outlying occurrences are cogenetic with the main UMV district mineralization (Heyl et al., 1959; Garvin, 1982; Heyl and West, 1982; Garvin, 1984b; Coveney et al., 1987; Garvin et al., 1987). This hypothesis is based on the apparent similarities in form, mineralogy, paragenesis, sulfur isotopic compositions, and fluid inclusion characteristics between mineralization in the main UMV district and the outlying minor occurrences. The significance of the hypothesis is that studies of the minor sulfide occurrences might provide valuable data which can be used to constrain interpretations of the size, geometry, hydrological, and geochemical nature of the fluid systems which precipitated both the main UMV orebodies and the outlying occurrences (Garvin et al., 1987).

The present contribution uses stable isotope (S, C, O, and H) and fluid inclusion data to: 1) determine the geochemical conditions under which minor occurrences in Iowa, Illinois and Wisconsin formed, 2) identify possible sources of ore forming constitiuents (i.e., fluids, sulfur and metals), and 3) use the above data, in conjunction with available geological, mineralogical, and paragenetic information, to evaluate whether the outlying occurrences are cogenetic with the main UMV district ores as has been proposed previously.

GEOLOGIC SETTING

The UMV district is bounded to the north and northeast by the Wisconsin Dome and Wisconsin Arch, respectively. Three outlying intracratonic basins flank the main UMV district and the outlying mineralized zone also; namely: the Michigan Basin to the east; the Illinois Basin to the south; and the Forest City Basin to the southwest. A fourth basin, the defined East-Central Iowa Basin (Bunker, 1981; Ludvigson et al., 1983; Bunker et al., 1985), coincides geographically with the main UMV district (Fig. 2). Several major fault and fold systems are present within and surrounding the main UMV district. The folds have been important as loci of fractures and faults which are mineralized or have acted as fluid pathways

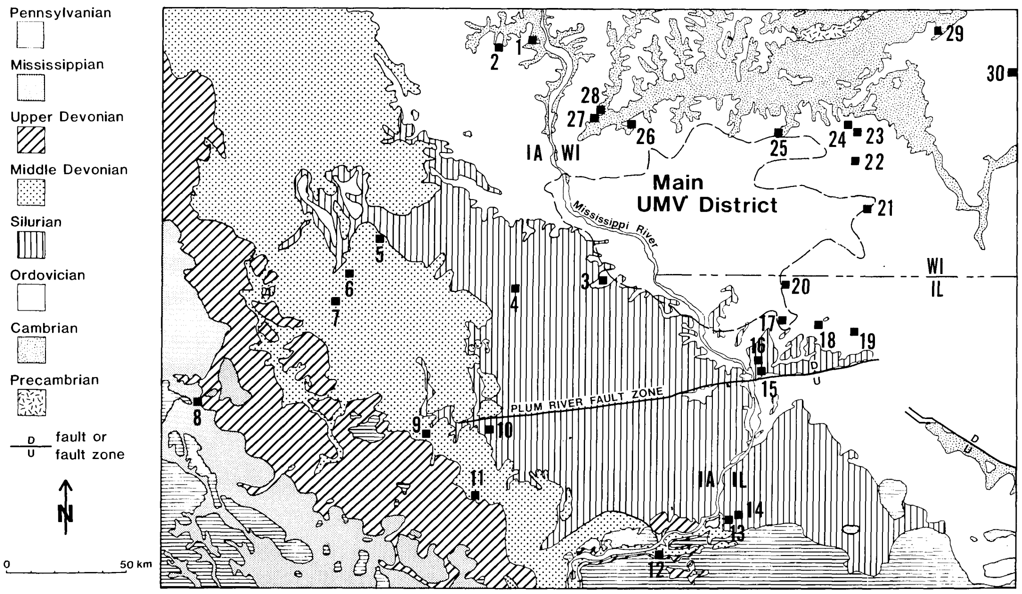


Figure 1. Generalized bedrock geologic map of part of the Upper Mississippi Valley with the location of outlying Upper Mississippi Valley occurrences (compiled after Bean, 1949; Hershey, 1969; Heyl and West, 1982; Willman et al., 1967). The numbered sulfide occurrences correspond to: (1) Lansing lead mines; (2) Mineral Creek mines; (3) Twin Springs Quarry; (4) G. Huber farm; (5) Fairbank Quarry; (6) Pint's Quarry; (7) Waterloo South Quarry; (8) Ferguson Quarry; (9) Four County Quarry; (10) Martin-Marietta Cedar Rapids Quarry; (11) Conklin Quarry; (12) Collinson Brothers Stone Quarry; (13) Midway Quarry; (14) NE Moline Quarry; (15) S. Remry farm; (16) Buckwalter farm; (17) Morseville digs; (18) Staderman gold prospect; (19) Yellow Creek lead mine; (20) Warren lead mines; (21) Exeter digs; (22) Hammersley Construction Quarry; (23) Camel Hill Quarry; (24) Speedway roadcut (County Road M) near Pine Bluff; (25) Demby-Weist mines; (26) Woodman lead mine; (27) Little Kickapoo Indian Caverns lead mines; (28) Plum Creek copper mine; (29) T. Rudd farm; (30) State Highway Quarry.

during mineralizing events. Heyl and West (1982) and Ludvigson et al. (1983) suggested that ore fluids which formed main UMV district mineralization followed the Plum River Fault Zone (PRFZ) and generated several minor outlying UMV occurrences (Figs. 1, 2).

The stratigraphic relationships of mineralization in the main UMV district have been described by Heyl et al. (1959) and Heyl et al. (1968b). The major orebodies occur predominantly in limestones and dolomites of the Middle Ordovician Platteville Formation and Galena Group (Heyl, 1968b). Silurian age rocks are mineralized only in the extreme southwest parts of the district, in Iowa.

Sulfide mineralization in the region surrounding the main UMV district occurs in Upper Cambrian through Mississippian strata in southeast Iowa (Fig. 3). Some sulfide mineralization is found in Pennsylvanian karst fills in rocks of Devonian age in parts of eastern Iowa and northwestern Illinois. The largest of the outlying occurrences are found in late Cambrian rocks and dolomites of the Prairie du Chien Group in northeastern Iowa (Lansing and Mineral Creek mines; Garvin, 1982) and along the northern fringe of the main UMV district in Wisconsin. To the south and west of the main UMV district, sulfides are exposed in progressively younger rocks. Pyrite and marcasite appear to be present in nearly every exposed unit of the Paleozoic section (Garvin et al., 1987); however, sphalerite is less widespread. Galena appears to be more common with proximity to the main UMV district.

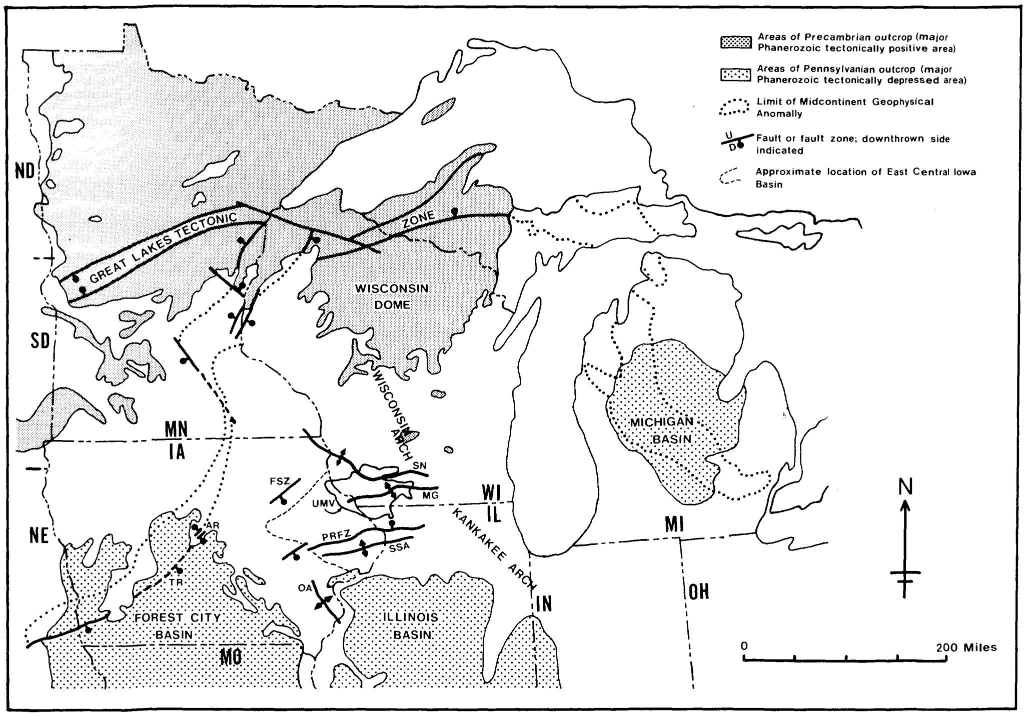


Figure 2. Generalized map of the major tectonic features of the north-central midcontinent region (modified after Bunker et al., 1985). Abbreviations used are: AR = Ames-Roland Horst; FSZ = Fayette Structural Zone; MG = Meekers Grove Anticline; OA = Oquawka Anticline; PRFZ = Plum River Fault Zone; SN = Synmagil Anticlinal Belt; SSA = Savanna-Sabula Anticlinal System; TR = Thurman-Redfield Structural Zone; UMV = Upper Mississippi Valley main ore district.

NATURE OF MINERALIZATION

The form of the mineralization in the outlying occurrences is controlled by the lithological, mineralogical, and structural characteristics of the host rocks. The degree of fracturing, jointing, and faulting and the porosity appear to be the most important controls on the morphology of the mineral occurrences. Carbonates are more often mineralized than shales or sandstones. These relationships appear to hold for both the main UMV district and the outlying minor occurrences.

Heyl (1983) noted that even though there are many forms of epigenetic mineralization, pitch-and-flat and gash-vein deposits are dominant in the main UMV district. The outlying occurrences have variable forms also and include: 1) vertical vein deposits (e.g., Lansing lead mine); 2) gash-vein deposits (e.g., Demby Weist mines); 3) disseminated breccia deposits (e.g., Mineral Creek mines); 4) vug linings (e.g., Pint's, Waterloo South, and Collinson Brothers Milan quarries); and 5) paleokarst replacement bodies (e.g., Plum Creek copper mines).

The mineralogy of the outlying minor occurrences is relatively simple when compared to the main UMV district (see Heyl et al., 1959; McLimans, 1977). Principal sulfides in the outlying occurrences are pyrite, marcasite, sphalerite and galena. The last mineral occurs as indiviual octahedra and cubes, or as crystalline masses. Pyrite occurs as cubes, cubo-octahedra, pyritohedra, and as colloform growths, whereas marcasite occurs almost exclusively as blades which are twinned and intergrown. Sphalerite generally forms as single euhedral to subhedral crystals or as small masses of intergrown crystals; colloform sphalerite is present at the Mineral Creek mines (Garvin, 1982). Although generally pale yellow to dark reddish-orange in color, sphalerite from several outlying occurrences (e.g., Collinson, Ferguson, Twin Springs and Camel

Hill quarries) is unusual in that individual crystals often have blue-green to purple colored cores. Strong color banding, like that present in sphalerite in the main UMV district (McLimans, 1977), is not recognized in the outlying sphalerites. Barite is found at the Ferguson, Pint's, Waterloo South, and Conklin quarries in Iowa and occurs as blades, which are commonly growth zoned, and as euhedral crystals of various habits. Minor amounts of millerite have been recognized at the Conklin Quarry (Garvin, 1984a). Rare chalcopyrite also occurs at the Collinson Brothers Quarry. Silver minerals, which commonly occur as blebs within galena in the main UMV district, have not been observed in the outlying occurrences.

The most common gangue mineral in the outlying occurrences is calcite. It commonly forms rhombohedra and scalenohedra and is usually white to honey-brown, and, in places, may be pink or green in color. Fluorite is present at Pint's and Waterloo South quarries. In addition, fluorite cubes have been found in the insoluble residues of carbonates from the Maynes Creek Member of the Hampton Formation (Ferguson Quarry) and the Eagle Center Member of the Little Cedar Formation (Glory Quarry near Cedar Falls; personal communciation J. Lemish, 1988). Quartz is present as an early constituent in most of the outlying mineral occurrences; however, it is usually not abundant.

PARAGENESIS

Garvin et al. (1987) classified the outlying mineral occurrences into three types based on paragenetic sequence: 1) occurrences in which sulfides are generally early and calcite is late; 2) occurrences in which calcite is early and sulfides are generally late; and 3) occurrences in which calcite is absent and, where more than one sulfide is present, iron sulfide is early and other sulfides are late. As Garvin et al. (1987) suggested, occurrences of the first type are most similar to the main UMV deposits, with regard to paragenesis. Furthermore, they suggested that the third paragenetic type is similar to the main UMV district deposits (with the exception of the lack of calcite) in that iron sulfide mineralization preceded deposition of galena and sphalerite.

Despite this classification, paragenetic sequences are known for only a relatively small number of the outlying occurrences. This is, in part, due to the sparse nature of sulfide mineralization in many localities and to the lack of obvious textural relationships between the minerals present. Paragenetic sequences for three localities that were not studied by Garvin et al. (1987) are compared in Figure 4 with the paragenetic sequence obtained by McLimans (1977) for the main UMV district.

FLUID INCLUSION STUDY

Fluid inclusion studies were undertaken by Coveney and Goebel (1983) and Coveney et al. (1987) on the Martin-Marietta Cedar Rapids Quarry and Conklin (River Products) Quarry as part of a regional fluid inclusion analysis of minor occurrences in the Midwest, predominantly in Kansas and Missouri. In contrast to the outlying UMV occurrences, several fluid inclusion studies have been undertaken on the main UMV district. Homogenization temperatures were collected by Newhouse (1933), Bailey and Cameron (1951), and McLimans (1977) on sphalerite and by these same workers in addition to Ericksen (1965) on calcite. Freezing point depressions of fluids in sphalerite and calcite were obtained by McLimans (1977) and Hall and Friedman (1963), respectively. In the present study homogenization and freezing point data was obtained from primary and secondary inclusions in sphalerite and calcite from twelve outlying occurrences (Fig. 5). These data were previously reported in Kutz and Spry (1987a,b).

Fluid inclusions contained in sphalerite and calcite in the outlying occurrences are predominantly two-phase liquid-vapor inclusions. A number of single-phase liquid inclusions are also present; however, these are generally secondary inclusions or are primary inclusions which show evidence of leaking. Determination of the primary and secondary nature of fluid inclusions were made according to the criteria of Roedder (1984). Since most samples did not show evidence of the direction of growth or were composed of intergrown crystal masses, any planar groups of inclusions were assumed to be of secondary origin. Inclusions which were

SYSTEM	SERIES	FORMATION	DOMINANT LITHOLOGY	KNOWN STRATIGRAPHIC DISTRIBUTION OF SULFIDE MINERAL OCCURRENCES
MISSISSIPPIAN	OSAGE	Warsaw Formation	Sh.	Ft. Madison area
		Keokuk Formation	Ls.	
		Burlington Formation	Ls.	
	KINDERHOOK	Gilmore City Formation	Ls.	
		Hampton Formation	Dol.	Ferguson Quarry
		English River Formation	Sh.	
		Maple Hill Formation	Sh.	
DEVONIAN	UPPER	Cedar Valley Formation	Ls.	Pints, Fairbank, Waterloo South, Four County quarries
	MIDDLE	Wapsipinicon Formation	Ls.	Collinson Brothers, Conklin quarries
		Otis Formation	Ls.	
SILURIAN	LUDLOVIAN	Bertram Formation	Dol.	
	WENLOCKIAN	Gower Formation	Dol.	Midway Quarry; Martin-Marietta Quarry
		Scotch Grove Formation	Dol.	G. Huber farm (?)
		Hopkinton Formation	Dol.	
	LLANDOVERIAN	Blanding Formation	Dol.	
		Tete des Morts Formation	Dol.	
		Mosalem Formation	Dol.	
ORDOVICIAN	UPPER	Maquoketa Formation	Sh.	Staderman gold prospect
		Dubuque Formation	Dol.	Mt. Carroll Mines, Yellow Creek
		Wise Lake Formation	Dol.	Mine, Morseville digs, S. Remry farm, Buckwalter farm
	MIDDLE	Dunleith Formation	Dol.	main district UMV ore deposits
		Decorah Formation	Ls.	Camel Hill Quarry, roadcut (Co.
		Platteville Formation	Ls.	Rd. M), St. Highway Quarry; Hammersly Const. Quarry
		Glenwood Shale	Sh.	
		St. Peter Sandstone	Ss.	
	LOWER	Shakopee Formation	Dol.	Mineral Creek, Woodman, Plum
		Oneota Formation	Dol.	Creek mines; Little Kickapoo Caverns digs
CAMBRIAN	UPPER	Jordan Sandstone	Ss.	T. Rudd farm (Doylestown Pb digs)
		St. Lawrence Formation	Dol.	Demby-Weist, Lansing mines

Figure 3. Generalized stratigraphic nomenclature and lithologies for Paleozoic rocks exposed in the study area (modified after Garvin et al., 1987; see Bunker this guidebook for new Devonian terminology).

interpreted to be of primary origin were usually individual inclusions separated from all other inclusions or were in non-planar clusters. Liquid-to-vapor ratios (L/V) of primary inclusions (and secondary inclusions where two phases are present) are extremely consistent on both local and regional scales. At room temperature, the vapor phase usually occupies five to ten percent of the inclusion volume for sphalerite and approximately five percent for inclusions in calcite. Although quartz and fluorite are present at several localities, these minerals did not yield inclusions suitable for analysis.

Heating and freezing measurements were determined using a Chaixmeca heating-freezing stage. This stage has been modified to include

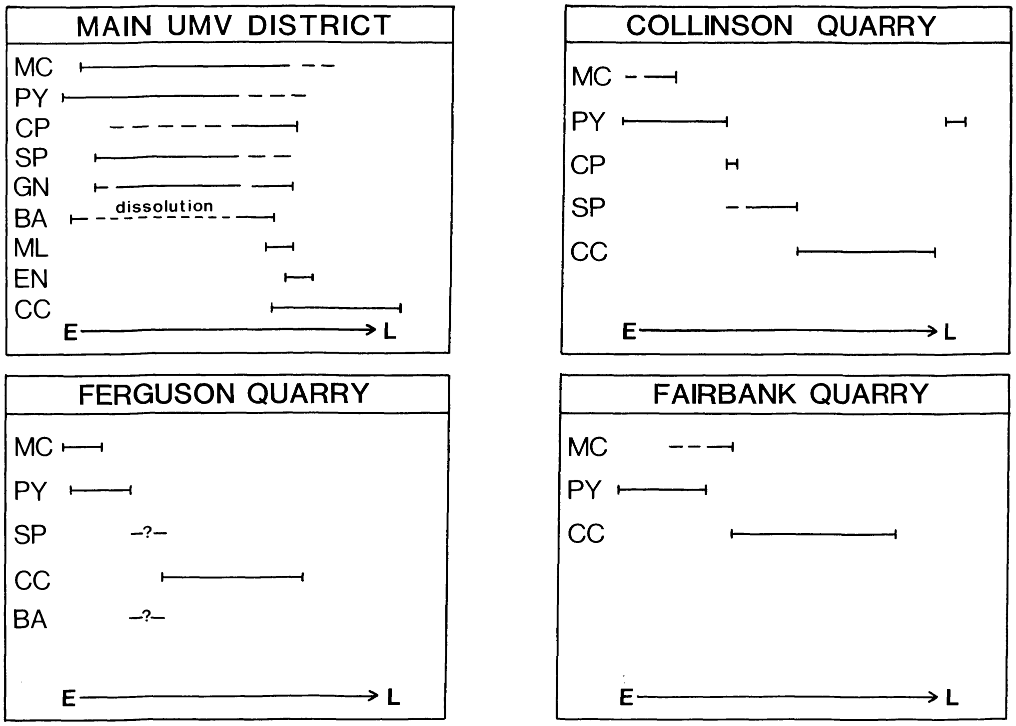


Figure 4. Paragenetic sequences for a) the main UMV district (McLimans,1977), b) Collinson Brothers Quarry, c) Ferguson Quarry, and d) Fairbank Quarry.

extra external insulation and a smaller sample chamber using the design of Cunningham and Carollo (1980). Two or three heating and freezing experiments were performed on every measured inclusion. The heating and freezing data are given in Kutz (1987) and shown in Figure 5. The precision of the homogenization temperature was $\pm 0.2^{o}$C or better and replicate measurements showed a reproducibility within $\pm 0.5^{o}$C.

Temperatures of homogenization (T_h) for primary inclusions in samples of sphalerite obtained from the outlying occurrences range from 57.4^{o} to 115.8^{o}C. In general, these temperatures are cooler than those recorded for the main UMV district. Newhouse (1933) and Bailey and Cameron (1951) reported ranges of 80^{o} to 105^{o}C and 75^{o} to 121^{o}C, respectively. Note, however, that the inclusions measured by these workers were in sphalerite from the later stages of sulfide mineralization. T_h obtained on secondary inclusions from sphalerite in outlying occurrences is 48.8^{o} to 75.1^{o}C and tends to be lower than T_h for primary inclusions from the same samples (Fig. 5). Some overlap of T_h does occur and, in some cases, secondary inclusions yielded higher T_h than primary inclusions from the same samples.

Homogenization temperatures for primary inclusions in calcite associated with outlying sulfide occurrences range from 37.8^{o} to 75.4^{o}C (Fig. 5). These temperatures are similar to the 50^{o} to 78^{o}C range reported by Bailey and Cameron (1951) for inclusions from middle calcite stage mineralization in the main UMV district and to the 46^{o} to 74^{o}C range reported by Erickson (1965) for calcites of various ages from the main UMV district.

McLimans (1977) and Coveney et al. (1987) have suggested a pressure correction of approximately 10^{o} to 12^{o}C should be added to T_h data for the main UMV orebodies and to minor UMV-type occurrences in the Midwest, respectively. Since the T_h data obtained in this study overlap with the data obtained by McLimans (1977) and Coveney et al. (1987) no pressure correction has been added.

Freezing experiments performed on primary and secondary inclusions in sphalerite from the outlying sulfide occurrences showed initial melting temperatures (T_e) lower than -38°C. Since the majority of the T_e measurements lie between -60° and -49°C, it is suggested that appreciable quantities of $CaCl_2$, and possibly $MgCl_2$, in addition to NaCl are present in the fluid inclusions. Coveney et al. (1987) suggested that T_e should be considered higher than the actual values of T_e because of difficulties involved with the onset of initial melting.

Final melting temperature (T_m) determinations of ice for primary inclusions in sphalerite from the outlying occurrences show that fluid salinity ranges from 15.6 to 23.8 equivalent weight % NaCl. This range overlaps with that reported by McLimans (1977) for sphalerite from the main UMV district (19.6 to >23 equivalent weight % NaCl). In contrast to the main UMV district, no daughter crystals have been recognized in outlying sphalerites. Secondary inclusions in outlying sphalerites tended to yield higher final melting temperatures than primary inclusions although, in several examples, primary and secondary inclusions in the same samples yielded indistinguishable values of T_m (Fig. 5).

Preliminary T_m measurements of inclusions in calcite yield values (-14.2° to -11.6°C) which are significantly different from that reported by Hall and Friedman (1963) for calcite from the main UMV district (-3.1° to -2.9°C). Corresponding salinity values are 15.6 to 18.1 and 4.8 to 5.1 equivalent weight % NaCl, respectively (Fig. 5). Significantly more salinity data on calcite is currently being obtained.

STABLE ISOTOPE STUDY

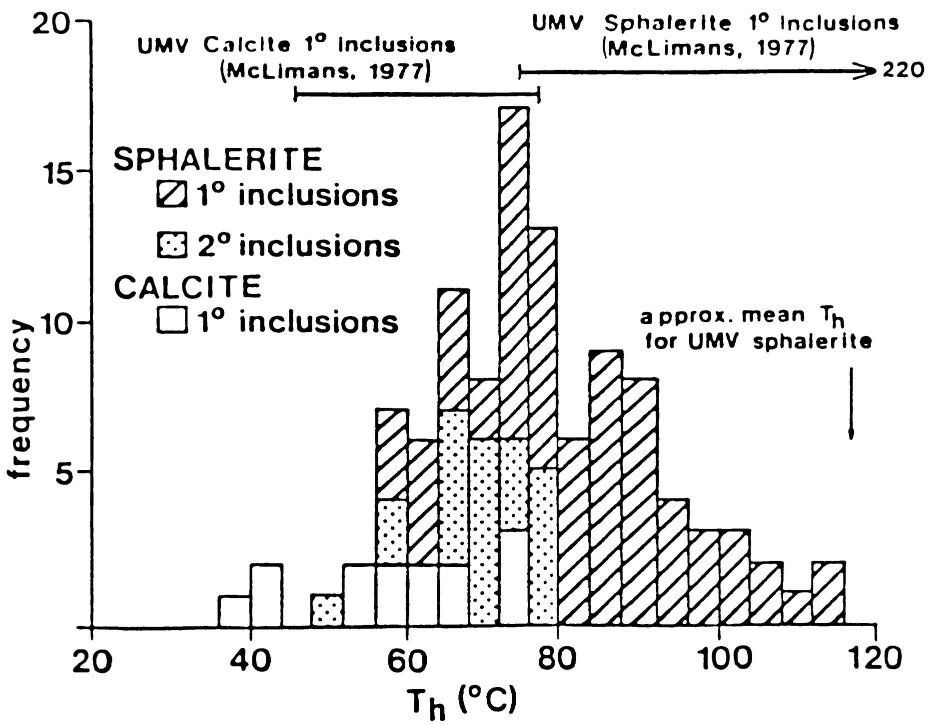
Stable isotope (S, C, O, and H) compositions were determined on samples of outlying sulfides, sulfates, carbonates (calcite and host limestones), carbonaceous shales, and fluids extracted from inclusions in sphalerite, calcite, and fluorite. The data collected from this study were initially reported in Kutz and Spry (1987a,b) and will be compared with the results obtained from three stable isotope studies (McLimans, 1977; Garvin et al., 1987; Garvin and Ludvigson, 1988) on the main UMV deposits and outlying occurrences. McLimans (1977) collected S, C, O and H isotope from the main UMV district and Garvin et al. (1987) undertook a reconnaissance sulfur isotope study of outlying UMV occurrences. Garvin et al. (1987) also reported data for three C and O isotope pairs from carbonates in the Martin Marietta Quarry. Garvin and Ludvigson (1988) have recently completed a detailed S, C, and O isotope study of karst-fill mudrocks at the Conklin Quarry.

Values of $\delta^{13}C$, $\delta^{18}O$, and δD were obtained in the laboratory of Dr. E. M. Ripley at Indiana University using a Finnigan Delta-E mass spectrometer with a 9 cm deflection radius whereas values of $\delta^{34}S$ were measured at the same facility on a 6" 60° sector Nuclide stable isotope ratio mass spectrometer. Analytical precision for values of $\delta^{13}C$ and $\delta^{18}O$ are generally better than ±0.05 per mil, for $\delta^{34}S$ between 0.05 and 0.10 per mil, and for δD ±1 per mil. Details of the analytical methods and individual isotope analyses are given in Kutz (1987).

Sulfur isotopes

Seventy-one sulfur isotope analyses were obtained on sulfides and sulfates from 20 outlying mineral occurrencs. These data are presented in Figure 6 along with those obtained by McLimans (1977) and Garvin et al. (1987). Data obtained in the present study show a range of sulfur isotope values for sulfides (-22.3 to 33.1 per mil) that is almost identical to the range (-18.9 to 35.1 per mil) reported by Garvin et al. (1987). This range contrasts to that (5.4 to 29.9 per mil) obtained by McLimans (1977) for sulfides from the main UMV district.

Although samples of pyrite, from this study, yielded sulfur isotopic compositions which span the entire range of $\delta^{34}S$ values (-22.3 to 33.1 per mil), sulfur isotopic compositions of sphalerite and galena are more restricted (-14.2 to 16.4 per mil and -8.8 to -4.0 per mil, respectively). The large variation in sulfur isotopic compositions of iron sulfides occurs within individual deposits at Collinson Brothers Quarry (-21.0 to 28.5 per mil), Ferguson Quarry (2.8 to 33.1 per mil), and Fairbank Quarry (-19.6 to -0.3 per mil), and was reported previously by Garvin et al. (1987) for the Martin-Marietta Cedar Rapids Quarry (1.0 to



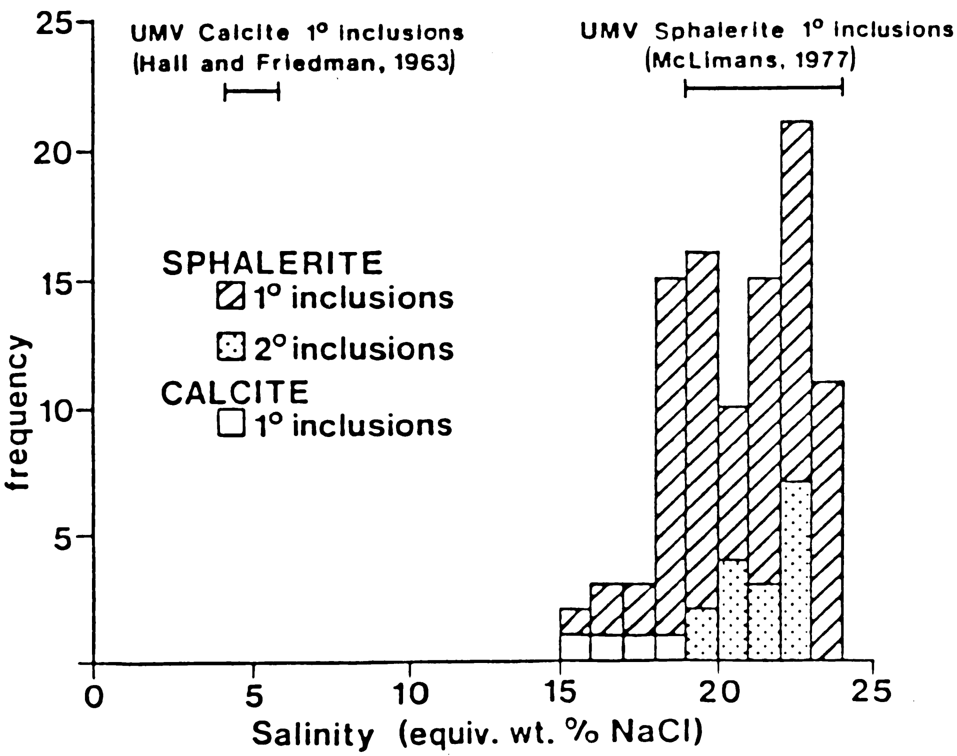


Figure 5. Histograms for data from fluid inclusion heating (top) and freezing (bottom) experiments.

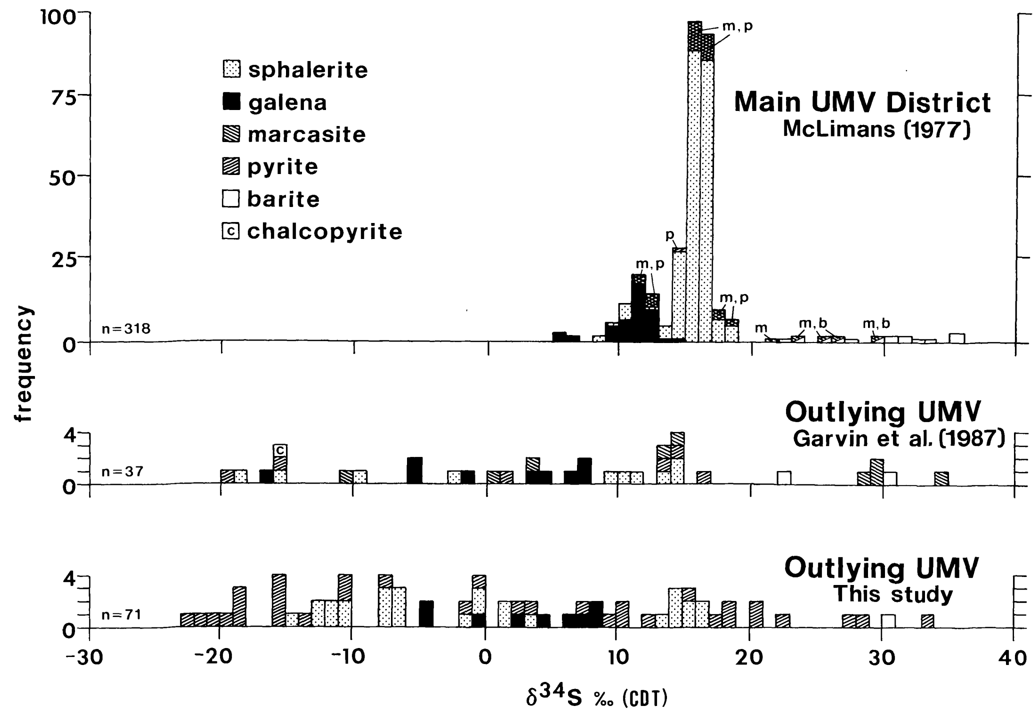


Figure 6. Composite sulfur isotope histograms for the main UMV district (McLimans, 1977), and outlying occurrences (Garvin et al., 1987, and this study); modified after Ludvigson (1988).

35.1 per mil) and the Mt. Carroll mines (-20.0 to 1.5 per mil). The lightest isotope value (-22.3 per mil) from this study was obtained from pyrite hosted by carbonaceous material in a karst fill at the Four County Quarry.

One sulfur isotopic value (30.1 per mil) of barite was determined and is consistent with values obtained for barite (22.4 and 30.7 per mil) from Conklin and Pint's quarries, respectively by Garvin et al. (1987) and the main UMV occurrences (22.3 to 35.9 per mil) by McLimans (1977).

Carbon and oxygen isotopes

Carbon and oxygen isotopic compositions were obtained on calcite associated with mineralization from nine outlying occurrences (Fig. 7). Values of $\delta^{18}O$ range from 21.3 to 28.5 per mil and are isotopically heavier than values (16.8 to 22.8 per mil) obtained from calcite associated with the main UMV district (McLimans, 1977). Only two samples analyzed in this study overlap with those determined for the main UMV district (Fig. 6). In contrast to the oxygen isotopic data, carbon isotopic compositions (-8.8 to 3.3. per mil) from the outlying occurrences overlap with those (-12.6 to -2.1 per mil) obtained by McLimans (1977) from the main UMV district. McLimans suggested that the more negative carbon isotope values may have been influenced by an influx of groundwater or by oxidation of methane or other organic compounds. Although organics and methane have not been detected in fluid inclusions from any outlying occurrences in Iowa, Illinois and Wisconsin (c.f. with the Jumbo mine, Kansas where petroleum was recognized; Blasch and Coveney, 1988), carbonaceous and organic-rich lamina are present in mineralized horizons at many of the outlying occurrences (e.g., Pint's, Waterloo South, Collinson Brothers, Four County, Ferguson, and Conklin quarries). Two carbon isotopic analyses of organic shale samples from the Ferguson and Four County quarries yielded very light $\delta^{13}C$ values of -25.1 and -26.0 per mil, respectively.

In order to determine whether the host rocks influenced the isotopic compositions of calcite associated with the outlying occurrences, carbon

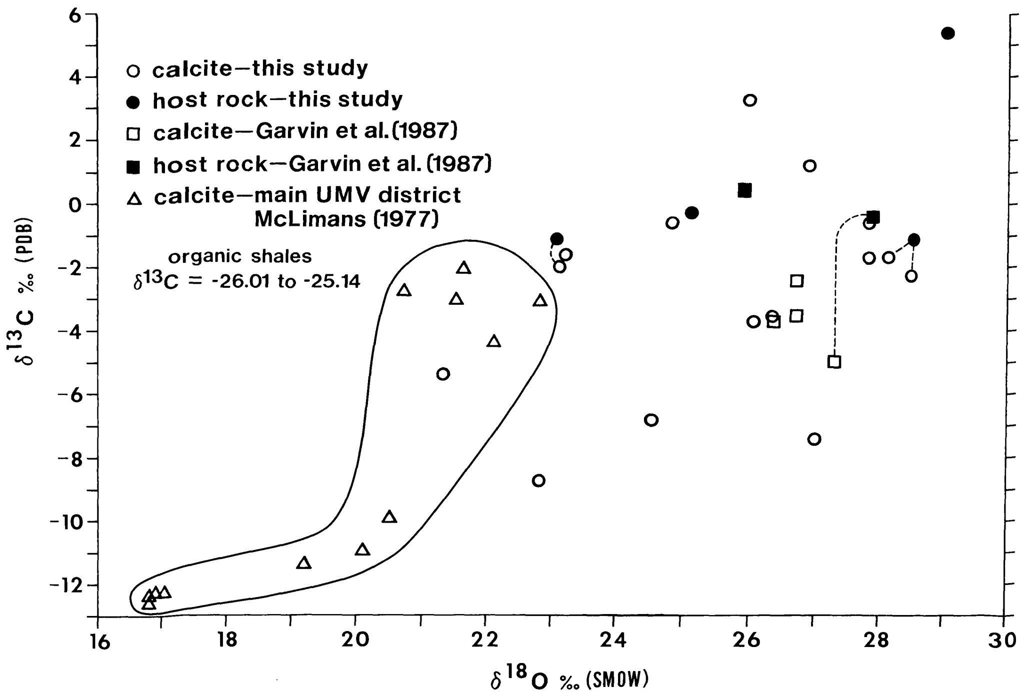


Figure 7. Plot of $\delta\,^{13}C$ versus $\delta\,^{18}O$ values for calcite and host rocks from the outlying sulfide occurrences. Carbon and oxygen isotope compositions of calcite from the main district are plotted in the enclosed area. Also shown are $\delta\,^{13}C$ values for organic shales from two outlying occurrences (see text for details); modified after Ludvigson (1988).

and oxygen isotopic compositions of some coexisting host rocks were also determined (Fig. 7). Calcite and coexisting dolomite from these localities display almost identical values of $\delta\,^{13}C$; whereas values of $\delta\,^{18}O$ for calcite are slightly lighter than those in coexisting dolomite. This relationship is consistent with data in Garvin et al. (1987). Carbon and oxygen isotope data from this study suggests that calcite may have formed by dissolution of the host carbonates during late diagenesis, and that isotopically light carbon may have been derived by local cycling from organic-rich strata.

Oxygen and hydrogen isotopes from fluid inclusions

Oxygen and hydrogen isotopic analyses were performed on fluids extracted from thirteen samples of calcite, sphalerite and fluorite from the Collinson Brothers, Fairbank, Ferguson, and Pint's quarries (Fig. 8). Fluids extracted from outlying calcites display considerable scatter in values of $\delta\,^{18}O$ and δD (-2.3 to 4.3 per mil and -83.2 and -18 per mil, respectively) and overlap with the one sample analyzed by McLimans (1977) from the main UMV district. Note that fluids extracted from two samples of sphalerite yielded values which are slightly lighter in δD than those obtained from sphalerite in the main UMV district (Fig. 8) The single analysis of fluorite from Pint's Quarry yielded the most isotopically light oxygen and hydrogen values ($\delta\,^{18}O$ = -8.5 per mil, and δD = -94.4 per mil).

DISCUSSION

Prior to the study by Garvin et al. (1987), minimal geochemical data were available to support whether the outlying occurrences were cogenetic with the main UMV district deposits. Previous investigations (e.g., Heyl and West, 1982; Garvin, 1984; Garvin et al., 1987), have made comparisons with geochemical data in McLimans' (1977) detailed study of the main UMV district.

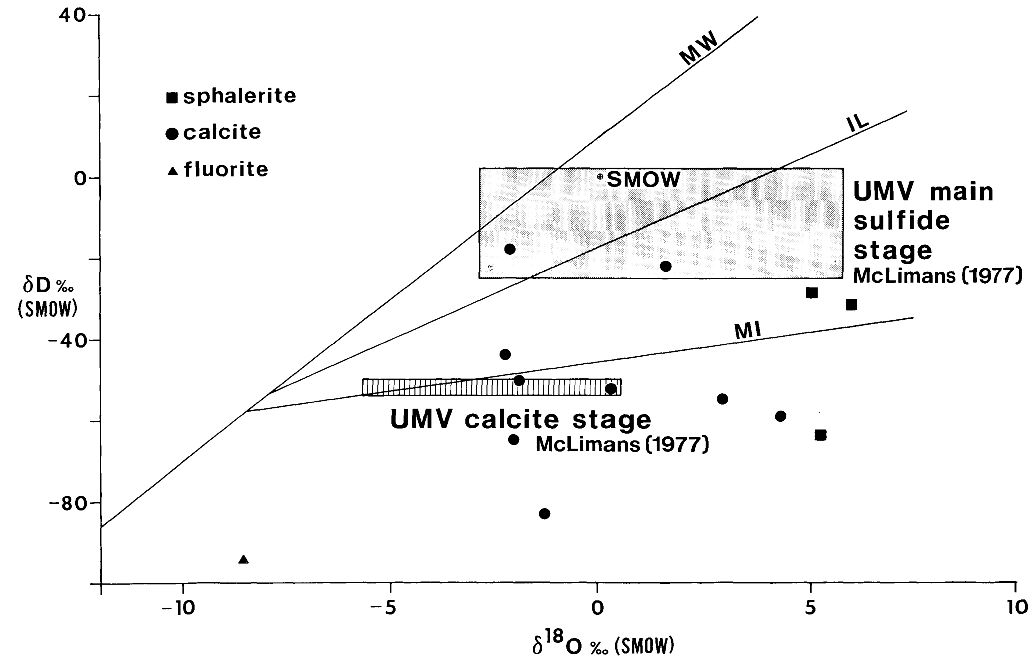


Figure 8. Plot of $\delta^{18}O$ versus δD for fluids extracted from inclusions in calcite, sphalerite, and fluorite from outlying mineral occurrences. Shaded boxes represent the range of δD and $\delta^{18}O$ values from the main UMV district sphalerite and calcite. Abbreviations used are: MW = meteoric water line; MI = Michigan Basin; IL = Illinois Basin; SMOW = Standard Mean Ocean Water.

It should be noted, however, that McLimans (1977) concentrated on the large pitch-and-flat deposits, and that geochemical data are not available on gash-vein, stockwork, bedded replacement, solution-collapse breccias, fissure veins and lodes, and ore-lined giant vugs or small sulfide encrusted caves that are present in the main UMV district. In view of these limitations, a discussion of the genetic relationship between the outlying minor base metal occurrences and the main UMV district should be restricted to the pitch-and-flat deposits.

Fluid inclusion data suggest that there are significant differences between fluids that formed the outlying UMV occurrences and the orebodies in the main UMV district. The mean T_h of fluids in sphalerite from the outlying occurrences is approximately 80°C and is significantly lower than the mean T_h of 120°C for sphalerite from the main UMV district. The highest temperature from an outlying occurrence is only 116°C. T_h for calcites from the outlying occurrences and the main UMV district overlap but preliminary salinity determinations on fluids in calcite from the outlying district indicate that they were considerably more saline. The absence of organics, CO_2, and daughter crystals in inclusions in the outlying occurrences and their presence in some inclusions in minerals from the main UMV district suggest further differences in fluid composition.

Garvin et al. (1987) classified the outlying minor sulfide occurrences into three groups based on sulfur isotopic compositions: 1) those with weakly to strongly positive values of $\delta^{34}S$ (1 to 30 per mil), 2) those with strongly negative values of $\delta^{34}S$ (-20 to -10 per mil), and 3) those with values of $\delta^{34}S$ which cluster around 0 per mil. The general contention of their study was that group 1 occurrences, which have parageneses most like the main UMV district, are cogenetic with main UMV sulfide mineralization. Their group 2 occurrences, which have parageneses most dissimilar to main UMV district orebodies, are diagenetic in origin with sulfur supplied through leaching of pre-existing sedimentary iron sulfides. The third group consisted solely of the Mt. Carroll lead mines and, although the paragenesis is similar to the main UMV district, Garvin et al. (1987) suggested that this occurrence formed from a mineralizing event that was distinct from those events that formed other outlying occurrences.

Despite this classification scheme, several factors suggest that three distinct types of outlying UMV occurrences cannot be distinguished with any degree of certainty. First, the values of $\delta^{34}S$ determined by Garvin et al. (1987) do not fall into three statistically distinct groups as they imply. The data collected in this study, in conjunction with their data suggest almost a continuum of values from -20 to 30 per mil (particularly for iron sulfides). Second, paragenetic data are limited and it is difficult to classify many deposits into the paragenetic sequences proposed by Garvin et al. (1987). Third, although paragenetic data are available for some deposits they do not necessarily fall into the isotopic classification of Garvin et al. (1987). With the exception of a minor amount of late pyrite at the Collinson Brothers Quarry, this occurrence, in addition to the Ferguson Quarry, fit into the paragenetic group 1 of Garvin et al. (1987), yet the sulfur isotopic characteristics of both occurrences are clearly different from the main UMV district ores and the sulfur isotope group 1 of Garvin et al. (see Kutz, 1987, Fig. 7). Furthermore, even though paragenetic data for the Lansing lead mines suggest classification into Garvin et al.'s (1987) group 1, sulfur isotopic values of -0.8 and -4.3 per mil from this study combined with the values of -16.9 and 5.1 per mil from Garvin et al.'s (1987) study, suggest a similarity in both isotopic and paragenetic data with the Mt. Carroll mines.

A comparison of sulfur isotopic compositions of the outlying occurrences (this study and Garvin et al., 1987) with those obtained by McLimans (1977) from the main UMV district reveal several generalities: 1) values of $\delta^{34}S$ for barite are strongly positive regardless of location or position in the paragenetic sequence; 2) outlying minor occurrences immediately adjacent to the main UMV district often have sulfur isotopic values within the range determined by McLimans (1977) for the main UMV district (e.g., George Huber farm, Woodman lead mine, and Little Kickapoo Indian Caverns lead mines), whereas deposits more than 60 km from the main district commonly have isotopic compositions lighter than those reported for the main UMV district (e.g., Waterloo South Quarry, Lansing lead mines, and Fairbank Quarry). Exceptions to this generalization are two quarries near Moline, Illinois, approximately 80 km from the main UMV district (Midway and Northeast Moline), which have sulfur isotopic ratios for sphalerite which are similar to those reported by McLimans (1977) for the main UMV district, and Speedway roadcut and Twin Springs Quarry, adjacent to the main UMV district, which have sulfur isotopic values for sphalerite that are lighter than those reported by McLimans (1977) for the main UMV district; and 3) as Garvin et al. (1987) suggested, more than one source of sulfur is needed to explain the observed sulfur isotopic values.

Likely sources of sulfur for the outlying occurrences are seawater sulfate and diagenetic (sedimentary) sulfide. Seawater sulfate isotopic compositions have varied from 17 to 30 per mil for the interval late Cambrian to early Mississippian (Claypool et al., 1979). In general, values of $\delta^{34}S$ for barite and some sulfides from the outlying occurrences lie within this range (Fig. 6). In addition, marine evaporites were periodically deposited during Paleozoic sedimentation and have been partially or completely dissolved. The Middle Devonian Wapsipinicon and Cedar Valley Groups (see Bunker, this guidebook) have been cited as possible sources for the outlying mineral occurrences (Garvin et al., 1987). Hansen (1983) reported values of $\delta^{34}S$ for gypsum from the Wapsipinicon Group which range from 18.7 to 21.3 per mil. These values are close to the lowest value of $\delta^{34}S$ for both outlying and main UMV district barites. More important, however, is the large number of sulfides from the outlying minor metal occurrences which have sulfur isotopic values close to the sulfate sulfur values reported by Hansen (1983). This observation supports the suggestion that evaporites may have supplied some sulfur to the outlying sulfide occurrences. The Middle Devonian evaporites may also have been a source of metals to these occurrences.

Pyrite (probably of diagenetic origin) is found in many of the carbonate strata of the Upper Mississippi Valley and in paleokarst fills within Devonian strata. Pyrite typically has values of $\delta^{34}S$ that are strongly variable and usually negative, and was produced by bacterial reduction in anoxic environments by the oxidation of organic carbon. The range of $\delta^{34}S$ values for sulfides within organic-rich rocks is -22.3 to -6.3 per mil. It is suggested that these sulfides may have produced isotopically light sulfur for late mineralizing events. The entire range of sulfur isotopic values observed for the outlying occurrences can be explained by a mixing model in which two sulfur-rich brines (an evaporite-derived sulfide-rich brine and a brine containing isotopically depleted sulfide sulfur) mixed at the site of deposition.

Viezer and Hoefs (1976) and Lohmann (1988) have shown that $\delta^{18}O$ of marine carbonates have varied through Phanerozoic time. Since $\delta^{18}O$ values of meteoric water are partly dependent upon the isotopic composition of the marine reservoir, the $\delta^{18}O$ values of meteoric water will also vary with age. There is an overlap in $\delta^{18}O$ and δD values of fluids in inclusions in sphalerite and calcite from the main UMV district and the outlying occurrences, however, data from the outlying occurrences are considerably more scattered (Fig. 8). If the main UMV district deposits and the outlying occurrences formed at the same time, differences in the $\delta^{18}O$ and δD values indicate a difference in fluid source or the effects of local rock-water interactions. However, if the main UMV district and the outlying occurrences did not form at the same time, differences in $\delta^{18}O$ values from the main UMV district and the outlying occurrences should be expected. Values of $\delta^{18}O$ from the outlying occurrences are significantly displaced from the meteoric water line and indicate that fluids which formed the outlying occurrences could not have been entirely meteoric in origin. Although few in number, these values are consistent with the hypothesis that fluids were derived primarily from a basinal brine. The data do not show which basin the fluids may have been derived from, however, they do indicate that it may not have been the same one which generated fluids for the main UMV mineralization. The large variation in δD suggests that a number of different fluids may have been responsible for the formation of the outlying occurrences.

Values of $\delta^{18}O$ for calcites from the outlying occurrences are generally enriched by one to seven per mil over the earliest main UMV district calcite stages. Enrichment of $\delta^{18}O$ is greater with respect to later stages of main UMV district calcite mineralization. Carbon isotopes tend to be lighter, overall, with respect to all stages of main UMV district calcite, however, most of the outlying calcite has values of $\delta^{13}C$ which are similar to values of $\delta^{13}C$ for McLimans (1977) stage I and II main UMV district calcite. Carbon isotopes and fluid inclusion T_h data indicate that calcite from the outlying minor base metal occurrences is similar to the two earliest stages of main UMV district calcite mineralization.

The range of values of $\delta^{13}C$ for the main UMV district calcite (-4.4 to -2.1 per mil for stages I and II; -12.6 to -10.0 per mil for stages III and IV) suggest that both marine and reduced carbon (from ground water or oxidation of organic compounds) may have been incorporated into the main UMV district calcite (McLimans, 1977). Carbon isotopic data for the outlying occurrences suggest that carbon was derived from the same type of sources as suggested by McLimans (1977) for the main UMV district mineralization. Most of the values of $\delta^{13}C$ for calcite from the outlying occurrences lie in the range of values typically exhibited by marine limestones (-4 to 4 per mil; Ohmoto and Rye, 1979). However, calcites with more negative values of $\delta^{13}C$ may have derived a light carbon component from reduced sources of carbon that are present in some of the host rocks. Because of the absence of igneous rocks in the vicinity of outlying occurrences, a contribution of reduced carbon from an igneous source is considered unlikely. In view of the similarity between the $\delta^{13}C$ and $\delta^{18}O$ of calcite and associated host rocks in some of the outlying occurrences, calcite was probably derived locally as a product of late diagenesis of the host rocks.

The main conclusion that can be derived from the current study is that there are two types of outlying UMV base metal occurrences (rather than three groups as suggested by Garvin et al. (1987)); those that occur near the margin of the main UMV district and which show mineralogical, paragenetic, stable isotope and some fluid inclusion characteristics that are similar to those associated with the main UMV deposits, and those that formed further away from the main UMV district and exhibit stable isotopic (particularly S), paragenetic and some fluid inclusion characteristics that are significantly different from those associated with the main

UMV district. The more distant deposits likely formed by more localized processes (e.g., diagenesis) which contrasts to the regional large scale processes that have been proposed by Ludvigson et al. (1983), Bethke (1986) and Leach and Rowan (1986) to have formed the large orebodies in the main UMV district.

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MINERALOGY, PARAGENSIS, AND STABLE ISOTOPIC COMPOSITIONS OF MINERAL DEPOSITS ASSOCIATED WITH LATE PALEOZOIC KARST FILLS IN JOHNSON COUNTY, IOWA

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INTRODUCTION

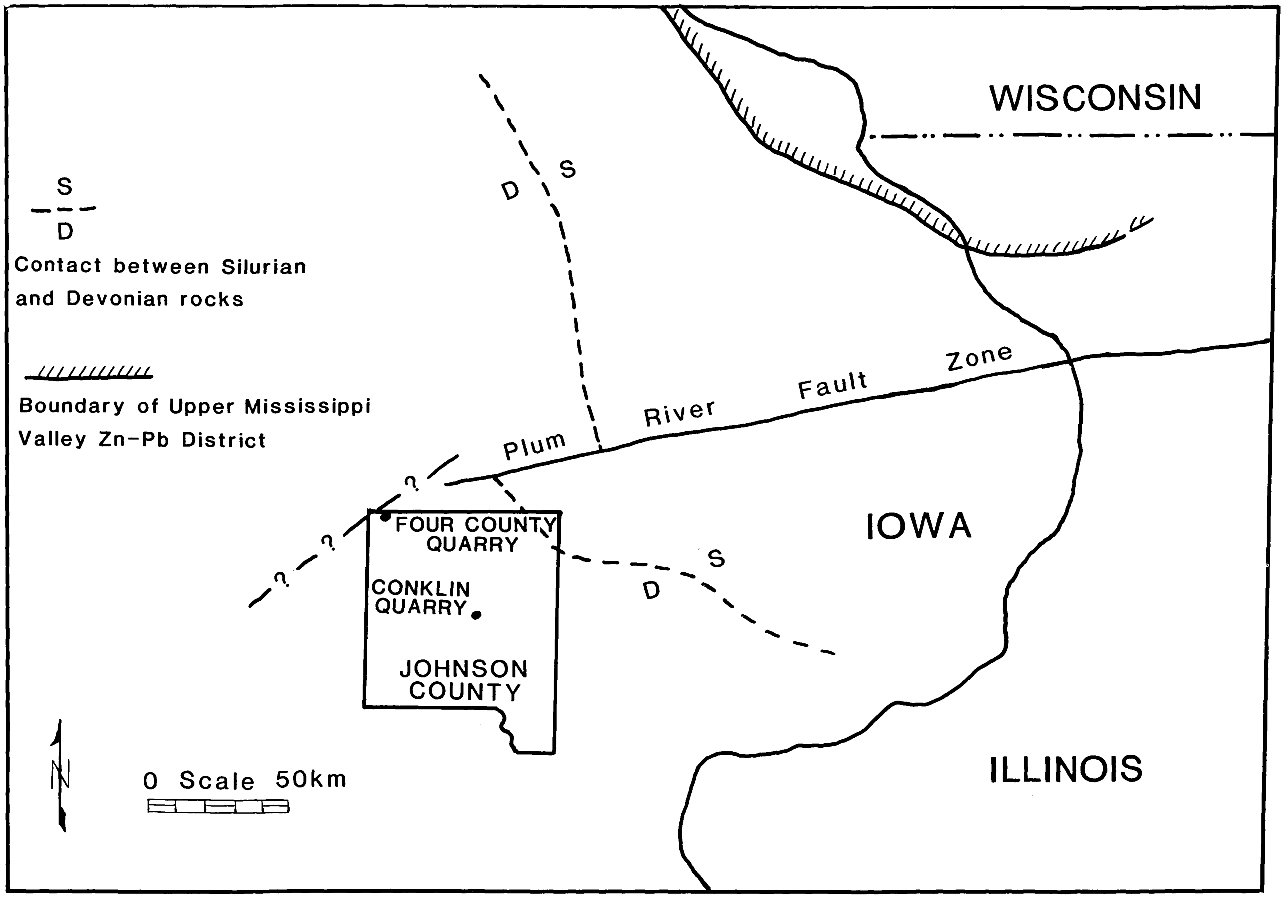
Minor sulfide-bearing mineral deposits are widespread in the upper Mississippi Valley region (Heyl, 1959; Heyl and West, 1982; Coveney and Goebel, 1983; Garvin et al, 1987; Spry and Kutz (this guidebook)). In eastern Iowa these occur principally as fracture fills in joints and minor faults, and as solution collapse breccia cements in carbonate rocks of Ordovician to Mississippian age (Ludvigson, 1976; Garvin, 1984a,b; Anderson and Garvin, 1984);(Fig. 1). Their location on the western fringe of the upper Mississippi Valley Zn-Pb District (UMV), and their structural and mineralogical similarities to UMV deposits has made them objects of interest in recent years (Heyl and West, 1982; Ludvigson et al, 1983). Reconnaissance studies of sulfur isotopic composition in some of these deposits (Garvin et al, 1985, 1987), demonstrate that some are not cogenetic with UMV deposits.

The existence of late Paleozoic karst fills in middle Paleozoic carbonate rocks in eastern Iowa has been known for many years (Udden, 1905, p. 406, Bunker et al, 1985). We have recently recognized that sulfide-bearing mineralization similar in many respects to other minor occurrences are found in paleokarsts that are also filled by organic-rich mudstones at several locations.

Detailed mineralogic, paragenetic, and stable isotopic analyses were carried out on two of these deposits. The deposit at the Conklin Quarry in central Johnson County (Fig. 1) was selected because there is a considerable body of available stratigraphic, petrographic, and mineralogical information upon which to base the analysis. In addition, there are fracture and vug-filling sulfide occurrences at Conklin which are not spatially related to karst fills, but bear mineralogic similarity to karst-associated minerals. The Four County Quarry, in the northwest corner of Johnson County (Fig. 1), was selected because sulfide mineralization in paleokarsts at the locality have mineralogic and paragenetic similarities to the mineralization at Conklin. A number of questions were raised: 1) how is karst associated mineralization related to disjunct sulfide mineralization at Conklin? 2) are the karst deposits at Conklin and Four County products of precipitation from the same fluid, i.e. did the mineralizing process operate on a regional or very localized scale? 3) what is the genetic relationship between these karst-associated deposits and main district UMV deposits? and 4) what might these deposits tell us about the nature and operation of ancient subsurface water flow systems in eastern Iowa?

TABLE 1. Isotopic Ratios For Sulfur in Karst-Related Mineral Deposits at the Conklin and Four County Quarries, Iowa.

Deposit	Sample No.	Mineral	δ^{34}So/oo (CDT)	Description
Conklin N.	CNI-9	pyrite	9.4	late nodular
	CNI-9	sphalerite	7.0	late nodular
	CNI-23	pyrite	2.3	early nodular
	CNI-23	pyrite	9.9	late nodular
	CNI-41	pyrite	-33.8	early nodular
	CNI-41	pyrite	11.6	late nodular
	CNI-44	pyrite	8.8	late nodular
	CNI-44	sphalerite	-2.0	late nodular
	CNI-45	pyrite	-21.6	early nodular
	CNI-45	pyrite	-3.6	late nodular
	CNI-27A	pyrite	-31.8	clast surface
	CNI-28C	pyrite	-30.9	clast fracture
	CNI-28C	pyrite	-30.9	clast fracture
	CNI-33A	pyrite	-21.4	clast fracture
	CNI-33A	pyrite	-22.8	clast fracture
	CNI-33A	pyrite	-17.8	clast fracture
	CNI-42	pyrite	-25.0	clast surface
Conklin S.	RC-1	pyrite	15.7	clast fracture
	RC-1	pyrite	-9.1	clast fracture
	RC-1	barite	44.3	clast fracture
	2RR-4	pyrite	13.6	clast fracture
	CS-3	marcasite	-19.3	complex nodule
	CS-3	marcasite	2.8	complex nodule
	CS-4	pyrite	-17.6	sandstone cement
	CS-4	pyrite	9.9	sandstone concretion
Four County	4C-27	pyrite	-25.0	early nodular
	4C-63	pyrite	-11.5	early nodular
	4C-63	pyrite	-30.9	late nodular
	4C-41	pyrite	18.4	clast surface-early
	4C-41	marcasite	9.8	clast surface-late
	4C-46	marcasite	15.2	clast fracture
	4C-50	marcasite	9.1	clast fracture-early
	4C-50	sphalerite	-10.7	clast fracture-late
	4C-60	pyrite	17.7	clast fracture



Data from Heyl et al (1959) and Bunker et al (1985)

Figure 1. Map showing locations of the Conklin and Four County quarries.

DESCRIPTION OF THE DEPOSITS

Conklin Quarry

Pennsylvanian karst fills have been exposed in numerous locations during the history of operation of the Conklin Quarry. Of these, the best exposed is near the north end of the quarry. The karst opening is about 20 meters across at the bedrock surface, becoming progressively narrower with depth (Fig. 2a). The fill truncates the Coralville and Little Cedar formations (see paper by Bunker in this guidebook). Karst-filling sediments are chiefly mudstones which exhibit chaotic stratigraphy. Variation in color from black to light brownish gray is likely due to variations in organic content. The darker layers locally contain abundant carbonized plant remains. Within the lower part of the fill are numerous foundered limestone clasts of Coralville and Little Cedar formations ranging in size from a few centimeters to several meters. Clast surfaces are pitted by solutional activity.

Near the south end of the quarry is another exposed karst fill. Here, the opening is smaller and less well defined. Karst-filling sediments consist of organic-rich mudstones and clean quartz sandstones. The sandstones are associated with adjacent Pennsylvanian channel deposits (Witzke, 1984, p. 21-23; Ludvigson, 1985).

The following types of karst-associated mineralization occur at Conklin: 1) pyritic nodules wholly within mudstones (Fig. 2b), 2) coatings on the surfaces of limestone clasts and fracture fillings within them (Fig. 2c), and 3) sandstone cements (these were observed only in the southern deposit). The nodules are concentrated in the lower part of the fill. A few small nodules and pyrite disseminations were observed in the upper part of the fill.

Minerals present in the deposits are: abundant pyrite and calcite with minor sphalerite, marcasite, barite, millerite and chalcopyrite. Calcite is most abundant in clast-penetrating fractures, but occurs locally in compactive fractures in pyritic nodules. Sphalerite also is found with nodules and clasts. Microscopically, sphalerite is in part birefringent, suggesting that wurtzite may also be present. Millerite and chalcopyrite are exclusively inclusions in calcite.

Paragenetic relations for Conklin mineralization are illustrated in Figure 3a. It is important to note that: 1) there is more than one generation of nodular pyrite. Evidence includes, concentric banding, differences in crystal size, and later pyrite filling fractures in earlier pyrite, 2) sulfide minerals and barite were deposited sequentially, and 3) sulfides are generally early and calcite late. Several samples show evidence of a fracturing event after the deposition of clast-associated pyrite and before calcite.

Four County Quarry

One large, and several small, karst fills have been exposed at this quarry. Mineralization has been observed only in association with the large fill. At the time of our sampling, the fill material was no longer in place, having collapsed into a large rubble pile beneath solution-fluted cavities in the south west wall of the quarry. Karst-filling material consists of organic-rich mudstone and clean quartz sandstone. Locally within the mudstone are carbonized wood fragments. Near the karst fill at the top of the bedrock surface are Pennsylvanian channel deposits of sandstone, containing locally abundant carbonized plant remains, siltstone, and Lingulid-bearing shale. These deposits are also locally mineralized.

The types of mineralization occurring at the Four County Quarry are: 1) pyritic nodules within mudstones, 2) clast coatings and fracture fills (Fig. 2d), and 3) pyritic cements.

Minerals present are: abundant pyrite, marcasite, and calcite, minor sphalerite, barite, and rare millerite and chalcopyrite. Pyrite is the only mineral that has been observed in the nodules. Millerite and chalcopyrite are exclusively inclusions in calcite.

Paragenetic relations for Four County mineralization are illustrated in Figure 3b. It is important to note that: 1) sulfides and barite were deposited sequentially, 2) pyrite is generally early and calcite late. There are at least two generations of nodular pyrite.

With regards to mineralogy and paragenesis, Conklin and Four County deposits are remarkably similar, with the exception that marcasite is much more abundant at Four County. There is also general similarity to UMV mineralogy and paragenesis. All minerals observed in the karst deposits also occur in UMV deposits. Like UMV, karst sulfides are early and calcite is late.

ISOTOPIC ANALYSIS

Procedures

Sulfur, carbon, and oxygen isotopic analyses were performed on samples from the two deposits. Twenty seven samples of pyrite, four sphalerites, three marcasites, and one barite were analyzed for sulfur, and thirteen fracture-fill calcites and three "whole rock" limestones were analyzed for carbon and oxygen. All samples were micro-drilled from polished slabs using dental carbide burrs in order to insure precise documentation of sample powders. Where possible, several analyses were made on the same rock sample, in order to investigate compositional variations between early and late pyrite, iron sulfide and sphalerite, sulfides and calcite, and calcite and host limestone.

All isotopic analyses were performed at the laboratory of Dr. E.M. Ripley, Indiana University, using standard procedures. Analytical uncertainty is less than +/-0.1 per mil for sulfur and +/-0.5 per mil for carbon and oxygen. Sample to sample reproducibility is generally within 0.5 per mil for sulfur and 0.1 per mil for carbon and oxygen. Sulfur values are reported as per mil deviations relative to Canyon Diablo Troilite (CDT), carbon to the PDB standard and oxygen to Standard Mean Ocean Water (SMOW).

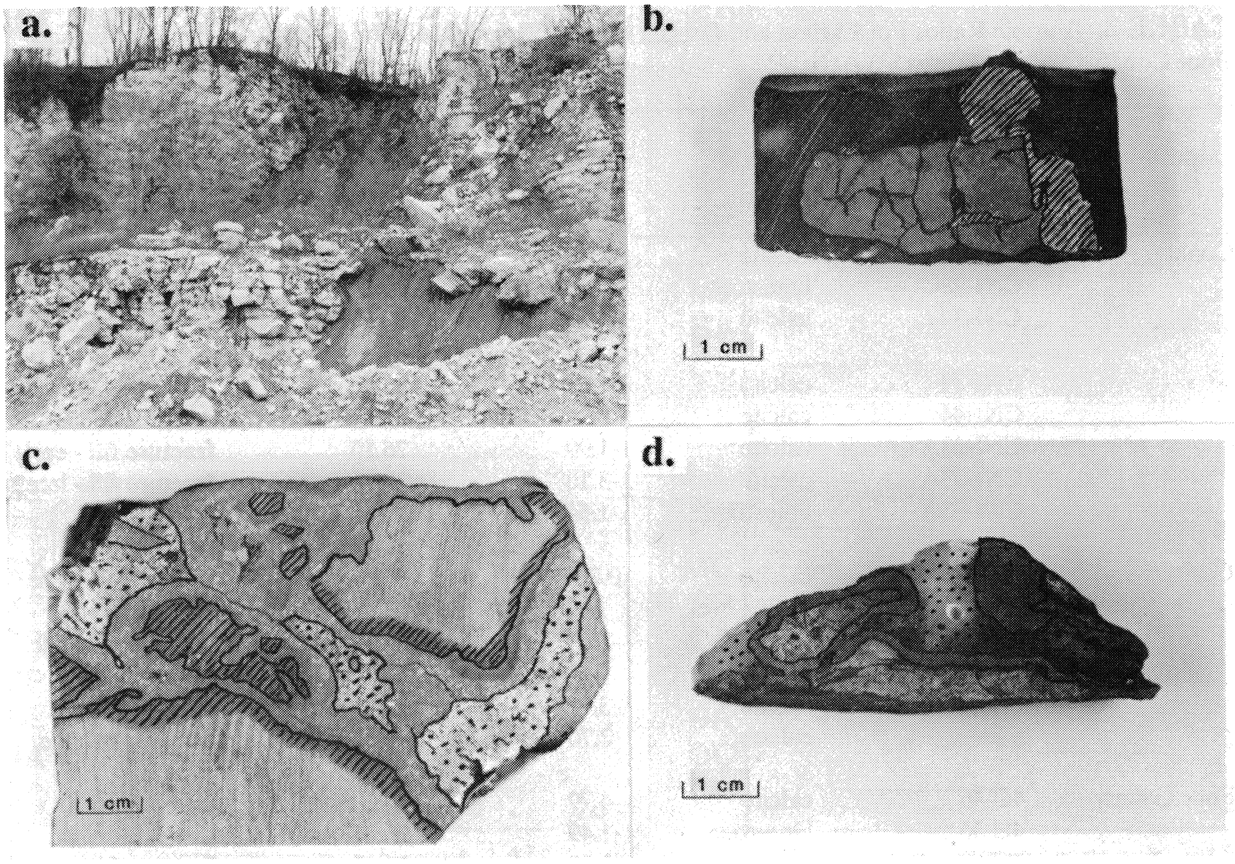


Figure 2. Photographs of paleokarst-related mineralization from the Conklin and Four County quarries.

a. Shale and mudstone-filled paleokarst at Conklin North, cutting through the Little Cedar and Coralville formations. Large foundered blocks of Coralville Formation in the foreground are weathering out of gray mudstones in the lower part of the fill.

b. Pyritic nodule from the Conklin North deposit. Medium gray - early pyrite; lined pattern - late pyrite; dark gray -dyed epoxide resin. Note compactive fractures.

c. Mineralization in clast-penetrating fracture, Conklin North deposit. Light gray - skeletal wackestone host rock; medium gray - fracture-filling pyrite; lined pattern - pyrite disseminated in limestone; stippled pattern - calcite containing inclusions of acicular millerite.

d. Mineralization in clast-penetrating fracture, Four County deposit. Mottled light gray - skeletal wackestone host rock; medium gray - fracture-filling pyrite; stippled pattern -calcite.

TABLE 2. Isotopic Ratios For Carbon and Oxygen in Karst-Related Mineral Deposits at the Conklin and Four County Quarries, Iowa

Deposit	Sample No.	Mineral	$\delta^{13}C$ ‰ (PDB)	$\delta^{18}O$ ‰ (SMOW)	Description
Conklin N.	CNI-28C	calcite	-2.98	27.15	fracture fill
	CNI-28C	limestone	1.21	24.28	
	CNI-34	calcite	-3.24	25.66	nodular sulfide impregnated
	CNI-34	calcite	-3.69	25.09	nodular sulfide free
	CNI-44	calcite	-2.44	26.98	nodular
	CNI-33A	calcite	-1.99	26.10	fracture fill - early?
	CNI-33A	calcite	-3.16	25.84	fracture fill - late?
	CNI-33A	limestone	-1.54	23.67	
Conklin S.	2RC-1	calcite	0.24	25.68	fracture - sulfide impreg.
	2RC-1	calcite	-3.86	25.37	fracture - sulfide free
	2RR-4	calcite	-3.78	23.08	fracture fill
	CS-4	calcite	-2.70	24.07	complex nodule
Four County	4C-46	calcite	-4.79	24.58	fracture fill
	4C-46	limestone	-1.49	25.07	
	4C-50	calcite	-2.03	24.07	fracture fill
	4C-60	calcite	-0.34	23.17	fracture

Results

Sulfur, carbon, and oxygen isotopic ratios for all deposits are reported in tables 1 and 2. Figure 4 illustrates variations and trends in sulfur isotopic compositions in relation to sulfide paragenesis. Figure 5 illustrates ranges of carbon and oxygen isotopic compositions of fracture fill calcites for all deposits and their relationship to each other and to the isotopic composition of host rock limestone. A study of these tables and illustrations reveals the following:

1. In the Conklin North deposit (CN) there is pronounced sulfur isotopic enrichment in late vs. early nodular pyrite (early pyrite mean = -17.7 per mil; late pyrite mean = 6.7 per mil).
2. CN sphalerite may be enriched or depleted in ^{34}S relative to pyrite in the same sample.
3. Clast-associated CN pyrites are strongly depleted in ^{34}S and the range of isotopic values is comparatively narrow (mean = -25.8 per mil).
4. ^{13}C in CN fracture fill calcites is slightly depleted relative to the limestone host.
5. ^{18}O in CN fracture fill calcites is slightly enriched relative to limestone host.
6. In the Conklin South deposit (CS) what appears to be nodular pyrite exhibits a wide range of $\delta^{34}S$ values (-19.3 to 2.8 per mil).
7. CS pyrite cement shows pronounced enrichment in ^{34}S from early to late (-17.9 to 9.9 per mil).
8. Early clast-associated CS pyrite is strongly enriched in ^{34}S (mean = 14.6 per mil).
9. $\delta^{13}C$ and $\delta^{18}O$ values for CS fracture-fill calcites are very similar to host rock values.
10. In the Four County deposits (FC) both early and late nodular pyrite are strongly depleted in ^{34}S (early pyrite mean = -18.2 per mil; late pyrite = -30.9 per mil).
11. FC clast-associated iron sulfides are strongly enriched in ^{34}S (mean = 14.0 per mil).

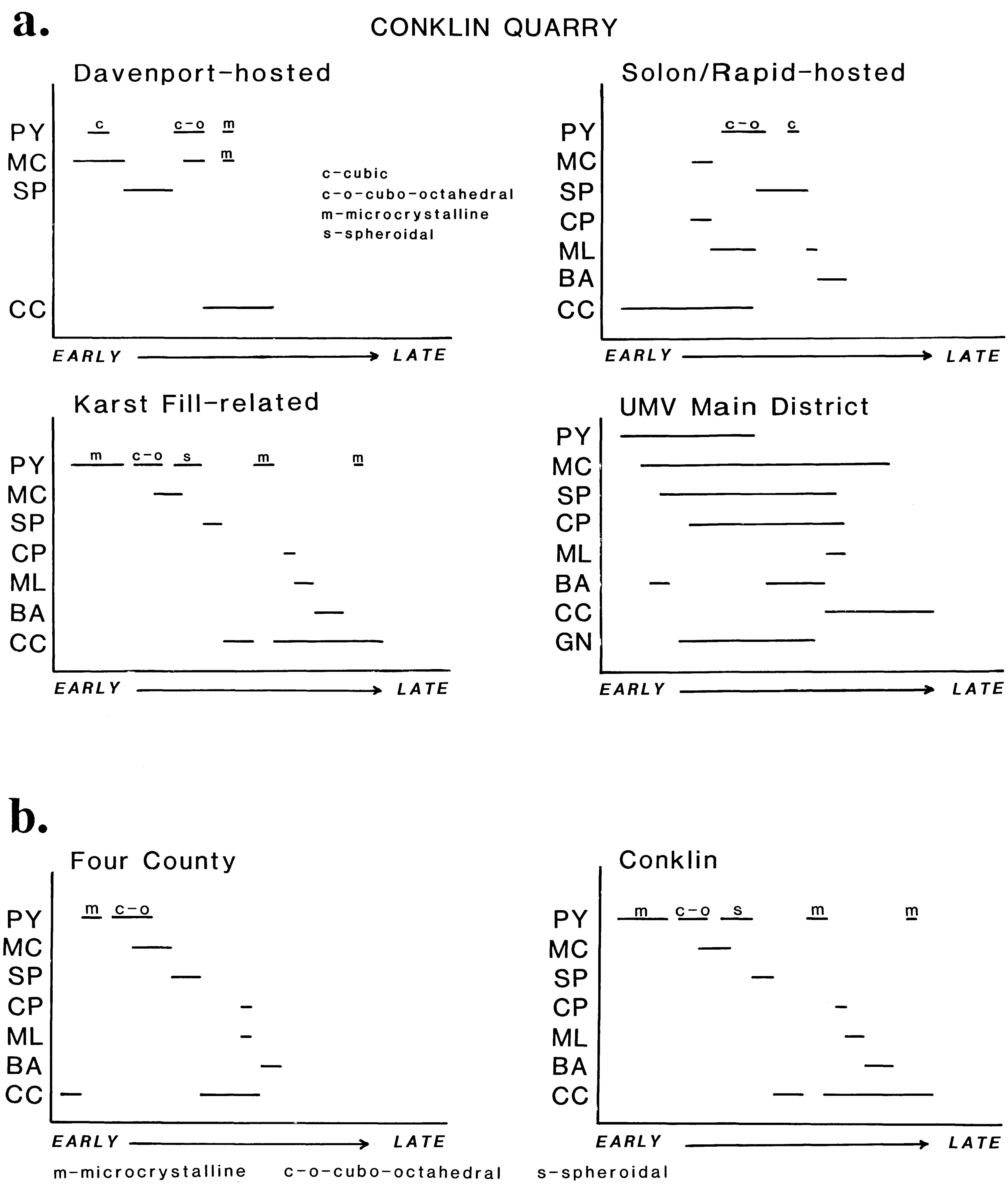


Figure 3. Mineral paragenesis in paleokarst-related deposits at the Conklin and Four County quarries.

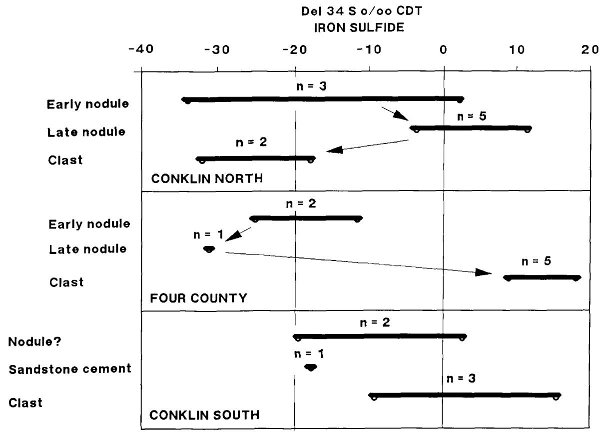


Figure 4. Ranges of sulfur isotopic ratios in iron sulfide minerals from paleokarst-related deposits, showing temporal trends.

12. FC clast-associated sphalerite is strongly depleted in ^{34}S (-10.7 per mil) relative to iron sulfide.
13. $\delta^{13}C$ and $\delta^{18}O$ values for FC fracture-fill calcites are very similar to host rock values.

Throughout the depositional history of each of the three deposits there are large shifts in sulfur isotopic composition. The directions of the shifts vary from one deposit to another. With respect to sulfur isotopic evolution, the only similarity among the deposits is that the earliest pyrite (nodular or sandstone cement) is strongly depleted in ^{34}S. For all deposits carbon and oxygen isotopic compositions of fracture-fill calcites are similar to each other, and to the composition of the host rock. These values also fall well within the range of isotopic compositions for Middle Devonian limestones in eastern Iowa (Ludvigson, 1988). It is significant that the isotopic compositions of these fracture-fill calcites are dissimilar to carbon and oxygen isotopic compositions of vein-filling calcites in the UMV Zn-Pb District, which are strongly depleted in both ^{13}C and ^{18}O (Fig. 5).

INTERPRETATIONS

The results of the mineralogic, paragenetic, and isotopic investigations of these Johnson County karst deposits permit the following interpretations:

1. Despite some mineralogic and paragenetic similarities, karst-filling and UMV deposits did not originate from the same fluid. This interpretation seemingly imposes geographic limits on the nature and extent of UMV mineralization in Iowa.
2. Despite mineralogic and paragenetic similarities, CN and CS karst deposits are not products of the same fluids.
3. The strong sulfur isotopic shifts in both negative and positive directions throughout the depositional history of each deposit indicate that multiple fluids were involved. This interpretation is supported by evidence for sequential precipitation of sulfides in each deposit. Differences in sulfur isotopic compositions between the Conklin North and Conklin South deposits indicate variations in fluid composition on a local scale.
4. Similarities in carbon and oxygen isotopic compositions between karst-related fracture- fill calcites and host limestone clasts indicate that the precipitating fluids were not hydrothermal. Depletion of ^{18}O in UMV calcites is generally considered a result of elevated fluid temperature.

The earliest nodular pyrites, which in all these deposits show strong depletion in ^{34}S, are likely

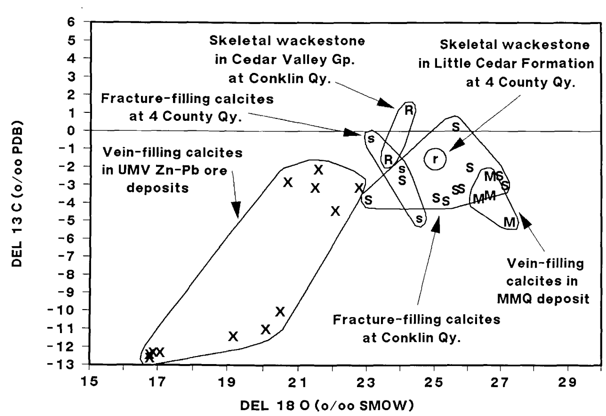


Figure 5. Carbon and oxygen isotopic ratios of Devonian limestones and fracture-filling calcites from the Conklin and Four County quarries, in comparison with sulfidic fracture calcites from other deposits in Paleozoic rocks of the UMV.

products of anoxic sulfidic diagenesis (Berner, 1981) through bacteriogenic reduction of sea water sulfate (Ohmoto and Rye, 1979). In the Conklin North deposit, the concentration of pyritic nodules in the lower fill along with limestone clasts may suggest that nodule formation preceded solution collapse. The remaining nodular and clast-associated sulfides might have originated through the recycling of early sedimentary pyrite by oxidizing groundwaters, with sulfate reduction and sulfide precipitation taking place in anoxic groundwater environments. The existence of multiple fluids might have resulted from periodic groundwater incursions related to eustatic sea level changes of several hundred meters during the deposition of the cyclothemic Pennsylvanian rocks that buried the sub-Pennsylvanian unconformity (Heckel, 1986). Variations in sulfur isotopic compositions within and between deposits reflect space/time differences in host rock chemistry and fluid compositions.

The similarity in carbon and oxygen isotopic compositions of fracture-fill calcites and associated limestone clasts suggests low fluid/rock ratios, causing host rock rather than fluid chemistry to control isotopic composition. Slight enrichments of ^{18}O in fracture-filling calcites relative to the host rock may indicate that the karst-related deposits were formed during a period when the $\delta^{18}O$ of the oceanic reservoir (and coeval meteoric water as well) was heavier than during the formation and diagenetic stabilization of the Devonian host rocks (eg. Lohmann, 1988). The suggestion of secular variations between Middle Devonian and Pennsylvanian waters is consistent with results reported in Veizer and Hoefs (1976) and Lohmann (1988).

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GEOCHEMISTRY OF FRACTURE-FILLING CALCITES IN THE PLUM RIVER FAULT ZONE; HYDROGEOLOGIC AND TECTONIC IMPLICATIONS

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INTRODUCTION

Pore fluids are active agents in the processes of rock deformation in faults in the shallow crust, and also play a significant role in triggering movement along these faults (Sibson, 1986). Fracture networks in brittle fault systems of the shallow crust, such as those exposed in Paleozoic cratonic sedimentary successions, thus have the potential to host aqueous mineral precipitates that preserve a geochemical record of the hydrogeologic systems present during active faulting (ibid.).

Stratigraphic and structural studies of the Plum River Fault Zone (PRFZ) of eastern Iowa and northwest Illinois have shown that the structure experienced multiple episodes of activity in Paleozoic time (Kolata and Buschbach, 1976; Ludvigson et al., 1978; Witzke, 1981; Bunker et al., 1985). A recently-completed petrologic study of deformed Paleozoic rocks from the PRFZ (Ludvigson, 1988) documented several overprinting stages of brittle deformation, and evaluated the environments and probable timing of fracture-filling mineralization in the fault. This paper discusses the data set for fracture-filling calcites, easily the most abundant and ubiquitous of the fault-healing minerals.

Deformed Paleozoic rocks have been investigated in five different study areas shown in Figure 1. Areas were chosen because of access to critical exposures and to assure wide geographic and stratigraphic coverage. Sampling was largely confined to carbonate facies characteristic of open marine deposition, in order to avoid interpretive confusion with diagenetic fabrics and processes associated with autobrecciation of supratidal to nonmarine carbonate strata. The calcite data are especially important because the mineral is ubiquitous as a fault rock-healing mineral phase in all of the study areas, and because the geochemical behavior of the calcite-water system in low-temperature sedimentary settings is well-known. Therefore, data on the calcites have the potential to yield insights into the paleohydrology of the PRFZ and the fundamental interrelationships between rock deformation and fluid circulation in the fault zone. Microscopic fabric, isotopic, and trace element data show that despite their development in marine carbonate strata, the fracture-filling calcites in the PRFZ were exclusively precipitated in meteoric phreatic environments, and that fault-related calcites formed during periods of marine inundation or hydrothermal mineralization apparently are lacking.

ALTERNATIVE HYDROLOGIC REGIMES FOR DEFORMATION IN CRATONIC FAULT SYSTEMS

Three possible environments for the precipitation of fracture-filling calcites in the PRFZ are shown in Figure 2. Each of these alternatives has been proposed to explain the origin of fracture calcites in ancient fault systems.

The Meteoric Phreatic Hypothesis

Calcites from the PRFZ occur as spars with blocky equant, poikilotopic, and spherulitic to columnar habits, and as microspar. These growth forms have been described from cements in non-marine settings that were precipitated from meteoric water (Chafetz et al., 1985). Modern marine carbonate muds are known to neomorphose to calcite microspar in contact with

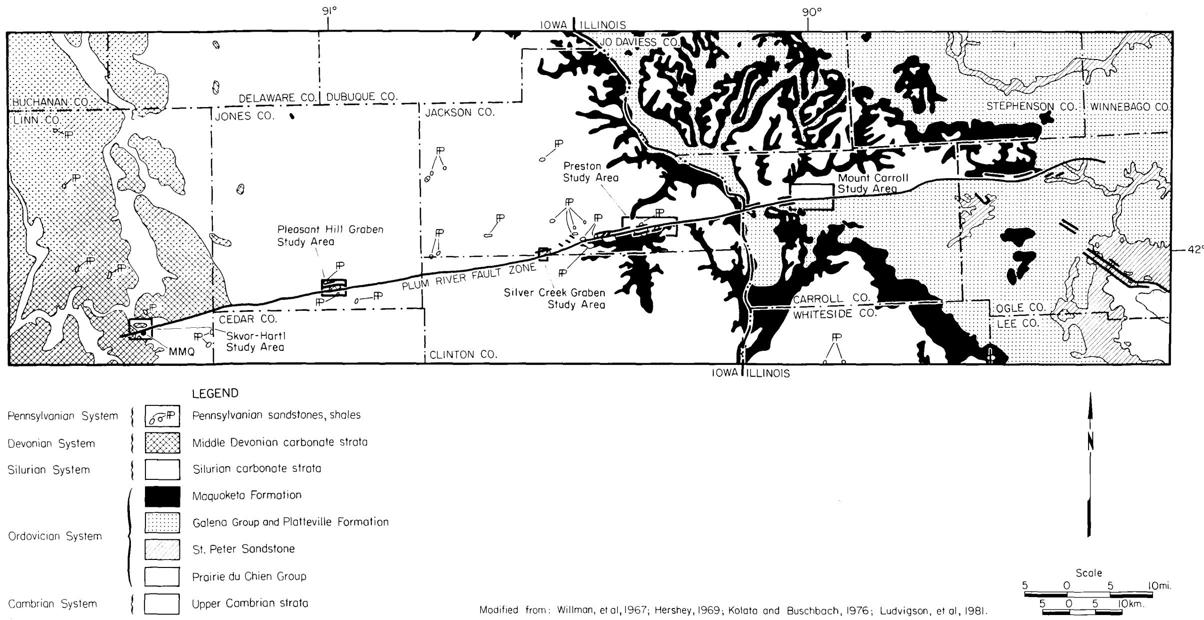


Figure 1. Bedrock geologic map of the Plum River Fault Zone, with locations of study areas examined by Ludvigson (1988).

meteoric water (Steinen, 1982). Relict meniscus or pendant fabrics in pore-filling cements, regarded as evidence of cementation in vadose environments, have not been observed. Microspar fills fractures that were previously lined by blocky equant spar, and is intercalated with spar. Thus the calcites formed in phreatic aqueous environments. The lack of crustiform fibrous growth habits, characteristic of precipitation of aragonite or high-magnesian calcite (HMC) cements in shallow marine environments from waters with high Mg/Ca ratios, indicates that the cements are primary precipitates of low-magnesian calcite (LMC) from dilute solutions of meteoric derivation (see Longman, 1980). Abundant ferric oxides coexist as cements and detrital inclusions, lending support to the meteoric phreatic hypothesis, because these phases are consistent with oxidizing continental weathering environments, and the general abundance of aqueous iron species in continental waters (Veizer, 1983).

A steady-state, meteoric gravity-driven phreatic groundwater flow system across the PRFZ is illustrated in Figure 2a. This hydraulic regime has been proposed to control the activity of continental fault systems (Costain et al., 1987). Several caveats are in order before this alternative is accepted as the paradigm for the spectrum of diagenetic environments in which the calcites were precipitated. Shallow marine aragonite and HMC cements are metastable, and are dissolved or neomorphosed in marine carbonates that are infiltrated by meteoric pore fluids. Care must be taken to assure that LMC cements are primary void fillings rather than pseudospars after earlier metastable marine cements. Neomorphic fabrics suggesting the former presence of early marine cements in ancient carbonate rocks are well-documented in the literature (i.e. Lohmann and Meyers, 1977; Mazzullo, 1980). No petrographic evidence for neomorphism of marine cementation was found in the cements of interest, and they are interpreted to be primary LMC precipitates from meteoric phreatic waters.

Isotopic Data

The covariation between carbon and oxygen isotopic ratios from 36 analyses of calcites from the PRFZ are compared in Figure 3 with examples from calcites deposited in several alternative hydrologic settings. Comparison reveals that the isotopic ratios of the cements in the PRFZ most closely correspond to those from

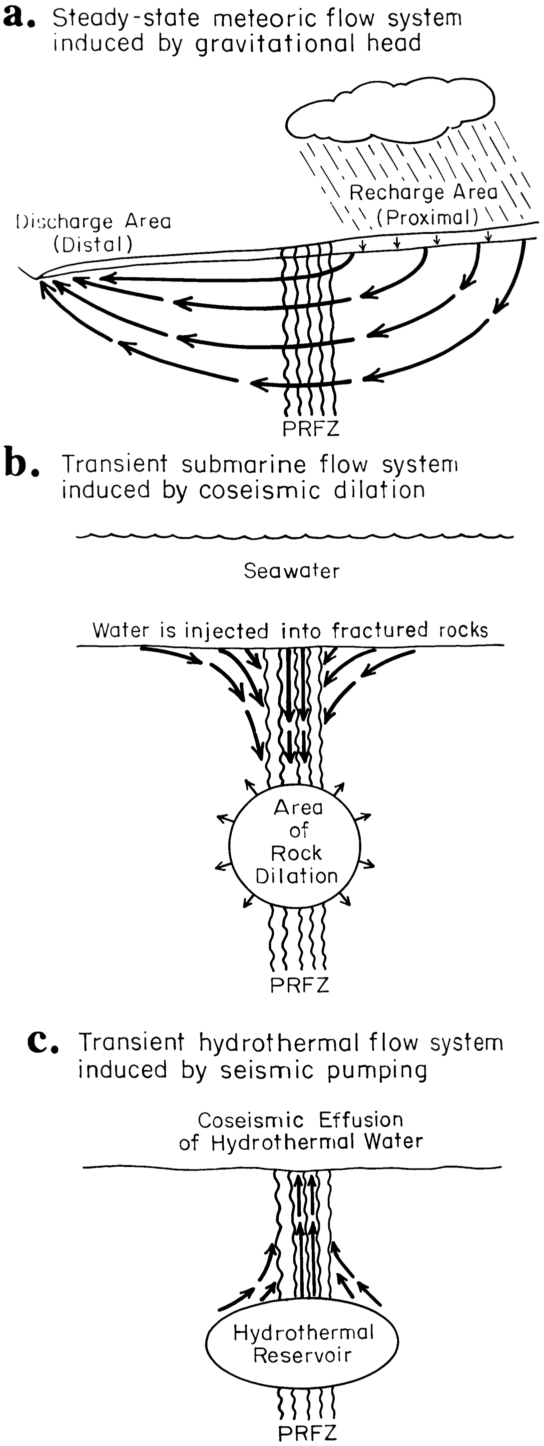


Figure 2. Alternative hydrogeologic settings for the precipitation of fracture-filling calcite cements along the Plum River Fault Zone. From Ludvigson (1988).

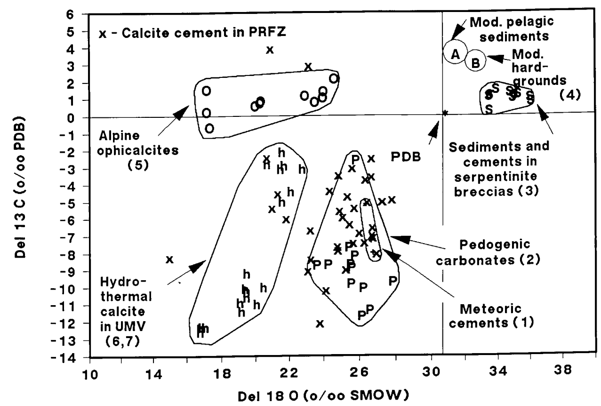


Figure 3. Carbon and oxygen isotopic ratios of fracture-filling calcite cements in the Plum River Fault Zone, in comparison to cements from various geologic environments. Sources of data are: 1, Dickson and Coleman, 1980; 2, Magaritz et al., 1981; 3, Bonatti et al., 1974; 4, Schlager and James, 1978; 5, Weissert and Bernoulli, 1984; 6, Hall and Friedman, 1969; and 7, McLimans, 1977. From Ludvigson (1988).

calcites deposited from meteoric waters in near-surface settings.

The main cluster of data from the calcites in the PRFZ form a field with similar orientation, and with similar ranges in $\delta^{13}C$ and $\delta^{18}O$ as that in the field for pedogenic and meteoric (i.e., fresh water) carbonates (Fig. 3). A field for vein-filling calcites formed in near-surface fresh waters (from Dickson and Coleman, 1980) conforms to an array with relatively constant $\delta^{18}O$ values and more highly variable $\delta^{13}C$ values. This field falls within a larger, but similarly-oriented field that encloses the data from Magaritz et al. (1981) for a group of pedogenic calcite nodules. Notably, both fields are depleted of heavier isotopes with respect to the PDB standard. These data arrays result from varying contributions by ^{13}C-depleted soil-gas CO_2 respired by aerobic and anaerobic bacteria that oxidize organic matter ($\delta^{13}C$ = -20 o/oo PDB) in the soil zone. Other contributions arise from rock matrix carbonates that are mobilized by weathering processes in the soil (generally marine carbonates with $\delta^{13}C \approx 0$ to 4 o/oo PDB), and unequal mixtures result in a characteristic wide range of possible carbon isotopic ratios. Oxygen isotopic ratios of calcites precipitated in meteoric environments are controlled by the $\delta^{18}O$ of the interstitial meteoric water (-10 to 0 o/oo SMOW in temperate to tropical coastal marine settings) and the temperature-controlled fractionation between the precipitating calcite and water (approx. +29 o/oo at 25°C). The meteoric phreatic hypothesis for the calcites in the PRFZ thus is shown to be compatible with the carbon and oxygen isotopic data.

Trace Element Data

A set of 96 electron microprobe analyses was collected on fracture calcites from various samples along the PRFZ. These data further show that the calcites as a whole are most reasonably interpreted as the products of cementation in near-surface environments saturated by waters of meteoric derivation.

Figure 4a shows a histogram of Mg contents measured in the calcite cements, along with superimposed ranges of Mg contents documented from elemental studies of modern carbonate sediments. These data confirm the conclusions based on petrographic observations of cement fabrics that all those examined are LMC.

The Mn and Sr distributions in the calcite cements in the PRFZ are portrayed in Figure 4b, following the conventions established by Brand and Veizer (1980), Veizer (1983), and Brand (1987). The theoretical range of Mn and Sr concentrations to be expected for LMC cements precipitated in equilibrium with sea water are shown in the rectangle in the upper left-hand corner of the diagram. Despite the limitations imposed by the sensitivity of the microprobe, it is evident that none of the calcites from the PRFZ would plot within the sea water field. A substantial number of the analyses had Mn and/or Sr concentrations well below the semiquantitative detection limit of the instrument (approx. 100 ppm), and these are plotted along the appropriate axes.

Significantly, all of the analyses are depleted in Sr with respect to the concentration expected from LMC precipitated in equilibrium with sea water. Plots for the calcites in Figure 4b broadly conform to an arcuate array that fans out to the lower right from the "sea water" field, indicating generally increasing Mn and decreasing Sr

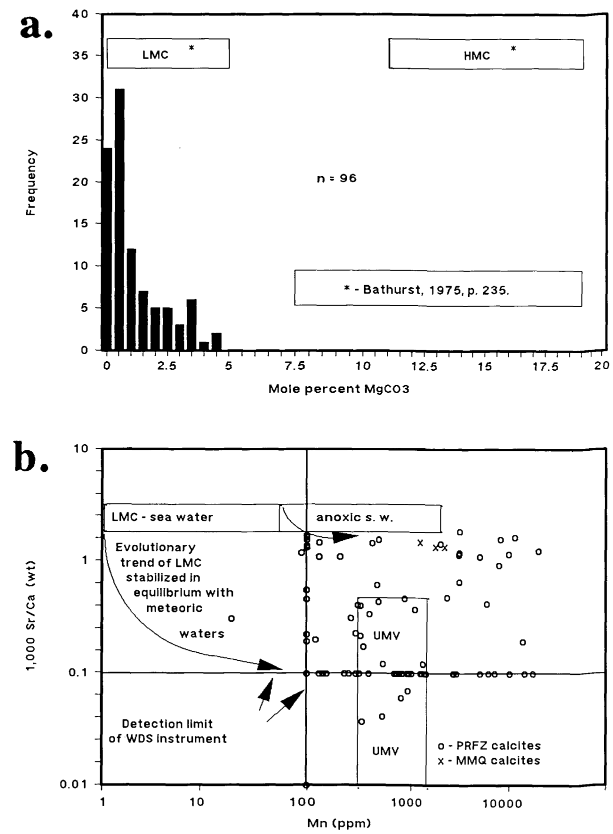


Figure 4. Element chemistry of fracture-filling calcites from the Plum River Fault Zone. a. Histogram of magnesium contents the calcites, in comparison to ranges defined for high-magnesian (HMC) and low-magnesian (LMC) calcite. b. Manganese and strontium distributions in the calcites, in comparison to fields for LMC in equilibrium with sea water, and data from vein-filling calcites from the Upper Mississippi Valley (UMV) zinc-lead district (Hall and Heyl, 1968). From Ludvigson (1988).

contents. This trace element pattern is typical for LMC cements in marine carbonate rocks that have precipitated in equilibrium with meteoric waters (Veizer, 1983), and decreasing Sr/Ca ratios are regarded as indicative of increasing water/rock interaction.

The widely varying Mn and Sr contents in the calcites from the PRFZ (Fig. 4b) yield information on the related ancient groundwater systems, including their timing with respect to the diagenetic stabilization of enclosing rocks. Continental waters are known to contain much higher Mn^{2+} concentrations than seawater (Veizer, 1983), and high concentrations develop in anoxic groundwater systems, where the reduction of detrital or authigenic manganese oxides provides an abundant source. Conversely, Mn^{2+} is unstable in oxidizing solutions, and precipitates as oxide phases (Mn^{3+}, Mn^{4+}) in near-surface oxygenated groundwaters. These controlling factors lead to a wide range of possible Mn contents in meteoric calcites, with higher concentrations associated with anoxic environments.

Original aragonite and HMC components in marine sediments contain high Sr contents, although the element is greatly depleted during diagenetic conversions to newly-formed LMC (Veizer, 1983). As the distribution coefficient for Sr in calcite is far less than unity, fluids involved in the diagenetic maturation of marine carbonate sediments should be expected to progressively scavenge dissolved Sr^{2+}. Mineralized fluids in the anoxic distal margins (Fig. 2a) of groundwater flow systems, whose solution chemistry results from prolonged contact with marine carbonates, thus have the potential to exhibit simultaneous enrichments in Mn^{2+} and Sr^{2+}. Some of the most Mn-rich calcites from the PRFZ are also relatively enriched in Sr. These calcites are interpreted to have formed in anoxic mineralized groundwaters located in the distal portions of a flow system (Fig. 2a) in diagenetically immature limestones. Mn^{2+} enrichment, although subject to fluid-rock partitioning as well, is more fundamentally controlled by redox chemistry, while Sr^{2+} concentration is relatively insensitive to Eh. Ancient marine limestones that have been fully converted to LMC do not yield appreciable quantities of Sr^{2+} to their contained groundwater systems (Kinsman, 1969). Therefore, fracture-filling calcites in the PRFZ with mutual Mn and Sr enrichments may have been formed before the diagenetic stabilization of the enclosing rocks was complete.

The Marine Phreatic Hypothesis

Consideration should be given to calcite cements precipitated in deeper marine realms, where submarine hardgrounds may be cemented by blocky equant LMC spars. Wilkinson et al. (1982) and Wilkinson et al. (1985) suggested that slow rates of precipitation may be of greater importance than Mg/Ca ratios of pore fluids in determining whether HMC or LMC cements are formed. The precipitation of submarine LMC cements may result from both the cooler temperatures of deeper bottom waters, and the diffusion-controlled transport of dissolved solutes resulting from negligible fluid exchanges between bottom sediments and overlying waters. It might be argued that that the LMC cements in the PRFZ precipitated from a mass of cool bottom-layer sea water that either diffused into, or was injected into previously lithified, dilationally fractured rocks beneath the floor of an epeiric sea. A transitory submarine flow system driven by coseismic dilation along the PRFZ is shown in Figure 2b. This interpretation is discounted, however, because of the isotopic and trace element chemistry of the fracture-filling calcites.

The isotopic ratios of Quaternary pelagic sediments and submarine LMC cements (fields A and B in Fig. 3) were characterized by Schlager and James (1978). Both fields are characterized by δ-values greater than the PDB standard. The higher oxygen isotopic ratios are believed to reflect equilibrium precipitation from cold marine bottom waters (3 to 10^{o}C, ibid.). Substantially similar results were obtained from breccia and fracture-filling pelagic sediments and aragonite/calcite cements in serpentinite breccias dredged from an oceanic fracture zone along the Mid-Atlantic Ridge (Bonatti et al., 1974). The carbonate phases were proposed to have been emplaced by the injection of bottom waters into oceanic basement rocks during deformation along an active transform fault (ibid.). They serve as modern mechanical and hydrologic analogues to the marine LMC hypothesis (Fig. 2b).

Data from ancient fault-related submarine cements was published by Weissert and Bernoulli (1984), with analyses of internal sediments and blocky equant spars filling neptunian dikes in Alpine ophicalcites, linear zones of altered ophiolites interpreted to be remnants of fossil oceanic fracture zones (ibid.; Bernoulli and Weissert, 1985). These data are also shown in Figure 3. The ophicalcites transect a regional metamorphic zonation, and $\delta^{18}O$ values in the calcites directly reflect the metamorphic facies of the enclosing rocks, with the lighter values in the higher-grade rocks. Original oxygen isotopic values have been thermally reset by regional metamorphism. The comparatively narrow range of $\delta^{13}C$ values, typically greater than the PDB standard, parallel other analyses of submarine cements, and differ from the field for near-surface meteoric phreatic environments and the data from the PRFZ (Fig. 3).

Marine bottom waters apparently are capable of precipitating LMC cements outside the "sea water" field as defined by Brand and Veizer (1980) and Veizer (1983). Brand (1987) found that unaltered LMC skeletal grains from dysoxic facies in a Pennsylvanian marl had Mn and Sr distributions that plotted within the rectangular field labelled "anoxic s. w." in Figure 4b. This trace element pattern results from the mobilization and upward diffusion of dissolved Mn^{2+} into the water column from the reduction of detrital Mn oxides in anoxic bottom sediments. Thus Mn-rich calcites can be precipitated from oxygen-depleted sea waters. Sr^{2+}, however, behaves as a highly-mobile conservative species that is insensitive to dissolved oxygen concentrations, and the relatively depleted Sr/Ca ratios in the calcites from the PRFZ indicates that they cannot be primary precipitates from sea water.

The Hydrothermal Hypothesis

Hydrothermal vein-filling calcites in the MVT lead-zinc deposits of the Upper Mississippi Valley grew as scalenohedral to rhombohedral spars with crystal diameters ranging from a few to tens of centimeters (Heyl et al., 1959). In thin section view, these spars could conceivably be mistaken for meteoric LMC spars precipitated in shallow, near-surface settings, although the presence of disseminated pyrite inclusions in the former (ibid.) could serve as a useful distinguishing characteristic. This alternative merits serious consideration in light of the recent suggestion by Sibson (1986) that near-surface brittle deformation along active fault zones may result in interseismic healing by hydrothermal fluids that are injected into higher crustal levels by seismic pumping. Although this phenomenon should generally be confined to faults in orogenic belts characterized by high geothermal gradients, hydrothermal calcites could occur if fault movements were contemporaneous with regional MVT mineralization. Hydrothermal mineralization is known to fill small-scale faults in Paleozoic rocks in the Upper Mississippi Valley zinc-lead district, and several authors have suggested that cogenetic sulfide mineralization is localized along the PRFZ (Heyl and West, 1982; Ludvigson et al., 1983). A transitory, seismically-induced hydrothermal flow system along the PRFZ is shown in Figure 2c. Isotopic and trace element data show, however, that this hypothesis is not tenable for the calcites in the PRFZ.

Data on the isotopic ratios of calcites associated with MVT zinc-lead deposits in the Upper Mississippi Valley (the only well-documented hydrothermal event affecting the Paleozoic rocks in the region) are shown in Figure 3. These calcites are notably depleted in ^{18}O compared with all the other populations except for the metamorphosed Alpine ophicalcites. The cluster of data from the PRFZ has ^{18}O contents that are enriched relative to the field of hydrothermal calcites by 2 to 6 o/oo (Fig. 3). The depleted values in UMV calcites result from smaller fractionations between calcite and water at higher temperatures. The depleted ^{13}C compositions of UMV calcites, especially those associated with the most ^{18}O-depleted cements, has been cited as possible evidence for interactions between hydrothermal fluids and thermally altered hydrocarbons (Sverjensky, 1981). These phenomena result in a field with similar $\delta^{13}C$ values, but grossly dissimilar $\delta^{18}O$ values to the data from the PRFZ. Accordingly, the hydrothermal hypothesis appears to be untenable.

The relatively restricted range of Mn and Sr contents measured from vein-filling calcites in the UMV zinc-lead deposits is shown in the "UMV" rectangle in Figure 4b. Many of the analyses of calcites from the PRFZ plot within this field, although a considerably wider range of values was detected. While the shared ranges of trace element contents could indicate identical environments of precipitation, this conclusion is negated by the observation that the most abundant iron-bearing minerals coexisting with the cements in the PRFZ are oxide phases, while those in the UMV cements are sulfide phases. Many of the cements in the PRFZ were precipitated in oxidizing environments that radically differed from the strongly reducing environments in which the UMV cements were precipitated. Among the calcites analyzed from the fault zone, only those from the sulfide mineral deposit at the Martin Marietta Quarry (MMQ deposit) coexist with sulfide minerals, yet these plot well outside the isotopic and trace element fields defined for UMV calcites (Ludvigson, 1988). These relationships further establish that none of the calcites examined from the PRFZ have identity with the vein-filling calcites of the UMV zinc-lead deposits, and further argue against the hydrothermal hypothesis for the calcites in the fault zone.

AN INTERPRETIVE SCHEME FOR THE METEORIC CALCITES IN THE PLUM RIVER FAULT ZONE

Calcites precipitated from meteoric groundwaters are deposited in several unique environments that impart unique geochemical signatures. A model is developed here to describe the spatial relationships between environmental zones that produce geochemically distinct calcites. A fundamental task in the formulation is the determination whether a single inclusive model or multiple models are most appropriate. Diagenetic systems that develop in dolostone terranes evidently are different enough from those in limestone terranes to warrant the formulation of two separate, but parallel models. Comparisons between the trace element distributions of calcites in juxtaposed limestones and dolostones (Ludvigson, 1988) showed that different covariate trends occur in the Mg and Fe contents of cements contained in each rock type. Examination of Mg distributions in LMC cements as grouped by host rock lithology (Fig. 5) shows that the calcites in limestones are characterized by concentrations between 1,000 and 6,000 ppm, whereas the cements in dolostones are characterized by concentrations of from 100 to

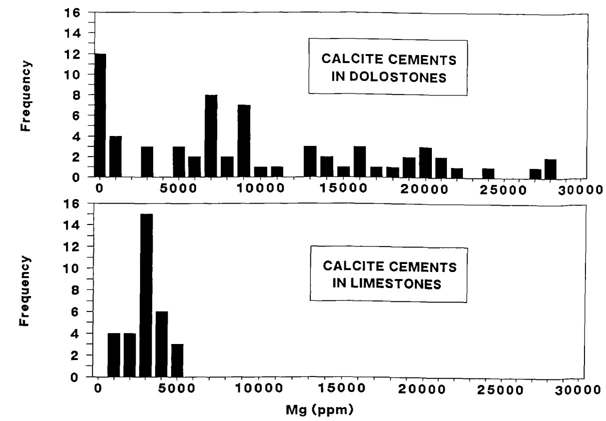


Figure 5. Histograms comparing Mg contents in fracture-filling calcites contained in limestones and dolostones from the Plum River Fault Zone. From Ludvigson (1988).

29,000 ppm.

The range of Mg contents observed in calcites from dolostones results from varying concentrations of dissolved Mg^{2+} in groundwaters flowing through weathering environments where the incongruent dissolution of dolomite (molar Mg/Ca ratio $\approx$ 1) drives the precipitation of LMC (molar Mg/Ca ratio = 0 to 0.05). The conversion of each mole of dolomite to calcite leads to a proportionate increase in the concentration of dissolved Mg^{2+} in the pore fluid. Since the distribution coefficient for Mg in calcite is less than unity, early-formed calcites in the dolostone terrane will incorporate negligible quantities of Mg, but downflow concentration of Mg^{2+} in the groundwater flow system is further enhanced. Dissolution/precipitation reactions in marine limestones, even those driven by dissolution of HMC (molar Mg/Ca ratio $\leq$ 0.2) result in: 1) a narrower total range of Mg contents, and 2) smaller concentrations of residual Mg^{2+} in newly-forming LMC.

While other factors, including changes in water/rock ratios and distribution (partition) coefficients as effected by rates of precipitation and sector zoning control the substitution of trace elements in the calcite lattice (Reeder and Grams, 1987), it is evident that the chemical composition of the rock matrix is a major factor in determining the Mg contents of fracture-filling calcites in the

PRFZ. Thus, two parallel models have been formulated by examining the generation of late diagenetic meteoric calcites in both dolostone and limestone terranes. The principal difference between the models is that groundwater flow systems in dolostone terranes are characterized by steeper downflow concentration of dissolved Mg^{2+}.

Configurations of Meteoric Groundwater Flow Systems

The overall configuration and flow geometry of both local and regional groundwater flow systems are principally governed by the configuration of land surface topography, with major surface drainage divides and base levels (discharge boundaries) serving as fundamental geographic boundaries. A generalized scheme for diagenetic zonation in the meteoric groundwater flow system is shown in Figure 6a. The relative concentrations of biologically-controlled dissolved gases (O_2, CO_2, CH_4, and H_2S) constitutes one of the most useful distinctions that can be made between different groundwaters. Bacterial conversions between oxidized and reduced gases in aqueous environments exert major control on the cycling of manganese and iron, and have been used to formulate a geochemical classification scheme for sedimentary environments (Berner, 1981). This scheme is applicable to environmental zones in meteoric groundwaters. To the maximum extent possible, terms and definitions of zones of phreatic diagenesis (Fig. 6) are based on those of Berner (1981).

The major difference between the diagenetic model shown here and Berner's (1981) paradigm results from differences in the distribution of sulfate salinity in meteoric and marine waters. Spatial and sequential relationships between zones of anoxic diagenesis in meteoric groundwater systems, as shown in Figure 6a, differ from those outlined by Berner (1981), who applied them mainly to environmental successions in submarine sediments. Of basic importance is the sequential ordering between oxic, anoxic nonsulfidic, and sulfidic environments in the downflow direction. The boundaries between diagenetic zones may be subhorizontal as shown, or may be steeply dipping, as has been documented in confined aquifer systems (Champ et al., 1979). The configuration shown in Figure 6a shows that oxic environments occur in shallow, near-surface groundwaters in proximal recharge areas, whereas anoxic environments occur in deeper groundwaters and distal discharge areas.

Another important spatial variation is the relative position of the oxic-anoxic transitional zone. The example shown in Figure 6a applies to many of the ancient environments recorded along the PRFZ, where oxic and oscillating oxic-anoxic meteoric phreatic environments are abundantly represented. Following discussions of specific zones of diagenesis refer to the configuration in Figure 6a, as applied to observations on fracture-filling cements in the PRFZ.

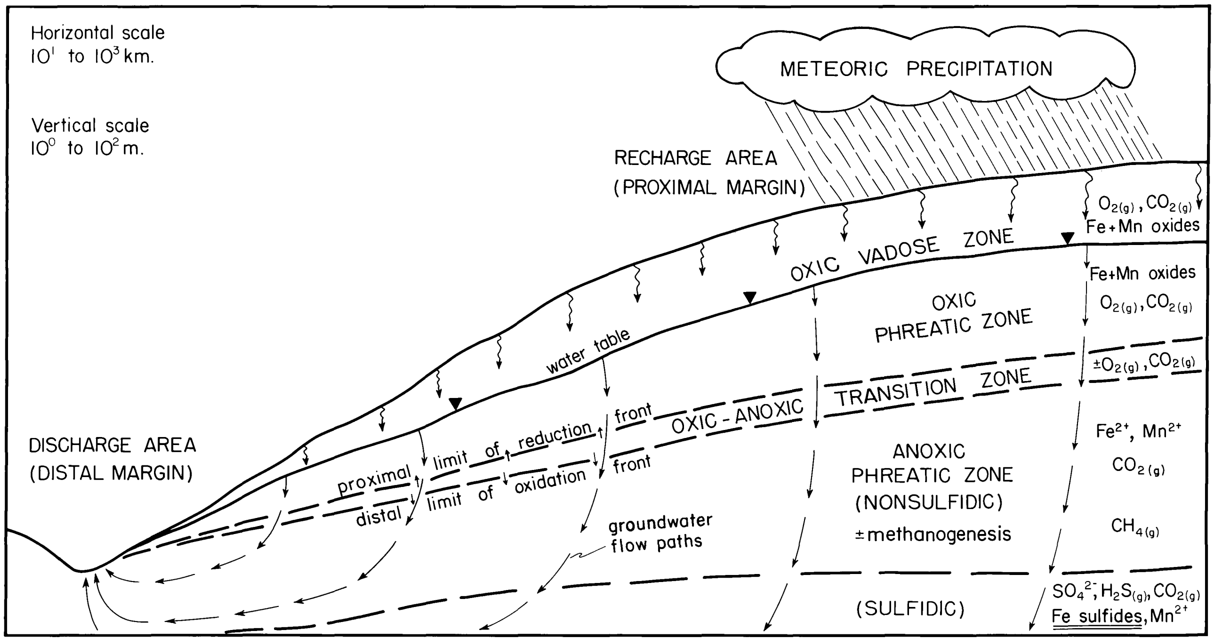
Vadose Zone

Soils in the vadose zone are sites of chemical weathering of rock-forming minerals, and the residual accumulation of their insoluble weathering products. Calcite is a common neoformed mineral in soils derived from many different parent rock materials, and its generation results from the universal biologic production of dissolved CO_2 gas in vadose soils. Vadose waters charged with CO_2 aggressively dissolve existing rock matrix carbonates, and calcite precipitation is initiated by CO_2 outgassing through pore networks, and through the evaporation of soil water (Rabenhorst et al., 1984).

Although calcite fabrics suggestive of vadose

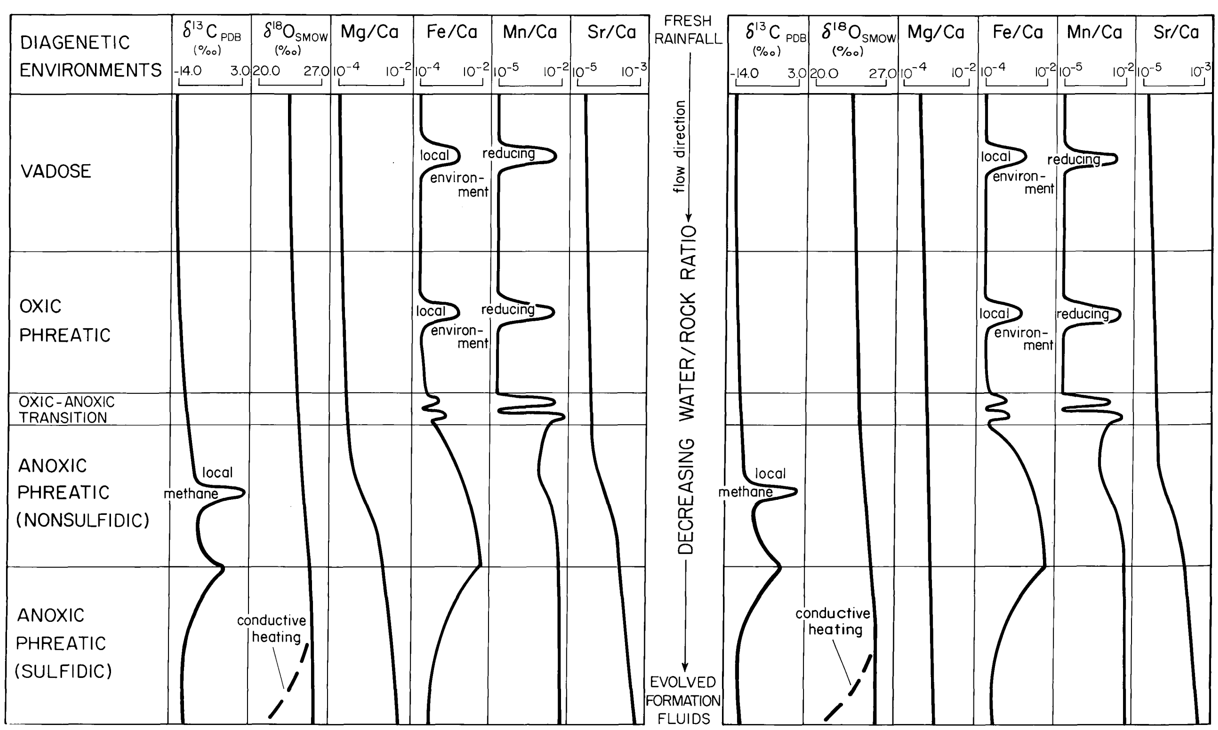
Figure 6. Conceptual model for the environmental zonation (Berner, 1981) of meteoric groundwater flow systems, as applied to the precipitation of fracture-filling calcites in the Plum River Fault Zone. a. Spatial distribution of zones of meteoric diagenesis. b. Variations in stable isotopic ratios and elemental ratios in calcites originating from groundwater environments in dolostone terranes. c. Variations in stable isotopic ratios and elemental ratios in calcites originating from groundwater environments in limestone terranes. From Ludvigson (1988).

a. Zones of meteoric diagenesis



b. Calcites precipitated in dolostone terrane

c. Calcites precipitated in limestone terrane



diagenesis were not observed in the rocks of interest, products of vadose diagenesis, including detrital inclusions of iron and manganese oxides and corroded dolomite fragments, are intimately associated with meteoric phreatic calcites in the PRFZ. Vadose diagenesis merits discussion, however, because soils are hosts to biological activity that imparts distinctive geochemical signatures to vadose waters as they infiltrate to the phreatic zone. The single most important geochemical characteristic of calcites precipitated from meteoric groundwaters, the development of depleted carbon isotopic ratios, originates from processes that operate within soils in the vadose zone. This vadose soil-formed signature is subsequently transferred to other portions of the groundwater system.

Gases in soils contain comparatively large concentrations of biogenic ^{13}C-depleted CO_2 gas generated by the respiration of root systems of vascular plants, and by the bacterial oxidation of organic carbon within the soil matrix. Biogenic CO_2 constitutes the major proportion of the dissolved carbonate reservoir in vadose zones, and pedogenic (vadose) carbonates thus are characterized by depleted carbon isotopic ratios (Rabenhorst et al., 1984; Figs. 6b, c). The oxygen isotopic ratios of calcites formed in freshly infiltrating ^{18}O-depleted meteoric waters differ from those formed in more downflow positions as a consequence of dissolution/precipitation reactions involving isotopically heavier marine carbonates in the rock matrix (Figs. 6b, c).

The distribution of Fe, Mn, Mg, and Sr in calcites can be used to identify cements that could only have precipitated at sites of proximate groundwater recharge, including the vadose zone and the adjacent oxic phreatic zone. The geochemical cycling of iron and manganese in meteoric groundwaters is principally controlled by aqueous redox reactions involving the dissolved gases O_2, CH_4, and H_2S, whose concentrations are governed by distinctive bacterial environments. The cycling of magnesium and strontium, however, are principally governed by dissolution/precipitation reactions with rock-forming minerals in the parent material of the soil.

Pore networks in the vadose zone are filled by mixtures of water and both oxygenated atmospheric and O_2-deficient, CO_2-rich indigenous gases. Geochemical signals acquired from vadose waters in the soil zone (depleted carbon and oxygen isotopic ratios, coprecipitation of Fe and Mn oxides) are at their maximum, and signals acquired by dissolution/precipitation reactions with the rock matrix (elevated concentrations of Mg and Sr) are at their minimum.

Manganese and iron oxides and hydroxyoxides are concentrated in the vadose zone as insoluble weathering products of unstable rock-forming minerals, and as a consequence of biological concentration and release from decomposing organic materials (Mount and Cohen, 1984; Figs. 6b, c). The virtual absence of dissolved Mn^{2+} in oxic vadose environments should result in the precipitation of Mn-free calcites that, lacking the major activating element, are nonluminescent under cathodoluminescent (CL) excitation. Contact with atmospheric oxygen dictates that insoluble Mn and Fe oxide phases are greatly predominant in the vadose zone, but concentrations of organic materials in the soil (i.e., fossil roots, fecal pellets from burrowing organisms) provide sites for the development of local reducing environments (i.e., "drab halos" surrounding fossil root traces), and facilitate the local mobilization of Mn^{2+} and Fe^{2+}. The heterogeneous mixture of localized environments can lead to the close juxtaposition of both iron and manganese oxide phases with soil-formed ferroan and manganoan calcites (Mount and Cohen, 1984; see Figs. 6b, c).

Dissolved species in groundwaters that are acquired from the dissolution of marine carbonate components in the rock matrix, notably Mg^{2+} and Sr^{2+}, are depleted in vadose waters compared to concentrations that develop in more downflow positions. Both elements have distribution coefficients of less than one with respect to the precipitation of LMC, further assuring their initial depletion in cements deposited in proximal recharge areas, and the development of positive concentration gradients along groundwater flow paths. Concentration gradients are further controlled by variations in rates of groundwater flow and rates of dissolution-precipitation reactions. Low Sr/Ca and Mg/Ca ratios are indicative of calcites formed in the proximal portions of meteoric groundwater flow systems, and are characteristic of vadose and oxic phreatic cements in both dolostone and limestone terranes (Fig. 6).

Oxic Phreatic Zone

After their infiltration into the phreatic zone, vadose groundwaters that were initially saturated with atmospheric oxygen pass into a diagenetic system that is closed with respect to any new inputs of dissolved oxygen. The oxic phreatic zone typically has a very limited extent in comparison to the total dimensions of the flow system. In most groundwater flow systems, dissolved oxygen will eventually be consumed to exhaustion by the oxidation of resident organic carbon in the phreatic environment. Median concentrations of dissolved organic carbon (about 0.7 ppm) in groundwaters evidently are sufficient to sustain oxygen-consuming aerobic bacterial activity in aquifer systems, ultimately leading to the development of anoxic phreatic environments (Champ et al., 1979).

Geochemical cycling processes in the oxic phreatic zone are similar to those described from the vadose zone. Located in the proximal reaches of the flow system, oxic phreatic environments are characterized by water-rock interactions with high water/rock ratios. Calcites precipitated in the oxic phreatic zone thus have geochemical signatures that are inherited from oxidation reactions in the vadose zone, while signals acquired from dissolution/precipitation reactions with preexisting rock matrix carbonates are minimal (Fig. 6). Therefore, calcite spars with depleted carbon isotopic ratios, coprecipitated Mn and Fe oxides, and low Mg/Ca and Sr/Ca ratios are characteristic of cements precipitated within the oxic phreatic zone (Fig. 6).

The Mn^{2+} ion is unstable in oxidizing environments. Thus oxidizing environments should be expected to produce Mn-free calcite cements that are nonluminescent under CL excitation. Nonluminescent, Mn-deficient calcites in the PRFZ are observed with some frequency (see Fig. 7, sites C, E, G and I). These cements are interpreted to have been deposited in oxic phreatic environments. Bulk oxidizing environments contain local sites of reduction around concentrations of organic material. These local environments serve as sites for the local activation and cycling of minor concentrations of dissolved Mn^{2+} and Fe^{2+} through the distal margins of the oxic phreatic zone (Figs. 6b, c). Only minimal concentrations of Mn, about 25 ppm, are necessary to activate CL luminescence in calcite (Mason, 1987). In light of the preferential inclusion of dissolved Mn^{2+} in the calcite lattice (magnified by up to 30 times), it is evident that luminescent calcites can be expected to precipitate in some generally oxic phreatic settings. Although it might be argued that simultaneous mobilization of Fe^{2+} and Mn^{2+} would extinguish CL luminescence (Fe being widely regarded as a CL-quenching element), the prediction of Mn-activated luminescence in oxic phreatic spars is strengthened by the observations that: 1) Fe/Mn ratios in oxic phreatic calcites could be suppressed by the metastability of Fe oxides relative to Mn oxides (Berner, 1981) at sites of reduction; 2) Fe/Mn ratios in calcites may be further suppressed by selective exclusion of Fe relative to Mn substitutions in the calcite lattice (Veizer, 1983); and 3) recent studies have failed to disclose any critical Fe concentration or Fe/Mn ratio that reliably extinguishes CL luminescence in Mn-bearing calcites (Mason, 1987).

Oxic-Anoxic Transition Zone

The zone of transition between oxygenated and oxygen-free groundwaters is one of great importance, because appreciable concentrations of dissolved Mn^{2+} and Fe^{2+} are restricted to environments lacking dissolved oxygen. In addition, this zone of transition corresponds to a fundamental boundary between CO_2-respiring aerobic and obligate anaerobic bacterial communities that further influence the chemical evolution of meteoric phreatic waters.

At any instant in time, the most distal limit of the oxic phreatic environment (dissolved $O_2 \geq 10^{-6}$ moles/liter; from Berner, 1981) corresponds to an idealized surface beneath which reducing environments are established. The position of the oxic-anoxic interface migrates as a redox front in response to changes in rates of groundwater recharge and flow velocity, and to changes in the rate of oxygen consumption in the proximal reaches of the flow system. Because these parameters are influenced by seasonality and more long-term climatic variations, sites straddling the oxic-anoxic interface are likely to

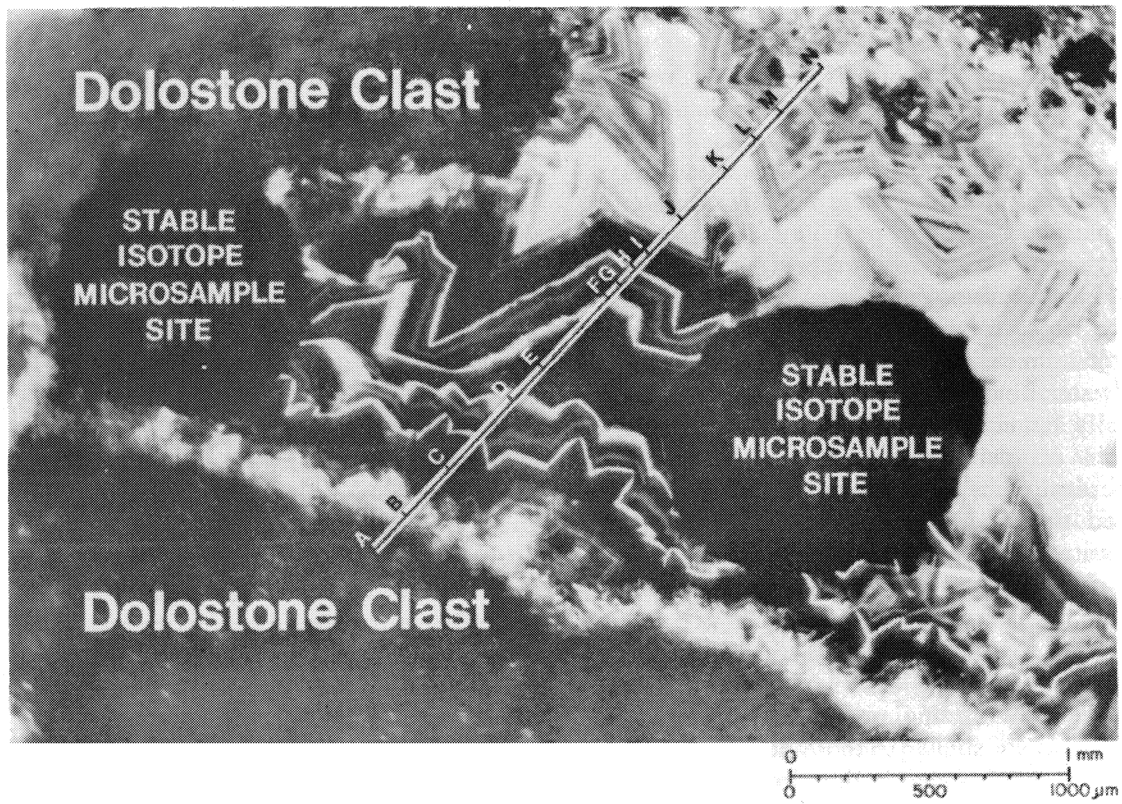
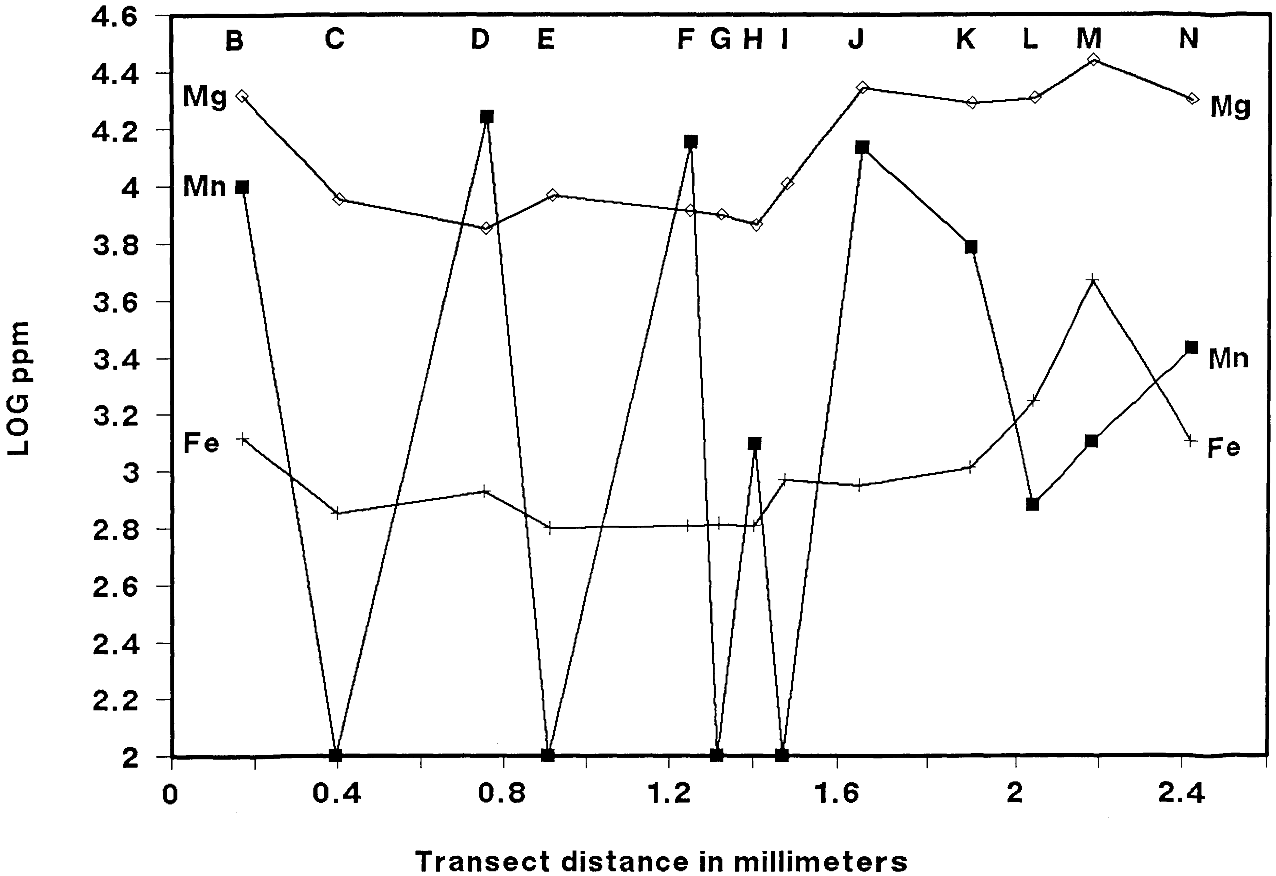


Figure 7. Electron microprobe transect across a pore-filling sequence of zoned blocky calcite from the Silurian Scotch Grove Fm in the southern boundary fault to the Silver Creek Graben (see description for STOP 6 in this guidebook). The left panel is a cathodoluminescence photomicrograph of the transect. Letters A through N mark sites of individual microanalyses, with scale to the lower right. The right panel is a plot of elemental abundances of iron, manganese, and magnesium in parts per million (weight) along the transect. From Ludvigson (1988).

experience oscillatory transitions between oxidizing and reducing environments.

The transition zone is repeatedly traversed by overlapping migrations of alternating oxidation and reduction fronts, with the zone's bounding limits corresponding to the most proximal migration of successive reducing fronts, and the most distal migration of successive oxidation fronts (Fig. 6a). If the position of the zone of oxic-anoxic oscillation is relatively static through time, that area becomes a sink for the diagenetic concentration of Fe and Mn-bearing authigenic minerals, as a consequence of local cycling between oxidized and reduced species. As oxidation fronts migrate downflow, concentration occurs in newly-forming oxide phases. Conversely, as reduction fronts migrate upflow, concentration may occur both by lattice substitutions in, and adsorption on the surfaces of newly-forming phreatic calcites (McBride, 1979). In time, these countervailing processes lead to the development of a spatially restricted transition zone that is capable of producing calcites with extreme manganese enrichments (up to several weight percent). Berner (1981, p. 362) noted that ferric oxides are metastable phases that may persist in anoxic environments, whereas Mn oxides do not. Consequently, the migration of reduction fronts through this environment leads to the preferential release of Mn^{2+} from the



concentrated oxide phases, thereby making Mn available for substitution in concurrently precipitating calcites.

Oscillatory migration of redox fronts through the oxic-anoxic transitional zone leads to the development of oscillatory Mn-zoning within indigenous calcite cements, with Mn-rich zones recording the passing of reducing fronts, and Mn-deficient zones recording periods of oxidation. Calcite spars with alternating darkly-colored Mn-rich zones and clear Mn-deficient zones have been described from rhizoliths in poorly-drained soils (Mount and Cohen, 1984), and have been observed in the PRFZ. These calcite cements are proposed to have precipitated in the zone of oxic-anoxic transition. Mount and Cohen (1984) noted that the oscillatory Mn-zoning effect in calcite cements may be further enhanced by the presence of Mn-rich terrigenous sediments in the depositional system. Groundwater flow systems in carbonate rock sequences that lack significant terrigenous clastic components would be expected to produce Mn cycling effects with lesser magnitudes.

The "manganese cycling effect" described above is the diagnostic characteristic of the oxic-anoxic transition zone, and can be recognized with CL petrography by the occurrence of calcite spars with finely-developed alternations between brightly luminescent and nonluminescent growth zones (Fig. 7).

Anoxic (Nonsulfidic) Phreatic Zone

Meteoric groundwater flow systems, by volume, are mostly comprised of anoxic fluids. The larger physical dimensions occupied by anoxic phreatic environments establish the potential for the generation of indigenous calcite cements that incorporate elevated concentrations of trace elements released from the reduction of resident oxide phases (manganese and iron), and also elements that are released from the dissolution of marine carbonate components in the rock matrix (magnesium and strontium).

The establishment of stable reducing environments in the anoxic phreatic zone leads to the permanent development of elevated Mn^{2+} and Fe^{2+} concentrations in the remainder of the flow system. Although positive downflow concentration gradients throughout the anoxic nonsulfidic zone might be generally expected for these elements, Mn^{2+} concentrations may actually be attenuated in the proximal margins of the zone. The extreme manganese enrichments and the oscillatory manganese cycling effect operating in the overlying oxic-anoxic transition zone may combine to export elevated Mn^{2+} concentrations to adjacent anoxic environments, and lead to the selective entrapment of large concentrations of Mn in the most proximal anoxic phreatic calcites. This will be the case if rates of Mn consumption by substitution and adsorption on calcite exceeds rates of Mn^{2+} generation in the anoxic zone, where oxide phases have been depleted by exposure to prolonged reducing environments (Figs. 6b, c).

The metastability of ferric oxide phases, as contrasted with the instability of manganese oxides in anoxic environments, could lead to generally more conservative behavior of dissolved Fe^{2+} throughout the oxic-anoxic transitional and anoxic nonsulfidic zones. Increasing downflow concentration of Fe^{2+} should be expected throughout the anoxic nonsulfidic zone, but major changes in the Fe/Mn ratios of calcites precipitated from anoxic nonsulfidic meteoric waters should be controlled principally by larger changes in Mn contents. Calcites precipitated under the influence of these phenomena would contain elevated concentrations of Fe and Mn, but with more variable Mn contents. Using CL petrography, these spars would be expected to be weakly to strongly luminescent, with oscillatory zoning of luminescent intensity controlled by variations in Mn contents. Cements fitting this description have been recognized in the PRFZ (Fig. 7, see sites B, J, K, L, M, and N), and are interpreted to have been deposited in the proximal margin of the anoxic nonsulfidic environment. Anoxic environments in the distal portions of carbonate aquifers that contain abundant sources of Mn may form CL nonluminescent calcites, where Mn contents greater than $\approx$50,000 ppm cause the self-quenching of luminescence (Mason, 1987).

The flow paths traversed by meteoric groundwaters as they pass through expansive anoxic realms are far longer than those in overlying oxidizing environments. This simple physical relationship, rather than any intrinsic character, dictates that geochemical signals derived from the dissolution of rock matrix carbonates (diagenetic systems with low water/rock ratios) are best developed in anoxic phreatic environments. These signals include increasing $\delta^{13}C$ and $\delta^{18}O$ values, and increasing Mg/Ca and Sr/Ca ratios (Figs. 6b, c). "Rock-dissolution" signals are maximized at the most distal portions of the flow system. The stoichiometry of rock-forming carbonate minerals assumes major significance in this portion of the flow system, and the major difference between calcites formed in dolostone terranes (Fig. 6b) and limestone terranes (Fig. 6c) is seen here. Dissolution of dolomite provides an abundant source of Mg^{2+}, and much greater Mg/Ca ratios are developed in the distal portions of dolostone aquifers than those in limestone aquifers.

Elevated Sr/Ca ratios also develop in the distal realm. Freshly-exposed limestones containing marine aragonite components should have the potential to generate phreatic groundwaters with the highest Sr/Ca ratios ($> 10^{-2}$), although this potential would be expected to progressively decrease through geologic time as the rock is completely stabilized to LMC (Kinsman, 1969, p. 505).

Anoxic nonsulfidic environments that contain dissolved or solid concentrations of organic carbon are populated by anaerobic methanogenic bacteria. Calcites that are precipitated in environments that are hosts to methanic fermentation may record two <u>possible</u> geochemical signatures. Biogenic CH_4 is strongly depleted in ^{13}C, and methanogenesis in environments with low water/rock ratios ($\approx$ closed system) produces strong residual ^{13}C-enrichments in the dissolved carbonate reservoir. Thus, growth zones in phreatic calcite cements with anomalously heavy carbon isotopic ratios probably record the local establishment of methanic environments (Fig. 6). Methanic fermentation results in extremely reducing conditions (Berner, 1981), and ferroan and manganoan carbonates can be precipitated in this environment, provided that sufficient sources of dissolved Fe^{2+} and Mn^{2+} are present and dissolved SO_4^{2-} is absent. Siderite and rhodochrosite are common diagenetic products in methanic environments that develop in terrigenous clastic terranes, where abundant sources of Fe and Mn are present. Local sites of methanogenesis that develop within the anoxic phreatic zone in carbonate aquifers, however, occur within a realm that probably has already attained the maximum possible concentrations of dissolved Fe^{2+} and Mn^{2+} (Figs. 6b, c). Further concentrations in carbonate terranes are probably limited by the availability of reducible oxide phases, and methanic calcite cements with higher Fe and Mn enrichments are unlikely in this setting (Figs. 6b, c).

Sulfidic Phreatic Zone

Anaerobic sulfate-reducing bacteria are energetically more efficient than methanogens, and the sulfidic zone develops wherever environments are favorable for their mutual habitation. The distribution of sulfate-reducers is limited, however, by the required presence of both dissolved sulfate and organic carbon. Dissolved SO_4^{2-} concentrations sufficient to support bacterial sulfate reduction are frequently developed in meteoric groundwater flow systems, particularly towards the distal portions (Champ et al., 1979). Increasing downflow concentration of SO_4^{2-} develops in aquifer systems by oxidation of sedimentary sulfides in the oxidizing portions of the flow system, and in some terranes by dissolution of evaporite minerals. The production of dissolved H_2S gas by sulfate-reducing bacteria has dramatic consequences for the Fe^{2+}/Mn^{2+} ratios in anoxic groundwaters. At sedimentary temperatures and pressures, Fe^{2+} has a much greater affinity for dissolved sulfide than Mn^{2+} (Berner, 1981), and the precipitation of iron sulfide phases results in the preferential removal of Fe^{2+} relative to Mn^{2+} (Figs. 6b, c).

The diagnostic characteristics of calcite cements formed in the sulfidic zone are the presence of low Fe/Mn ratios, coprecipitated sulfide minerals, and depleted carbon isotopic ratios. The predominance of Mn over Fe substitutions leads to the expectation that with the use of CL petrography, calcite cements from sulfidic environments should be uniformly luminescent. Brightly luminescent calcites with coprecipitated sulfide phases have been observed along the PRFZ, and are interpreted to have formed in sulfidic phreatic waters (eg. the sulfide deposit at MMQ, see Ludvigson, 1988, and papers by Garvin and Ludvigson, and Spry and Kutz elsewhere in this guidebook).

The position of the zone of sulfide generation is dependent on the position of the oxic-anoxic interface and the concentration gradient of SO_4^{2-} in the aquifer system. The development of a sulfidic environment in the distal portions of a carbonate aquifer would be expected to reverse the proximal to distal trend toward higher ^{13}C values in newly-forming calcites, particularly in settings where voluminous biogenic H_2S production leads to a decrease in Fe^{2+} concentrations. Other geochemical signals (oxygen isotopic ratios, Mg/Ca and Sr/Ca ratios) are not intrinsically related to sulfate reduction, and are controlled by the position of the anoxic sulfidic zone within the flow system. Sulfidic environments may be developed in more proximal anoxic settings in carbonate aquifers that contain interstratified evaporites. In most other settings in carbonate rocks, the distal position of the sulfidic zone dictates the coincidence of "rock dissolution" signals (increasing $\delta^{13}C$ and $\delta^{18}O$ contents, and Mg/Ca and Sr/Ca ratios; Figs. 6b, c).

HYDROLOGIC AND TECTONIC IMPLICATIONS

The growth of fracture-filling minerals in the PRFZ preserves a record of the geochemical evolution of ancient meteoric groundwater flow systems. These fluid flow systems existed beneath ancient erosion surfaces whose development was coeval with episodes of active faulting along the PRFZ. Regional paleogeographic reconstructions of syn-deformational erosion surfaces recorded by major Paleozoic unconformities along the PRFZ (Bunker et al., 1985; 1988) can be used to make predictions about the regional geometry of subsurface fluid flow systems in which fracture-filling minerals were precipitated. The predictions can be compared with environmental interpretations derived from petrographic and geochemical data on fracture-filling minerals to evaluate their mutual compatibility.

Published regional stratigraphic syntheses and field studies along the fault zone indicate that the development of two major Paleozoic unconformities can be confidently predicted to have temporally overlapped with periods of tectonic deformation along the PRFZ. These are the unconformities separating Sloss' (1963) Tippecanoe and Kaskaskia cratonic sequences (the unconformity at the base of the Devonian System in the Midcontinent), and separating his Kaskaskia and Absaroka sequences (the unconformity at the base of the Pennsylvanian System). For a discussion of tectonism associated with the sub-Devonian unconformity, see the paper by Bunker in this guidebook. A synthesis of the ancient regional fluid flow system and fracture-filling mineralization associated with the development of the sub-Pennsylvanian unconformity is presented below.

The Fluid Flow System Beneath the Sub-Absaroka Erosion Surface

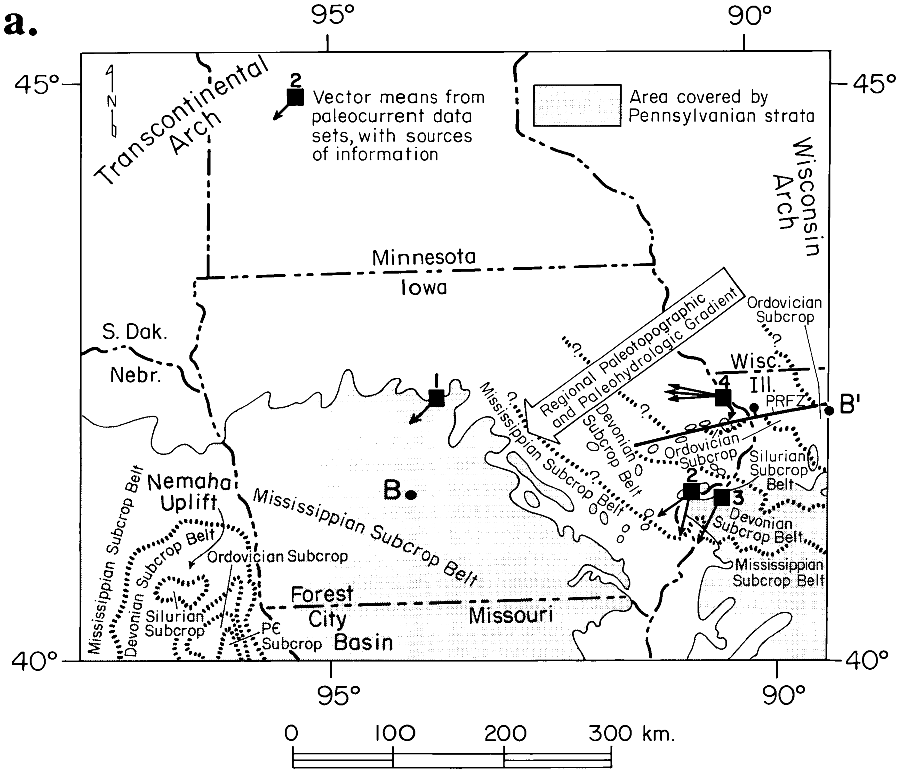
The paleogeography, paleotectonics, and paleorelief of the sub-Absaroka surface in the area of the PRFZ have been discussed by Bunker et al. (1985; 1988). The North American cratonic interior was subjected to a dramatic regional structural reorganization during the period of erosion preceding the deposition of Pennsylvanian rocks. The centers of the mid-Paleozoic North Kansas and East-Central Iowa basins were uplifted relative to adjoining areas (see introduction to Part 1 of this guidebook). A paleogeologic map of the sub-Absaroka surface reveals that rock sequences on the order of hundreds of meters in thickness were eroded from uplifted former basinal areas forming newer positive structural elements, including the Nemaha Uplift and Wisconsin Arch (Fig. 8). To the immediate south of the PRFZ, active uplift of the Savanna-Sabula Anticlinal System is indicated by subcrop patterns on the sub-Absaroka surface (Fig. 8). The Forest City Basin developed as an asymmetric depocenter with a north-south trending axis bordering a faulted mutual boundary with the Nemaha Uplift (Fig. 8).

Analysis of regional paleotopographic trends on the sub-Absaroka erosion surface in the area of the PRFZ is complicated by the development of local paleorelief on the order of 15 m, and because stratigraphic relationships between the Pennsylvanian rocks straddling the fault zone are poorly known (Ludvigson, 1985). Paleocurrent data from fluvial deposits in the oldest Pennsylvanian rocks of the Quad Cities area (Caseyville Formation; Isbell, 1985; Ludvigson and Swett, 1987) record enigmatic northward draining paleoslopes that conflict with the southwesterly paleoslopes indicated from younger Pennsylvanian rocks in the Upper Mississippi Valley region. These data may record local structural or topographic influences that have not been recognized elsewhere in the region. Vector means of paleocurrent data sets from Middle Pennsylvanian fluvial deposits in the Iowa area conform to the southwestward-sloping continental surface suggested by many other studies (Fig. 8). The configuration of the ancient land surface would thus have directed regional subsurface fluid flow from an area of high gravitational potential along a recharge boundary possibly on the axis of the Wisconsin Arch to a poorly-defined discharge boundary somewhere to the southwest.

Figure 8b shows a cross-section illustrating the reconstructed geology and paleoslope of the sub-Absaroka surface along a line between points B and B'. The cross-section traverses an approximately 415 km distance from near the axis

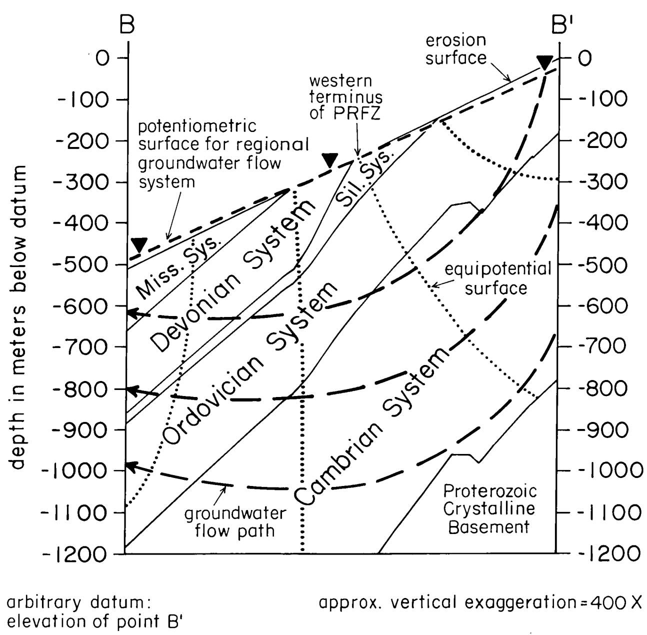
Figure 8. Paleohydrology of the groundwater flow system beneath the ancient continental surface buried by Sloss' (1963) Absaroka Sequence. a. Paleogeologic map of the sub-Absaroka surface in the region of the PRFZ (after Bunker et al., 1985; 1988). Sources of paleocurrent data are: 1, Lemish et al., 1981; 2, Fitzgerald, 1977; 3, Isbell, 1985; 4, Ludvigson, unpublished data, 1987. b. Westward-flowing fluid migration in the groundwater flow system beneath the ancient sub-Absaroka erosion surface. From Ludvigson (1988).

a.



95°
90°
45°
40°
Transcontinental Arch
Vector means from paleocurrent data sets, with sources of information
Area covered by Pennsylvanian strata
Wisconsin Arch
Minnesota
Iowa
S. Dak.
Nebr.
Regional Paleotopographic and Paleohydrologic Gradient
Ordovician Subcrop
Wisc.
Ill.
PRFZ
B'
Mississippian Subcrop Belt
Devonian Subcrop Belt
Ordovician Subcrop
Silurian Subcrop Belt
Devonian Subcrop Belt
Mississippian Subcrop Belt
Nemaha Uplift
B
Mississippian Subcrop Belt
Devonian Subcrop Belt
Ordovician Subcrop
Silurian Subcrop
PЄ Subcrop
Forest City Basin
Missouri
0 100 200 300 km.

b.



B
B'
erosion surface
western terminus of PRFZ
potentiometric surface for regional groundwater flow system
Miss. Sys.
Devonian System
Sil. Sys.
Ordovician System
Cambrian System
equipotential surface
groundwater flow path
Proterozoic Crystalline Basement
depth in meters below datum
0
-100
-200
-300
-400
-500
-600
-700
-800
-900
-1000
-1100
-1200
arbitrary datum: elevation of point B'
approx. vertical exaggeration = 400 X

of the Wisconsin Arch (recharge boundary to the regional groundwater flow system) to an arbitrary constant flux boundary (discharge boundary not specified) along a line bordering the south side of the PRFZ. A first-order approximation of the gradient of the paleoslope (1.24 m/km) was calculated from the rate of northeastward stratigraphic thinning of Lower and Middle Pennsylvanian strata in the Forest City Basin (Bunker et al., 1988) along a line paralleling the cross-section line. The calculated total paleotopographic relief along line B-B' is greater than 500 m. Paleogeologic reconstructions were based on the position of geologic contacts in Figure 8, and published and unpublished isopach and subsurface structure maps on file at the Geological Survey Bureau in Iowa City.

Major confining units in portions of the Devonian and Ordovician systems assure that a southwesterly-flowing confined regional flow system was established with a geometric configuration something like that shown in Figure 8. The area along the PRFZ was located at the proximal margin of this flow system, and Silurian and Devonian carbonate rocks exposed at the ancient erosion surface would be expected to host unconfined local flow systems controlled by the local topographic configuration.

Extensive paleokarst was developed in carbonate rocks exposed at the sub-Absaroka erosion surface, as a consequence of humid tropical weathering environments and local paleotopographic relief in excess of 30 m (Bunker et al., 1985; Ludvigson, 1985). The presence of authentic Pennsylvanian speleothems in paleokarsts up to 6 m below the general level of the sub-Pennsylvanian erosion surface, noted from exposures in the Quad Cities area (R. C. Anderson, 1988, pers. comm.) demonstrates that vertically extensive vadose zones were developed in the proximal reaches of the flow system. These relationships lead to the expectation that fracture-filling minerals in the PRFZ that were precipitated from groundwaters circulating beneath the sub-Absaroka erosion surface would record geochemical environments characteristic of the proximal margin of the flow system. These could include vadose environments and meteoric phreatic environments straddling the oxic-anoxic boundary.

Calcite Mineralization Beneath the Sub-Absaroka Erosion Surface

The predictions formulated in the preceding discussion apparently are confirmed by the recognition of a distinctive generation of fracture-filling and replacive mineralization related to the development of the sub-Absaroka unconformity along the PRFZ. Integrated field, petrographic, and geochemical data have been presented by Ludvigson (1988) to support arguments that vein-filling calcites in faults from STOPS 4, 6, and Sunday STOP 2 on this field conference were precipitated from phreatic oxic and anoxic nonsulfidic groundwaters that infiltrated from the ancient sub-Absaroka erosion surface.

A distinctive compositional zonation in sub-Absaroka calcites is well documented from faults at the Silver Creek Graben (Fig. 7, STOP 6). Early CL luminescent spars are succeeded by alternating luminescent and nonluminescent spars, and progressively lower Mg and Fe contents are recorded by the growth sequence. This zonation is interpreted to record a transition from an early anoxic nonsulfidic environment, through the oxic-anoxic transition zone, with latest deposition in oxidizing environments. Based on the CL microstratigraphy observed in coexisting calcites and Fe oxides in other localities, it appears that the "manganese cycling effect" is universally present (Fig. 9), and that oscillating luminescent-nonluminescent spars are bracketed either by uniformly luminescent or nonluminescent spars. These sequences evidently record groundwater evolution traversing the oxic zone-transitional zone-anoxic zone boundaries. Although all of these indicate calcite precipitation in the proximal reaches of the groundwater flow system, differing evolutionary histories are recorded at different localities (Fig. 9).

The Role of Fluids in the Tectonic Evolution of Cratonic Faults

Regional stratigraphic syntheses of the Paleozoic history of the North American craton have previously established an empirical relationship between periods of widespread continental exposure and periods of epeirogenic tectonism (Sloss, 1963; Bunker et al., 1988). A rationale is advanced here for the hypothesis that cratonic fault systems are prone to reactivation and increased rates of strain during periods of continental emergence, using the PRFZ as a general example. This rationale integrates perspectives on the crustal evolution of the PRFZ, the paleohydrology of Paleozoic deformational episodes along the PRFZ, and the body of published evidence that fluid-rock interactions play a significant role in the triggering of seismogenic fault rupture.

Crustal Evolution

Episodic Paleozoic deformation along the PRFZ implies that the structures exposed at the land surface overlie a zone of weakened Proterozoic crust that was subject to renewed shear failure during periods of epeirogenic tectonism. Although no samples of basement rocks have been retrieved from the PRFZ, regional studies of the Precambrian basement in the Midcontinent region (Nelson and DePaolo, 1985) have shown that the continental crust beneath the PRFZ was accreted to the North American protocontinent during the 1.9 to 1.8 Ga Penokean Orogeny, with later anorogenic plutonism occurring at 1.4 Ga (including the UPH granite of Northern Illinois discussed in papers by Anderson, and by Ludvigson and Millen elsewhere in this guidebook). The angular deviation between the trend of the PRFZ and the regional northeast-southwest trend of the Penokean Orogen in Iowa ranges between 30^{o} and 60^{o}, and it is probable that the PRFZ is a rejuvenated Proterozoic fault system (see paper by Anderson in this guidebook).

The brittle deformation of Paleozoic rocks in the PRFZ records their passive response to stick-slip fault behavior that almost certainly was triggered in healed faults in the Precambrian basement rocks, at crustal depths of at least several kilometers. Paleohydrologic interpretations of the PRFZ, when viewed from the perspectives gained from studies on the role of aqueous fluids in seismogenic fault mechanisms, suggests that a genetic relationship may exist between regional fluid transport and basement fault reactivation.

Aqueous Fluids and the Strength of the Continental Crust

The exclusive involvement of meteoric pore fluids in the healing of ancient brittle microstructures in the marine sedimentary rocks exposed along the PRFZ could signify a linkage between cratonic fault movement and groundwater migration in the upper continental crust. Costain et al. (1987,1988) have suggested that episodes of increased groundwater recharge that temporarily raise regional water tables may initiate periods of intraplate seismicity by the diffusive transmission of transient increases in fluid pressure to depths of several kilometers in the continental crust. Studies of reservoir-induced seismicity indicate that pore-pressure fronts can migrate through crystalline basement rocks at velocities of 10^1 to 10^2 km/year (Talwani and Acree, 1985), and suggest that surface-controlled hydrogeologic phenomena can cause the migration of pore-pressure perturbations into the Precambrian basement. From a geologic perspective, the transmission of pore-pressure fluctuations from the continental surface into the basement crustal rocks may be viewed as a virtually instantaneous process.

Pore-pressure diffusion apparently plays a dual role in promoting the brittle failure of crustal rocks (Talwani and Acree, 1985). The purely mechanical effects of pore-pressure on crack propagation and the reduction of friction are well known in the geological literature. Recent research advances in rock mechanics have also shown that purely chemical effects of aqueous fluids further lower the strength of rock-forming silicate minerals (Kirby, 1984). Pressure-induced fluid motion through rocks in the continental crust thus can promote a variety of chemical processes that contribute to their further weakening (ibid.). The strength of clay-rich gouge, characteristic of many active fault systems in crystalline rock terranes, is believed to be especially sensitive to variations in pore pressure (Talwani and Acree, 1985).

Factors Controlling the Evolution of Cratonic Faults

Zoback and Zoback (1981) have discussed the phenomenon of multiple reactivations of intraplate fault systems in cratonic continental environments, and related the occurrence to

Figure 9. Photomicrographs of fracture-filling calcites from the Plum River Fault Zone interpreted to have formed beneath the ancient sub-Pennsylvanian (sub-Absaroka) erosion surface.

a. Detrital quartz grains (white) engulfed in fracture-filling calcite with coprecipitated ferric oxides, in brecciated sparsely fossil-moldic dolostone of the Waubeek Member, Silurian Scotch Grove Formation. This is a good example of probable Morrowan (Early Pennsylvanian) siliciclastic detritus infiltrating into the fracture pore network of near-surface rocks beneath the unconformity. See discussion from Sunday STOP 2. Sample NOJQ-1, Plum River Fault Zone at Oxford Junction, Jones County, Iowa. Plane polarized light. Field of view is 12.8 mm.

b. Detrital quartz grains (white) forming geopetal floor, with spherulitic/columnar calcites (gray) filling pore space, and siliciclastic mudstone (black) filling remainder of pore. Dolomite stained by ferric oxides to the upper left and upper right. Note zonation in spherulitic spar marked by dark rings of ferric oxide, in the upper left. Sample NOJQ-1. Plane polarized light. Field of view is 3 mm.

c. Cathodoluminescent micrograph of zoned blocky calcite spar cementing dilational breccia in the Welton Member, Silurian Scotch Grove Formation, at STOP 4. Early nonluminescent zone (oxic phreatic environment) is succeeded by oscillatory luminescent-nonluminescent zones (manganese cycling effect, oxic-anoxic transitional environment). Dolostone clasts are dull gray. Sample MF-3-S1. Field of view is 3.7 mm.

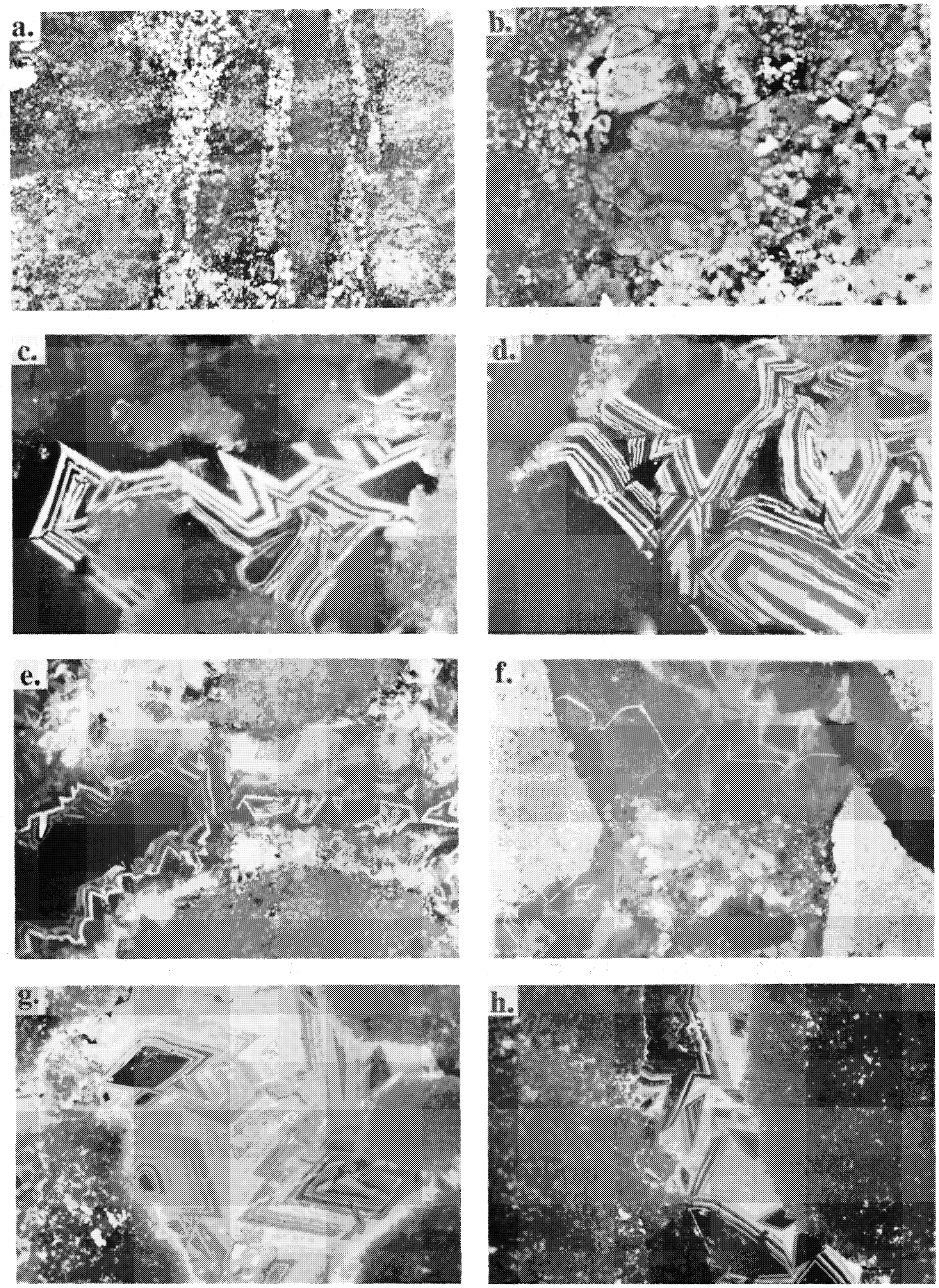
d. Cathodoluminescent micrograph of zoned blocky calcite spar in dilational breccia. Early nonluminescent zone is succeeded by oscillatory alternation between brightly luminescent and nonluminescent spar. Dolostone clasts are dark gray. Sample MF-3-S1. Field of view is 3.7 mm.

e. Cathodoluminescent micrograph of zoned blocky calcite spar cementing breccia in Welton Member of Scotch Grove Formation at south boundary fault to the Silver Creek Graben (STOP 6). Early luminescent zone (anoxic nonsulfidic environment) is succeeded by oscillatory luminescent-nonluminescent growth zones (oxic-anoxic transitional environment) and nonluminescent zone (oxic phreatic environment). Dolostone clasts are dark gray. Sample SCO-SBF-S2. Field of view is 3.7 mm.

f. Cathodoluminescent micrograph of nonluminescent blocky calcite spar (oxic phreatic environment) with a narrow growth band of luminescent spar (oxic-anoxic transition) cementing breccia in the Waubeek Member of the Scotch Grove Formation at the north boundary fault to the Silver Creek Graben (STOP 6). The calcite spar engulfs a geopetal accumulation of internal detritus including ferric oxides (black) and corroded dolomite rhombs (white). Ferric oxide cement (black) coats a dolostone clast (white) to the right. Sample SCO-NBF. Field of view is 3.7 mm.

g. Cathodoluminescent micrograph of zoned blocky calcite spar filling a veinlet in the Welton Member of the Scotch Grove Formation. Early nonluminescent spar (oxic phreatic environment) is succeeded by oscillations between luminescent and nonluminescent spar (oxic-anoxic transition) and luminescent spar (anoxic nonsulfidic environment). Dolostone wallrock is dark gray. Sample BCR-3-S2. Field of view is 3.7 mm.

h. Cathodoluminescent micrograph of zoned vein-filling calcite. Early nonluminescent spar (oxic phreatic environment) is followed by alternating luminescent-nonluminescent spar (oxic-anoxic transition) and luminescent spar (anoxic nonsulfidic environment). Dolostone wallrock is dark gray. Sample BCR-3-S2. Field of view is 3.7 mm.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long-term reorientations of the stress field in the shallow crust. According to their hypothesis, ancient continental fault systems are susceptible to reactivation only when the stress field is properly oriented with respect to pre-existing structures, and stress reorientations resulting from plate tectonic interactions may initiate brittle failure along old planes of weakness in the continental crust. The hypothesis of Zoback and Zoback (1981) relates periods of fault reactivation to changes in the magnitude of shear stress along ancient faults over geologic time. As noted by Costain et al. (1987, 1988), however, hydraulic effects reducing the shear strength of crustal rocks may trigger short-term episodes of renewed fault movement. Thus the evolution of cratonic faults could be controlled by the interactions of two independent processes affecting: 1) the concentration of shear stress in the basement rocks along a fault, and 2) the shear strength of those rocks.

The preceding considerations lead to the conclusion that global plate tectonic interactions operating on time scales of perhaps 10^5 to 10^7 years probably control the gross timing and geometry of epeirogenic deformation in the craton by inducing changes in the magnitude and orientation of the regional stress field. Climatically and eustatically-controlled hydrogeologic processes operating on time scales of perhaps 10^0 to 10^6 years probably control temporal variations in the heterogeneous strength of the continental crust, and may determine which of the properly-oriented paleostructures in the basement will ultimately be reactivated as cratonic faults, and when that failure occurs. Temporal variations in the hydraulically-influenced shear strength of stressed crustal rocks are proposed to operate as the specific triggering mechanism for the reactivation of cratonic faults.

The exclusive association of meteoric waters with multiple Paleozoic deformation/healing sequences in the near-surface rocks from the PRFZ is noteworthy because the surrounding region was flooded by epeiric seas over half of the approximately 158 million year duration of the Middle Ordovician to Pennsylvanian interval (B. Witzke, 1988, pers. comm.). The apparent temporal relationship between deformation and continental exposure could be fortuitous, but is consistent with the hypothesis that the sedimentary history of deposition and erosion along the PRFZ influenced its tectonic history, and that surface-controlled hydrologic processes in the continental crust could serve as a major factor in determining the specific timing of episodes of fault reactivation.

The "hydraulic-triggering" hypothesis outlined above predicts that the development of major cratonic erosional unconformities during periods of global eustatic drawdown should coincide with major periods of basement fault reactivation in the interiors of continents undergoing epeirogenic deformation. Geometric relationships between cratonic faults and the directions of regional fluid transport may assume special significance during these erosion cycles, and may help to further explain why certain preexisting basement structures have been especially prone to failure while others were not. The regional westward gravity-driven fluid migration proposed for the sub-Absaroka erosion surface (Fig. 8) exemplifies this relationship. Since the PRFZ trends at a relatively small acute angle to the regional paleohydrologic gradient proposed for this episode, it appears that zones of enhanced fracture permeability proximal to the fault would have been favorably oriented for conducting rapid conduit fluid flow. Given that equipotential surfaces in the flow system were approximately perpendicular to the fault plane in both map and cross-section view (Fig. 8), it is apparent that the proposed geologic and hydrologic configuration probably focused deeply circulating fluid flow through the Precambrian basement rocks along the PRFZ. Periods of fault-parallel fluid flow would be expected to increase the access of migrating fluids to deeper crustal levels than periods of fluid stagnation or cross-fault flow. Thus, the "hydraulic-triggering" effect, resulting from the fluid-induced weakening of the crust, would be expected to be most effective during regional fault-parallel fluid migration.

At the suggested velocities for pore-pressure diffusion in the continental crust, the geohydrologic consequences of even minor erosional episodes on the craton (durations on the order of 10^4 years) could conceivably trigger episodes of fault seismicity by temporarily lowering the strength of the continental crust below a critical threshold at which failure occurs, and an increment of stress-relieving fault strain is initiated.

Marine transgressions resulting from major eustatic sea level rises on the order of hundreds of meters could produce hydrostatic pressures of greater magnitudes than those that result from gravity-driven meteoric groundwater flow systems. In addition, the isostatic effects of crustal loading-unloading by sea level changes in epeiric seaways should also be considered. Isostatic changes following the retreat of interior seaways could also help trigger crustal readjustments during the intervals of continental exposure. Nevertheless, the capacity of regional gravity-driven groundwater flow systems to induce fluid flow through continental basement rocks, and deliver short-term fluctuations (seasonal and climatic variations) in hydrostatic pressure through the crust by pore-pressure diffusion, suggests that periods of continental exposure are more likely to have corresponded with periods of hydraulically-reduced shear strength in basement rocks. For this reason, erosional cycles leading to the development of cratonic unconformities may have coincided with the failure of ancient fault zones that were located in the appropriate stress fields. Fracture-filling calcites in the PRFZ formed in meteoric groundwaters; the apparent absence of marine cements further suggests that primary movements along the PRFZ occurred during periods of continental exposure.

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PART 3

Field Guide

FIELD TRIP STOPS:

SATURDAY	1	2	3	4	5	6		
SUNDAY							**1**	**2**
Pennsylvanian								
Spoon-Abbot fms					X			
Caseyville Fm					X			X
Devonian								
Cedar Valley Gp						X		
Wapsipinicon Gp						X		
Silurian								
Gower Fm								
Scotch Grove Fm	X			X		X		X
Hopkinton Fm	X			X				
Blanding Fm							X	
Tete des Morts Fm							X	
Mosalem Fm				X			X	
Ordovician								
Maquoketa Fm		X	X	X	X		X	
Dubuque Fm		X					X	
Wise Lake Fm		X						

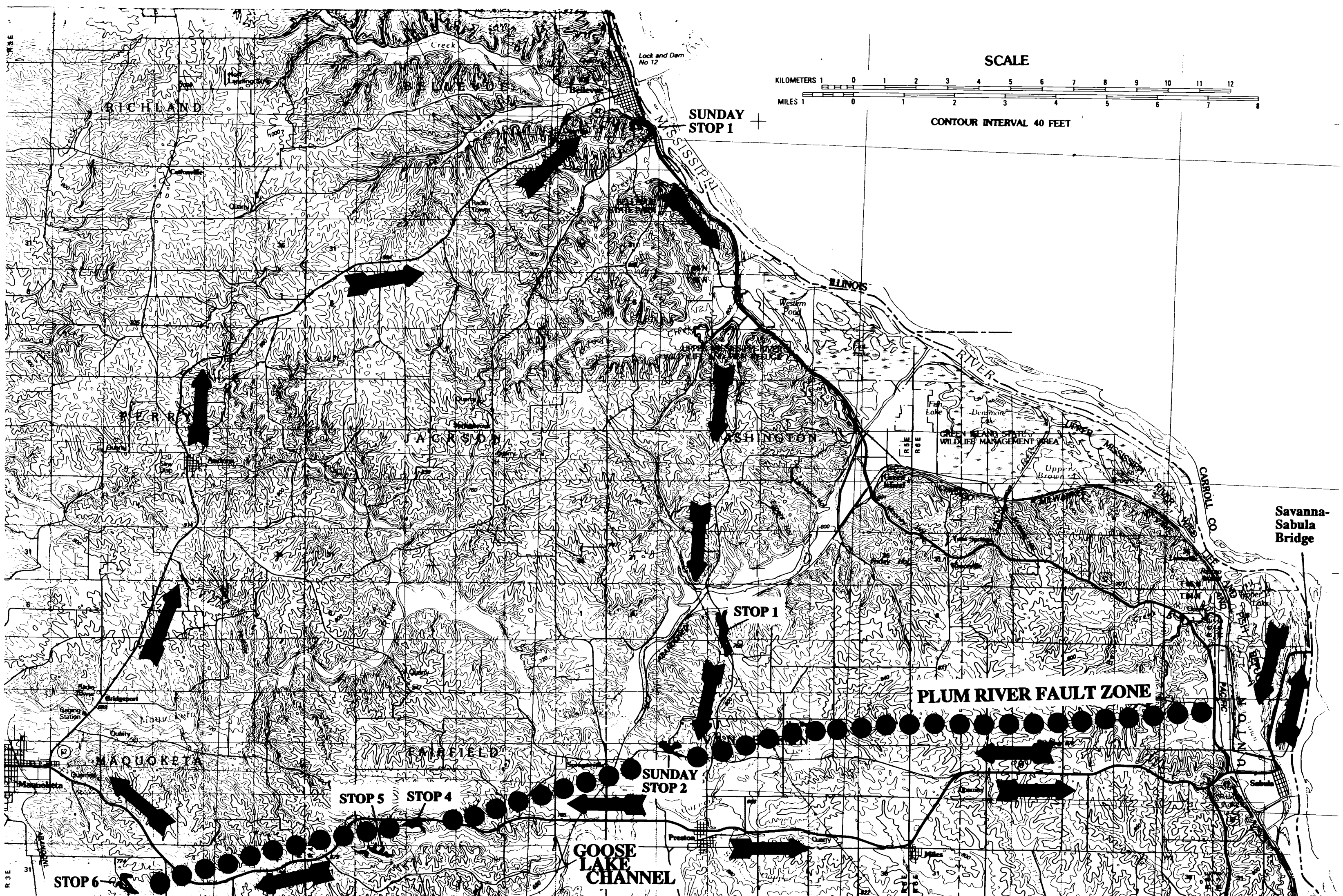

Figure 1. Location of field trip route and stops in Jackson County, Iowa. From the Jackson County 1:100,000-scale topographic map.

FIELD TRIP GUIDE AND STOP DESCRIPTIONS

The first three field trip stops on Saturday morning are intended to aquaint you with the stratigraphy and depositional history of the Ordovician and Silurian rocks exposed along the Plum River Fault Zone (PRFZ) in this area. If you take the time to closely examine these successions and note their lithic character, the following stops will be much more rewarding for you. Later in the afternoon, faulted sections of these same rock units will be examined, along with exposures of Devonian and Pennsylvanian rocks. The field trip route through portions of Jackson County, Iowa and Carroll County, Illinois are shown on Figures 1 and 2.

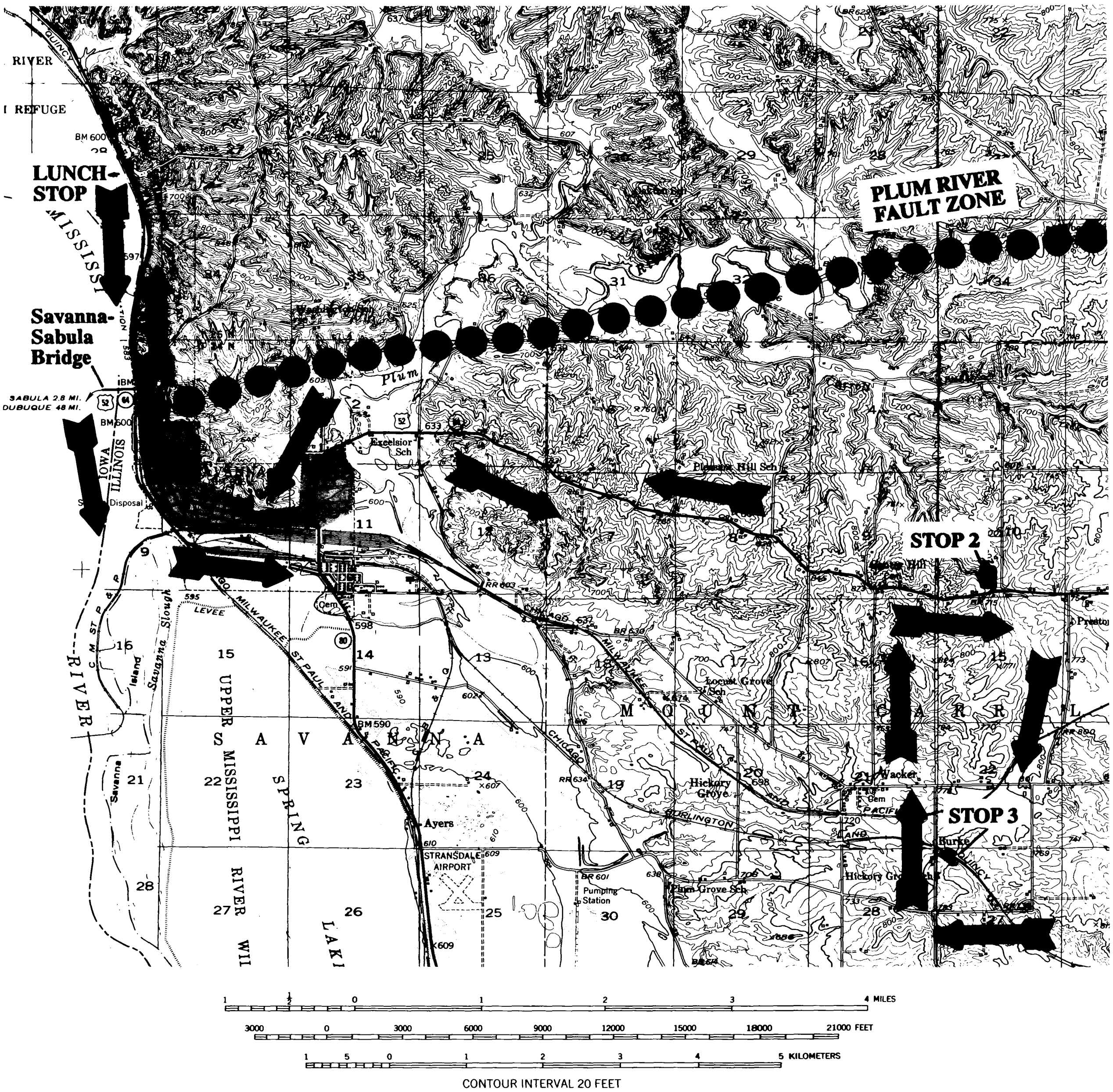

Figure 2. Location of field trip route and stops in Carroll County, Illinois. From the Savanna Illinois-Iowa 15 minute Quadrangle.

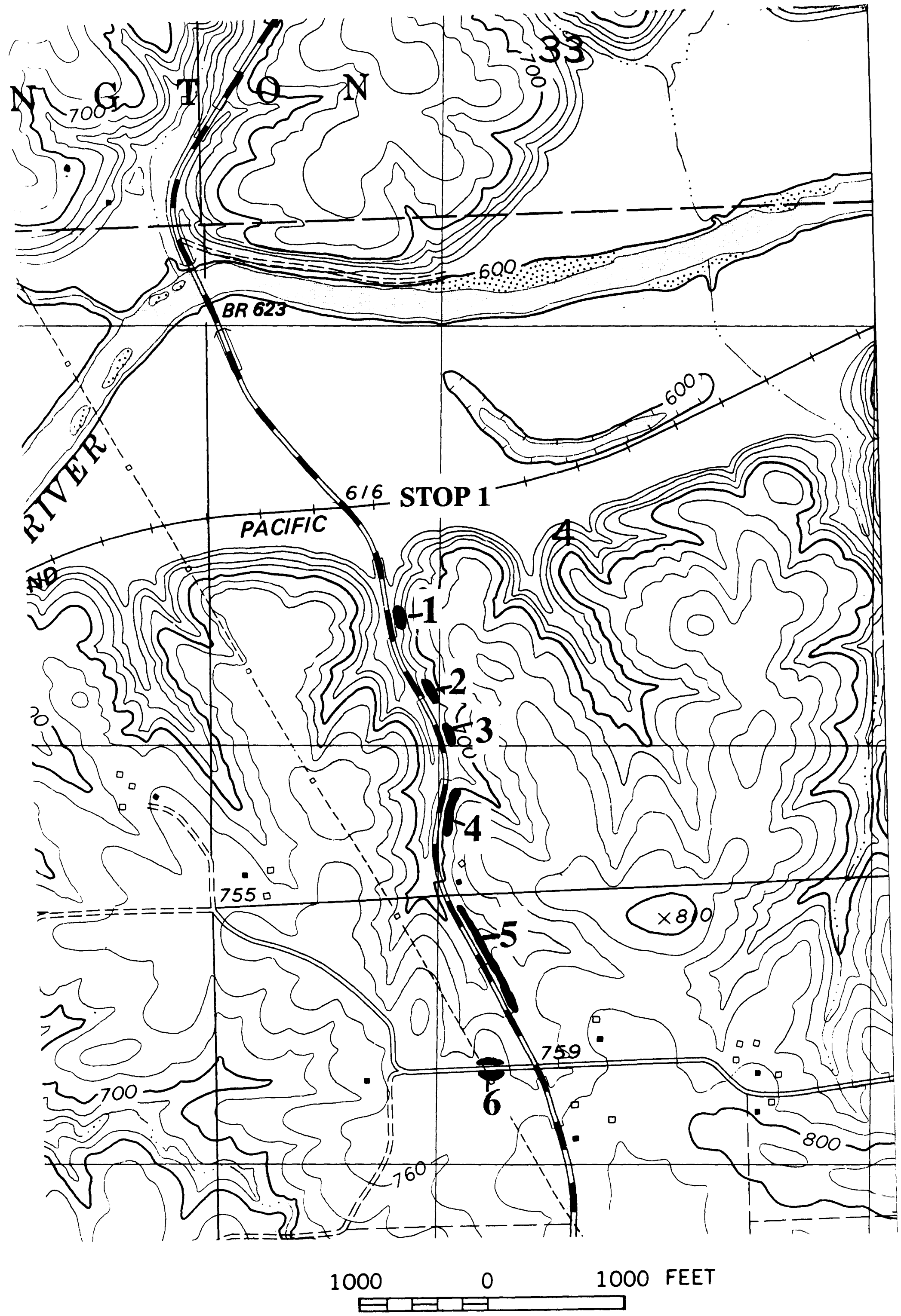

Figure 3. Location of field trip STOP 1. Numbers 1 to 6 refer to localities with measured sections in the following figure. From the Preston, Iowa 7.5 minute Quadrangle.

SATURDAY

STOP 1. The Silurian Hopkinton and Scotch Grove formations.

Greg Ludvigson and Brian Witzke

At this stop we will walk up section through a series of roadcuts to study the stratigraphy, paleontology, and depositional history of the Lower Silurian rocks exposed to the north of the PRFZ. The following two stops will examine Ordovician successions to the south of the structure. Basal Silurian rocks that have been faulted out of view along the PRFZ will be examined on Sunday morning.

Figure 3 shows the six different localities that were used to construct the full stratigraphic section shown in Figure 4. The Hopkinton and Scotch Grove formations, exposed here along the south bluff line of the Maquoketa River, are the same rock units so spectacularly exposed in the cliffs at Mississippi Palisades State Park in Illinois, but here they can be viewed at leisure on slopes that are manageable for a group like this. The entire exposed sequence consists of open marine carbonates that were deposited in subtidal environments. Markes Johnson's paper in this guidebook discusses the depth-related succession of benthic communities in these and correlative strata from other cratonic basins. The main point to remember is that the vertical changes in benthic communities were principally controlled by eustatic changes in sea level, and that synchronous deepening and shallowing trends have been recognized in Silurian cratonic basins on several different continents. It is also worth noting that these rocks were deposited in consistently deeper water than correlative strata from the Michigan and Williston basins. The section we are walking through accumulated near the axial center of the East-Central Iowa Basin, a mid-Paleozoic structural feature whose history is briefly outlined in the introductory section to Part 1. A short description of the section, noting some of the more prominent features follows. A more thorough accounting of lithic and paleotological features is included in the section description.

At locality 1, the Sweeney Member of the Hopkinton is exposed in some large rotated blocks on the steep hillslope. Note that regardless of the position of these blocks, some of the silicified laminar stromatoporoids in these dolomites have been overturned. The dominant megafossils in the Sweeney are tabluate colonial corals and stromatoporoids. The <u>Stricklandia lens progressa</u> (loc. 2) that we collected from this unit (Fig. 4) demonstrates that this "deep water" indicator occurs interstratified with associations of corals and stromatoporoids in the Sweeney Member. Disarticulated echinoderm debris, preserved as skeletal molds and dolomite-replaced unit crystals, is a major skeletal component throughout the section. Nodular to bedded cherts are also a prominent feature.

At locality 2, the coral-stromatoporoid community is replaced upward by dolostones with conspicuous concentrations of the brachiopod <u>Pentamerus oblongus</u>, which are characteristic of the Marcus Member. Wackestone-packstone fabrics with reworked shells are developed in this unit, although some pentamerids are preserved in life position.

At the top of locality 2, and better exposed at localities 3 and 4, are the massive, cliff-forming dolostones of the Farmers Creek Member. These porous, vuggy beds are conspicuously rich in echinoderm debris, and scattered throughout are the green algae <u>Cyclocrinites</u>, in this unit.

At locality 4, the contact between the Farmers Creek and the overlying Picture Rock members is exposed. Notice that the Picture Rock is characterized by a coral-stromatoporoid community, as was the Sweeney Member at the base of the section. As below, laminar stromatoporoids commonly are silicified.

Locality 5 exposes the contact between the Hopkinton and Scotch Grove formations. The basal unit of the Scotch Grove, the Johns Creek Quarry Member, is a thin-bedded dense dolomite less than one meter thick. The overlying Welton Member, better exposed at locality 6, is a richly fossiliferous unit, with echinoderm skeletal debris preserved as both large unit dolomite crystals (in the lower part) and as moldic porosity. Crinoid and cystoid calyces are <u>very</u> common in these richly fossiliferous strata.

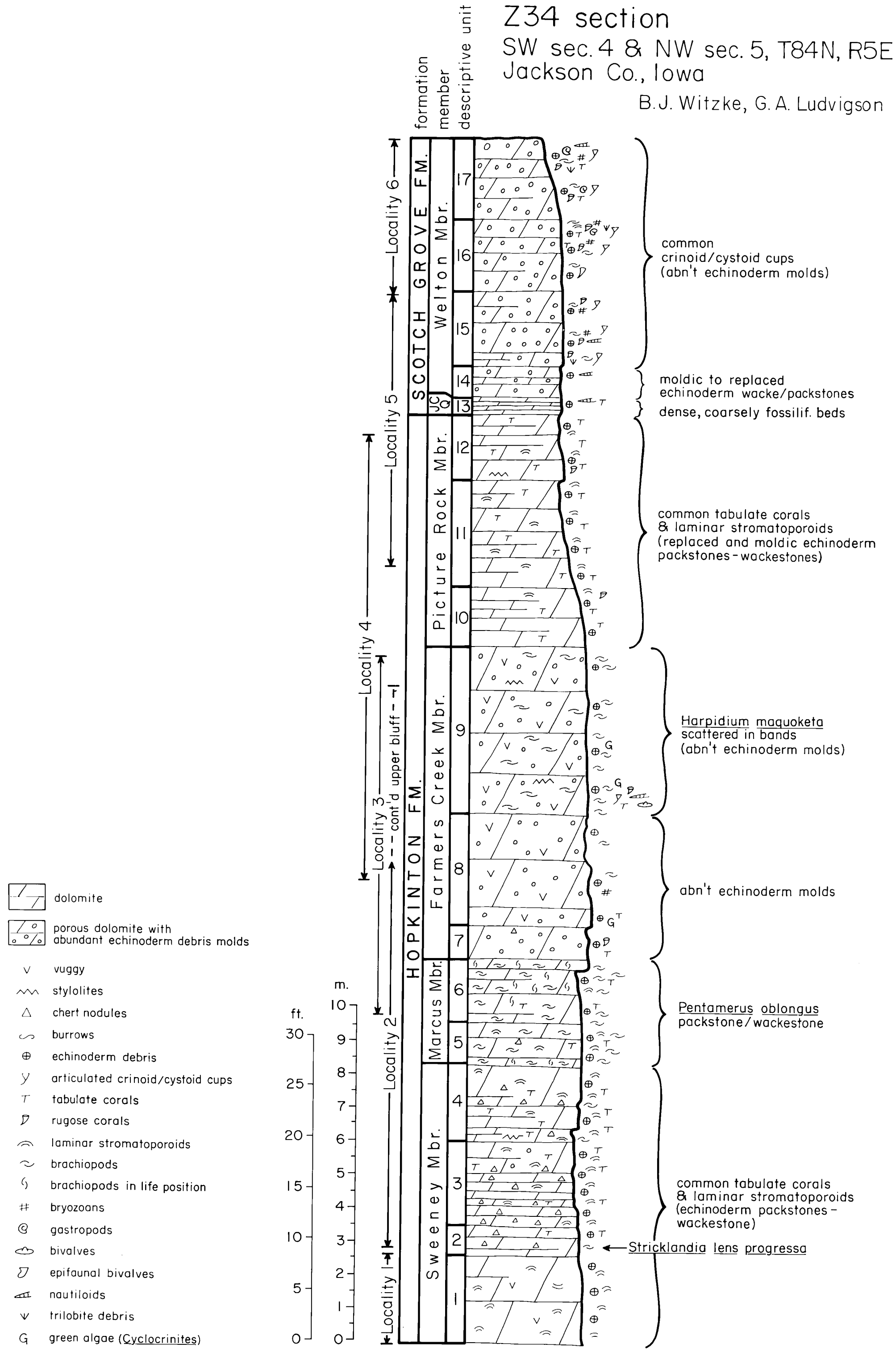

Figure 4. Graphic log of the composite stratigraphic section at STOP 1. The county road Z-34 section.

Z34 SECTION, SILURIAN SEQUENCE
SW sec. 4 & NW sec. 5, T84N, R5E, Jackson Co., Iowa
description by B.J. Witzke and G.A. Ludvigson

SILURIAN
SCOTCH GROVE FORMATION
WELTON MEMBER
UNIT 17
Dol., xf-vf xlln, abundant echinoderm molds, very porous, moldic wackestone to packstone fabrics; thick-bedded, recessive upward; abundantly fossiliferous, echinoderm calyces common (Eucalyptocrinites, Siphonocrinus, Caryocrinites), common brachiopods (Atrypa, Eospirifer, Macropleura?, Ferganella, Protomegastrophia, Fardenia, Leptaena, etc.); additional fauna includes small tabulate corals (Halysites), solitary rugosans (Porpites, small cup corals), bryozoans (fenestellid, flat branching), gastropods, nautiloids, trilobites (Bumastus); 2.45 m thick.
UNIT 16
Dol., xf-vf xlln, abundant echinoderm molds, very porous, moldic wackestone to packstone fabrics; medium to thick-bedded; abundantly fossiliferous, echinoderm calyces common (Eucalyptocrinites, Calliocrinus, Siphonocrinus, Periechocrinus, Myelodactylus, Caryocrinites, etc.); brachiopods (Atrypa, Eospirifer, Cyrtia, meristellid, Ferganella, Protomegastrophia, Fardenia, Resserella, Isorthis, Dolerorthis, etc.), small tabulate corals (Halysites), solitary rugosans (Porpites, small cup corals), small stromatoporoids, gastropods, fenestellid bryozoans, trilobites (phacopid); 2.15 m thick.
UNIT 15
Dol., xf-f xlln, abundant echinoderm molds, very porous, moldic wackestone to packstone fabrics; in beds 20 to 90 cm thick, overhanging at base; abundantly fossiliferous, echinoderm calyces common (Eucalyptocrinites, Calliocrinus, Lampterocrinus, Myelodactylus, Caryocrinites, etc.); brachiopods (Atrypa, gypidulinid, Ferganella, Protomegastrophia, Fardenia, etc.), solitary rugosans (Porpites, small cup corals), fenestellid bryozoans, nautiloids, trilobites (Encrinurus); 2.2 m thick.
UNIT 14
Dol., xf-vc xlln, dense, scattered to abundant echinoderm molds, interbedded crinoidal moldic wackestones and dolomite-replaced crinoidal packstones, crinoidal debris becomes finer upward, part vuggy; in beds 30 to 65 cm thick, ledges at base, recessive upward; several small nautiloids noted in top 30 cm; 95 cm thick.
JOHNS CREEK QUARRY MEMBER
UNIT 13
Dol., xf-vf xlln, dense, sparsely fossiliferous (scattered small echinoderm molds) wackestone fabrics, becomes slightly more porous and skeletal moldic upward, scattered small vugs (1 cm) in upper half; in four to five beds, prominent bedding surfaces ("upper quarry beds") at top and middle, 3 cm bed near middle is recessive; fossils noted include small tabulates (Favosites, Halysites), nautiloid (Dawsonoceras); 47 cm thick.

HOPKINTON FORMATION
PICTURE ROCK MEMBER
UNIT 12
Dol., vf-f xlln in lower part, f-c xlln in upper part, echinoderm-moldic wackestone to dolomite-replaced crinoidal packstone fabrics, stylolitic; medium bedded, prominent ledge former; scattered moldic to silicified laminar stromatoporoids, tabulate corals to 30 cm (Favosites, Halysites, Syringopora), cup corals; 1.8 m thick.
UNIT 11
Dol., vf-m xlln, interbedded dolomite-replaced crinoidal packstone and skeletal moldic wackestone; medium to thick bedded, benched back top of section 4, recessive near top; scattered moldic to silicified laminar stromatoporoids (thin), tabulate corals to 40 cm (Favosites, Halysites, Syringopora), large horn coral noted near middle; 3.2 m thick.
UNIT 10
Dol., f-m xlln, dolomite-replaced crinoidal packstone to wackestone fabrics; thin to medium-bedded, irregularly bedded, basal contact abrupt; scattered moldic to silicified laminar stromatoporoids, tabulate corals (Favosites, Halysites, Syringopora), silicified horn coral noted near top; 1.8 m thick.
FARMERS CREEK MEMBER
UNIT 9
Dol., xf-f xlln, porous, common to abundant small echinoderm molds (larger crinoid molds noted 1-2.3 m above base), vuggy (Swiss-cheese rock), moldic wackestone to packstone fabrics stylolites scattered in lower 80 cm; thick-bedded to massive, cliff former; echinoderm debris conspicuous, bands of small (1-2 cm) pentamerids (Harpidium) scattered throughout but become generally more abundant upward, calcareous green algae (Cyclocrinites 2-3 cm diam.) scattered through; additional fauna noted in lower 1 meter includes brachiopods (Platystrophia, rhynchonellids, Stricklandia), cystoids (Gomphocystites), cup corals, tabulate corals (Favosites, Halysites), stromatoporoids, bryozoans, gastropods, bivalves, nautiloids;

about 5.0 m thick.

UNIT 8

Dol., xf-f xlln, porous and vuggy as above, abundant small echinoderm molds, no pentamerids noted; massive to thick-bedded (0.5-1.5 m), overhanging ledge at base, cavernous vugs above; non-crinoidal fauna is sparse and includes Cyclocrinites (1-2.5 cm diam.), small Favosites, brachiopods (Atrypa, rhynchonellids), fenestellid bryozoans; 3.25 m thick.

UNIT 7

Dol., xf-vf xlln, porous, abundant small crinoid molds; massive, vertical cliff-former, overhanging (with upper unit 6), basal contact sharp; nodular chert band locally at top; non-crinoidal fauna sparse and includes small Favosites and cup coral; 1.0 m thick.

"MARCUS" MEMBER

UNIT 6

Dol., xf-vf and vf-m xlln, dominated by Pentamerus-moldic packstones but includes small crinoid-moldic and crinoid-replaced fabrics; Pentamerus oblongus wackestones to packstones primarily composed of reworked shells and disarticulated valves but horizons every 10 to 50 cm with Pentamerus in life position, isolated valves may be convex up or convex down; most Pentamerus 1-3 cm but larger specimens to 5-6 cm; rare chert nodules 30 cm from top; medium to thick-bedded, generally slightly recessive but top 25 cm locally overhanging (with unit 7); scattered to common tabulate corals (Halysites) and laminar stromatoporoids (part silicified) in lower to middle beds; gastropod (3 cm) noted near top; 1.65 m thick.

UNIT 5

Dol., xf-vf and vf-f xlln, basal 12 to 15 cm is Pentamerus oblongus packstone to wackestone with reworked shells, some in life position (shells to 9.5 cm); remainder of unit is dominated by small crinoid-moldic to crinoid-replaced wackestone to packstone with scattered valves of Pentamerus, locally with Pentamerus packstones (some in life position) in middle to upper part, valves convex up and convex down; intervals with sparse Pentamerus display scattered to common laminar stromatoporoids (silicified), tabulate corals (Halysites, Favosites, Syringopora), cup corals; medium-bedded; rare chert nodules in middle bed; 1.15 m thick.

SWEENEY MEMBER

UNIT 4

Dol., vf-f xlln and f-c xlln, common small crinoid debris molds, becomes more coarsely xlln upward and includes crinoid-replaced packstones; scattered nodular chert bands noted at 20 cm, 40 cm, 1.1 m, 1.6 m above base and at top; argillaceous stylolites in lower part; medium-bedded in lower half, single bed top half, rubbly and recessive in lower bed, becomes cliff-former above (in part overhanging); abrupt contact at top where Pentamerus from overlying unit 5 are locally "rooted" into unit 4; scattered to common silicified laminar stromatoporoids throughout, generally increasing in abundance upward; scattered tabulate corals (Halysites, Syringopora) to 20 cm, corals locally overturned; additional fauna includes brachiopods (Atrypa 55 cm down from top, indet. rhynchonellid in lower half) and horn corals; pentamerid brachiopods (?Pentamerus to 4 cm) rare and locally noted near base, 1.15 m above base, and 25 cm below top; 2.38 m thick.

UNIT 3

Dol., vf-m xlln and vf-c xlln, dolomite-replaced crinoidal packstone fabrics common, some crinoid grains are silicified; scattered nodular chert bands noted at base, 15 cm, 35 cm, 65 cm, 1.1 m, 1.55 m, and 1.65 m above base; medium-bedded (20-50 cm thick), ledge-former at base becomes rubbly and recessive above; scattered to common silicified laminar stromatoporoids (10-30 cm wide), scattered moldic to silicified tabulate corals (Halysites, Favosites, Syringopora) to 15 cm; common cup and horn corals 90 cm below top; 2.5 m thick.

UNIT 2

Dol., vf-f xlln and vf-c xlln, interbedded dolomite-replaced crinoidal packstones; in two to three beds, rubbly and poorly exposed at base, more resistant above; two nodular chert bands; scattered tabulate corals (Halysites, Favosites), cup corals, and silicified stromatoporoids; stricklandiid brachiopods (Stricklandia lens progressa) noted; 70 cm thick.

UNIT 1

Dol., vf-m xlln, small crinoid debris molds, common vugs (decrease upward); in two thick beds; scattered silicified laminar stromatoporoids (2-10 cm wide), generally increasing in abundance upward, some stromatoporoids overturned in upper half; small cup coral noted near top; 2.6 m thick.

STOP 2. The Ordovician Galena and Maquoketa groups at the Mount Carroll Quarry; Mississippi Valley-Type lead-zinc deposits.

THE SLOPES IN THIS QUARRY ARE DANGEROUS!

PLEASE FOLLOW THE INSTRUCTIONS OF THE FIELD TRIP LEADERS

HARD HATS AND BOOTS ARE REQUIRED!

Greg A. Ludvigson, Dennis R. Kolata, and Tim M. Millen

The Mount Carroll Quarry provides excellent exposures of the Ordovican rocks that crop out to the south of the Plum River Fault Zone in Carroll County, Illinois. These include the Wise Lake and Dubuque formations of the Galena Group, and the basal unit of the Maquoketa Group, the Scales Formation (Fig. 5).

Although all of the Galena Group carbonates are pervasively dolomitized at this locality, fresh exposures of unoxidized rocks still reveal many important features. The Wise Lake-Dubuque succession largely consists of lime mudstones to wackestones that were deposited in deep subtidal open marine environments below normal effective wave base. Occasional thin grainstone units (3 to 5 cm thick) are generally regarded as lag deposits after rare large-magnitude storm events. Pyrite-blackened hardground surfaces are further marked in the quarry walls by the development of secondary ferric oxide stains (unit 1, Fig. 5), and by their irregular sculpted surfaces. These surfaces record the cessation of carbonate deposition for an indefinite period. There is disagreement over the importance of chemical vs. mechanical erosion in the formation of the hardground surfaces. Regardless, the relief was developed on lithified carbonates on the sea floor, and borings and attached benthic organisms are common features. As noted by Delgado (1983), both pyrite and apatite are common authigenic minerals in hardgrounds from the Galena Group.

While the general environmental aspects of the Galena Group apparently are well-known, specific interpretations of eustatic sea level cycles remain controversial. The paper by Witzke and Kolata (this guidebook) presents two alternative interpretations of the relative sea level curve through the Wise Lake Formation. A major part of the interpretive difference revolves around the significance of the apparent upward appearance and disappearance of dasycladacean Receptaculitid algae (units 1 to 3, Fig. 5). Does this phenonenon record benthic environments that alternately were above and below the photic zone?

The overlying Dubuque Formation largely consists of carbonate/shale interbeds, and probably records the deepening of the epeiric sea. Dubuque carbonates are notably crinoid-rich, but are marked by the upward disappearance of green algae, bryozoans, and grainstone interbeds. Likewise, the appearance of abundant acritarchs, chitinozoans, Tasmanites, Lingulid brachiopods, and pervasive early sulfidic diagenesis are suggestive of oxygen-stressed benthic environments (Ludvigson, 1987).

The top of the Dubuque is sculpted by a major hardground surface with cavities up to 30 cm in width and depth at this locality (units 13 and 14, Fig. 5). These cavities are completely filled and overlain by a pyrite-cemented phosphorite containing concentrically-laminated ooids, pellets, and nodules, and phosphatized diminutive fossils. The basal Maquoketa phosphorite is well exposed on a benched surface at the top of the northern highwall of the quarry. Field trip participants who wish to visit exposures of the Maquoketa Formation should follow the field trip leaders so that they will not endanger themselves or other participants. Shales immediately overlying the basal phosphorite contain Lingulid brachiopods. Three feet (0.9 m) above the phosphorite is a one foot (30 cm) thick, hard, brownish-black shale bed with Chondrites burrow networks. This is the

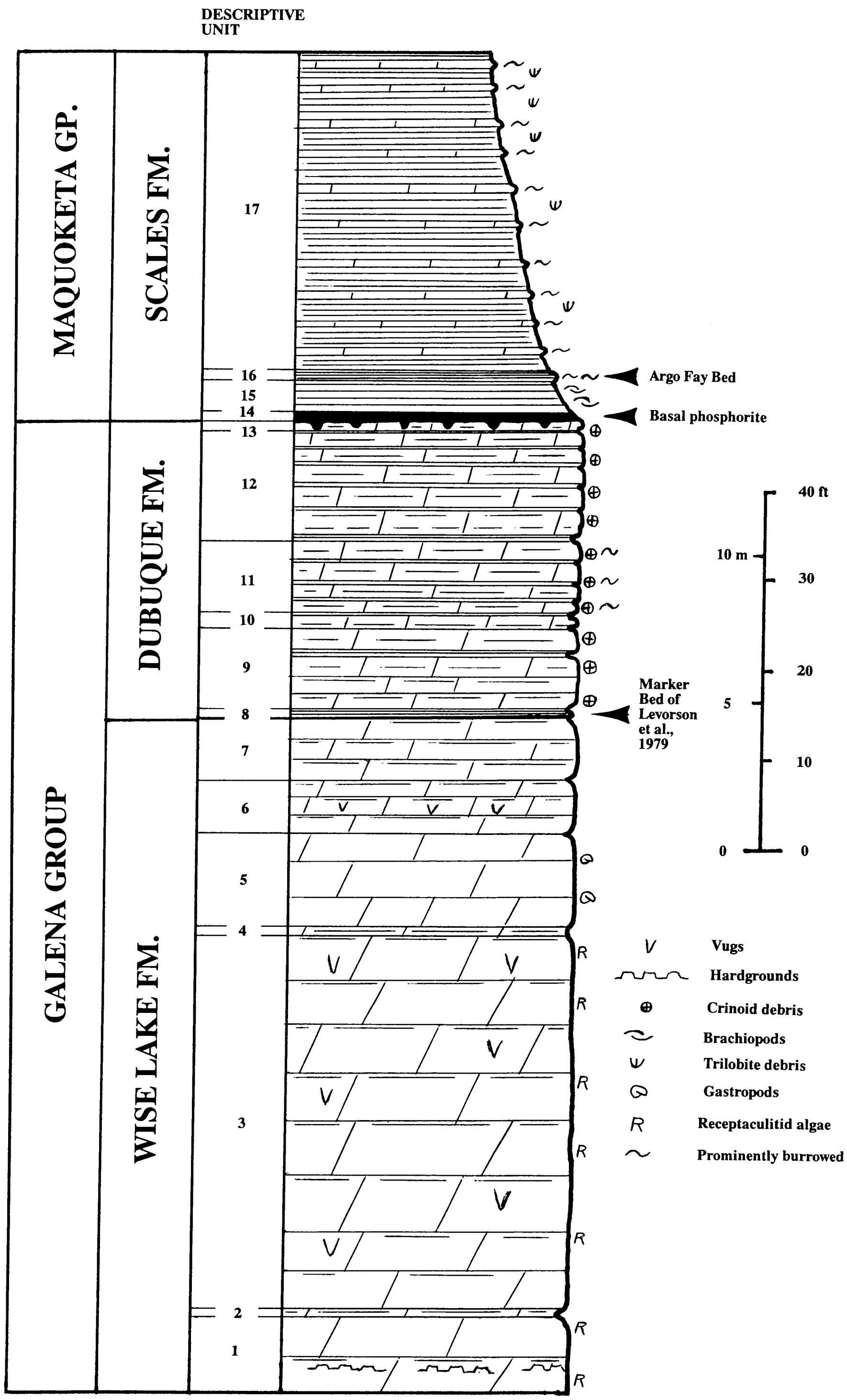

Figure 5. Graphic log of the stratigraphic section at STOP 2. The Mount Carroll Quarry.

Argo-Fay Bed at its type locality, a petroleum source bed that here contains 16% TOC. Carbonaceous fragments of graptolites and chitinozoans have been noted from the Argo-Fay bed, although most of the organic material is apparently of algal origin (Kolata and Graese, 1983). Witzke and Kolata (this guidebook) interpret the Argo-Fay bed as the maximum transgressive deposit of a major T-R cycle in the lower Maquoketa, and note that the unit only occurs in the the central portions of the East-central Iowa Basin and passes laterally into shallower facies, suggesting that basinal sudsidence was initiated in the Late Ordovician. Another 35 feet (11 m) of trilobite-bearing interbedded shale and argillaceous dolomudstone is exposed above the Argo-Fay Bed. These units are exposed on benches cut back from the basal Maquoketa phosphorite, and along a ramp cut down to that surface. In the following field trip stop we will examine the rock succession in the upper part of the Maquoketa Group.

Galena Group rocks at the Mount Carroll Quarry are cut by high angle veinlets filled by pyrite, galena, and sphalerite. Isotopic studies of these minerals, reported by Ludvigson and Millen (this guidebook) show that the veinlets at this locality are coeval with lead deposits that were mined in this area. Vertical to steeply northward-dipping mineralized fractures in the quarry strike N75^{o}W to N80^{o}W, and are up to several centimeters thick. Cubo-octahedral pyrite and marcasite are the earliest minerals, followed by complex intergrowths of galena and sphalerite. Early galenas are cubic in form, whereas later galenas are octahedral. Sulfur isotopic temperatures of sulfide mineral pairs from this quarry range between 107^{o} to 295^{o} C. As discussed by Ludvigson and Millen (this guidebook), these deposits might have precipitated from basinal fluids that were driven northward as a consequence of tectonic and hydrologic processes operating along the convergent Ouachita continental margin in Late Paleozoic time.

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MOUNT CARROLL SOUTHWEST SECTION

Quarry on north side of U. S. Route 52
2 1/2 miles (4 km) southwest of Mount Carroll, Carroll County, Illinois
SW NE SW 10, T24N, R4E, Wacker 7.5 minute Quadrangle
Type section of Argo-Fay Bed.
Description by D. R. Kolata
(updated from Willman and Kolata, 1978, and Kolata and Graese, 1983).

CINCINNATIAN SERIES

MAQUOKETA GROUP

SCALES FORMATION

UNIT 17

Shale, light olive-ray (5Y 6/1) to olive-gray (10 YR 5/4), silty, conchoidal fractures; asaphid trilobites occur throughtout but are most abundant in upper 10 ft (3 m); interbeds of very argillaceous , very fine-grained dolomite in even beds 2 to 6 in. (5 to 15 cm) thick; long sinuous tracts 1/16 in. (2 mm) wide on bedding planes in dolomite; 35 ft. (10.6 m) thick.

UNIT 16

Shale, hard, brownish black (5 YR 2/1) to black (N 1), carbonaceous, fissile; upper 4 to 5 in. (10 to 12.5 cm) contain flattened moderate yellowish brown (10 YR 5/4) burrows from 1/32 to 1/15 in. (1 to 2 mm) wide assignable to Chondrites; total arganic carbon 16 percent (Argo-Fay Bed, type section); 1 ft. (0.3 m) thick.

UNIT 15

Shale, olive-gray (5 Y 4/1), silty conchoidal fractures, contains abundant linuloid brachiopods (Leptobolus sp.); 3 ft. (0.91 m) thick.

UNIT 14

Phosphorite, light brownish gray (5 YR 6/1) to brownish gray (5 YR 4/1), poorly sorted, bioturbated, silty, friable, pyritic; contains phosphatic oolites, pellets, and nodules and phosphatized diminutive fossils (Palaeoneilo? fecunda, Nuculites neglectus, Michelinoceras sociale, Plagioglypta iowensis, Liospira sp., Leptobolus sp. and Onniella sp.) upper 2 to 3 in. (5 to 7.5 cm) cemented by pyrite; 1 ft. (0.3 m) thick.

GALENA GROUP (108 ft. or 32.4 m)

DUBUQUE FORMATION (32 ft. 11 in. or 9.9 m)

UNIT 13

Dolomite, argillaceous, fine to medium grained; 6 to 8 in. (15 to 20 cm) even beds; thick brown shaly partings; abundant recrystallized crinoidal debris on bedding surfaces; top marked by pitted, ferruginous, cryptocrystalline phosphate encrusted surface with cavities up to 12 in. (30 cm) wide and 12 in. (30 cm) deep filled with overlying phosphorite; 1 ft. (0.3 m) thick.

UNIT 12

Dolomite, as above but in 10 in. to 12 in. (25 cm to 30 cm) even beds; prominent reentrant at top; 3 in. (7.6 cm) shaley bed at base; 12 ft. 3 in. (3.7 m) thick.

UNIT 11

Dolomite, as above but 8 in. to 10 in. (20 cm to 25 cm) beds; thin shaley partings; Chondrites and Paleophycus common; 7 ft. 10 in. (2.4 m) thick.

UNIT 10

Dolomite, dense, massive; 4 in. (10 cm) shale at top; 2 in. (5 cm) shale at base; 1 ft. 9 in. (0.52 m) thick.

UNIT 9

Dolomite, like unit 11 above; 4 in. to 5 in. (10 to 13 cm) shale bed 2 ft. (0.61 m) below top; 3 ft. 8 in. (1.1 m) thick.

UNIT 8

Dolomite, prominent bed set off by shaley partings; widely traced marker bed of Levorson et al. (1979); 4 in. to 6 in. (10 to 15 cm) thick.

WISE LAKE FORMATION (74 ft. or 22.5 m)

UNIT 7

Dolomite, relatively pure, gray; few thin, reddish brown shale partings; 7 ft. 3 in. (2.1 m) thick.

UNIT 6

Dolomite, gray; slightly more argillaceous than above or below; 12 in. to 24 in. (30 to 61 cm) beds; thin, brown shale partings; vuggy near middle; 5 ft. 6 in. (1.7 m) thick.

UNIT 5

Dolomite, pure, gray; 1 ft. to 2 ft. (30 cm to 60 cm) beds; numerous moldic molluscan fossils; 10 ft. 6 in. (3.2 m) thick.

UNIT 4

Dolomite, two 5 in. (12.7 cm) beds with thin, brown shaley partings; 10 in. (25.4 cm) thick.

UNIT 3

Dolomite, pure, gray, vuggy; 2 ft. to 3 ft. (0.62 to 0.91 m) beds; very thin, shaley partings; Receptaculites common; 41 ft. 2 in. (12.5 m) thick.

UNIT 2

Dolomite, slightly argillaceous; thin, dark brown to gray partings; 3 in. to 4 in. (7.6 to 10.2 cm) thick.

UNIT 1

Dolomite, like unit 3 above; prominent hardground surface 66 in. (1.67 m) below top; 8 ft. 6 in. (3.1 m) thick.

STOP 3. Upper portion of the Maquoketa Group at the Wacker railroad cut.

Greg Ludvigson and Brian Witzke

The Wacker railroad cut is one of the best exposures of the Brainard Shale succession anywhere in the Upper Mississippi Valley. Shales in the Late Ordovician Maquoketa Group represent the distal tongues of an immense mass of fine-grained terrigenous sediments that dispersed westward across the craton from the Taconic orogen (Witzke, 1980). As discussed by Witzke and Kolata (this guidebook), the Brainard is a shallowing-upward sequence that includes early deposition on a deep subtidal sea floor that only rarely was winnowed by storm events. Dysoxic-anoxic environments prevailed during early Brainard deposition, and the only evidence for benthic activity is the development of bioturbated fabrics, and occasional trilobite-bearing beds.

These strata pass upward into units that are increasingly skeletal rich, with diverse benthic faunas. At least periodic development of oxic environments on the sea floor seems to be indicated in these deposits. Upper portions of the section contain a variety of brachiopods, echinoderms, bryozoans, and trilobites. Especially noteworthy are the packstone units near the top, thinly interbedded with shales containing large Prasopora bryozoans (unit 17, Fig. 6). Some Prasopora colonies are encrusted with crinoid holdfasts and bored with worm dwelling tubes.

Carbonate interbeds dispersed through the section best display the sedimentary fabrics in these rocks, and apparently formed through several mechanisms. Lenticular carbonate beds in the lower part of the section developed as nodular concretions in unfossiliferous, laminated, sparsely burrowed shales. Some carbonate interbeds in the lower part of the section also contain thin packstone beds that may represent storm deposits (top of unit 4, Fig. 6). The lenticular skeletal packstone bed at the base of unit 11 (Fig. 6) is exposed on a bench surface, and formed as a set of starved ripples of coquina enveloped in a shale matrix. At the top of the section are predominantly carbonate units (units 13 and 15, Fig. 6) with burrowed wackestone-packstone/grainstone fabrics. These units evidently record brief periods of increased carbonate production, probably reflecting increased shallowing.

The lithologies exposed in the rock succession at Wacker can be used to provide a perspective on the approximate stratigraphic position of natural exposures of Maquoketa strata that we will be visiting this afternoon. As you might guess, good natural exposures of the Maquoketa are very rare. Nevertheless, field geologists can use the fabrics and faunal composition of carbonate interbeds weathering out of natural exposures to recognize stratigraphic position within the Maquoketa sequence. This discrimination has proven to be especially useful in our mapping experience along the PRFZ.

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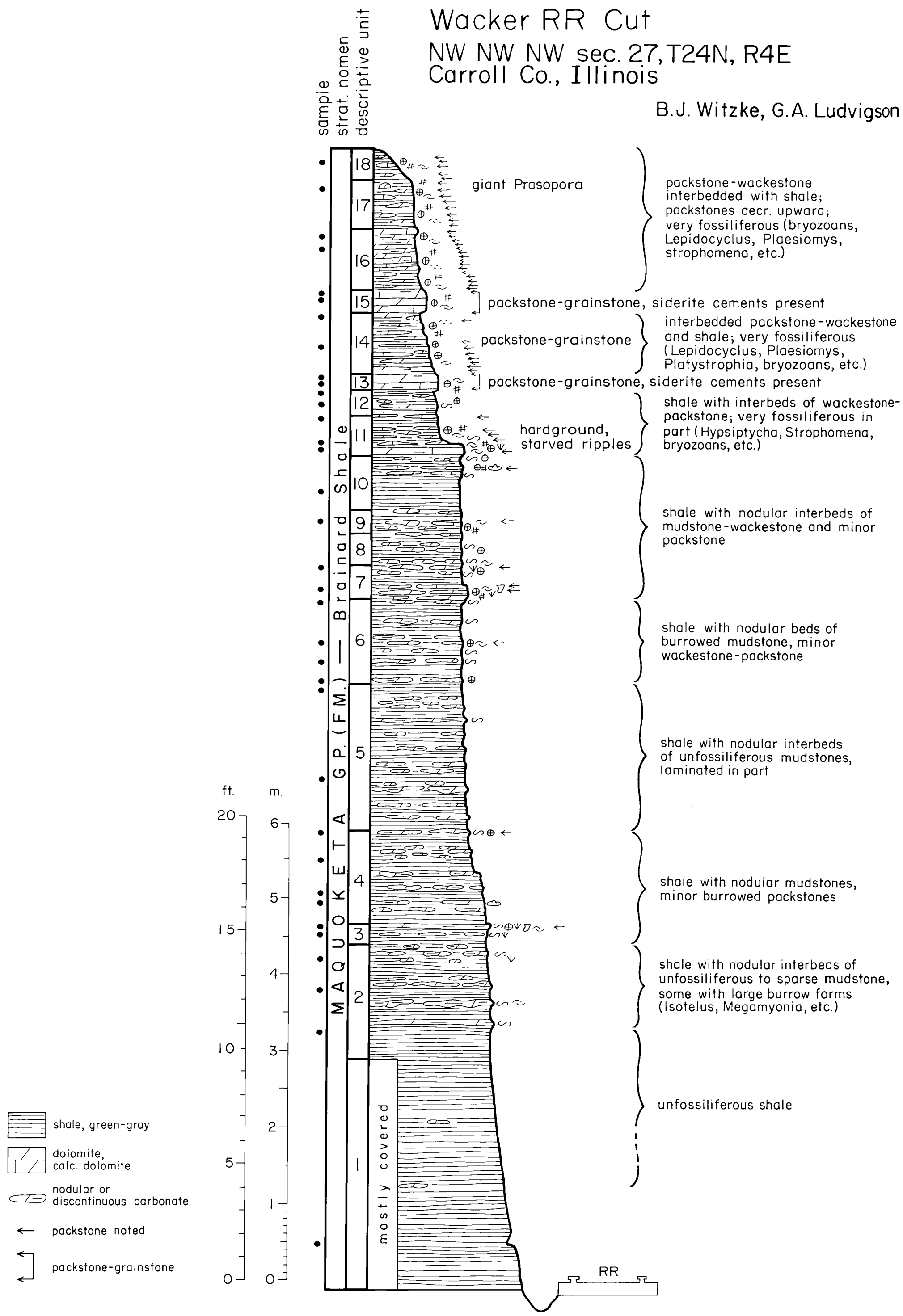


Figure 6. Graphic log of the stratigraphic section at STOP 3. The Wacker railroad cut.

WACKER RAILROAD CUT

NW NW NW sec. 27, T24N, R4E, Carroll Co., Illinois
description by B.J. Witzke and G.A. Ludvigson

UPPER ORDOVICIAN (RICHMONDIAN)

MAQUOKETA GROUP

BRAINARD SHALE

UNIT 18

Shale, green-gray, dolc., thin skeletal wacke-packstone beds (calc. dol.) in upper part; top bench of section; chunky shale fabrics at top may be soil, C-horizon, overlain by thin drift; lower part contains abundant loose branching trepostome bryozoans; carbonates in upper half with common bryozoans and brachiopods (Lepidocyclus, Platystrophia, large lingulids to 5 cm); 40 cm thick.

UNIT 17

Shale, green-gray, dolc., interbedded with thin skeletal wackestone and packstone beds (calc. dol.), shale and dolomite roughly equal proportions, becomes more shaley upward; slope forming unit, benched at top; top 20 cm with conspicuous large hemispherical bryozoans (Prasopora) to 16 cm diameter, some overturned, Prasopora colonies encrusted in part with small crinoid holdfasts and bored with worm dwelling tubes on upper surface; abundantly fossiliferous--branching trepostome bryozoans (Hallopora, Homotrypa, etc.), ctenostome bryozoans encrusting brachiopods and trepostomes, echinoderm debris; common brachiopods include in decreasing order of abundance Lepidocyclus, Plaesiomys, Strophomena, Platystrophia, Glyptorthis, Onniella, Opikina, craniid (on Prasopora); 65 cm thick.

UNIT 16

Interbedded shale and skeletal packstone interburrowed to interbedded with wackestone (calc. dol.), unit is carbonate dominated, packstone-filled burrows occur in shale; interval is slope former, ledge-forming packstone 30 cm above base and top; lower 30 cm is very echinoderm-rich, with notable decrease in brachiopod abundance compared to underlying units; observed fauna includes bryozoans (branching trepostomes, encrusting ctenostomes), abundant echinoderm debris, brachiopods (Lepidocyclus, Hypsiptycha, Plaesiomys, Lingula); 80 cm thick.

UNIT 15

Dol., calc., prominent carbonate ledge in three beds with thin shaley partings between; lower bed is skeletal packstone with wackestone burrows; middle bed is a skeletal packstone to grainstone, part porous, with siderite cements; upper bed is a burrowed packstone to grainstone with siderite cements, part porous; abundant echinoderm and brachiopod debris, part abraded; bryozoans rare to absent; 30 cm thick.

UNIT 14

Thinly interbedded shale and skeletal wacke-packstone (calc. to v. calc. dol.), carbonate beds dominated by packstone, grainstone bed with siderite cements noted in middle part, carbonate lenses 1-4 cm thick, shale beds 1-10 cm thick; slope-former, in part poorly exposed; very fossiliferous, abundant loose fossils weathering out along slopes, brachiopods, bryozoans, echinoderm debris most common groups, some grains are abraded; crinoid debris includes Carabocrinus plates, barrel-shaped columnals of Cupulocrinus; bryozoans include branching trepostomes, Prasopora, and encrusting forms; brachiopods abundant, in decreasing order of abundance, Lepidocyclus, Plaesiomys, Platystrophia, Hypsiptycha, Strophomena, Glyptorthis, Thaerodonta, Hesperorthis, Onniella, craniid (encrusting); additional fossils include trilobites (calymenids), conical worm tube (encrusting bryozoan), and straight-shelled nautiloids; 95 cm thick.

UNIT 13

Skeletal packstone to grainstone, in part swirled or burrowed with wackestones, carbonates are calc. dolomite, grainstones in part porous with common siderite cements, grains are abraded in part; unit in two beds, prominent ledge former; dominated by echinoderm and brachiopod grains, minor bryozoans; brachiopods include Lepidocyclus, Hypsiptycha, Onniella, resserellid; 20 cm thick.

UNIT 12

Shale, green-gray, dolc., with minor discontinuous nodular argillaceous carbonate lenses in top 25 cm; carbonates include sparse skeletal wackestones (slightly calc. dol.) with skeletal material concentrated in burrows, echinoderm debris present (incl. Cupulocrinus); 35 cm thick.

UNIT 11

Shale, as above, interbedded with thin 1-3 cm thick discontinuous beds of skeletal packstone (sl. calc. to calc. dol.), shale with some packstone-filled horizontal burrows; prominent 0-12 cm thick skeletal packstone (calc. dol.) at base (top of major bench above RR cut), packstone interburrowed with wackestones (also in sheltered voids); basal packstone displayed as arcuate starved megaripple bedforms with crests spaced at 1.2 to 2.4 m intervals, trending N30°E, basal packstone locally capped by bored hardground surface; basal carbonate with abundant fossils, especially brachiopods and echinoderm debris, brachiopods include in general decreasing order of abundance Hypsiptycha, Strophomena, Lepidocyclus, Plaesiomys, Onniella, Thaerodonta, Glyptorthis, Platystrophia, Diceromyonia, Opikina; echinoderm debris includes dendrocrinids, carabocrinids, Cupulocrinus, rhombiferan plates; bryozoans include common branching trepostomes, rare Prasopora, sheet-like forms, and ctenostomes (encrusting brachiopods, trilobites); worm tubes (Cornulites) and trilobite debris (ceraurids, calymenids) scattered; 50 cm thick.

UNIT 10

Shale, green-gray, dolc., with three discontinuous thin nodular carbonate (calc. dol.) bands in top 25 cm; lower carbonate is argillaceous burrowed skeletal wackestone with "red specks" and small echinoderm debris; middle carbonate is a skeletal packstone with siderite void fills, echinoderm debris, bryozoans, bivalves (Ambonychia); upper carbonate is wackestone with packstone burrow fills; 73 cm thick.

UNIT 9

Shale, as above, with three 1-5 cm thick nodular carbonate bands, carbonates (calc. dol.) are primarily argillaceous sparse skeletal mudstones to wackestones, minor skeletal packstone (echinoderm, bryozoan, brachiopod); 29 cm thick.

UNIT 8

Shale with four 1-6 cm thick nodular carbonate (calc. dol.) bands, carbonates are horizontally burrowed and include small "red specks" and scattered small echinoderm debris (mudstone to wackestone); 43 cm thick.

UNIT 7

Shale with five discontinuous nodular carbonate bands 2-6 cm thick (argillaceous sl. calc. dol.), some "red specks", primarily skeletal wackestone to mudstone, burrowed in part; includes skeletal packstone in lower half; fauna includes echinoderm debris, brachiopods (fine-ribbed orthids and strophomenids), small branching bryozoans, bivalves (Ambonychia), trilobite debris (incl. encrinurids); 43 cm thick.

UNIT 6

Shale with seven discontinuous nodular carbonate bands 1-6 cm thick (0 cm, 35 cm, 45 cm, 57 cm, 65 cm, 86 cm, 114 cm above base); carbonates dominated by burrowed argillaceous mudstone (dol.) with scattered "red specks" (ferruginized echinoderm debris?); third carbonate is skeletal wackestone (echinoderm debris, brachiopod) with minor packstone-filled burrows (fauna includes Thaerodonta, Megamyonia, Strophomena); 1.14 m thick.

UNIT 5

Shale with ten discontinuous nodular carbonate bands 1-6 cm thick (19 cm, 34 cm, 56 cm, 79 cm, 89 cm, 1.1 m, 1.31 m, 1.49 m, 1.69 m, 1.85 m above base), carbonates are dense, unfossiliferous argillaceous mudstones (dol.), most of which are finely laminated and unburrowed; fourth dol. bed from top displays horizontal burrows on the surface (this bed may be continuous); 2.0 m thick.

UNIT 4

Shale with six discontinuous nodular carbonate bands 1-5 cm thick (25 cm, 33 cm, 51 cm, 68 cm, 97 cm, 1.22 m above base); carbonates are dense, unfossiliferous, unburrowed, argillaceous mudstones (dol.), second bed contains rare infaunal bivalves (Paleoneilo); top-most carbonate bed is mudstone containing a 3 cm thick skeletal packstone (sl. calc. dol.) with echinoderm debris and brachiopods, packstone disrupted by mudstone-filled burrows; 1.22 m thick.

UNIT 3

Shale with two carbonate bands in lower half, unfossiliferous to sparsely fossiliferous burrowed mudstone (dol.), pyritized trilobite debris noted; prominent carbonate ledge at top, 6-13 cm thick, upper carbonate is a burrowed argillaceous mudstone with a conspicuous but discontinuous packstone interval internally 1-4 cm thick, packstone is disrupted by mudstone-filled burrows, packstone includes echinoderm and trilobite debris, small bryozoans, brachiopods (fine-ribbed orthids and strophomenids, Megamyonia, Strophomena, Thaerodonta, ?Holtedahlina), epifaunal bivalves (Ambonychia), infaunal bivalves (Paleoneilo); 30 cm thick.

UNIT 2

Shale, green-gray, dolc., with six nodular to discontinuous carbonate bands 1-9 cm thick (50 cm, 75 cm, 98 cm, 1.06 m, 1.26 m, 1.4 m above base); carbonates are dominated by unfossiliferous argillaceous mudstones, burrowed in part; very large horizontal to subhorizontal burrow traces (1-6 cm diameter) conspicuous along bottom surfaces of some carbonates (some of these resemble large trilobite resting traces); skeletal fauna preserved in some carbonate beds (skeletal mudstone to sparse wackestone) includes brachiopods (small lingulids, Megamyonia, ?Tetraphalerella, Thaerodonta, Diceromyonia), small bryozoans (encrusting on brach, small stick-like form), trilobites (including one specimen of an Isotelus pygidium 13 cm wide); 1.5 m thick.

UNIT 1

Shale, green-gray, dolc., featureless, poorly exposed (mostly covered, trenched in part), may include minor argillaceous dolomite nodules; no burrows or fossils observed; about 3.0 m thick.

STOP 4. The Plum River Fault Zone: exposures of cataclastic dolostones from the Maquoketa, Hopkinton, and Scotch Grove formations.

Greg Ludvigson

As the bus passes west of Preston Iowa on Hwy. 64, turn to Figure 7 to follow our progress on the geologic map by Chao (1980). We will be stopping at some creek bed exposures along the southern margin of the Plum River Fault Zone. As the bus traverses the upland ridge through the map area, note the influence of the Silurian-Ordovician contact on the landscape. The ridgetop straddles the contact as we approach the fault zone from the southern, upthrown side.

The creek bed and valley wall exposures are located on the Nolting and Miller farm properties, where deformed carbonates from the Brainard Shale and the Hopkinton and Scotch Grove formations are in fault juxtaposition (Fig. 8), suggesting relative vertical displacements of about 100 m (330 ft.) within the fault zone. Chao (1980) identified rocks of the Silurian Gower Formation (Fig. 7) in graben fault blocks in the PRFZ that suggest local structural relief of 150 m (500 ft.) in the near-surface rocks of this area. This locality has had a long history of geologic investigations. Savage (1906, p. 607) noted the structural anomaly indicated by the exposure of Maquoketa strata in this creek bed. Aten and Herzog (1977) recognized that the Silurian rocks exposed in the north wall of the valley are in fault contact with the Brainard Shale, and noted deformational fabrics along the contact. The Nolting Farm was visited ten years ago during the 42nd Tri-State Geological Field Conference (Ludvigson et al., 1978) and was used to illustrate several important features of the Plum River Fault Zone. Chao (1980) mapped the geologic structure of the fault in the area surrounding this locality (Fig. 7).

Friable, microfractured skeletal-moldic dolostones of the Welton Member of the Scotch Grove Formation are intermittently exposed with their undeformed equivalents at the Nolting farmstead, also to its south, and in abandoned quarry workings to the west on the Miller Farm property. Petrographic studies of these rocks require prior stabilization by vacuum-injected epoxide resin, but are informative. The deformation of these rocks included the crushing of skeletal molds, and the development of a pervasive network of closely-spaced dilational microfractures cutting across all fabric constituents. Fracture porosity and surviving skeletal molds were filled by calcite and ferric oxide cements that now are preferentially weathering out of these rocks in the modern environment (Fig. 9a,b). The cathodoluminescence petrography of these calcite cements is illustrated and discussed in the paper by Ludvigson (this guidebook, see fig. 9c,d in that paper). These fracture-filling cements are interpreted to have precipitated from groundwaters that infiltrated from the ancient sub-Pennsylvanian erosion surface in this area. The abundant ferric oxide phases and oscillatory cathodoluminescent zonation in the coexisting calcites ("manganese cycling effect" of Ludvigson, this guidebook) indicates that these deformed rocks were healed in near-surface phreatic groundwater environments straddling the oxic-anoxic redox boundary, in the upper part of the groundwater flow system.

To the south of the Scotch Grove exposures on the Nolting Farm property is a narrow exposed block of the Sweeney Member of the Hopkinton Formation no wider than 100 feet (30 m) along a north-south line. These rocks were freshly exposed by road grading along a north-south farm lane in 1978, and are delineated by the abundance of chert nodule float in the colluvial exposures cut by the grade. Fresh rock exposures in the road bed revealed the unit to be a coral-stromatoporoid bearing dolostone with chert nodules (compare to units 2, 3, and 4 at STOP 1). These rocks are cut by zones of penetrative cataclastic deformation, best revealed by the fragmentation and smearing of chert nodules. Although the rocks are less well exposed than they were a decade ago, close inspection of the remaining exposures should reveal subtle evidence of cataclasis. Polished, slabbed samples of these cohesive microbrecciated fault rocks are available for inspection, and Figure 9c illustrates the microscopic fabric of these rocks.

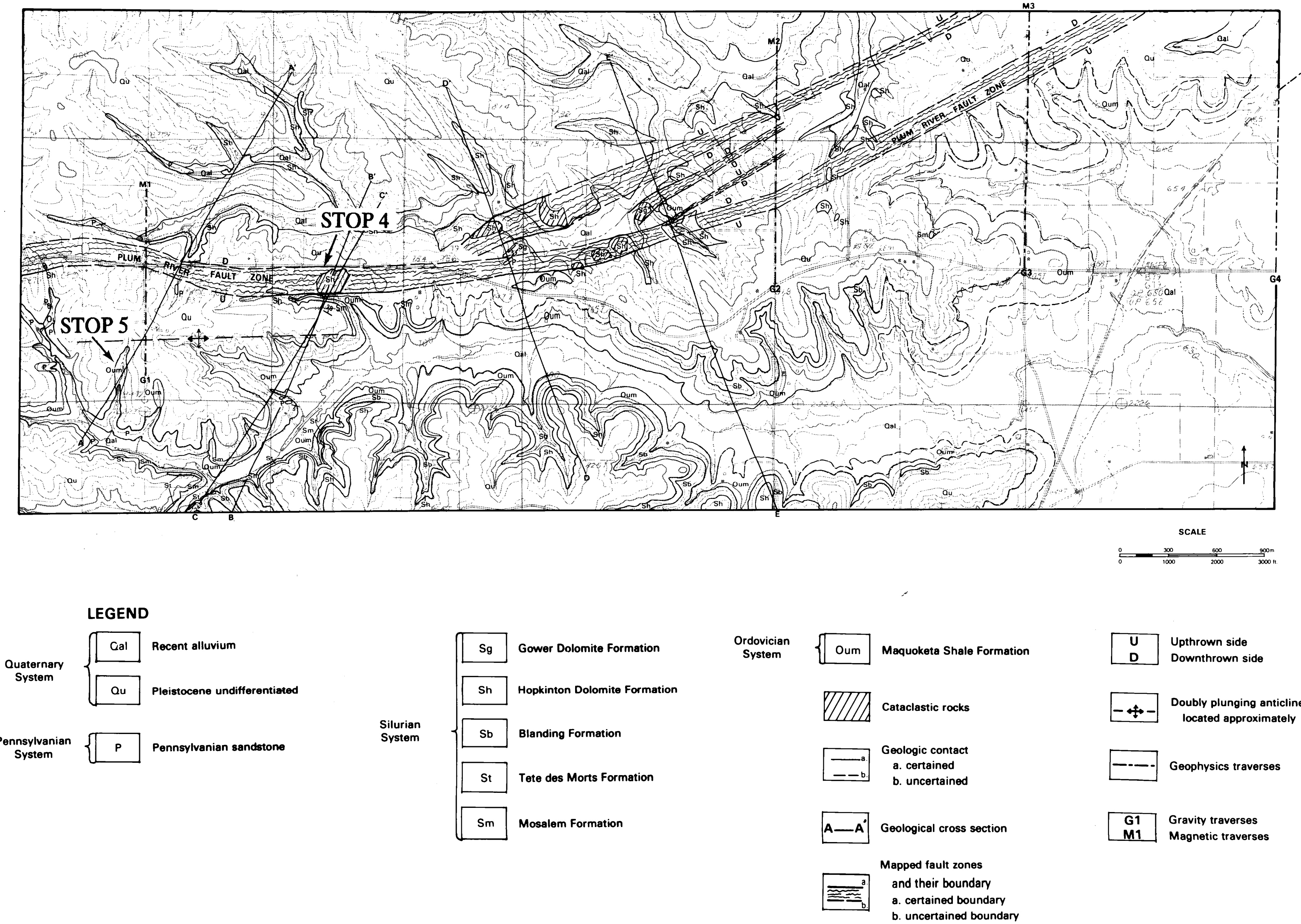
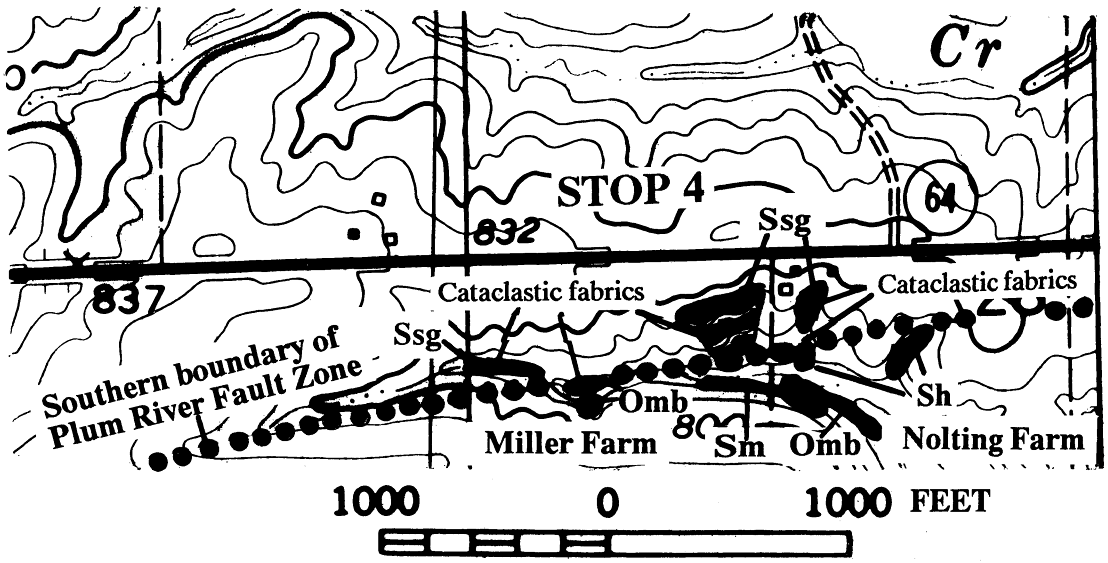


Figure 7. Geologic map of the area around STOP 4 and STOP 5. From Chao (1980). Rocks mapped as Hopkinton Formation include both Hopkinton and Scotch Grove strata.



LEGEND

SILURIAN SYSTEM
- Ssg - Scotch Grove Formation
- Sh - Hopkinton Formation
- Sm - Mosalem Formation

ORDOVICIAN SYSTEM
- Omb - Maquoketa Formation

Figure 8. Location map of important bedrock exposures at STOP 4. Deformed Ordovician and Silurian rocks in the Plum River Fault Zone at the Nolting and Miller farms.

To the south of the Hopkinton exposures, and extending about 270 m (900 ft.) further to the west, are a series of stream bed and stream bank exposures of the Brainard Shale. The packstone/grainstone fabrics in carbonates interbedded with the shale, as well as the basal Silurian rocks exposed on the south valley wall, indicate that these strata belong to the uppermost portion of the Brainard. The dark color of these freshly-exposed dolostones results from the partial replacement of skeletal grains by sedimentary sulfides, indicating extensive sulfidic diagenesis during the early history of these rocks. At the western end of the Brainard exposures, the edge of the zone of cataclastic deformation is exposed in the northern stream bank. The fault rocks formed in darkly-colored, sparsely fossiliferous dolomudstones with scattered echinoderm debris and abundant sedimentary sulfides. These cataclasites are very hard cohesive microbreccias with a groundmass of dolomitic gouge (grain sizes down to 10 μm diameter). Figure 9(d,e,f) shows the microscopic fabrics of these microbreccias, and illustrate some of the processes by which rock masses are altered by cataclasis. The principle change in these rocks is a reduction in grain size, accomplished by the plucking of clasts from shear fracture walls during active sliding, and accompanied by intragranular fracturing and rigid body rotation of clasts (Engelder, 1974). Cataclasites evidently form in anastomosing networks of shear fractures, where intersections are sites of intense fracturing and grain comminution (House and Gray, 1982; Fig.

9d). With continuing fault strain, the volume of gouge in the rock steadily increases (Fig. 9e) until the only remnants of undeformed rock are clasts that float in the gouge matrix (Fig. 9f.). This conversion has consequences for the mechanisms of ongoing fault strain. As noted by Mitra (1984), coarser grains continue to fracture, but ductile deformation occurs in the gouge matrix by a combination of diffusional creep and grain boundary sliding. This strain softening effect--the dominance of diffusional processes in water-saturated gouges (Wojtal and Mitra, 1986), assures that future fault strains are accomodated in prexisting zones of cataclastic deformation. Ludvigson (1988) reported that the presence of anastomosing microstylolites in cataclasites from the PRFZ, further indicating that fault strains in these rocks were accomodated by diffusion-controlled pressure solution.

Ludvigson et al.(1978) and Bunker et al (1985) have noted wide belts of rock deformation in the PRFZ, up to 600 m (2000 ft.) in width (Bunker et al., 1982). As illustrated in Figure 9, however, these include microbreccias formed by cataclasis (c,d,e,f) and dilational processes (a,b). Sibson (1986) has noted that rock dilation in shallow crustal faults results from the transient formation of steep fluid pressure gradients during seismogenic fault rupture. The rock fabrics illustrated in Figure 9(a,b) are probably best understood as products of hydraulic fracturing. The "brittle infrastructure" (Sibson, 1986) of shallow crustal faults is likely to include rock products of faulting in multiple environments, and thus a heterogeneous mixture of deformational fabrics is likely to result.

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Figure 9. Rock fabrics of deformed carbonates at the Nolting and Miller farms.

a. Calcite veinlets (white) cutting fossil-moldic dolostones (gray) in the Welton Member, Scotch Grove Formation. Miller Farm Quarry. Plane polarized light. Field of view is 12.8 mm.

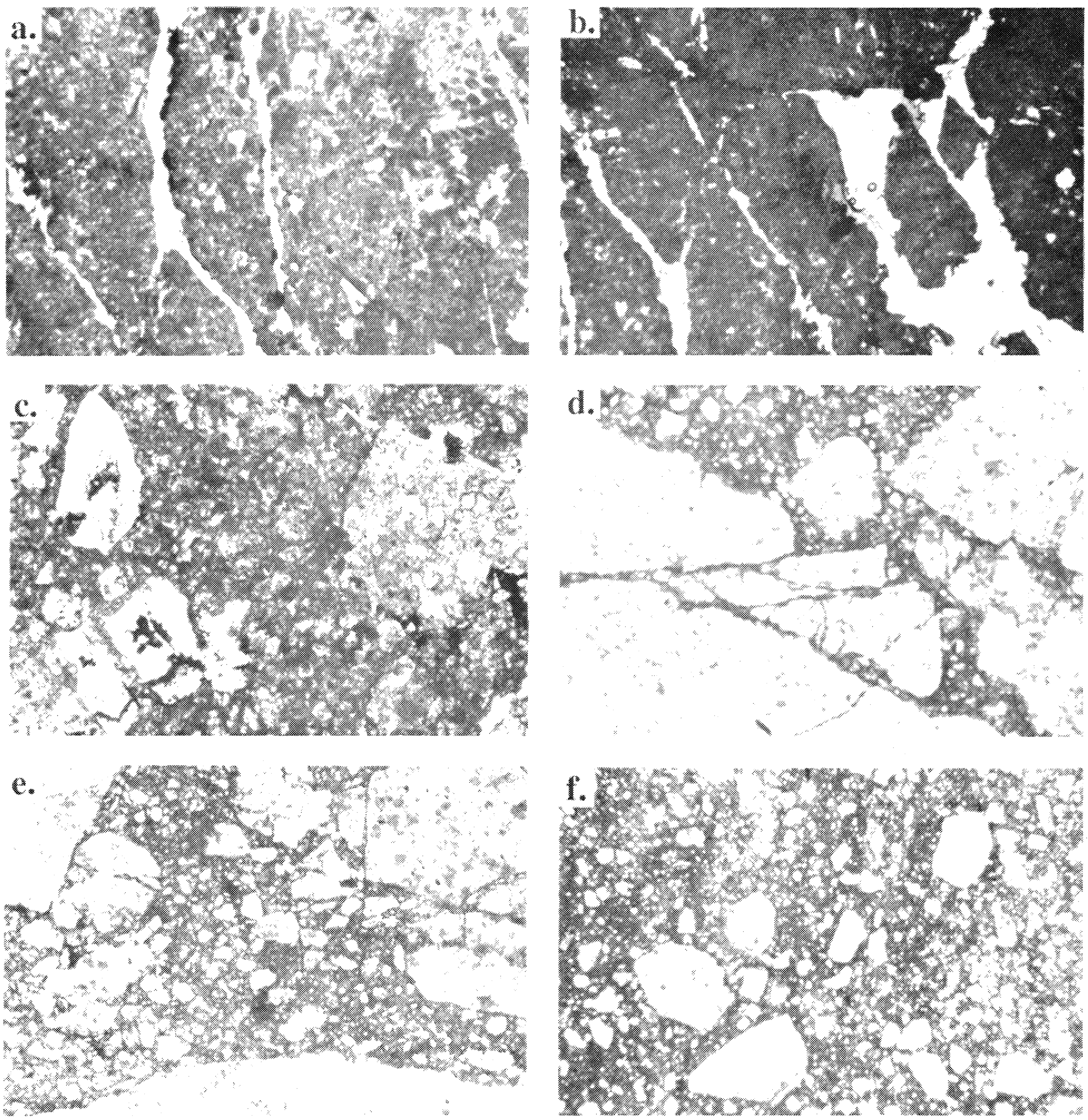
b. Calcite cement (white) filling dilational breccia in dolostones (dark gray) of the Welton Member, Scotch Grove Formation. Miller Farm Quarry. Plane polarized light. Field of view is 12.8 mm.

c. Cataclasite in cherty dolostone of the Sweeney Member, Hopkinton Formation. Chert clasts (white, to left) and dolostone clast (light gray, to upper right) float in a matrix of dolomitic gouge with ferric oxide cements (dark gray). Nolting Farm. Plane polarized light. Field of view is 3 mm.

d. Cataclasite in dense fine-grained dolostone of the Maquoketa Formation. Dolostone (light gray) is cut by microfractures filled by dolomitic gouge (dark gray). Miller Farm. Plane polarized light. Field of view is 3 mm.

e. Cataclasite in dense fine-grained dolostone of the Maquoketa Formation. Microfault filled by dolomitic gouge (dark gray) separates masses of undeformed dolostone (light gray). Miller Farm. Plane polarized light. Field of view is 3 mm.

f. Cataclasite in dense fine-grained dolostone of the Maquoketa Formation. Dolostone clasts (light gray) float in a matrix of dolomitic gouge (dark gray). Plane polarized light. Field of view is 12.8 mm.


a.
b.
c.
d.
e.
f.

STOP 5. Pennsylvanian-Ordovician Unconformity Along the Plum River Fault Zone

Greg Ludvigson

After a short drive further west (Fig. 7), we will walk southward into the valley of a tributary stream to Sugar Creek on the Stewart Farm property, where an unconformity between Pennsylvanian sandstones and the Ordovician Brainard Shale is exposed. Pennsylvanian rocks in the area straddle the PRFZ (Fig. 10), and evidently post-date major vertical movements along the structure (Fig. 11). The sandstones we will be visiting were deposited in a channel complex incised into the Maquoketa Formation along a doubly-plunging, east-west trending anticline bounding the south side of the PRFZ (Figs. 7, 11).

Ludvigson (1985) described the sedimentary facies and framework grain mineralogy of the Pennsylvanian rocks at this locality (then described as the D. Banowetz Farm Outlier). The sandstones occur in two lithogenetic suites. The first consists of poorly-sorted, quartz-pebble conglomeratic, medium- to coarse-grained sandstones, predominantly with trough cross stratification (Fig. 10, DBF-1A, DBF-2). The second consists of well-sorted, fine- to medium-grained sandstones, predominantly with planar cross stratification (Fig. 10, DBF-4). The basal sandstones in both suites contain an abundance of coarse-grained detritus derived from underlying Silurian and Ordovician rocks that were eroded from the paleochannel walls (Fig. 11, see DBF-4, DBF-5, DBF-6). Calamites impressions and small carbonized plant debris are abundant throughout. The sandstones evidently are fluvial in origin. Paleocurrent data from both suites indicate westward sedimentary transport (Fig. 10).

Petrographic studies of the sandstones (Ludvigson, 1985) revealed that the two suites are also characterized by compositional differences. A standard QFL plot shows significant overlap between the populations, all of which are quartz-rich units with substantial quantities of lithic detritus. Examination of lithic grain populations reveals, however, that the quartz-pebble conglomeratic sandstones plot at or near the sedimentary rock fragment end-member, while the fine- to medium-grained planar cross-stratified sandstones contain a significant component of metamorphic lithic detritus, principally foliated quartz-muscovite tectonites. Disregarding the effects of contamination by local shale and dolostone detritus, the compositional fields of the sandstones from this area (Fig. 12) are comparable to those determined from the data reported by Isbell (1985) from confirmed successions of Morrowan (Caseyville Formation), Atokan (Abbott Formation), and Desmoinesian (Spoon Formation) units in the Quad Cities, Illinois-Iowa area. The upward trend toward increasingly immature sandstones in the Lower to Middle Pennsylvanian rocks of the Upper Mississippi Valley has been reported by many workers (see summary in Introduction to Part 1), and Ludvigson (1985) suggested that both Morrowan (quartzarenites) and Atokan-Desmoinesian (feldspathic litharenites) units are preserved at this locality.

The identification of Caseyville strata has now been confirmed by the miospore flora contained in the carbonaceous mudstone at the top of section DBF-2 (Fig. 10; see the paper by Nations in this guidebook). Regional considerations dictate that the feldspathic litharenites must be considered to be Atokan-Desmoinesian in age, and therefore the channel complex at this field trip stop contains multiple fills of fluvial sediments that were deposited by streams with different sedimentological characteristics. The Caseyville stream, while transporting quartz-rich detritus, carried a heavier bed load with rapidly changing discharge conditions, while the later stream(s) carried a lighter bed load under conditions of more constant discharge. While the significance of possible paleoclimatic influences cannot be ignored, these changes are also suggestive of a temporal transition from proximal to distal fluvial regimes, and generally decreasing stream gradients. It is also quite possible that these sandstones record some information on the chronology of late Paleozoic uplift in the Upper Mississippi Valley (see Introduction to Part 1).

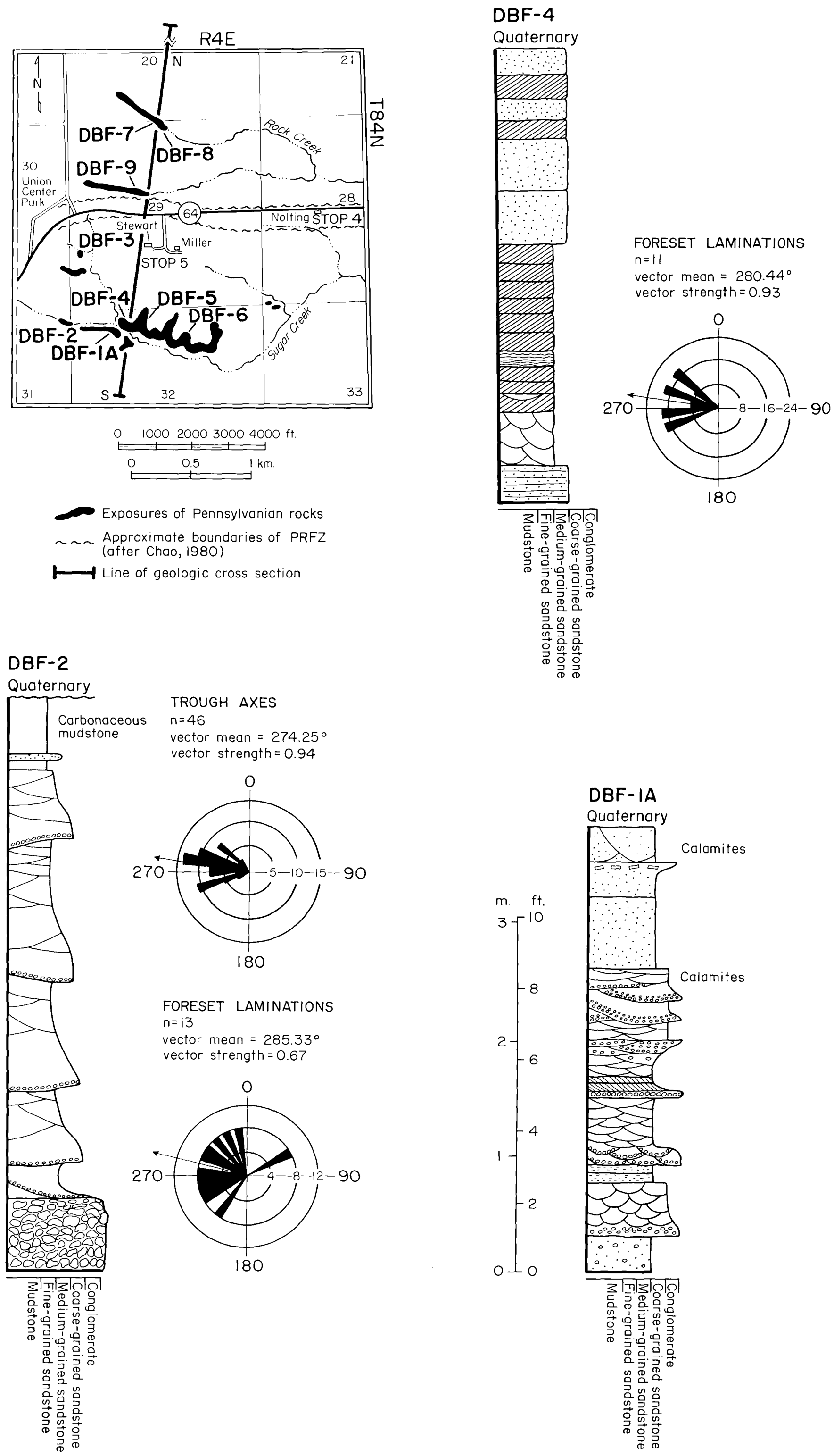


Figure 10. Location map, graphic logs of measured stratigraphic sections, and paleocurrent data from Pennsylvanian sandstones at STOP 5. The Stewart Farm and neighboring properties.

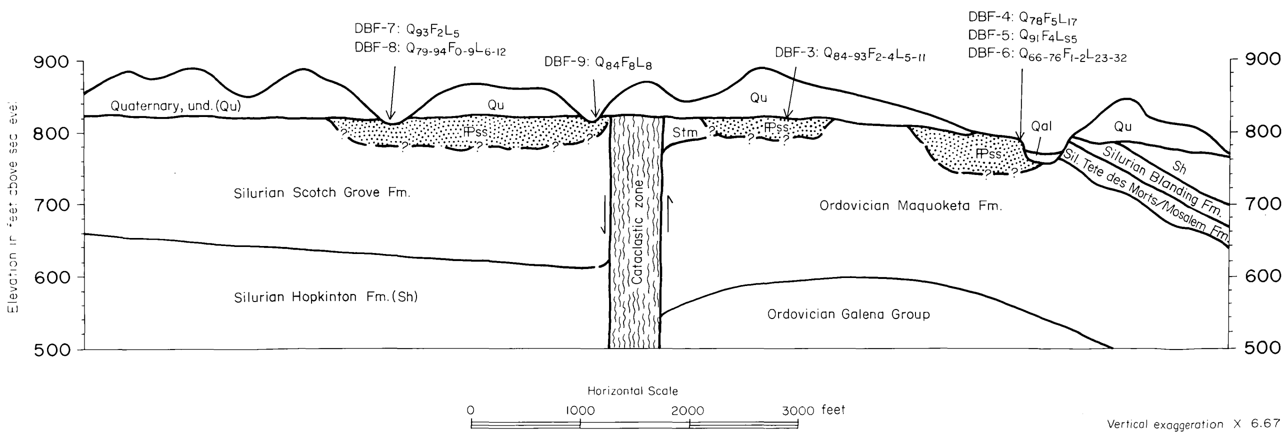


Figure 11. North (left) to south geologic cross-section across the Plum River Fault Zone at STOP 5. Line of section is shown in the preceding figure. Pennsylvanian sandstones post-date major movements along the fault system. From Ludvigson (1985).

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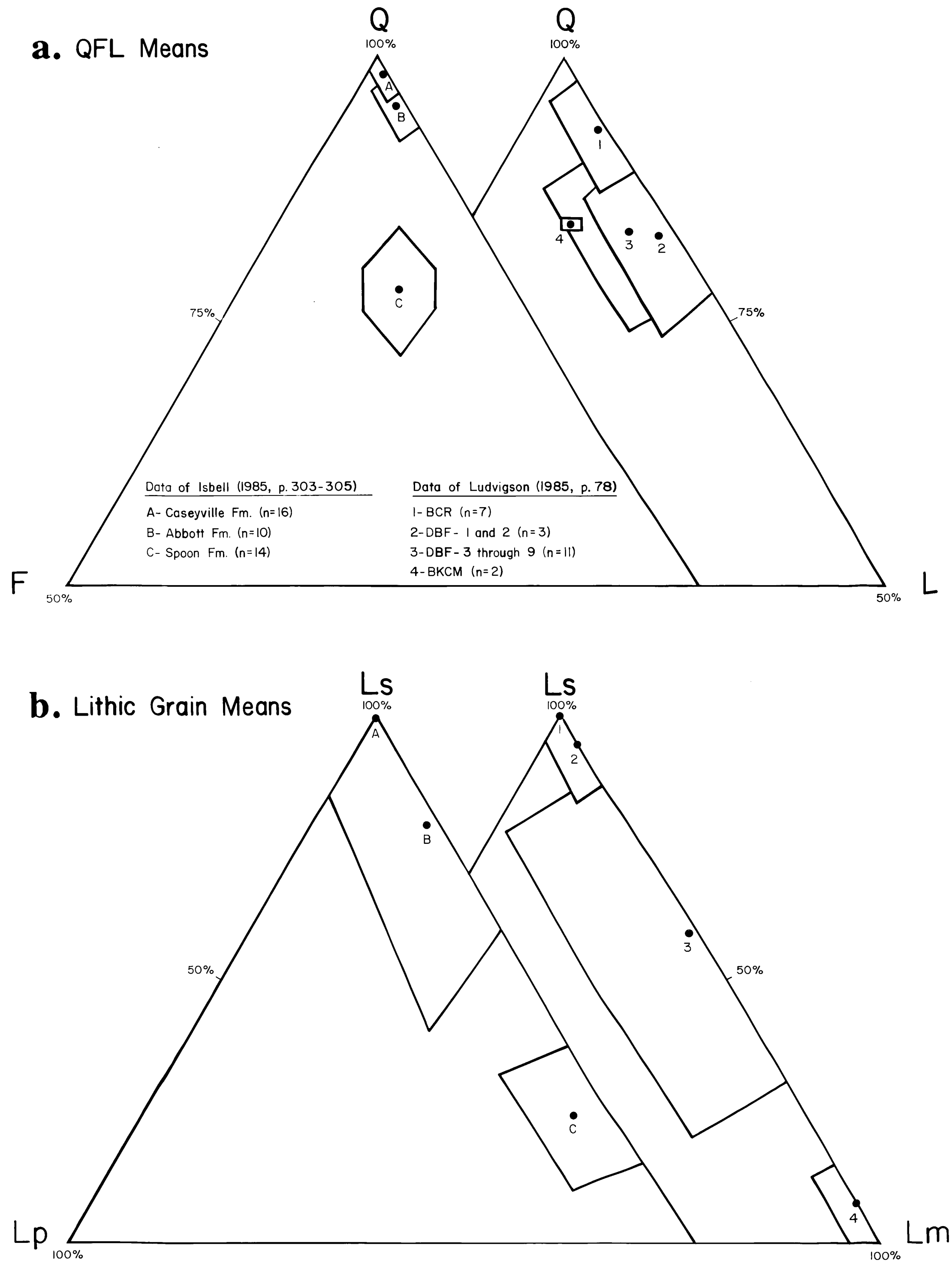


Figure 12. Compositional data on framework grains from Pennsylvanian sandstones. a. QFL (quartz:feldspar:lithic grains) plot. Points are mean values of sample populations (n is number of observations). Polygons enclose standard deviation bars for each grain parameter along the appropriate axis. b. LsLpLm (sedimentary:plutonic:metamorphic lithic grains) plot. Construction is the same as above.

STOP 6. Fault-related carbonate diagenesis in Devonian and Silurian rocks at the Silver Creek Graben

Greg Ludvigson

The Silver Creek Graben is a 150 m (500 ft.) wide block of Middle Devonian limestones that are in fault juxtaposition with Silurian dolostones of the Scotch Grove Formation. This important Devonian outlier was discovered by Dorheim (1953). The bedrock structure at this locality can be recognized by closely inspecting roadcut and natural exposures during a walk along a north-south meandering gravel road (Fig. 13) following the valley of a tributary stream to Silver Creek. The faulted Devonian strata are structurally preserved some 65 km (40 mi.) to the east of the updip erosional edge of the Devonian outcrop belt in eastern Iowa.

An interpretation of the shallow geologic structure along the north-south roadside transect is shown in Figure 14. The graben is evidently cut into multiple fault blocks. The south boundary fault to the graben (SCO-SBF, Figs. 13 and 14) is exposed in the valley walls to the east of the road, and in float to the west of the road. The fault rocks are rusty-colored, brecciated, sparsely to abundantly fossil-moldic dolostones of the Welton Member of the Scotch Grove Formation, with pervasive ferric oxide and calcite cements in the gouge matrix. These rocks abut gray-colored Devonian limestones, chiefly fossiliferous open marine skeletal wackestones/packstones of the Little Cedar Formation of the Cedar Valley Group, but also narrow slices of unfossiliferous lime mudstones of the Wapsipinicon Group (see paper by Bunker in this guidebook).

The Little Cedar Formation is exposed in roadcuts and natural exposures to the west side of the gravel road, and natural exposures and float blocks to the east side. Conodont, brachiopod, and coral faunas collected from the rocks on the east side (Bunker, this guidebook) indicate placement within the lower portion of the Rapid Member, and confirm the northward dip of the fault block containing the Cedar Valley rocks (Fig. 14). Structural attitudes observed in these rocks are difficult to decipher, possibly because of mass wasting processes on the hillsopes, and possibly because of local structural deformation. The Cedar Valley rocks are cut by a plethora of calcitic veins and fissure-fills; stylolites at high angles to bedding have also been observed. The fissures are filled both by spars and geopetal (now rotated to various orientations) unfossiliferous lime muds preserved as microspar. Petrographic and geochemical data presented in following discussions suggest that the fracture-filling calcites in the Cedar Valley strata at this locality were emplaced during syn-depositional movements along the PRFZ.

Further to the northwest, intensely-fractured, unfossiliferous gray lime mudstones of the Davenport Member of the Pinicon Ridge Formation (Wapsipinicon Group) are exposed in the road bed and ditch to the east of the gravel road. Stratigraphic considerations indicate that a fault (not exposed) separates these rocks from the Cedar Valley exposures to the southeast (Fig. 13 and 14).

Fault rocks from the northern boundary fault to the Silver Creek Graben are exposed on the east side of the gravel road, both in the road ditch and in natural exposures (SCO-NBF, Figs. 13 and 14). These are brecciated, sparsely fossil-moldic dolostones of the Waubeek Member of the Scotch Grove Formation, with abundant ferric oxide and calcite cements. Judicious use of the geologic hammer is advised for those participants who are not used to looking at dolomites in the field. Fault rocks and their undeformed protoliths are not necessarily easy to distinguish on weathered outcrop, but freshly-broken rock surfaces usually are diagnostic.

Dolostones of the Waubeek Member of the Scotch Grove Formation (upper part) are exposed in abandoned quarry workings at the northern end of the transect. These are the youngest Silurian rocks that will be visited during this field conference. The stratigraphic position of these rocks indicate a net down-to-the-north fault throw across the Silver Creek Graben, which is bounded to the south by lower Scotch Grove strata (Fig. 14). Approximately 7.5 m (25 feet) of sparse small-crinoid-moldic dolowackestone is exposed at the quarry. Witzke (1981, p.84) collected abundant Rhipidium brachiopods (large

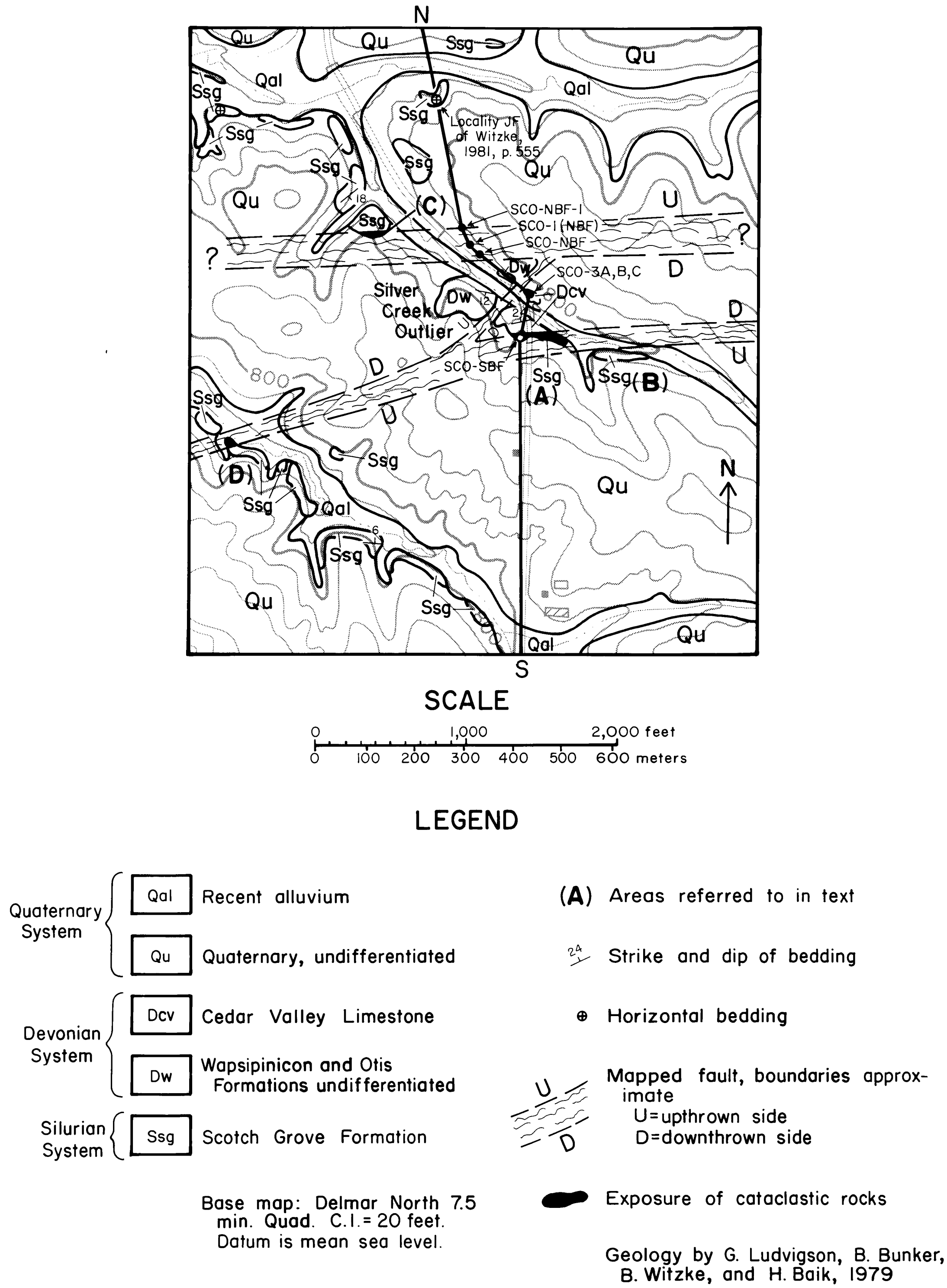


Figure 13. Bedrock geologic map of the Silver Creek Graben. Modified from Bunker et al. (1985) and Ludvigson (1988b).

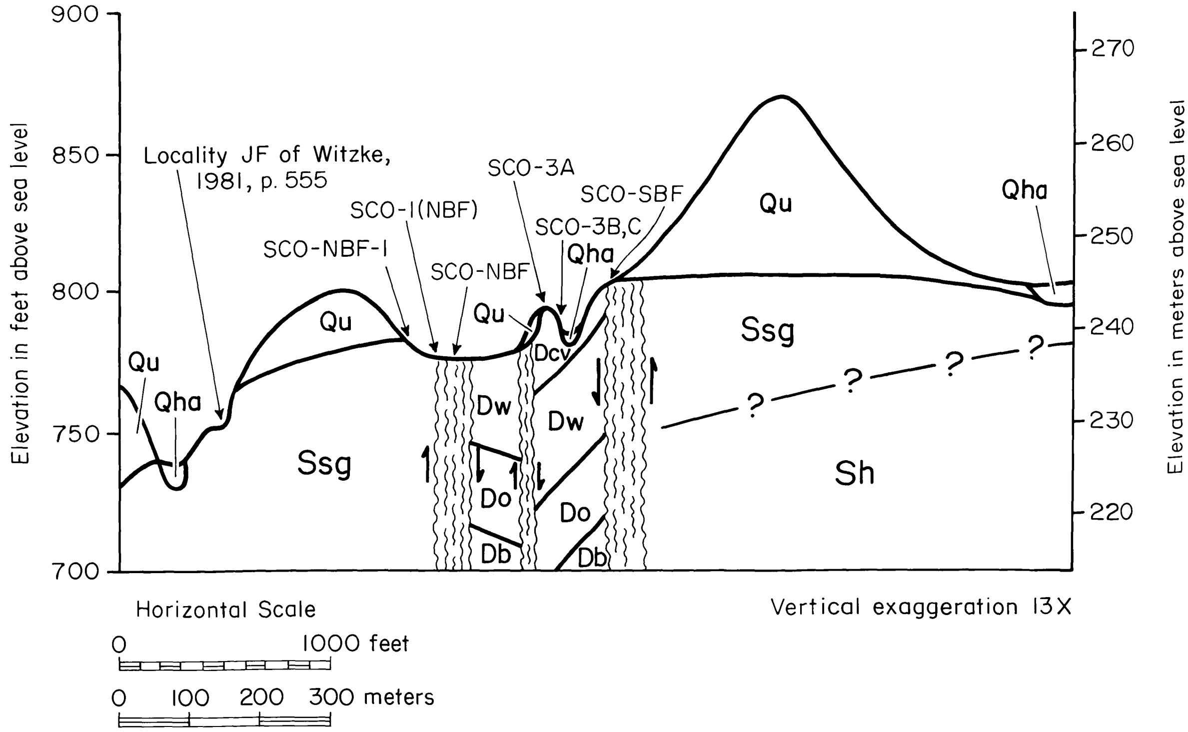


Figure 14. North (left) to south geologic cross-section across the Silver Creek Graben. Line of section is shown in the preceding figure. From Ludvigson (1988b).

Wenlockian pentamerids) from this site, and included it as a reference locality (JF).

Elemental studies of fracture-filling calcites collected from the Silver Creek Graben and its two bounding faults clearly show two distinctive covariant trends in Fe/Ca and Mg/Ca ratios (Fig. 15). These data indicate that fracture calcites in the graben and those in the faults are two separate generations, each with distinctive chemical environments of precipitation. Arguments are presented in Ludvigson (this guidebook) and in the discussion for Sunday STOP 2 that the cements in the boundary faults were precipitated from groundwaters that infiltrated beneath the regional sub-Pennsylvanian erosion surface. Cathodoluminescent micrographs of zoned calcite cements from the boundary faults at this locality are shown in figure 9 (e,f) of Ludvigson (this guidebook). The fracture calcites in the Devonian rocks, however, appear to be products of earlier Devonian deformational episodes along the PRFZ (Ludvigson, 1988a,b).

Rock fabrics and stable isotopic data from microdrilled samples of fractured Cedar Valley strata are shown in Figures 16 and 17. Composite fissure-fills, described as neptunian dikes in Figure 16, were formed by incremental openings of dilational fractures, with each increment filled by a distinctive generation of internal sediments. Multiple generations of cross-cutting calcite veinlets are shown in Figure 17. All of the labelled fabric constituents in Figures 16 and 17 were analyzed for isotopic and trace element chemistry in order to interpret environments of diagenesis.

Multiple electron microprobe analyses of the fabric constituents in the Cedar Valley rocks are shown in a covariant cross-plot of Mn and Sr contents in Figure 18. They cluster in two populations: one with elevated Mn contents with Sr at or below the detection limit of the microprobe (100 ppm), and another with elevated concentrations of both Mn and Sr. The first population fits a trace element pattern that is typical of marine carbonates that stabilized in

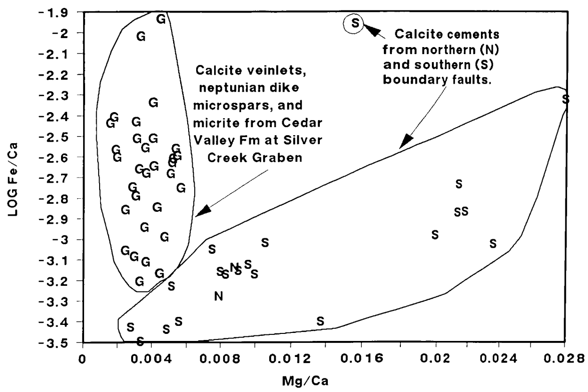
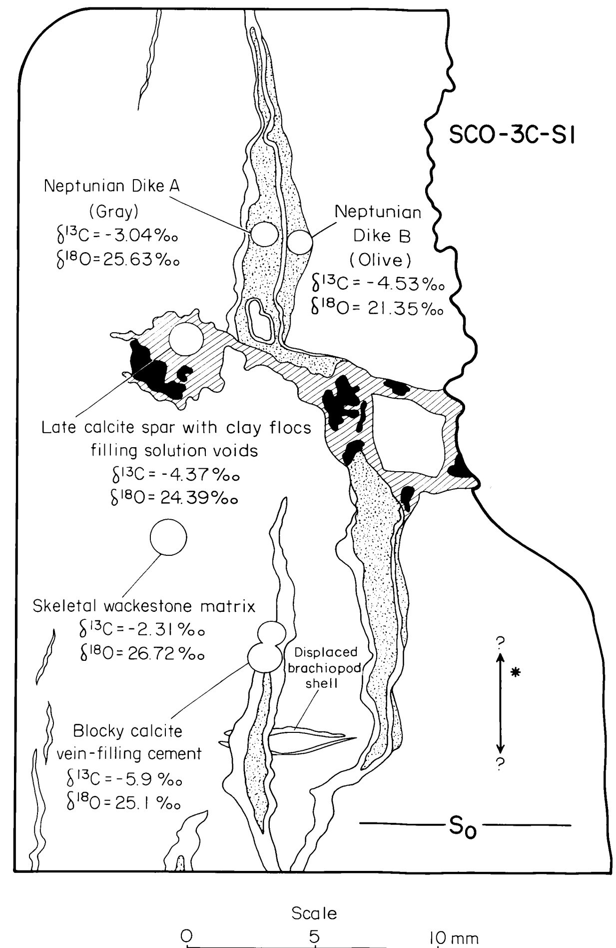


Figure 15. Cross plot of Fe/Ca and Mg/Ca ratios in fracture-filling calcites at the Silver Creek Graben. From Ludvigson (1988b).

equilibrium with shallow meteoric waters near the site of recharge (see paper by Ludvigson, this guidebook). The second population, with elevated concentrations of both Mn and Sr, is indicative of marine carbonates that stabilized in equilibrium with evolved meteoric waters remote from the site of recharge (ibid.). Moreover, the elevated Sr contents suggest that the veinlets formed before the diagenetic conversion from metastable marine carbonates to stable low-magnesian calcite had been completed in the enclosing host rocks (ibid.). This suggests that the veinlets might not post-date their host rocks by great expanses of geologic time.

Carbon and oxygen isotopic data from the rocks at this locality are shown in Figure 19. With respect to the constituents in the Cedar Valley rocks, micrites in the skeletal wackestones have the least depleted values, suggesting that they most faithfully record the composition of carbonates precipitated in equilibrium with Middle Devonian sea water. Two other clusters are apparent: a first cluster with $\delta^{18}O$ values that are slightly depleted with respect to the micrites, but with highly variable and depleted $\delta^{13}C$ values, and a second cluster that is significantly depleted in both ^{18}O and ^{13}C with respect to the micrites. The micrite data and the neighboring cluster fit an isotopic pattern that is typical for marine carbonates stabilized by early meteoric diagenesis. This so-called "inverted J" trend (Lohmann, 1982) is a mixing curve that connects fabric elements whose isotopic chemistry was most influenced by the composition of the original marine carbonate phases (micrites), and fabric elements whose isotopic chemistry was most influenced by the composition of the meteoric water (veinlet #1). The second cluster, with significantly depleted $\delta^{13}C$ and $\delta^{18}O$ values, is interpreted as a population of fabric constituents (later generation vein-filling calcite cements and fissure-filling lime muds) that precipitated in burial environments at temperatures perhaps up



* Sample was collected from an exposure of vertically-dipping strata. Facing direction is not known.

Figure 16. Drawing of fabric relationships in a compound fissure-fill from the Little Cedar Formation at Silver Creek Graben. Circles show the positions of drilled stable isotope microsamples. From Ludvigson (1988b).

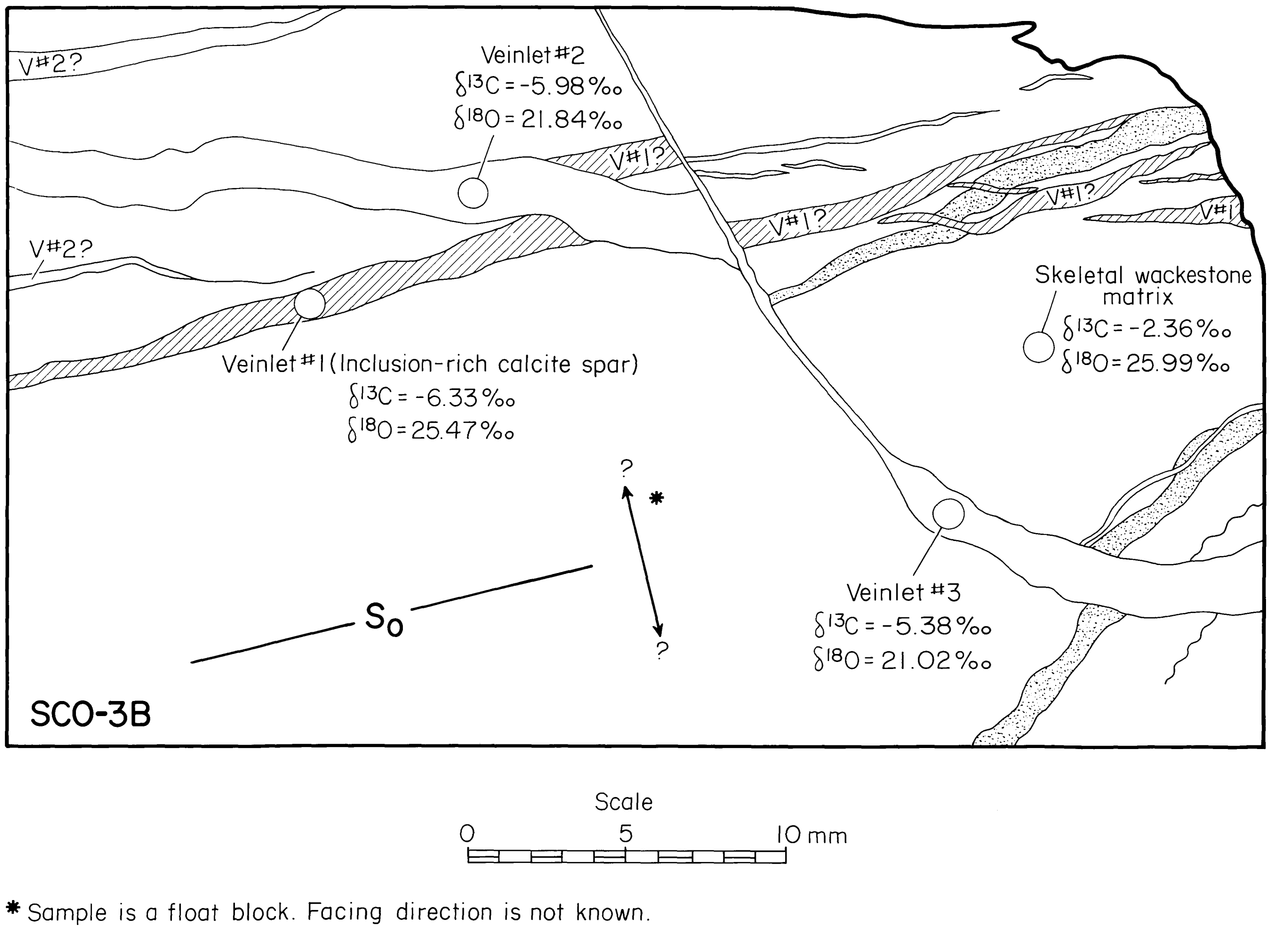


Figure 17. Drawing of fabric relationships between cross-cutting calcite veinlets in the Little Cedar Formation at Silver Creek Graben. Circles show the positions of drilled stable isotope microsamples. From Ludvigson (1988b).

to 15° C warmer than those in the first cluster. In other words, the late veinlets formed in a burial environment, and the late fissure-filling muds, which had been emplaced earlier in the rock's history, were neomorphosed to microspar in a burial environment. Note that all of the veinlets have elevated Mn and Sr contents suggestive of deposition by evolved meteoric fluids, while only the later veinlets were precipitated at elevated temperatures. Note also that the late fissure-filling microspars do not have elevated Sr contents. This observation is in accord with the suggestion by Lohmann (1978) that neomorphosed fabric constituents in limestones retain their original trace element compositions, while their isotopic ratios are reset by the environment of neomorphism.

The fissure fills at this locality pose a dilemma: while fissure-filling sediments are well-known fabric constituents in carbonates from supratidal to nonmarine environments, and are also known from carbonate mound facies that are subject to periodic exposure, here they occur in a thick sequence of subtidal carbonate strata. In addition, similar features from this stratigraphic interval have not been found in other localities. Figure 20 shows an interpretation of the stratigraphic origin of fissure-filling sediments in Cedar Valley rocks at Silver Creek Graben. The stratigraphic column is a measured section of Cedar Valley strata from the same facies tract as this locality. At the top of the section is an interval

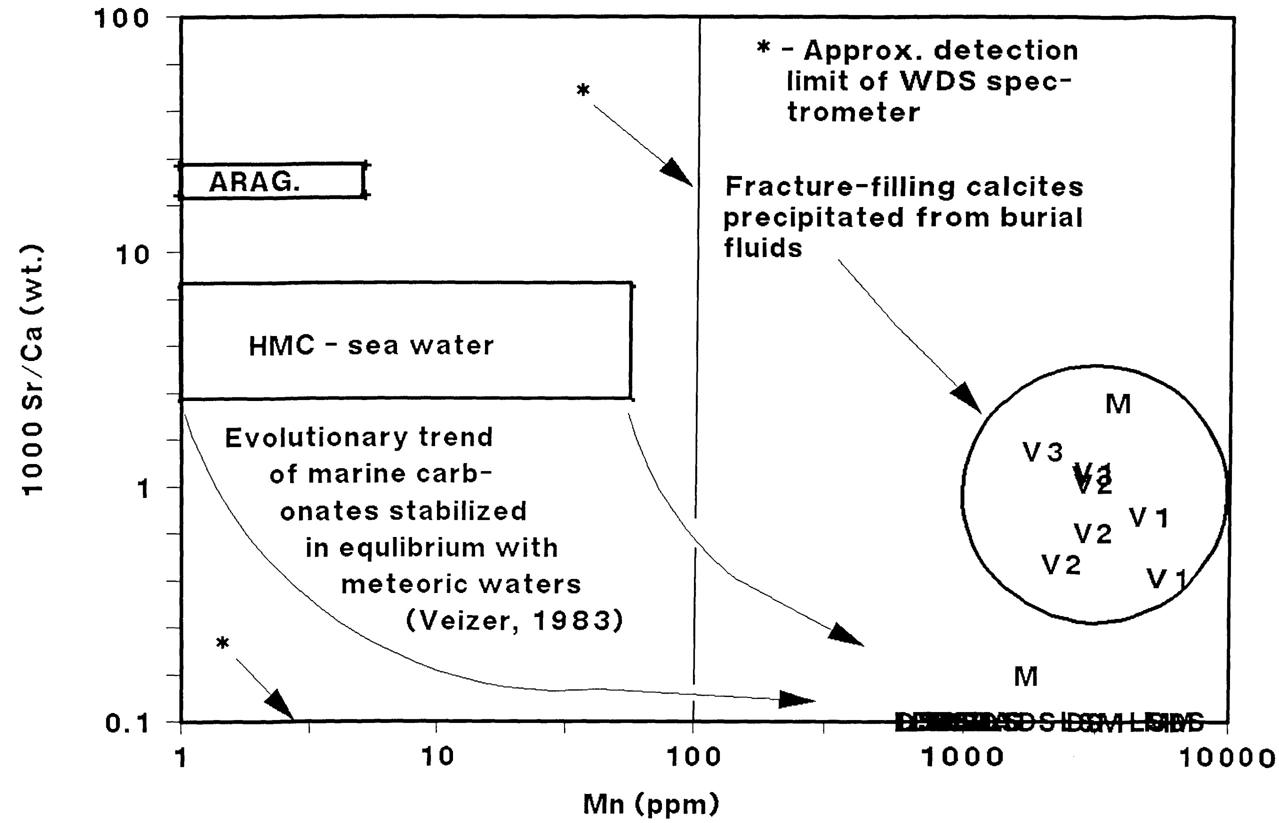


Figure 18. Cross plot of strontium and manganese contents in calcitic fabric constituents in the Little Cedar Formation at Silver Creek Graben. Rectangles to the left show the theoretical ranges of concentration in marine aragonite and high-magnesian calcite. The dense cluster of points to the lower right shows the position of most fabric constituents shown in figures 16 and 17. Other fabric constituents are: M, micrite; V1, veinlet #1; V2, veinlet #2; and V3, veinlet #3.

of fenestral, stromatactis-bearing calcilutites known to contain abundant internal sediments preserved as microspar (Kettenbrink, 1973). Further similarity between the internal sediments at this locality and those from the fenestral calcilutites in Figure 20 is established by the mutual presence of abundant illitic clays (Kettenbrink, 1973; Ludvigson, 1988b). Illite was first detected in the microspars from Silver Creek Graben by EDX spectra, and confirmed by SEM photomicroscopy showing rounded detrital illite inclusions embedded in the microspar. Note that all of the strata intervening between the exposed interval at Silver Creek Graben and the proposed source interval are marine deposits (Fig. 20). Since the fissure-filling sediments are enclosed by early meteoric calcite spars (Fig. 16) and must have infiltrated through meteoric pore fluids, none of the intervening strata are reasonable candidates for the internal sediments in the fissures at Silver Creek Graben. Thus the most proximate source was an exposure surface at the top of the Coralville Formation.

How could the internal sediments have infiltrated through several tens of meters to fill the fractures at this locality? The fabrics of the fissures provide some answers. Fabric relationships indicating multiple openings of extension fractures, as in Figures 16 and 17, show that rock dilation was a continuing, ongoing process. Rock samples with abundant fissure fills are dilational breccias that were filled by multiple increments of sediment infiltration. These relationships, and the structural context of this locality, indicate that the fissures are of tectonic origin. The purely dilational character of the

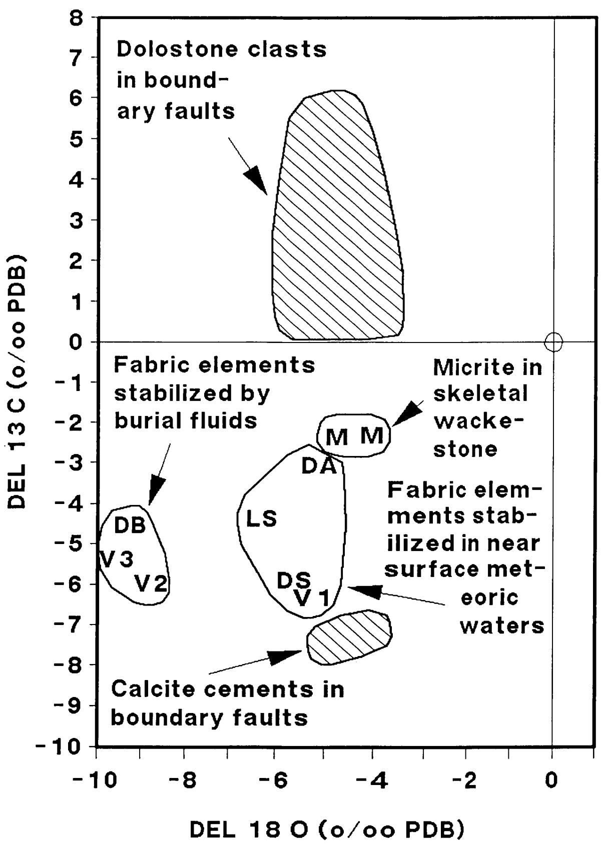


Figure 19. Plot of carbon and oxygen isotopic ratios from fabric constituents shown in figures 16 and 17. Ruled areas show fields occupied by carbonate constituents from Silurian rocks in the boundary faults. Fabric constituents are: M, micrite; DS, blocky calcite vein-filling cement lining neptunian dike; DA, neptunian dike A (gray microspar); DB, neptunian dike B (olive microspar); V1, veinlet #1; V2, veinlet #2; V3, veinlet #3; LS, late calcite spar with clay flocs. Modified from Ludvigson (1988b).

movements suggests that cracks were opened by hydraulic fracturing, as a consequence of transient fluid pressure disturbances during coseismic movements along the PRFZ (e.g. Sibson, 1986). This deformational episode apparently occurred during the Late Givetian exposure of a southeastward-prograding shallow marine carbonate shelf (see paper by Bunker, this guidebook).

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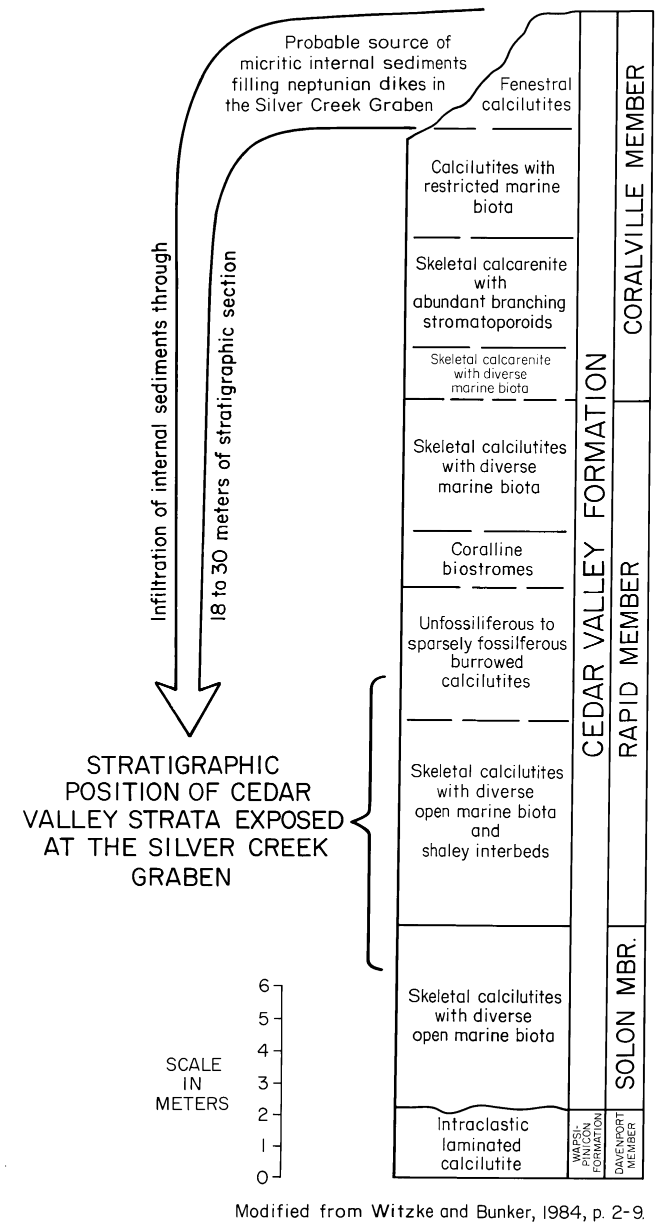

Figure 20. Interpretation of the stratigraphic origin of internal sediments from the Coralville Formation filling fissures in the Little Cedar Formation (Solon and Rapid Members). For clarification of stratigraphic terminology see paper by Bunker (this guidebook). From Ludvigson (1988b).

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SUNDAY

STOP 1. Exposures of the Silurian-Ordovician contact at Bellevue State Park. The Mosalem, Tete des Morts, and Blanding formations

Greg Ludvigson and Brian Witzke

The basal Silurian successions being examined this morning were not exposed at the previous stops, except for a brief glimpse at STOP 4. For the most part, these rocks have either been eroded from the upthrown side of the PRFZ, or buried by younger strata on the downthrown side. The Mosalem, Tete des Morts, and Blanding formations (Fig. 21) buried a major erosional unconformity that separates the Ordovician and Silurian systems of the Upper Mississippi Valley (see paper by Witzke and Kolata, this guidebook). The deepest known incision into the sub-Silurian surface in this region occurred in the general area of this field trip, although the broader configuration of the drainage network remains unclear. The Mosalem, Tete des Morts, and Blanding formations successively buried the sub-Silurian erosion surface, and achieve their greatest aggregate thickness in the areas of the deepest incision, as seen here.

Rock exposures at Bellevue State Park and immediate environs provide access to approximately 68 m (223 ft.) of Ordovician and Silurian strata. The uppermost beds of the Dubuque Formation (Fig. 21, also see STOP 2) are exposed in the creek bank and riffles in the bed of Mill Creek, adjacent to Potter's Mill. The Maquoketa Formation is poorly exposed for the most part at this locality, although some sections are accessible along the cliff face.

Exposures of the Mosalem Formation are accessible along the cliff face, and large talus blocks below the cliff are also instructive for examining sedimentary features, including spectacular burrow traces and ripplemarks on bedding surfaces. Upper portions of the cliff section are accessible by trail from the overlook parking lot in the park. Pulpit Rock, a rock spire of Tete des Morts Formation, is located to the north of the parking lot. To the south, the Indian Mound trail leads to a series of old dimension stone quarries that were benched into a southward-sloping ridge along the edge of the cliff. Excercise extreme caution if you choose to examine these exposures!

One of the most remarkable aspects of the Mosalem sequence at this locality is the succession of rhythmic couplets of resistant and recessive beds, each several to tens of centimeters in thickness. The origins of these features (differences in grain size or volume of siliciclastic inputs? possible tidal influences on sedimentation? differences in mode of carbonate diagenesis?) are not known because modern, comprehensive sedimentological and petrologic studies of the the Mosalem Formation have not been undertaken. Is anyone here looking for a research project?

Faunas from the lower Mosalem include lingulid brachiopods, graptolites, and <u>Lecthaylus</u> soft-bodied worms in laminated to ripple laminated units. <u>Chondrites</u> burrow networks are developed throughout, except for some finely-laminated units, and early sulfidic diagenesis is manifest by pyrite nodules. These strata pass upward into argillaceous dolostones with trilobites and articulate brachiopods, and finally units with echinoderm debris, brachiopods, gastropods, and corals. Ripple lamination and <u>Chondrites</u> burrows are characteristic features. Oxygen-stressed to anoxic environments in restricted <u>shallow water</u> are indicated in the lower Mosalem, whereas deeper more normal marine sedimentation characterized the upper part of the unit.

The Tete des Morts is a massive cliff-forming unit that forms overhanging ledges near the top of the cliff section at Bellevue State Park. The lowermost beds include horizontally laminated strata (unit 15, Fig. 21), but most of the Tete des Morts is a vuggy dolostone mottled by microporous networks that resemble thalassinoid-type burrows. Faunal content is largely obscured by diagenetic alteration, but echinoderms, laminar stromatoporoids, tabulate and cup corals, and brachiopods are noted.

The bedded cherts at the top of the cliff section at Bellevue are included in the Blanding

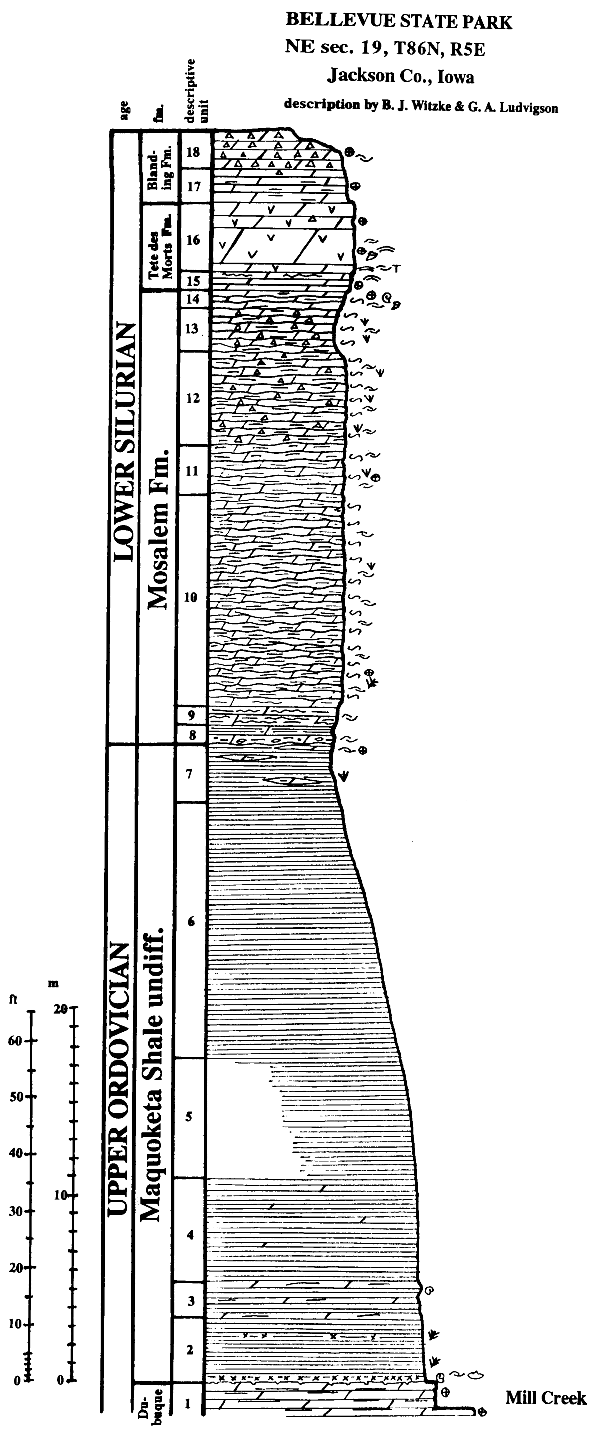

Figure 21. Graphic log of the stratigraphic section at Sunday STOP 1. Bellevue State Park.

Formation. This dolostone unit is notable for the abundance of nodular and bedded cherts 5 to 10 cm thick. Both the dolostone and chert interbeds contain open marine faunas, including echinoderm debris and brachiopods.

BELLEVUE STATE PARK SECTION

bluff and roadcut section along Hwy 52 south of park entrance
SE SE NE NE and E 1/2 SE NE sec 19, T86N, R5E
basal section along Mill Creek NW NE sec 19, T86N, R5E
Jackson County, Iowa
description by B.J. Witzke and G.A. Ludvigson
(supplemental descriptions by R.A. Davis, 1965, M.S. thesis, Univ. Iowa)

LOWER SILURIAN

BLANDING FORMATION

UNIT 18

Dol., xf-f xlln, even-bedded 10-20 cm; very cherty, nodular chert bands and bedded cherts 5-10 cm thick, chert is chalky to smooth, bedded cherts are primarily smooth chert; echinoderm debris and indet. brachiopods noted in cherts; about 2.0 m thick.

UNIT 17

Dol., xf-vf xlln, dense ledges, slightly argillaceous, in beds 7 to 20 cm thick; scattered nodular chert in upper part; echinoderm debris noted; sharp contact at base; these are the "lower quarry beds" of Calvin; 1.8 m thick.

TETE DES MORTS FORMATION

UNIT 16

Dol., vf-f and vf-m xlln, vuggy to very vuggy, mottled with microporous networks (probably large thalassinoid-type burrows); in beds 30 cm to 1.2 m thick, prominent cliff-former, forms overhanging ledges in main cliff section; small fossil debris molds scattered, f-m xlln fabrics probably replaced crinoid debris; generally non-cherty but some silicified grains and chalcedony-linings present, rare small chert nodules in upper part; macrofossils are scattered, most noteworthy in the lower half, and include silicified laminar stromatoporoids (to 25 cm wide), tabulate corals (favositids), cup corals, and indet. brachiopods; 3.6 m thick.

UNIT 15

Dol., vf-m xlln, denser than above, in beds 7-28 cm thick, burrow networks are more coarsely xlln; slightly porous but vugs are rare; upper bed with prominent horizontal laminae; small crinoid debris represented by molds and replaced grains; silicified laminar stromatoporoid noted near middle; 98 cm thick.

MOSALEM FORMATION

UNIT 14

Dol., xf-vf and vf-m xlln, slightly argillaceous, in beds 5-20 cm thick, becomes generally thicker bedded upward, beds have undulose surfaces separated by thin shaley partings (1-10 mm thick), faintly laminated in part with some probable ripple laminae; Chondrites burrows scattered to common through (less common than underlying units); m xlln skeletal packstone bed near top, grains silicified in part; silicified skeletal debris scattered to common along some bedding surfaces and laminae; fossils include brachiopods (Dalmanella, orthids, atrypids, rhynchonellids), echinoderm debris, gastropods, corals (3 cm horn coral noted near base); 1.0 m thick.

UNIT 13

Dol., xf-vf xlln, slightly argillaceous to argillaceous, in beds 2-15 cm thick, undulose bedding surfaces separated by shaley partings 1 mm to 3 cm thick, argillaceous laminae and some probable ripple laminae present; unit becomes generally less argillaceous and shaley upward; Chondrites burrows scattered to abundant; small nodular cherts in irregular bands (1-5 cm thick), chert follows horizontal burrow networks in part, cherts are primarily chalky and argillaceous but some nodules are cored by smooth cherts in upper 1 m; skeletal debris scattered to common along some bedding surfaces and laminae, in part silicified; fauna includes common to abundant brachiopods (Dalmanella cf. edgewoodensis, Plectatrypa, Strophonella, etc.), common disarticulated trilobites (Calymene, proetid, odontopleurid), scattered echinoderm debris (columnals), scattered gastropods and bivalves (Pterinea), rare bryzoans (small cryptostomes), rare corals (small cup coral); 2.45 m thick.

UNIT 12

Dol., xf-vf xlln, argillaceous, in beds 2-15 cm thick, wavy bedding surfaces separated by shaley beds 1-4 cm thick; faint argillaceous laminae; common to abundant burrow mottles (primarily Chondrites); small chalky chert nodules scattered through unit; scattered silicified skeletal debris includes brachiopods (?Dalmanella, others) and trilobite fragments; 4.8 m thick.

UNIT 11

Dol., xf-vf xlln, argillaceous, thin to medium interbedded arg. dol. (2-15 cm) and shaley dol.(4-7 cm), wavy bedded, less argillaceous beds form resistant ledges along weathered cliff face; beds with faint argillaceous laminae in part, some display probable ripple laminae; interval with common to abundant burrow mottles (primarily Chondrites but including larger horizontal burrows); 50 cm below top is a thin f xlln dol (possibly fine skeletal packstone); scattered skeletal grains in unit (some silicified), including trilobite debris, sparse indet. brachiopods, rare echinoderm debris; 2.7 m thick.

UNIT 10

Dol., xf-vf xlln, argillaceous, silty in part, interbedded lighter-colored less argillaceous bands and darker-colored shaley units, wavy bedded 3-20 cm thick, lighter bands form more resistant ledges on weathered cliff face; faint argillaceous laminae through most of unit, may include some ripple laminae; common to abundant burrow mottles, primarily Chondrites but including larger horizontal (2 mm to 2 cm diameter) and subvertical burrows (in part weathering out as resistant dolomite in more argillaceous matrix), some burrows are dark gray to black with carbonaceous material; scattered bands of vf-f xlln dol 1-3 cm thick form thin resistant ledges in lower to middle part, in part pyrite-cemented (oxidized to ferric oxides); unfossiliferous to sparsely fossiliferous through much of unit, although some scattered articulate brachiopods (rhynchonellids) and other indet. skeletal debris (possible Cornulites-like tube noted) are present; basal portion of unit contains early Llandoverian graptolites (Diplograptus cf. modestus, see C.A. Ross, 1964, Jour. Paleon., v. 38, p. 1107) and other fauna (Lecthaylus soft-bodied worms, Sphenothallus tubes, Metaconularia, Lingula, bivalves, ceratiocarids; T.J. Frest, 1988, pers. comm.); unit forms prominent cliff face along Hwy 52, upper part of unit difficult to access; about 11.3 m thick.

UNIT 9

Dol., m. brown to brown-green, very argillaceous to shaley, slightly carbonaceous, less resistant than overlying unit but forms base of cliff-forming sequence, lower 35 cm is blockier and more dolomitic than upper 60 cm (shalier); entire interval is prominently and finely laminated (laminae < 1 mm), argillaceous laminae are flat and continuous on outcrop; top 15 cm has two lighter-colored dolomitic bands 4-6 cm thick; unit is generally unburrowed, although scattered burrow traces and indeterminate carbonaceous debris (probably including graptolite blebs) are seen along some laminae; macrofossils are scattered to common in a few thin horizons, primarily inarticulate brachiopods (Lingula cuneata, orbiculoids); 95 cm thick.

UNIT 8

Shale and dol.; upper 40 cm is dolomitic shale, green-gray to brown-gray, plastic when wet, chunky, contains thin platy flags of argillaceous dolomite, faint laminae, poorly exposed as recessive at base of cliff section; lower 60 cm is a reddish-brown argillaceous dolomite, silty to sandy in part, contains scattered clasts of phosphatic nodules and Maquoketa shale fragments, reworked phosphatic and dolomitic skeletal grains, basal portion with abundant iron oxides, ripple marks present in interval, gradational with upper shaley beds; unit is poorly exposed, description of lower 60 cm largely derived from R.A. Davis (1965, M.S. thesis, Univ. Iowa); 1.0 m thick.

UPPER ORDOVICIAN

MAQUOKETA SHALE undifferentiated

(based on regional thickness patterns the upper Maquoketa Shale has been erosionally truncated at Bellevue; preserved section is probably equivalent to the Elgin-Clermont-Fort Atkinson sequence in northeast Iowa; Brainard Shale equivalents are apparently absent)

UNIT 7

Shale, green-gray, dolc, unit is poorly exposed at present, contains some interbeds (1-8 cm) of argillaceous dolomite, mostly unfossiliferous to sparsely fossiliferous (3 cm Isotelus cranidia noted); skeletal wackestone to packstone locally at top with echinoderm debris, bryozoans, and brachiopods (incl. Thaerodonta); about 3.2 m thick.

UNIT 6

Shale, green-gray, dolc, slumped and poorly exposed over much of section but most of unit is exposed at south end of roadcut, plastic when wet; no carbonate interbeds noted; unfossiliferous shale with pyrite (limonite) concretions (to 5 cm) in some horizons; 13.8 m thick.

UNIT 5

Shale, green-gray, dolc; completely covered in Bellevue section; shale-dominated sequence represented in nearby well sections; about 6.2 m thick.

UNIT 4

Shale, green-gray, dolc, with thin interbeds of unfossiliferous argillaceous dol.; mostly slumped and overgrown, but intermittently exposed in cut bank of Mill Creek (also units 2 & 3); about 5.4 m thick.

UNIT 3

Shale, brown-gray to green-gray, with interbeds of dense, argillaceous dol.; upper ledge contains phosphatized diminutive gastropods; 2.2 m thick.

UNIT 2

Shale, m. brown to brown-gray, dolc, blocky, organic-rich (especially in lower part), laminated in part, scattered graptolites (Orthograptus truncatus peosta); basal 6 cm is phosphorite, pyrite-cemented in part, composed of concentrically-laminated apatite pellets (< 1 mm), contains scattered irregular apatite clasts (to 5 cm), common phosphatized molds of mollusc-dominated diminutive fauna (gastropods, bivalves, Plagioglypta, Septemchiton, 3-D graptolite molds, etc.); thin phosphatic horizon near top; base overlies irregular phosphatized discontinuity surface (hardground) at top of Dubuque Fm.; this unit encompasses the lower part of the "brown shaly unit" of Brown & Whitlow (1960, USGS Bull. 1123A); 3.2 m thick.

GALENA GROUP

DUBUQUE FORMATION

UNIT 1

Dol., vf-m xlln, crinoidal, medium-bedded ledges separated by thin dolc shale partings (1-2 cm); exposed as ledges in lower banks and streambed of Mill Creek near Potter's Mill; about 1.5 m thick.

STOP 1B. Exposures of Lower Silurian strata along Highway 62 west of Bellevue; Tete des Morts, Blanding, and Hopkinton formations

Brian Witzke and Greg Ludvigson

A series of roadcut and quarry exposures along Hwy 62 about 2 miles southwest of Bellevue completes our examination of the Silurian stratigraphy in the Jackson County area. The basal part of these exposures overlaps with the Bellevue State Park section, and the upper beds approximately overlap the basal Z34 roadcut section visited yesterday. The Hwy 62 exposures display an easily accessible sequence spanning the entire Blanding Formation and most or all of the Sweeney Member of the Hopkinton Formation (Fig. 21b). The lower units are included in the Tete des Morts Formation, which forms the upper cliff face at Bellevue State Park. The Tete des Morts is a thick-bedded dolomite, vuggy in part, and contains moldic to replaced crinoid debris and scattered laminar stromatoporoids and tabulate corals.

The basal portion of the overlying Blanding Formation was termed the "lower quarry beds" during the early years of the Iowa Geological Survey under Sam Calvin. The thin beds split into easily quarried layers, and the interval was widely quarried in eastern Iowa during the 1800s for foundation stone. The "lower quarry beds" are dense and argillaceous and contain a sparse fauna of brachiopods, bryozoans, crinoid debris, corals and other fossils. The remainder of the Blanding Formation is characterized by thin to medium beds of dolomite with scattered corals and stromatoporoids; argillaceous content generally decreases upward. The most characteristic feature of the Blanding is an abundance of chert, which occurs as discrete beds of chert (to 20 cm thick), bands of nodular cherts, and scattered chert nodules. The abundance of chert serves to distinguish the Blanding from all other Silurian rock units in the Jackson County area. The type section of the Blanding is designated to the north in Jo Daviess County, Illinois (Willman, 1973).

The upward change from very cherty dolomite to more resistant non-cherty and sparsely cherty dolomite marks the boundary between the Blanding Formation and the overlying Hopkinton Formation. The portion of the Hopkinton sequence exposed in the quarries and roadcuts along Hwy 62 is assigned to the Sweeney Member, a rock unit characterized by very fine to medium crystalline dolomite with a conspicuous fauna of silicified to moldic tabulate corals, solitary rugosans, and laminar stromatoporoids. Crinoid debris is ubiquitous and occurs as molds or as silica or dolomite replacement; medium to coarse crystalline dolomite lenses or layers in the interval are dolomitized crinoidal packstones. The Sweeney is much less cherty than the underlying Blanding, but scattered nodules occur in the middle part (with common to abundant chert nodules in unit 18).

The type Sweeney section (16.8 m thick) is designated at Mississippi Palisades State Park in Carroll County, Illinois (Willman, 1973), about 18 miles (29 km) southeast of the Bellevue West section. Willman (1973) observed a zone with the brachiopods Stricklandia (= "Microcardinalia") and Pentamerus oblongus about 4.5 to 7.6 m (15-25 ft) above the base of the Sweeney in the type area, and Pentamerus and Stricklandia are noted at a position about 7.5 m (24 ft) above the base of the Sweeney at the Bellevue West section (unit 20). Johnson (1983) apparently excluded Pentamerus-bearing beds from the Sweeney Member in Iowa (which he listed as 9.5-10 m thick), and observed common to abundant Stricklandia lens progressa in the mid Sweeney. Nevertheless, well-preserved Pentamerus oblongus occur within the Sweeney interval at Bellevue West. Intervals bearing Pentamerus oblongus vary significantly in thickness in eastern Iowa, ranging from 2.8 to 9.8 m (9-32 ft) thick; Pentamerus-bearing units interfinger with Sweeney-like lithologies in the lower Hopkinton at many localities in eastern Iowa over a stratigraphic range nearly as thick (Bunker et al., 1985). Willman (1973) defined the contact with the overlying Marcus Formation in Illinois (Marcus Mbr of Hopkinton in Iowa) at the base of a 1.5 to 4.5 m (15-25 ft) massive coquinoid accumulation of Pentamerus shells; this contact is not exposed at the Bellevue West section (but was seen at Z34 section). However, packed accumulations of pentamerids become noteworthy in the upper Sweeney along Hwy 62 (e.g., unit 22). If Willman's definition of the Sweeney is not preserved, the base of an expanded Marcus Member could be drawn to encompass the Pentamerus-bearing strata in the upper roadcut section.

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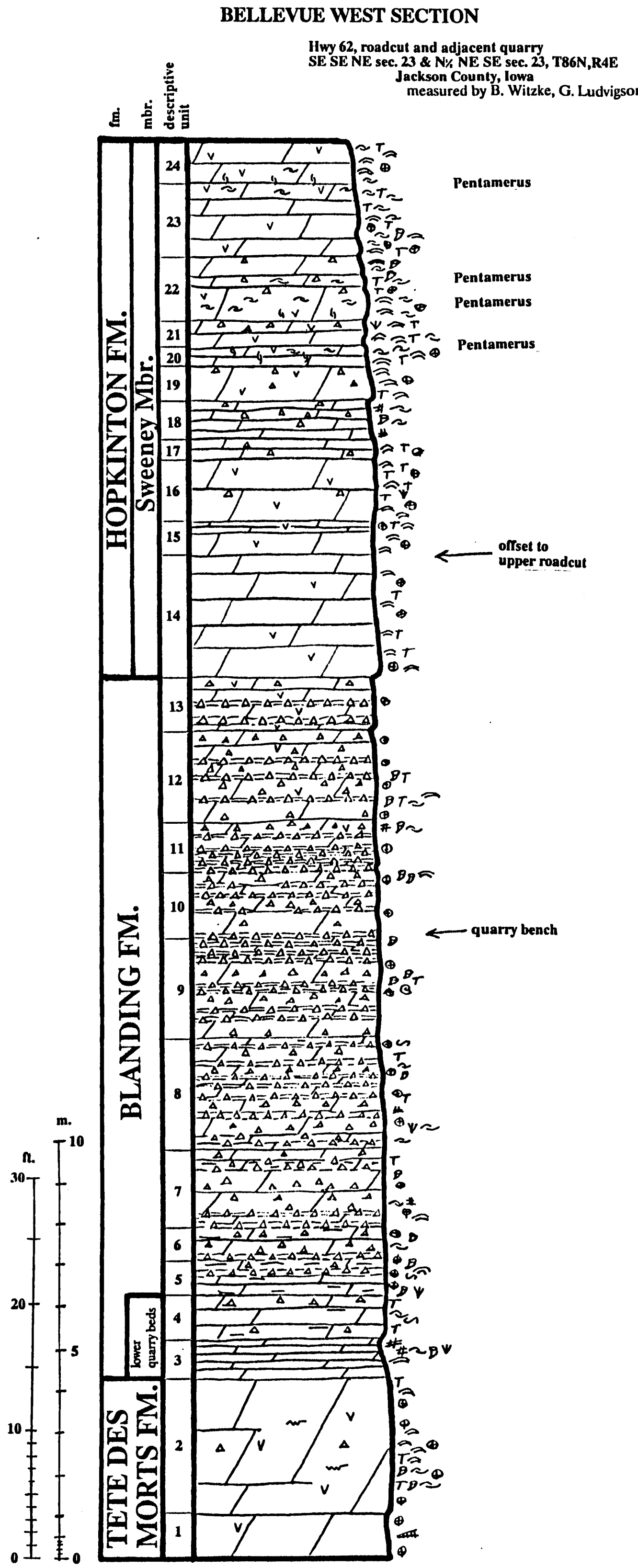

Figure 21b. Graphic section of Silurian stratigraphic sequence exposed along Hwy 62 west of Bellevue; Sunday STOP 1B. (symbols as in Fig. 4).

BELLEVUE WEST SECTION

roadcuts and adjacent quarry along Hwy 62
about 2 miles southwest of Bellevue
SE SE NE sec 23 & N1/2 NE SE sec 23, T86N, R4E, Jackson Co., Iowa
measured by B.J. Witzke and G.A. Ludvigson (8/1988)

SILURIAN SYSTEM

HOPKINTON FORMATION

SWEENEY MEMBER

UNIT 24

Dol., xf-f, vf-m xlln, fossil moldic, vuggy; crinoid debris; common Pentamerus at base (2-7 cm), some in life position; Pentamerus decreases in abundance upward, only scattered valves in upper 40 cm; silicified laminar stromatoporoids in middle to upper part; Favosites in middle part; Halysites (some silicified) in middle and upper parts; 1.0 m thick.

UNIT 23

Dol., xf-f, vf-m xlln (more finely xlln 45-100 cm up), part microporous mottled, fossil moldic; divided into three intervals: lower 45 cm recessive at top, silicified laminar stromatoporoids, Halysites, indet. brachiopod, crinoid debris, domal stromatoporoid overturned; 45-100 cm interval marked by bedding surface at top, crinoid debris, large horn corals in lower half, silicified laminar stromatoporoids, overturned Halysites near base, a few scattered Pentamerus present; top 75 cm is rubbly and vuggy, lower part has a few scattered Pentamerus valves, Pentamerus scattered to common above, crinoid debris, Favosites, common vugs at top; 1.75 m thick.

UNIT 22

Dol., xf-f, vf-m xlln, fossil moldic; divided into three intervals marked by thin recessive units at top of each containing scattered chert nodules; lower 85 cm with scattered vugs, massive single bed, scattered to common Pentamerus valves and shells along some horizons (1.5-4 cm shells) forming packstones along some horizons, some Pentamerus in life position 22-26 cm up, Stricklandia noted, silicified laminar stromatoporoids scattered (to 70 cm wide at 40 cm up), silicified Favosites in upper part (to 30 cm), silicified horn coral at 60 cm up; interval 85-112 cm up, silicified to moldic Halysites scattered, scattered Pentamerus valves (forms discontinuous packstone layer along one horizon), indet. silicified rhynchonellids; top 45 cm, includes some coarse xlln dol., few scattered Pentamerus valves in middle, includes laminar stromatoporoids, cup and horn corals, indet. small brachiopods, Halysites; 1.57 m thick.

UNIT 21

Dol., xf-vf xlln, scattered f-m xlln, small moldic to replaced crinoid debris, drusy quartz vug linings present, scattered to common chert nodules in top 30 cm (prominent nodular band 15-20 cm down), cherts are in part skeletal moldic, recessive at top; lower half has silicified fossils, laminar stromatoproids, Halysites, Syringopora, cup corals, a few scattered Pentamerus and ?Stricklandia valves; upper half with silicified laminar stromatoporoids (to 40 cm diam), Halysites (to 20 cm), trilobite (Stenopareia); 62 cm thick.

UNIT 20

Dol., xf-f xlln, replaced to moldic crinoid debris, scattered chalcedony and drusy megaquartz void linings, rubbly bedded unit; silicified laminar stromatoporoids, Halysites scattered through; silicified domal Heliolites near base; top 30 cm has three bands (laterally discontinuous) with large Pentamerus (to 8 cm long) and scattered Stricklandia lens progressa, some in life position, packed in part, part silicified; 45 cm thick.

UNIT 19

Dol., xf-f, vf-m xlln, common corals and stromatoporoids, replaced to moldic crinoid debris, forms overhanging ledge at base, scattered chert nodules 40 cm up and near top, chalcedony void fillings; common silicified laminar to domal stromatoporoids, moldic to silicified Halysites (to 8 x 25 cm) scattered (overturned noted 15 cm up), Favosites common (to 25 cm), Syringopora present in lower 15 cm (50 cm wide); 85 cm thick.

UNIT 18

Dol., xf-vf xlln, dense,, irregularly bedded with common to abundant nodular cherts (especially top 50 cm), unit is recessive; lithology contrasts with over and underlying units; scattered fossil molds include fenestellid bryozoans, cup corals, brachiopods (Hesperorthis); 90 cm thick.

UNIT 17

Dol., xf-vf xlln, similar to below but less coarsely xlln, thin bedded, scattered chert nodules (especially in middle); silicified laminar stromatoporoids, Halysites, Syringopora; 48 cm thick.

UNIT 16

Dol., xf-vf, f-m xlln, small vugs scattered, in 1 to 2 beds with rubbly recessive horizon containing scattered nodular cherts near midpoint, rubbly at top; moldic to replaced small crinoid debris; common to abundant silicified laminar stromatoporoids, Syringopora colony 65 cm up (15 x 50 cm), overturned domal stromatoporoids noted at 45 and 65 cm down, silicified Halysites noted at top and 45, 60, and 73 cm down, silicified horn coral at 65 cm down, trilobite (Stenopareia) in chert band; 1.5 m thick.

UNIT 15

Dol., xf-f xlln, small crinoid debris molds, massive in lower part, becomes irregularly bedded with scattered vugs in top 30 cm, scattered chert nodules, some chalcedony void fills; scattered silicified laminar stromatoporoids through, top 20 cm with Halysites, Favosites, Syringopora (to 25 cm diam), brachiopod (delthyridid aff. Howellella); 80 cm thick.

UNIT 14

Dol., xf-f, vf-m xlln, moldic to replaced small crinoid debris, microporous mottling, medium bedded, forms overhanging ledge above cherty Blanding; lower 1.2 m accessible in north quarry-roadcut section, remainder of unit accessible in south quarry area; scattered to common silicified laminar stromatoporoids and Favosites; about 2.9 m thick.

BLANDING FORMATION

UNIT 13

Dol., xf-vf, vf-m xlln, ledge-former at base, vuggy, 15 cm above base in a prominent 5-20 cm thick chert bed (thickest in Blanding section), 5 cm thick chert bed 60 cm up, nodular cherts in thinner bedded dol in upper 25 cm; fossil moldic, crinoid debris, silicified laminar stromatoporoids; 1.3 m thick.

UNIT 12

Dol., xf-f, vf-m xlln, more skeletal moldic than below, becomes more coarsely xlln upward; small crinoid debris, scattered small vugs; continuous beds of smooth to chalky chert (most are sparsely fossiliferous) noted 50 cm (5 cm thick, displays packstone fabrics), 80 cm (3 cm thick), 1.05 m (5-10 cm), 1.45 m (5-12 cm), 1.95 m (grades to nodular band) above base; nodular cherts noted in lower 20 cm and 65 cm, 90 cm, 1.3 m (to 15 cm thick), 1.65 m above base; silicified laminar stromatoporoid and tabulate coral at 45 cm up, 50 cm up with indet. brachiopods and cup corals, 80 cm up with moldic Favosites; 2.15 m thick.

UNIT 11

Dol., xf-vf xlln, includes f-m xlln top 50 cm (includes microporous mottlings), moldic to replaced small crinoid debris; prominent beds of smooth chert 5-14 cm thick at base, 18 cm, 35 cm, 53 cm (grades to nodular band), 86 cm, 1.05 cm (to band) above base, and top (to band); fossils include indet. brachiopods and bryozoans (20-30 cm down), small cup corals (top 50 cm); 1.2 m thick.

UNIT 10

Dol., xf-vf xlln, prominent argillaceous bedding surface at top, slightly argillaceous with scattered argillaceous partings, argillaceous content generally increases upward; bedded cherts at base, 95 cm up (8 cm thick), nodular cherts and nodular bands at 40 cm, 60 cm, 70 cm, 80 cm, 1.2 m, 1.3 m, and 1.4 m above base; cherts are smooth and sparsely fossiliferous; top 25 cm with silicified laminar stromatoporoids, horn corals, Favosites; 1.6 m thick.

UNIT 9

Dol., xf-vf xlln, scattered crinoid debris; nodular chert bands at base and 20 cm, 80 cm 90 cm, 1.45-1.55 m above base; bedded cherts at 45 cm, 65 cm (10 cm thick), 1.05 m, 1.15 m, 1.25 m (6 cm), 1.9 m (6 cm), 2.05 (7 cm), 2.2 m above base; fossils include gastropod 1.1 m up, cup corals 1.2 m up, packstone bed at 1.45 m up with cup and horn corals (to 2.5 cm diam), cup coral at top; 2.4 m thick.

UNIT 8

Dol., xf-vf xlln, scattered small crinoid debris molds; top 1.0 m has probable packstone interbeds with coarse xlln replaced crinoid debris; top 14 cm forms prominent argillaceous recessive bed, rubbly and burrowed; argillaceous streaks at 45 and 90 cm up; pyrite nodules 2-3 cm diam at 2.1 m above base; prominent chert beds at 14 cm, 28 cm (8 cm thick), 76-90 cm (8-14 cm thick), 1.25 m, 1.5 m (6 cm), 1.7 m (10 cm), 2.15 m (10 cm), 2.55 m (6 cm) above base; chert nodules and nodular bands at 38 cm, 50 cm, 95 cm, 1.06 m, 2.0 m above base; fossils include rhynchonellid at base, indet. trilobite at 40 cm up, Resserella at 45 cm up, fenestellid bryozoan at 90 cm up, silicified Favosites (3 x 25 cm) at 1.25 m up, strophomenid brachiopod at 2.0 m up, indet brachiopod and silicified Favosites (14 cm diam) at 2.3 m up; 2.75 m thick.

UNIT 7

Dol., xf-vf xlln, dol is slightly more coarsely xlln than below; slightly argillaceous, wispy argillaceous streaks at 15-30 cm above base, scattered pyritic blebs and nodules; chert beds at base and 30 cm, 75 cm, 1.45 m (7 cm thick), 1.6 m (5 cm) above base; nodular cherts and nodular bands (smooth chert with chalky to chalcedonic rims) at 20 cm, 40 cm, 45 cm, 60 cm, 90 cm, 1.0 m, 1.15 m, 1.75 m above base; fossils include scattered crinoid debris molds, fenestellid bryozoans, atrypid brachiopods, silicified laminar stromatoporoids in lower 50 cm; scattered cup corals and silicified Favosites in top 65 cm; 1.9 m thick.

UNIT 6

Dol., xf-vf xlln, slightly argillaceous, wavy argillaceous streaks in top 25 cm; prominent chert bed (5-12 cm thick) at base; scattered chert nodules throughout with nodular chert bands at 20, 25, 45, and 55 cm above base, cherts are smooth, light to dark brown, part with chalky rinds, may preferentially replace burrow networks; slightly fossiliferous with indet.

brachiopod 20 cm up, rhynchonellid and crinoid debris near middle, small cup coral 50 cm up; 75 cm thick.

UNIT 5

Dol., xf-vf xlln, slightly argillaceous to argillaceous, lower 24 cm has thin beds of xf-m xlln dol with scattered small crinoid debris molds separated by wavy argillaceous partings; middle 33 cm is less argillaceous in 2 or 3 beds with scattered small fossils molds and pyrite blebs, contains subvertical burrows (1 x 7 cm); top 35 cm becomes more argillaceous with wavy argillaceous partings every 1 to 2 cm; interval contains 4 or more nodular chert bands, preferentially replacing burrow networks, top 18 cm laterally becomes single chert bed; silicified cup coral at base, scattered silicified fossils at top include small cup coral, Resserella, strophomenid; 82 cm thick.

"lower quarry beds"

UNIT 4

Dol., xf-vf xlln at base, xf-f xlln above, relatively dense, scattered small fossil molds, argillaceous (especially at base), includes argillaceous streaks and stylolites above, argillaceous wavy-bedded aspect; dark colored burrow mottles scattered throughout, dispersed pyritic blebs; in 3 or more beds; irregular bands of nodular cherts at 25-30 cm and 80 cm above base, scattered cherts top 15 cm, incipient silicified zones or diffuse cherts scattered; prominent bedding surface at top; sparsely fossiliferous with scattered indet. brachiopod molds (orthids?), upper half with small Halysites, small cup coral, laminar alveolitid coral (30 cm diam), illaenid trilobite (smooth free cheek with large eye stalk); 1.05 m thick.

UNIT 3

Dol., xf-vf xlln, much denser than below, vf-m xlln at base, becomes more dense upward; in 10 to 15 beds, 3 to 13 cm thick, separated by thin argillaceous partings that become more abundant upwards; upper 17 cm beddings surfaces become more wavy and argillaceous; contrasts markedly from underlying lithologies; 55 cm above base is planar hardground surface (locally with 1 cm relief), pyritic to iron-oxide stained, scattered hardground clasts to 1.5 cm above surface; scattered iron-oxide stained horizontal burrows; scattered fossils include silicified laminar stromatoporoid 55 cm up, 65 cm up is a layer with common brachiopod molds (brachiopods include Fardenia, Resserella, indet. strophomenids, with fenestellid and stick bryozoans, cup and horn corals, Calymene trilobite fragments); scattered small crinoid molds throughout; top 10 cm with scattered fenestellid bryozoans; 1.0 m thick.

TETE DES MORTS FORMATION

UNIT 2

Dol., xf-vf, vf-m xlln, scattered vugs (1-10 cm diam, part with chalcedony linings); moldic to replaced small crinoid debris (in part forming thin dolomitized packstone beds), scattered silicified crinoid debris; argillaceous stylolitic streak at 1.1 m up, stylolite at 2.1 m up, thin argillaeous parting at top; one massive unit, laterally with one or two bedding splits; a few scattered chert nodules 1.6 m above base; scattered corals and stromatoporoids; silicified laminar stromatoporoids at 1.1 m, 1.3 m (to 35 cm diam), 1.6 m, 1.9 m, 2.7 m above base; overturned domal stromatoporoid (7 cm diam) overgrown by laminar stromatoporoid at 1.3 m above base; Favosites at 75 cm, 1.3 m, 1.4 m above base and at top (to 55 cm diam); rhynchonellid brachiopod and cup coral at 75 cm up, silicified horn corals at 1.1 m up, lissatrypid brachiopod at 1.3 m up; 3.2 m thick.

UNIT 1

Dol., vf-m xlln, single bed; some coarse xlln (replaced crinoid debris) along horizons (probable packstone lenses); scattered vugs, partly burrowed, zones of moldic porosity (small crinoid debris); nautiloid fragment in middle (3 cm diam); 1.1 m thick.

STOP 2. Calcitized Dolostones of the Silurian Scotch Grove Formation from an Ancient Weathering Profile Beneath the Early Pennsylvanian Caseyville Formation

Greg Ludvigson

Combined geologic mapping and petrologic studies of rocks along the PRFZ in the area north of Preston, Iowa (Figs. 22 and 23), provide key information on the origin of ubiquitous coprecipitated ferric oxide and calcite cements healing brittle deformation in the fault system. An apparent curvature of the PRFZ in this field area (Fig. 22) coincides with the development of a 1 km-wide belt of rock deformation exposed in structurally depressed rocks of the Scotch Grove Formation (Fig. 24a). Pennsylvanian quartz pebble conglomeratic sandstones in this field area rest unconformably on deformed Scotch Grove strata (Fig. 24b). The calcitic ("dedolomitized") alteration of Scotch Grove dolostones beneath the unconformity are the subject of the last field trip stop.

The calcitic alteration is best exposed at the Carstensen Farm and neighboring properties (Figs. 22, 23, 24b), and extends about 15 m (approx. 50 ft.) below the position of the Pennsylvanian-Silurian unconformity. Alteration can be studied along roadcut and natural hillslope exposures at the BCR locality (Fig. 25a). The distribution of altered fabrics is heterogeneous, and relatively unaltered dolostones are closely juxtaposed with altered equivalents in some areas. Petrographic studies of the composition of calcitized dolostones from the area indicate, however, that a systematic vertical zonation is developed (Fig. 25a). The overlying Pennsylvanian sandstones are exposed in large float blocks on the upland. These quartz-pebble conglomeratic sandstones closely resemble those from the Lower Pennsylvanian Caseyville Formation discussed at STOP 5 (DBF-1A, DBF-2, Fig. 10), and petrologic data from this locality also establish strong compositional similarities to the Caseyville (see BCR population in Fig. 12).

The alteration of Scotch Grove dolostones is manifest by the upward replacement of dolomite by calcite spars and opaque iron and manganese oxides (Fig. 25a). Enhanced porosity in the zone of alteration (Fig. 25a) is developed by the preferential weathering of calcite spars in the modern weathering environment. Oxide phases occur as detrital inclusions and cements that are intergrown with calcite. The mineralogical zonation of this ancient weathering profile is also paralleled by a vertical zonation in the fabrics of calcite spars (Fig. 25b), with poikilotopic spars at the top of the profile being replaced by blocky equant spars at depth. Partially-calcitized dolomite rhombs and spherulitic/columnar spars are minor fabric constituents that are sporadically distributed throughout (Fig. 25b).

Calcite veinlets are dispersed throughout the altered rocks, and contain abundant geopetal accumulations of detrital ferric oxides, corroded dolomite rhombs, and rounded quartz sand grains, some with quartz overgrowths. Polished slabs of representative rock fabrics will be available for study and discussion. The geopetal debris in the veinlets is interpreted as material that infiltrated from the erosion surface, including some unlithified siliciclastic detritus that <u>might</u> have been derived from the same sediment mass as the quartz sandstones that overlie the unconformity. Similar accumulations of siliciclastic detritus are widely dispersed in the oxide/calcite cements along the PRFZ (Ludvigson, 1988; see fig. 9a,b in the paper by Ludvigson in this guidebook).

Carbon and oxygen isotopic ratios of carbonate constituents at this locality (Fig. 26) fall in two distinct clusters. The ^{13}C-depleted field includes poikilotopic calcite spars from the upper part of the profile, which have the most negative $\delta^{13}C$ values (Figs. 25b, 26). These data are consistent with results that would be expected from phreatic calcites whose isotopic chemistry was strongly influenced by the chemistry of freshly infiltrating meteoric waters in a diagenetic system with high water/rock ratio. Blocky and spherulitic spars from lower in the profile have less depleted $\delta^{13}C$ values, reflecting increased contributions from the host dolostone (Figs. 25b, 26). Note that the isotopic ratios of blocky spars from this cluster are very similar to those from blocky calcite spars healing the boundary faults at the Silver Creek Graben (Fig. 19).

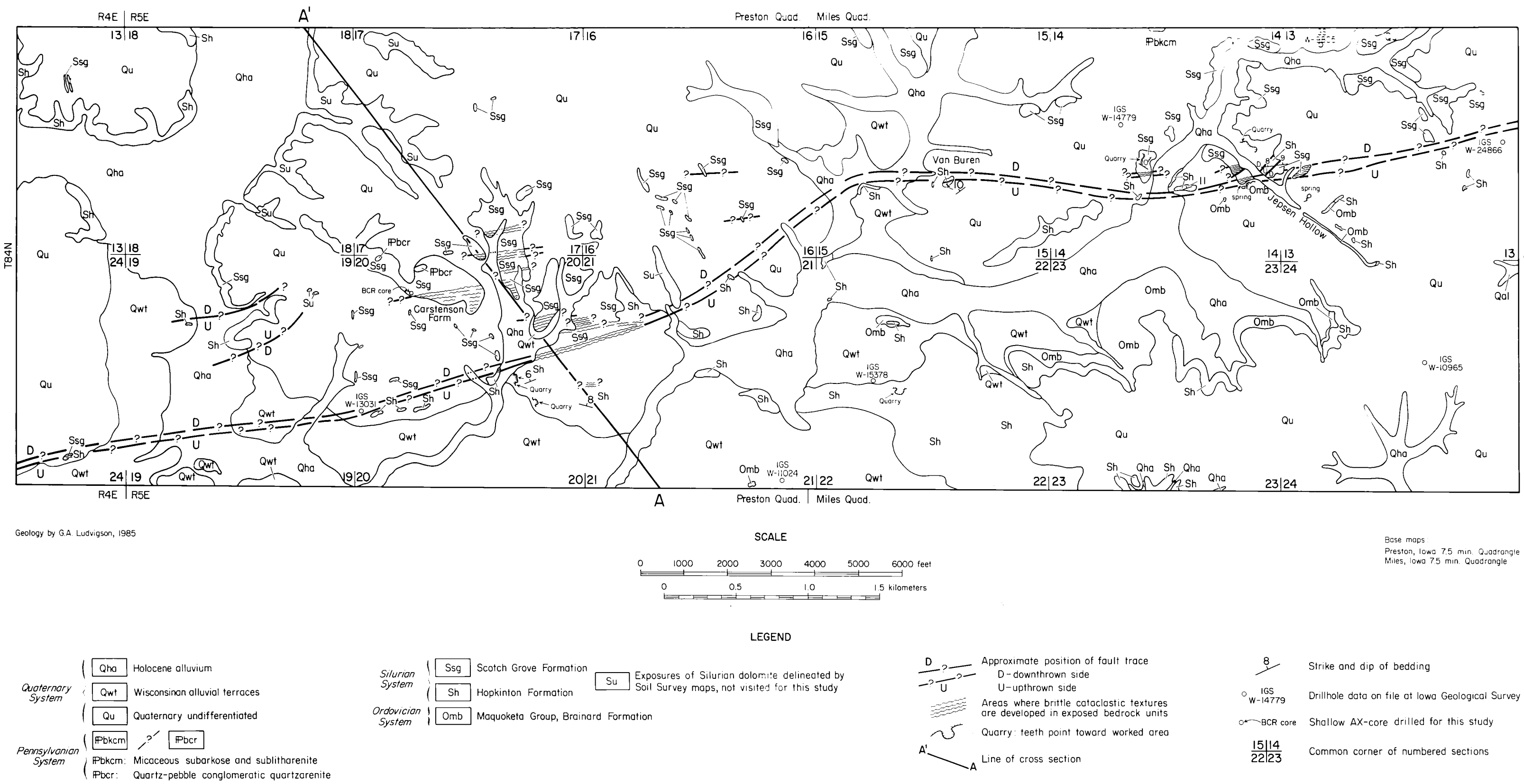

Figure 22. Geologic map of the Plum River Fault Zone north of Preston, Iowa. From Ludvigson (1988).

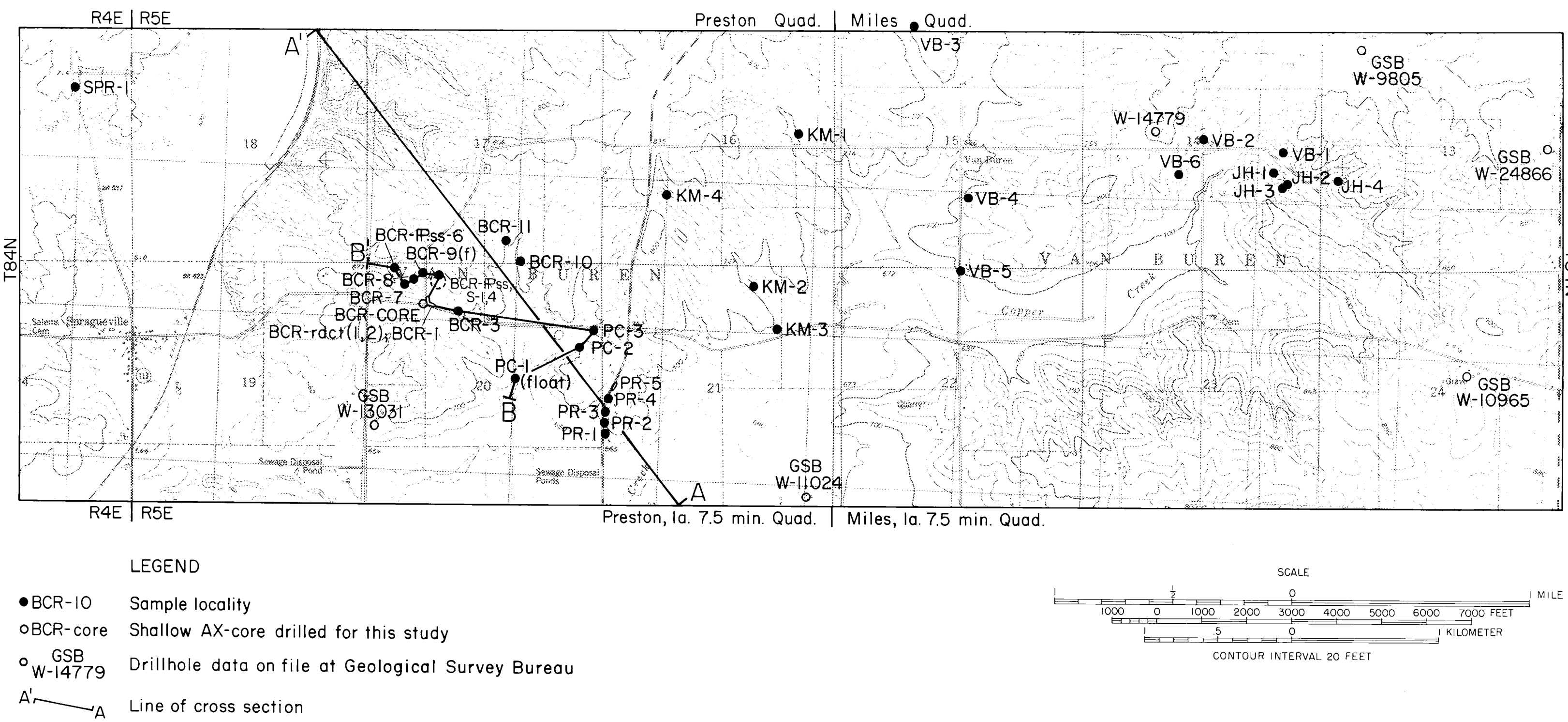

Figure 23. Sampling localities along the Plum River Fault Zone north of Preston, Iowa. From Ludvigson (1988).

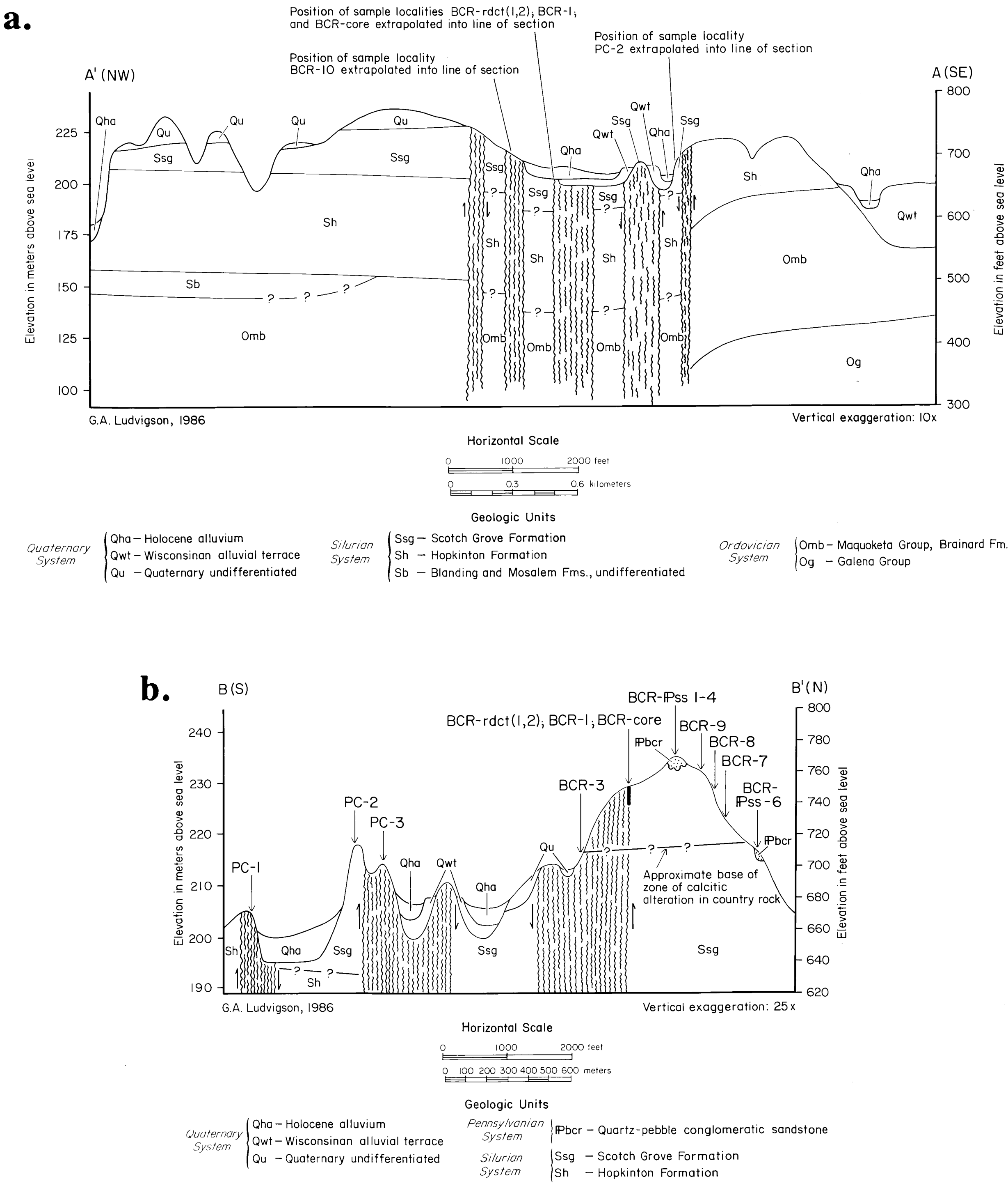

Figure 24. Geologic cross-sections across the Plum River Fault zone north of Preston, Iowa. a. Northwest-southeast cross-section, line of section shown in Fig. 22. b. Northwest-southeast cross-section through sampling localities, line of section shown in Fig. 23. From Ludvigson (1988).

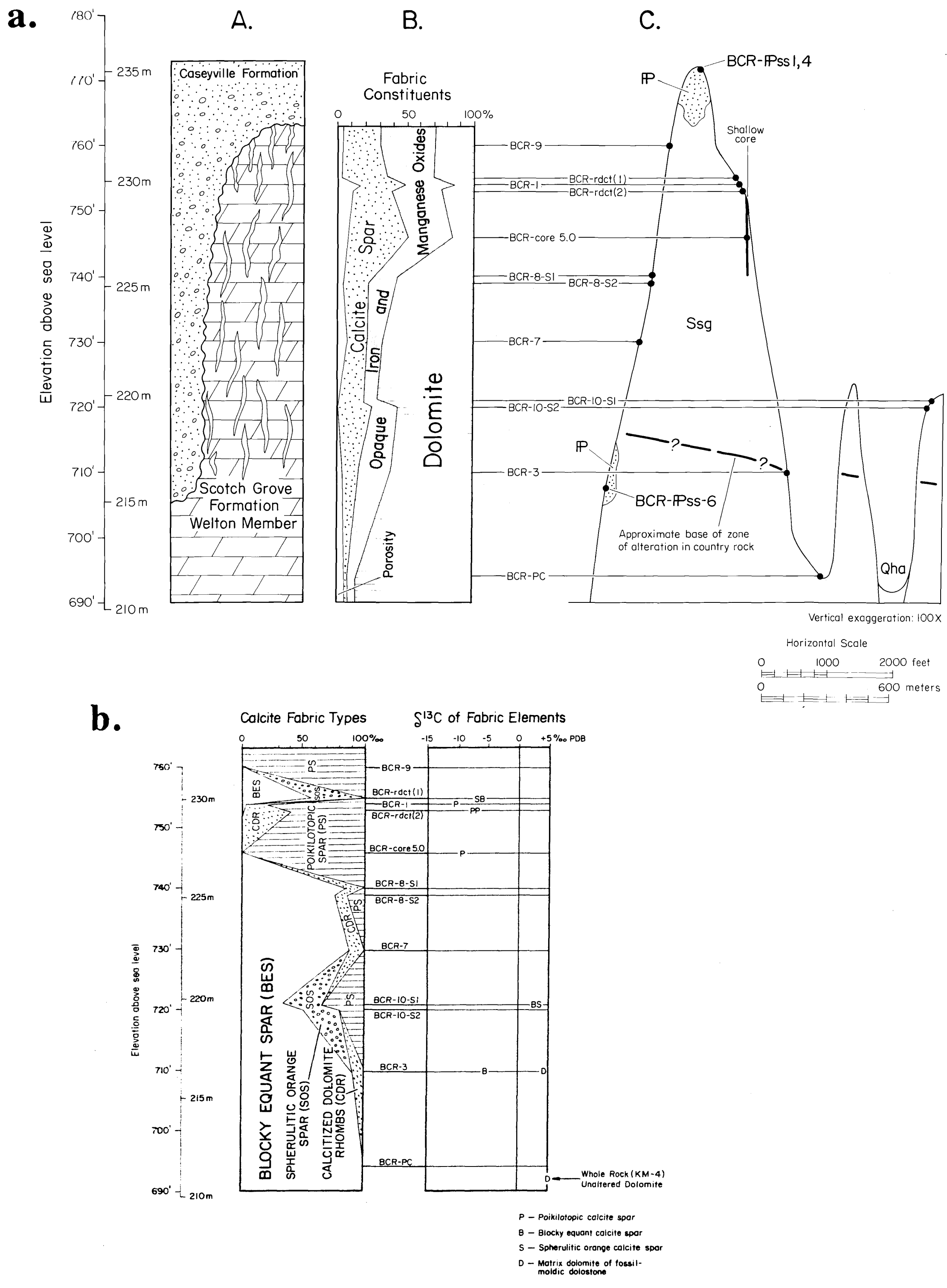

Figure 25. Compositional variations in the ancient weathering profile at Carstensen Farm. a. Reconstruction of the stratigraphic section, with compositional variations documented by modal analyses of rock samples collected across a landscape transect of the exhumed weathering profile. b. Vertical variations in the abundance of calcite fabric types, and carbon isotopic ratios of carbonate phases in the ancient weathering profile at Carstensen Farm.

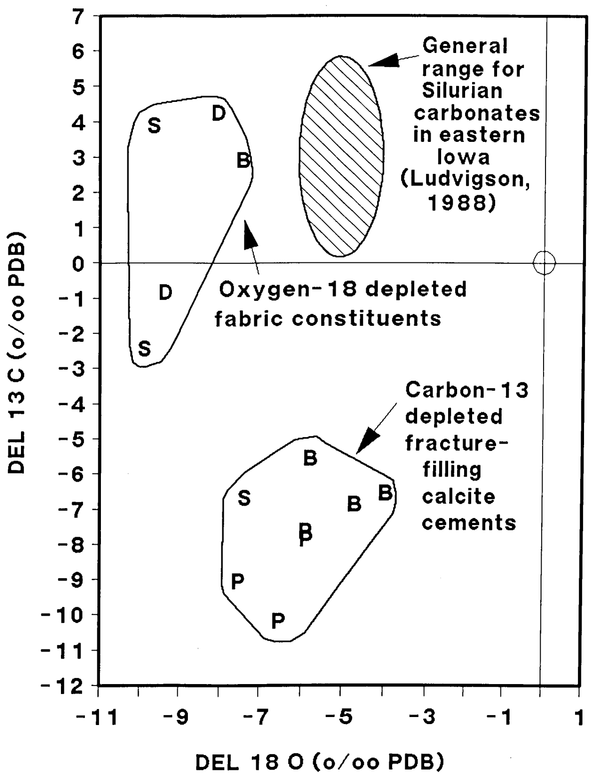

Figure 26. Plot of carbon and oxygen isotopic ratios of carbonate constituents from calcitized dolostones in the Preston study area. Fabric constituents are: P, poikilotopic calcite spar; B, blocky equant calcite spar; S, spherulitic to columnar calcite spar; D, dolomite. Modified from Ludvigson (1988).

Some calcite spars and dolomites sampled from this locality have $\delta^{18}O$ values that are significantly depleted from the range determined from Silurian limestones and dolostones at other localities in eastern Iowa (Fig. 26). Additional sampling is needed to better understand these anomalous results. It is quite possible that ferric hydroxide contaminants in the powdered microsamples (2-5 mg) contributed isotopically light oxygen during the generation of the CO_2 gas for isotopic analysis, and led to erroneous results. Data on the isotopic systematics of goethite indicate that ferric hydroxides that are copreciptated with calcite would have both structural and nonstoichiometric oxygens with $\delta^{18}O$ values that are significantly lower than that of calcite (Yapp, 1987).

Covariant trends of Mg/Ca and Fe/Ca ratios in the calcite spars (Fig. 27) are consistent with the interpretive framework discussed above. For a more complete discussion of the criteria for interpretation, see the paper by Ludvigson elsewhere in this guidebook. Poikilotopic spars are the purest of the calcites, and precipitated from oxidizing groundwaters (lowest range of Fe contents) that were least influenced by the incongruent dissolution of dolomite (negligible Mg contents). As with the isotopic data (Fig. 26), high water/rock ratios are indicated.

Blocky equant spars precipitated from both oxidizing and reducing groundwaters, and the most Fe-rich of these spars generally were associated with diagenetic settings with the lowest water/rock ratios (elevated Mg contents indicating progressive dissolution of dolomite). Cathodoluminescent micrographs of growth zonation showing the trend from oxidizing to reducing environments of precipitation in these spars are shown by Ludvigson (this guidebook, see fig. 9g,h in that paper). Note that the trace element trend for the blocky calcite spars at this

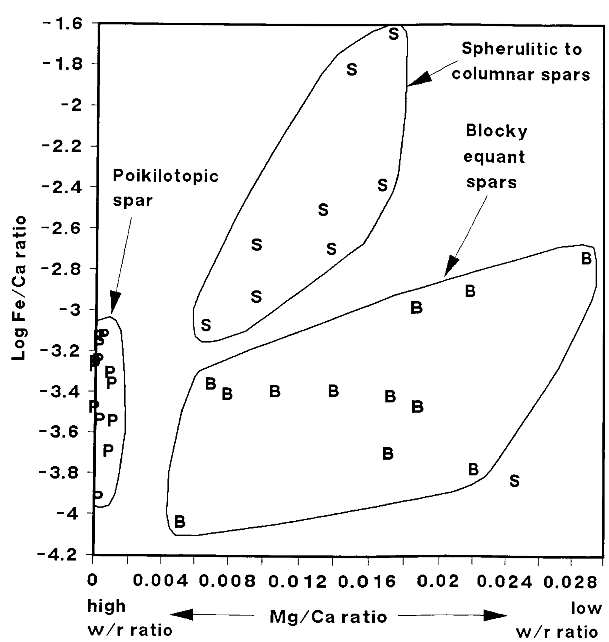

Figure 27. Cross plot of Fe/Ca and Mg/Ca ratios from calcite spars in the ancient weathering profile at Carstensen Farm. Modified from Ludvigson (1988).

locality (Fig. 27) compares very closely to that from the blocky equant spars healing the boundary faults at the Silver Creek Graben (Fig. 15). These calcites and their coexisting ferric oxides are ubiquitous to the PRFZ, always heal the latest deformation, and are interpreted as cements that precipitated from groundwaters that infiltrated beneath the sub-Pennsylvanian erosion surface (Ludvigson, 1988).

The spherulitic to columnar spars are also widely distributed along the PRFZ, and appear with the blocky spars. Their higher Fe contents are indicative of precipitation in anoxic settings. Environmental interpretations of these distinctive spars are provisionally withheld pending further study.

This stop concludes the field conference, and we hope that it has been a rewarding one for you. Have a safe trip back home!

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