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1985 Number 13

THE PLUM RIVER FAULT ZONE AND THE STRUCTURAL AND STRATIGRAPHIC FRAMEWORK OF EASTERN IOWA

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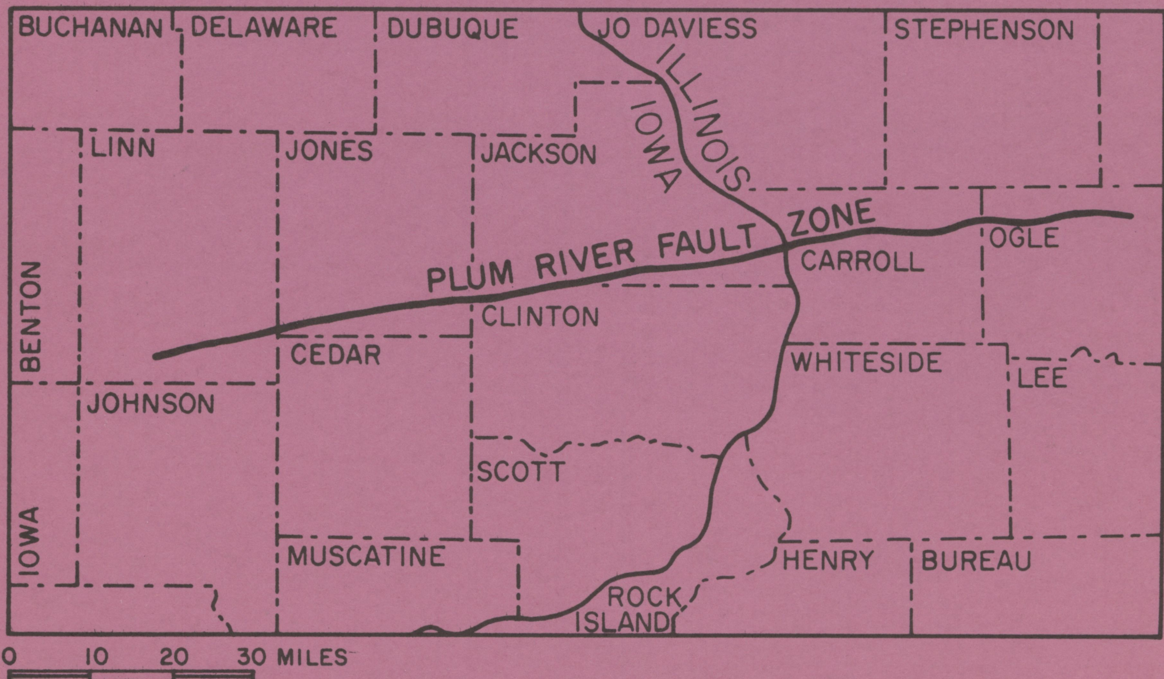

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FOREWORD

The Midwest in general, and Iowa specifically, is largely acknowledged to be very stable in a geologic sense. Evidence of large deformations or major movements along fractures is rare in Iowa. The Plum River Fault Zone, as detailed in this report, is a notable exception. Aggregation of information from work on several projects in eastern Iowa during the last thirteen years, by Survey staff and by university graduate students, has provided adequate documentation to release a report on this significant structural feature.

Knowledge of the extent of the Plum River Fault Zone, and related movements, helps to explain variation in the distribution and thickness of associated rock units. More importantly, it helps us to understand and evaluate the distribution and availability of our precious groundwater resources in east-central Iowa, which are so important to existing and potential commercial, industrial, agricultural, and residential water supplies. This work significantly increases our geologic understanding of this part of the Midwest and provides information which is both useful and necessary to manage and develop Iowa's natural resources.

Donald L. Koch
State Geologist and Director
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THE PLUM RIVER FAULT ZONE
AND
THE STRUCTURAL AND STRATIGRAPHIC FRAMEWORK
OF
EASTERN IOWA

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ABSTRACT

The Plum River Fault Zone is a 112 mile (180 km) long, east-west trending zone of high-angle faulting in east-central Iowa and northwest Illinois. The north side of the fault zone is downthrown, with documented net vertical displacements of Silurian strata up to 270 feet (70 m). Detailed geologic field mapping has shown that the internal structure of the Plum River Fault Zone is characterized by a complex pattern of intersecting high angle faults which have graben and horst fault-block relationships. Major faults within the Plum River Fault Zone are recognized by the occurrence of zones of brittle cataclastic deformation, and the maximum known width of the fault zone in Iowa is 3900 feet (1.2 km). Within the fault zone, vertical displacements of up to 500 feet (150 m) have been interpreted from Paleozoic rocks exposed in adjacent fault blocks, and vertical displacements on the Precambrian basement surface of up to 1100 feet (335 m) have been estimated from combined gravity and magnetic traverses across the fault zone.

Recently completed investigations and revisions of the Paleozoic stratigraphy of eastern Iowa, along with detailed investigations of the Plum River Fault Zone, have led to a comprehensive reevaluation of the structural geology and Phanerozoic tectonic history of eastern Iowa. Descriptions and criteria for the recognition of exposed Phanerozoic stratigraphic units are presented in this report, and a new Silurian unit, the Scotch Grove Formation, is formally proposed. A newly recognized regional structure, the Fayette Structural Zone of northeast Iowa, is also defined.

The Phanerozoic tectonic history of eastern Iowa is reviewed within the context of Sloss' (1963) unconformity-bounded cratonic sedimentary sequences. During the deposition of the Sauk Sequence (Upper Cambrian-Lower Ordovician) in eastern Iowa, sediments accumulated in a north-south oriented, southward plunging trough termed the Hollandale Embayment. Significant structural reorganization in the central midcontinent region occurred during deposition of the Tippecanoe (Middle Ordovician-Silurian) Sequence. The earlier north-south structural grain in eastern Iowa was supplanted by an east-west structural grain, and subsidence was initiated in the East-Central Iowa Basin, whose axis paralleled the Plum River Fault Zone. During the Early Silurian, maximum subsidence in this structural and depositional basin was 18.6 miles (30 km) to the north of the Plum River Fault Zone, whereas during the Middle Silurian maximum subsidence occurred along the north edge of the fault zone. Earliest Kaskaskia (Middle Devonian-Mississippian) deposition reoccupied the East-Central Iowa Basin, with penecontemporaneous faulting along the Plum River Fault Zone. A significant structural reorganization in the midcontinent region began in the Late Devonian, and prior to initial deposition of sediments of the Absaroka Sequence (Pennsylvanian), the East-Central Iowa Basin was uplifted and deeply eroded. Pennsylvanian deposition in east-central Iowa was preceded by major faulting along the Plum River Fault Zone, uplift of the Savanna-Sabula Anticlinal System, and the development of the present regional structural geometry. The physical relationships of Pennsylvanian deposits to the Plum River Fault Zone are not known with sufficient precision to preclude up to 33 feet (10 m) of post-Pennsylvanian displacement. Historic seismic data are inadequate to evaluate the potential for seismic hazard associated with the Plum River Fault Zone.

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INTRODUCTION

Detailed geologic investigations in northwestern Illinois led to the recognition of the Plum River Fault Zone, originally defined as a narrow belt of high angle faults trending roughly east-west for approximately 60 miles (97 km) through northwestern Illinois and east-central Iowa (fig. 1a; Kolata and Buschbach, 1976). The Plum River Fault Zone was named for exposures along the Plum River near Savanna, Illinois, which first revealed the structure. It is the principal element in an east-west trending belt of structural deformation formerly termed the Savanna-Sabula Anticline (Cady, 1920).

As originally defined, the western terminus of the fault zone was in the area to the south of the town of Maquoketa, Jackson County, Iowa. Detailed field and subsurface studies in Iowa, however, indicate that the fault zone continues westward approximately 50 miles (80 km) from the interpreted southern Jackson County terminus to an area south of Cedar Rapids in Linn County, Iowa (fig. 1b). Throughout its extent, the north side of the fault is downthrown, with documented net vertical displacements ranging from 170 to 270 feet (52-70 m).

This report reviews the regional geologic setting and history of geologic investigations pertaining to the Plum River Fault Zone. The structural geology of the fault zone and the stratigraphy of bedrock units exposed along it are described in detail. Finally, this report interprets the Phanerozoic tectonic history of eastern Iowa and evaluates the potential for neotectonism along the Plum River Fault Zone.

PREVIOUS STUDIES

Much of the history of geologic investigations related to the Plum River Fault Zone is summarized by Kolata and Buschbach (1976, p. 3). In eastern Iowa some of the most valuable contributions leading to the mapping of the westward extension of this structure have come from detailed studies of the stratigraphy of the Devonian and Silurian carbonate sequences. Until recent years, Silurian rocks in Iowa have been very poorly understood, and the precise stratigraphic position of most areas of exposed bedrock and subsurface materials was not known. Recent studies by Philcox (1970a, b, 1972), Johnson (1975, 1977a) and Witzke (1976, 1980b, 1981a, b) have led to the recognition of a regionally consistent lithostratigraphy.

Bunker and Ludvigson (1977) and Ludvigson et al. (1978, p. 25) recognized the relationship between the geometry of units formerly included in the Middle Devonian Wapsipinicon Formation and the tectonic history of the Plum River Fault Zone. The stratigraphy of Wapsipinicon Formation and older Devonian units in eastern Iowa is complicated by syn- and post-depositional deformation. The general stratigraphy and complications related to these units were described by Norton (1920). Church (1967) and Sammis (1978) described the geometry, petrology, and depositional environments of these units.

In many places the recognition of the fault zone has been greatly aided by careful geologic observations by previous workers. Norton (1895a) noted Devonian and Pennsylvanian outliers in the Silurian terrane of eastern Iowa. The bedrock geology of Jackson County was mapped by Savage (1906), who noted the location of many critical outcrops. A Devonian outlier which delineates the location of the fault zone in southern Jackson County was discovered by a

b

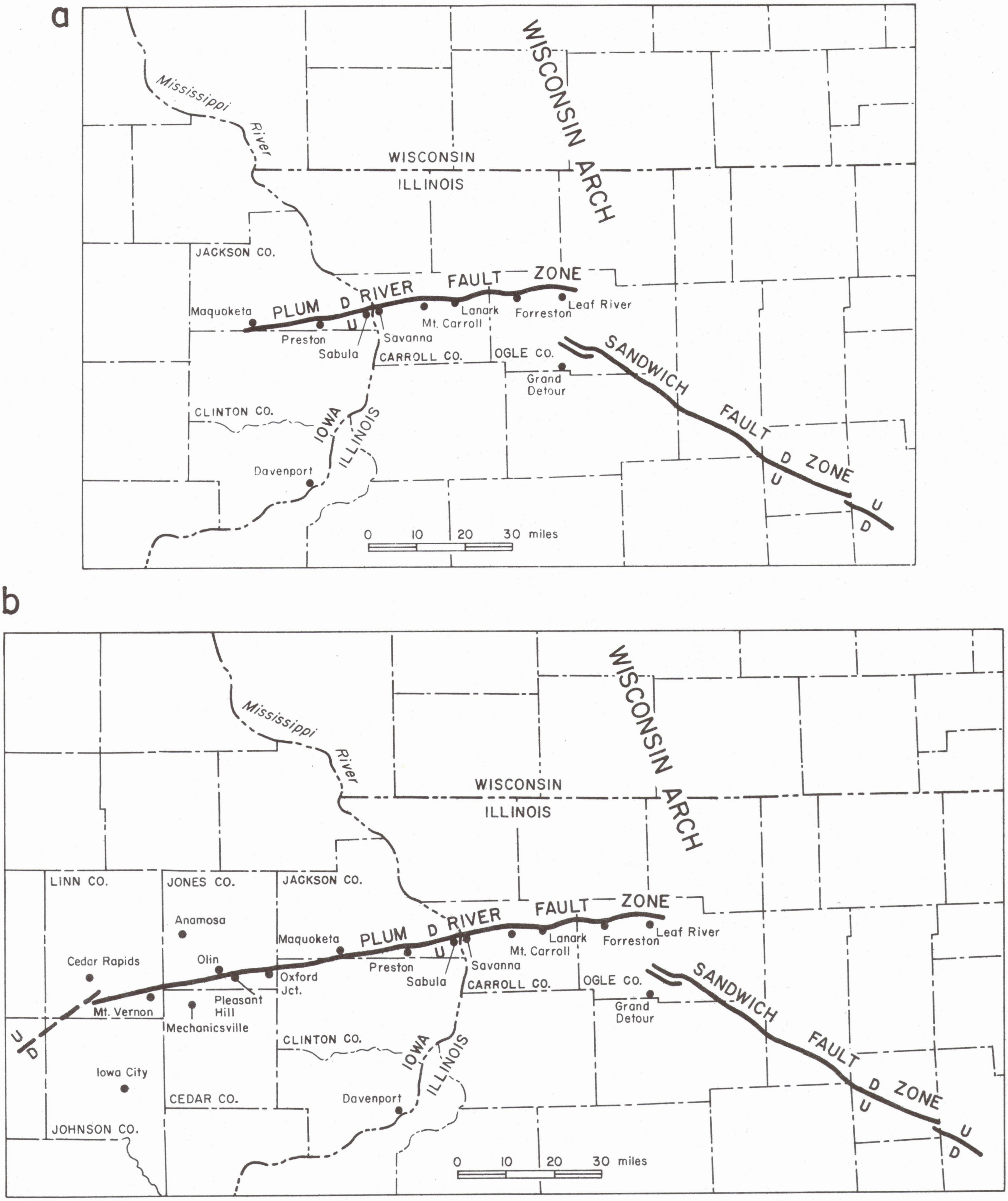


Figure 1. Location and extent of the Plum River Fault Zone: (a) as determined by Kolata and Buschbach (1976); (b) as defined by this paper.

Dorheim (1953). Dow and Mettler (1962) mapped an area of structural deformation along the fault zone in southern Linn County.

Kolata and Buschbach (1976) named and defined the Plum River Fault Zone, and Aten and Herzog (1977) described the geology of the fault zone in Jackson County, Iowa. The westward extension of the Plum River Fault Zone, recognition of extensive cataclastic deformation, and the significance of mid-to late-Paleozoic outliers along the fault zone were first discussed by Bunker and Ludvigson (1977). These aspects were further elaborated by Ludvigson et al. (1978).

Svoboda et al. (1980) described the detailed structure of selected portions of the fault zone, based on field mapping integrated with gravity and magnetic investigations. Svoboda (1980) interpreted gravity and magnetic data along the fault zone to deduce the detailed structure. Chao (1980), Saribudak (1980), and Baik (1981) field-mapped portions of the Plum River Fault Zone, and made supplementary gravity and magnetic investigations to interpret the detailed structure. Ludvigson (1980) described and interpreted the petrology of altered Silurian rocks from the fault zone in Jackson County, Iowa. Witzke (1981a, b) described the relationship between the Plum River Fault Zone and the Silurian stratigraphy of eastern Iowa. Cumerlato (1983) collected and interpreted the data from a 2.3-mile (3.8 km) seismic reflection profile across the fault zone in southern Jones County.

REGIONAL GEOLOGIC SETTING

The Plum River Fault Zone is located in the central stable interior region of the North American continent (fig. 2). During the Phanerozoic eon, tectonism in this region has been solely epeirogenic in style. Late Precambrian, Paleozoic, and Mesozoic sedimentary rocks in the midcontinent region rest unconformably on a cratonic basement composed of igneous and metamorphic Precambrian rocks. Phanerozoic shallow marine to non-marine sediments were deposited intermittently on the craton in a series of major transgressive-regressive marine cycles (fig. 3; Sloss, 1963). The areal distribution and thickness of each cycle was controlled, in part, by the location and trend of broad intracratonic structures that apparently were reactivated in tectonic pulses synchronous with the craton-wide marine transgressions (Ham and Wilson, 1967).

Positive intracratonic structures have been uplifted episodically, exposing pre-existing rocks to erosional truncation. Phanerozoic structural elements that have experienced maximum relative uplift are delineated by the present-day outcrop areas of Precambrian rock (fig. 2). There, the entire sedimentary sequence is missing due to erosional bevelling and/or non-deposition during several periods of Phanerozoic tectonism.

Intracratonic basins, which are best represented by the present day outcrop areas of Pennsylvanian rock (fig. 2), have been subject to episodic subsidence and, due to longer periods of marine inundation and structural preservation, have retained the most representative Phanerozoic rock record in the region. The succession of sedimentary rocks in the southern portion of the Illinois Basin attains thicknesses in excess of 14,000 feet (Willman et al., 1975, p. 19) in its deepest portion.

The location and trend of the Plum River Fault Zone appears to fit into the regional pattern of major faults and folds in the central midcontinent (fig. 2). Most of these structures occur in a configuration arrayed concen-

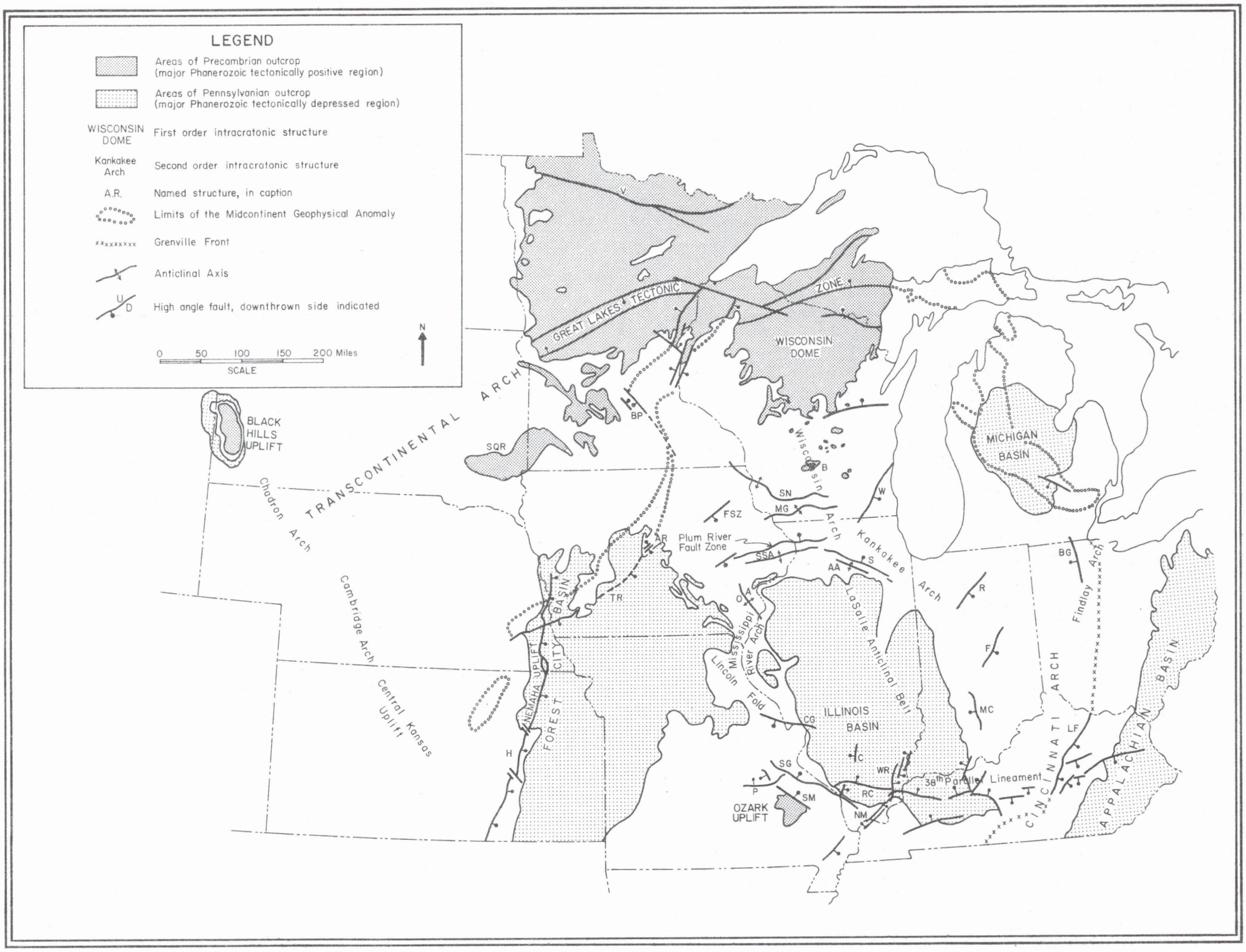


Figure 2. Generalized map of major tectonic features of the north-central midcontinent region. Abbreviations utilized for specific structures are: AA, Ashton Arch; AR, Ames-Roland Horst; B, Baraboo Syncline; BP, Belle Plaine Fault; BG, Bowling Green Fault; C, Centralia Fault; CG, Cap Au Gres Faulted Flexure; F, Fortville Fault; FSZ, Fayette Structural Zone; H, Humboldt Fault; LF, Lexington Fault; MC, Mt. Carmel Fault; MG, Meekers Grove Anticline; NM, New Madrid Fault Zone; OA, Oquawka Anticline; P, Palmer Fault System; R, Royal Center Fault; RC, Rough Creek Fault System; S, Sandwich Fault; SG, St. Genevieve Fault; SM, Simms Mtn. Fault; SN Snymagil Anticlinal Belt; SSA, Savanna-Sabula Anticline; TR, Thurman-Redfield Structural Zone; V, Vermilion Fault; W, Waukesha Fault; WR, Wabash River Fault Zone.

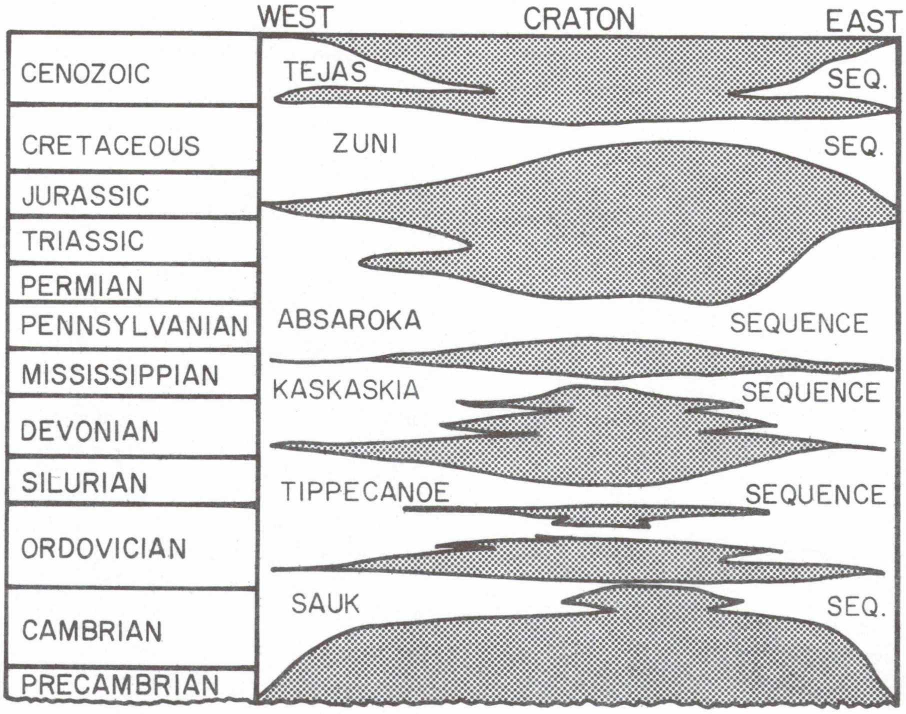


Figure 3. Time-stratigraphic relationships of the cratonic sequences on the North American continent. Shaded areas represent the major unconformities recorded in the rock record across the continent. The more complete depositional sequences are preserved at the continental margins (modified from Sloss, 1963, p. 110).

tric and marginal to the intracratonic arches, domes, and basins. Where adequate data exists, many of these Phanerozoic structural elements can be shown to have Precambrian antecedents that may have played important roles in the later development of the larger regional structures (Rudman et al., 1965). In Iowa, the relationship is best illustrated by the faulted southern margin (Thurman-Redfield Structural Zone) of the Keweenawan (late Proterozoic) Midcontinent Rift System indicated by the outline of the Midcontinent Geophysical Anomaly in figure 2. The Thurman-Redfield Structural Zone has a history of repeated Phanerozoic activity (Coons et al., 1967). The Thurman-Redfield Structural Zone and the Humboldt Fault Zone form the northern and western boundaries, respectively, to the deepest portions of the Middle Pennsylvanian Forest City Basin (fig. 2).

Although the significance of faulting along the Plum River Fault Zone has only recently been recognized, the presence of structural deformation in northwestern Illinois and east-central Iowa has been noted by many previous workers. A regional structure contour map drawn on top of the Galena Group (Middle and Upper Ordovician) (fig. 4) delineates several prominent structural features in eastern Iowa:

1. the Plum River Fault Zone of east-central Iowa and northwestern Illinois;
2. the broad east-west trending Savanna-Sabula Anticlinal Sys-

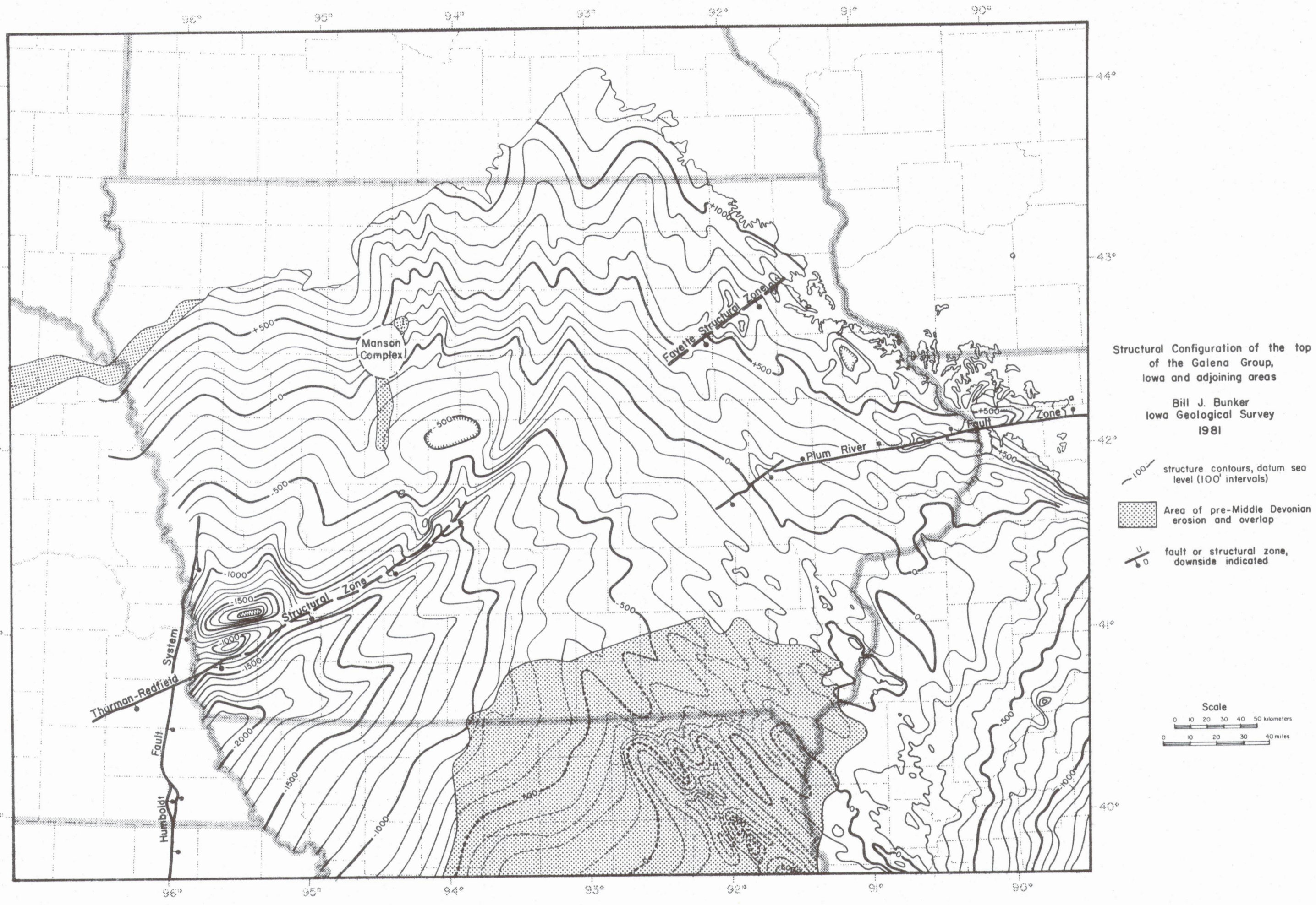


Figure 4. Structure contour map on top of the Galena Group (Middle Ordovician) of Iowa and parts of adjoining states (adapted from Bunker, 1982). Shaded areas on the map show regions of pre-Middle Devonian uplift and erosion of Galena Group rocks preceding the Middle Devonian transgressive overlap.

tem, which parallels the Plum River Fault Zone and is coincident with the southern uplifted side of the fault;

3. the northwest-southeast trending Lincoln Fold System of northeastern Missouri and southeastern Iowa;
4. the Mississippi River Arch, a structural saddle between the Savanna-Sabula Anticlinal System to the north and the Lincoln Fold System to the south, separating the Forest City Basin of southwestern Iowa from the Illinois Basin to the southeast;
5. the Oquawka Anticlinal System in southeast Iowa, paralleling the Lincoln Fold System and cross-folded across the crest of the northeast trending Mississippi River Arch; and
6. the Fayette Structural Zone of northeastern Iowa, strongly

delineated on the aeromagnetic map of Iowa (Zietz et al., 1976). Basement faulting has been interpreted by Heitzman (1972) and Gilmore (1976, p. 14) along the trend of this feature. Structural deformation has been noted in the near-surface Paleozoic rocks of the area by previous workers (Calvin, 1898, p. 220; Savage, 1905, p. 499; and Howell, 1921, Plate IV), but detailed work on the structural zone has not yet been undertaken.

Mapping of bedrock geologic units in the area of the Plum River Fault Zone has been hampered by generally poor bedrock exposures. Pleistocene glacial, glacio-fluvial, and aeolian deposits blanket the bedrock surface over most of the region. Local exceptions occur along isolated bedrock highs, near the valleys of major streams, and in the highly dissected area near the Mississippi River in Jackson County. However, even with the limited bedrock exposures, a general conception of the bedrock geology can be obtained. In east-central Iowa the fault has juxtaposed predominantly Silurian strata against Ordovician strata in its eastern extent, Silurian strata against Silurian strata in its central extent, and Devonian strata against Silurian strata in its western extent (fig. 5). Where rocks of different systems have been faulted into juxtaposition, structural anomalies have already been documented (Savage, 1906; Dow and Mettler, 1962; Kolata and Buschbach, 1976; Ludvigson et al., 1978). Inability to reliably distinguish Silurian units at the surface and in the subsurface delayed recognition of the regional continuity of the structure until recent years.

STRATIGRAPHY

An understanding of the Paleozoic stratigraphy of eastern Iowa is essential for interpreting the structural history and present-day structure of the area. The stratigraphic synopsis presented in this section outlines criteria for recognizing various Middle Ordovician through Quaternary stratigraphic units in outcrop and in the subsurface of eastern Iowa. For the most part, stratigraphic terminology follows previously established usage at the Iowa Geological Survey, although significant revisions of and additions to Paleozoic stratigraphic nomenclature in eastern Iowa is included (see Table 1). This includes the formal definition of a new Silurian formation in eastern Iowa. Cambrian nomenclature is included in Table 1 for completeness but is not discussed in this report. A map of the general study area (fig. 6) is included to assist the reader in locating the various Iowa counties and other geographic locations mentioned in this section. The type localities of certain Paleozoic stratigraphic units are also shown on the map.

The Phanerozoic sedimentary record in eastern Iowa, and in the North American continental interior in general, can be subdivided into a series of six major sedimentary rock sequences separated by major inter-regional unconformities (Sloss, 1963; fig. 3). The sequence terminology of Sloss (in particular the Sauk, Tippecanoe, Kaskaskia, and Absaroka sequences) is utilized in portions of the stratigraphy and structural history chapters of this report, primarily because the defined sequences are lithogenetic packages which afford a convenient way to contrast major episodes of sedimentation and erosion in the midcontinent area.

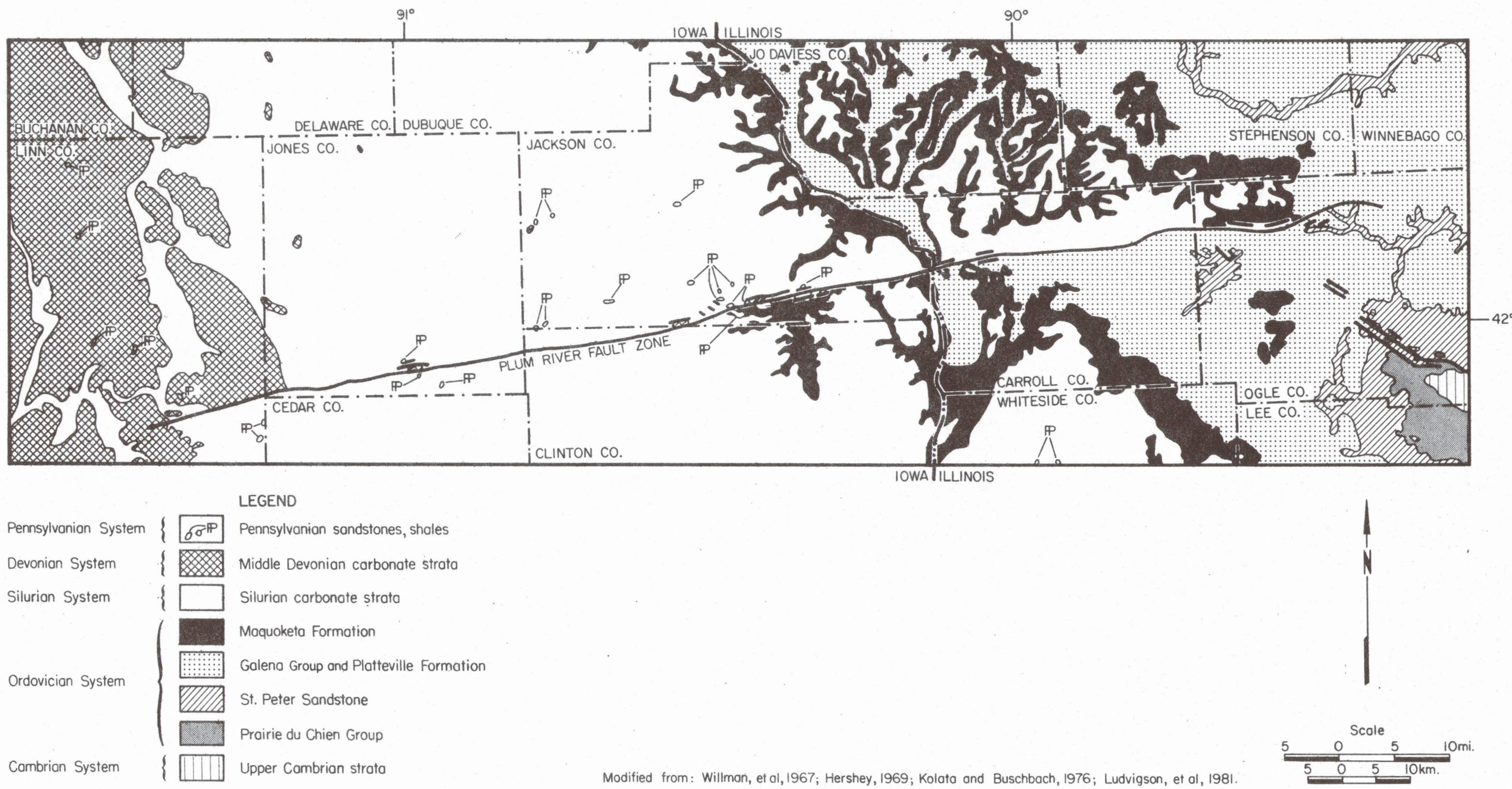


Figure 5. Regional bedrock geologic map of the area bordering the Plum River Fault Zone. The northwestern terminus of the Sandwich Fault Zone of northwest Illinois is shown in the lower right-hand side.

SEQUENCE (Sloss, 1963)	SYSTEM	SERIES	PREVIOUS IOWA GEOLOGICAL SURVEY ROCK-STRATIGRAPHIC NOMENCLATURE: Group	Formation	Member	ROCK-STRATIGRAPHIC NOMENCLATURE THIS REPORT: Group	Formation	Member	LITHOLOGIES: Dominant	Secondary
ABSAROKA	PENNSYL-VANIAN	Desmoinesian-Atokan	Cherokee Gp.				Spoon Fm. & M. Penn. undiff.		ss.	sh.
		Morrowan					Caseyville Fm.		ss.	sh.
		Chesterian					U. Mississippian undiff.--karst filling		sh.	
KASKASKIA	DEVONIAN	Upper					U. Devonian undiff.--karst filling		sh.	
		Middle		State Quarry Fm.			Cedar Valley Fm.	State Quarry Mbr.	ls.	
				Cedar Valley Fm.	Coralville Mbr.			Coralville Mbr.	ls.	
					Rapid Mbr.			Rapid Mbr.	ls.	sh./cht.
					Solon Mbr.			Solon Mbr.	ls.	
				Wapsipinicon Fm.	Davenport Mbr.		Wapsipinicon Fm.	Davenport Mbr.	ls.	dol.
					Spring Grove Mbr.			Spring Grove Mbr.	dol.	ls.
					Kenwood Mbr.			Kenwood Mbr.	sh.	dol.
					Otis Mbr.		Otis Fm.	Cedar Rapids Mbr.	ls.	dol.
					Coggon Mbr.			Coggon Mbr.	dol.	cht.
					Bertram Mbr.		Bertram Fm.		dol.	sh.
TIPPECANOE	SILURIAN	Ludlovian		Gower Fm.			Gower Fm.	see figs. 8 and 9	dol.	
		Wenlockian		Hopkinton Fm.			Scotch Grove Fm.		dol.	cht.
		Llandoverian					Hopkinton Fm.		dol.	cht.
				Kankakee Fm.			Blanding Fm.		dol.	cht.
				Edgewood Fm.			Tete des Morts Fm.		dol.	
							Mosalem Fm.		dol.	sh.
	ORDOVICIAN	Upper		Maquoketa Fm.	Neda Mbr.		Maquoketa Fm.	Neda Mbr.	sh.	ironst.
					Brainard Sh.			Brainard Sh.	sh.	dol.
					Ft. Atkinson Mbr.			Ft. Atkinson Mbr.	dol.	cht.
					Clermont Sh.			Clermont Sh.	sh.	
					Elgin Mbr.			Elgin Mbr.	sh.	dol.
				Galena Fm.	Dubuque Mbr.	Galena Gp.	Dubuque Fm.		dol.	sh.
					Stewartville Mbr.		Wise Lake Fm.		dol.	ls.
		Middle			Prosser Mbr.		Dunleith Fm.		ls.	dol.
				Decorah Fm.	Ion Mbr.		Decorah Fm.	Ion Mbr.	ls.	sh.
					Guttenberg Mbr.			Guttenberg Mbr.	ls.	sh.
					Spechts Ferry Sh.			Spechts Ferry Sh.	sh.	ls.
				Platteville Fm.	McGregor Mbr.		Platteville Fm.	Quimbys Mill Mbr.	ls.	
								McGregor Mbr.	ls.	dol.
					Pecatonica Mbr.			Pecatonica Mbr.	dol.	
				Glenwood Sh.		Ancell Gp.	Glenwood Sh.		sh.	ss.
				St. Peter Ss.			St. Peter Ss.		ss.	
SAUK		Lower	Prairie du Chien Gp.	Shakopee Fm.	Willow River Mbr.	Prairie du Chien Gp.	Shakopee Fm.	Willow River Mbr.	dol.	ss.
					New Richmond Ss.			New Richmond Ss.	ss.	dol.
				Oneota Fm.			Oneota Fm.		dol.	cht.
	CAMBRIAN	Upper		Jordan Ss.			Jordan Ss.		ss.	
				St. Lawrence Fm.			St. Lawrence Fm.		dol.	
				Franconia Fm.			Lone Rock Fm.		ss.	sh.
				Ironton Ss.			Wonewoc Fm.		ss.	
				Galesville Ss.						
				Eau Claire Fm.			Eau Claire Fm.		ss.	sh.
				Mt. Simon Ss.			Mt. Simon Ss.		ss.	

Table 1. Paleozoic stratigraphic nomenclature of east-central Iowa.

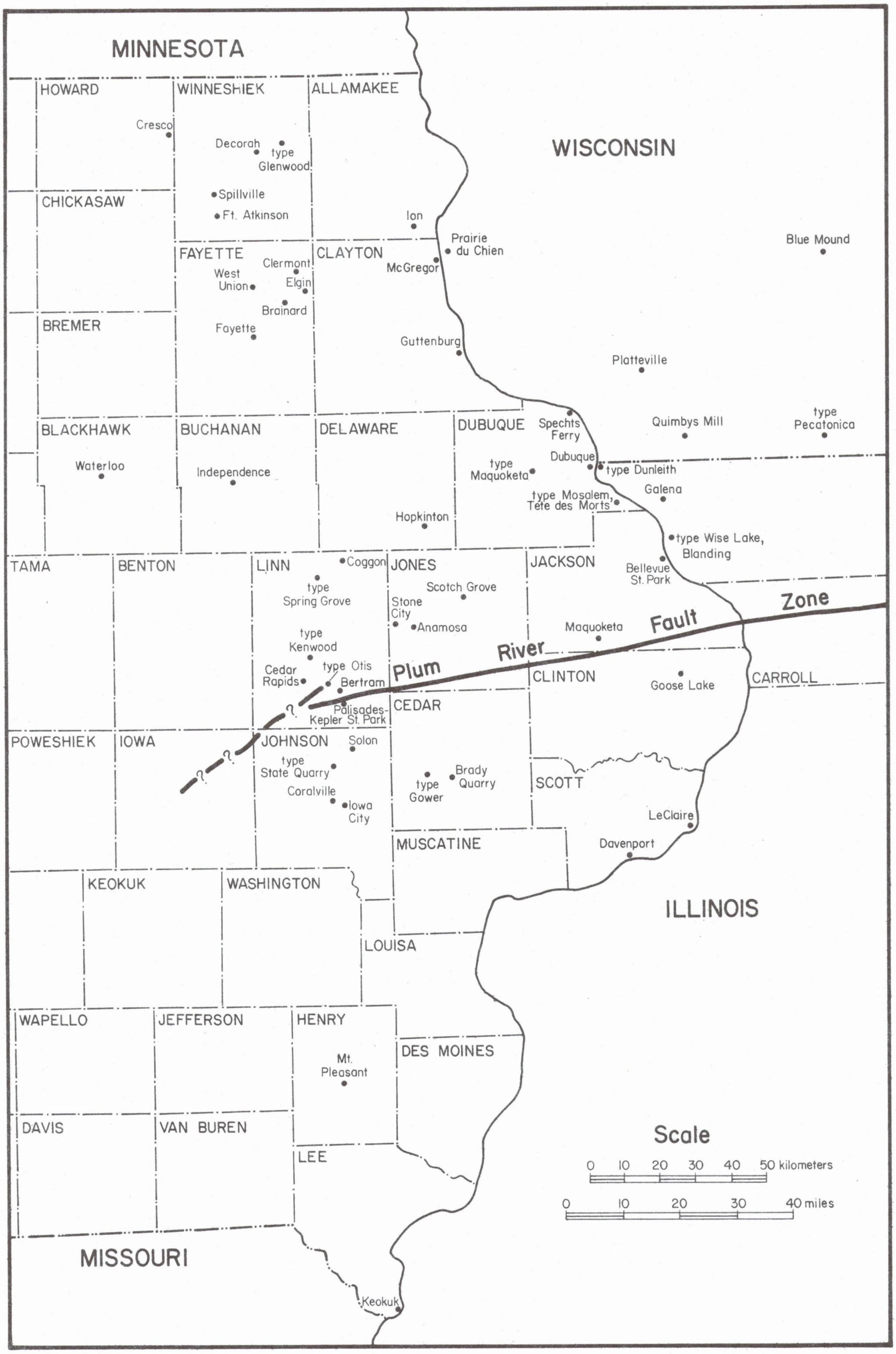


Figure 6. General study area map. Type localities of certain Paleozoic stratigraphic units discussed in report are also shown.

Ordovician System

The oldest rocks that crop out along the Plum River Fault Zone in east-central Iowa belong to the Upper Ordovician Maquoketa Formation. All formations included in the Tippecanoe Sequence (Middle and Upper Ordovician and Silurian) of Iowa are considered in this chapter, but Cambrian and Lower Ordovician (Sauk Sequence) stratigraphic units are not discussed, in part due to the lack of detailed subsurface control. The description of sub-Maquoketa Tippecanoe stratigraphic units (i.e., St. Peter Sandstone through Galena Group) in east-central Iowa is based on examination of well cuttings and cores in the area of the Plum River Fault Zone and on outcrops immediately to the north in Dubuque and northern Jackson counties. Age relationships of Ordovician strata in Iowa are outlined by Sweet and Bergstrom (1976) and Witzke (1980a).

St. Peter Sandstone

A regional unconformity separates rocks of the Sauk Sequence (Upper Cambrian-Lower Ordovician) from those of the overlying Tippecanoe Sequence in Iowa and other parts of the Midcontinent. The initial transgression of the Tippecanoe seas into eastern Iowa was marked by deposition of the Middle Ordovician St. Peter Sandstone (Chazyan). In most of the study area the St. Peter rests directly on the Shakopee Formation (Lower Ordovician, upper Prairie du Chien Group), but in Jackson County, the St. Peter locally overlies a thinned Oneota Formation interval (Lower Ordovician, lower Prairie du Chien Group). The sections where the St. Peter overlies the Shakopee Formation in east-central Iowa generally range in thickness from 30 to 60 feet (9-18 m), rarely ranging up to 100 feet (30 m), and are characterized by friable, very fine to coarse (usually very fine to medium) grained quartzarenite; the grains range from angular to rounded (usually subrounded to rounded).

The St. Peter where it overlies the Oneota Formation is considerably thicker, ranging from about 250 to 345 feet (75-105 m). These thick St. Peter sections are probably fillings of older valleys and karstic terranes developed on the Prairie du Chien carbonate surface during the long erosional episode separating Sauk and Tippecanoe deposition. Farther north in eastern Iowa, in Dubuque and Clayton counties, thick St. Peter Sandstone sections locally rest directly on Upper Cambrian units where the entire Prairie du Chien sequence was erosionally removed prior to St. Peter deposition. The thick St. Peter sections in east-central Iowa are composed primarily of friable, very fine to medium (some coarse) grained sandstone, although the basal 100 feet (30 m) or so locally includes green to reddish brown shales, dolomite and chert clasts (Prairie du Chien lithologies), and quartz pebbles. The lower shaly to conglomeratic interval is referrable to the Readstown (Kress) Member of the St. Peter Sandstone (fig. 7). The remainder of the St. Peter Sandstone in east-central Iowa is assigned to the Tonti Member (Templeton and Willman, 1963). At three well points in Clinton-Jackson counties, where thick St. Peter overlies the Oneota Formation, an 80 to 135 foot (24-41 m) thick interval of green, pink, and reddish brown sandy shale, sandstone, and chert residuum occupies a position directly above Cambrian rock units and beneath the upper portion of the Oneota Dolomite. Although this interval occupies the general position of the Oneota Formation, it is characterized by Readstown Member lithologies. This interval probably represents a St. Peter filling of karst openings within the Oneota sequence.

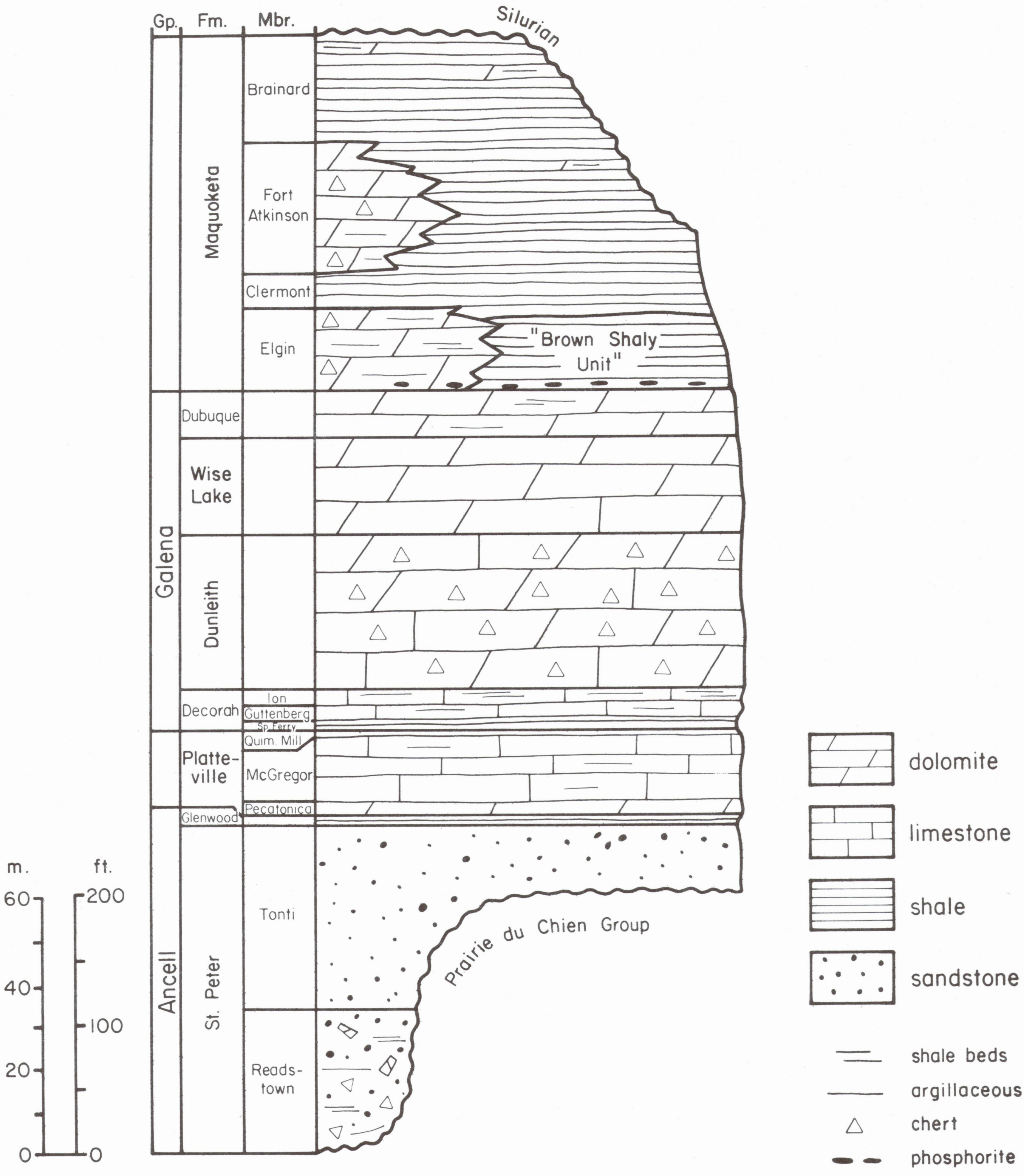


Figure 7. Generalized Middle and Upper Ordovician stratigraphic sequence of east-central Iowa.

Glenwood Shale

The Glenwood Shale overlies the St. Peter, possibly with minor disconformity, and is primarily characterized in the study area by green noncalcareous feldspathic shales, in part sandy or with phosphatic (apatite) pellets. Brown shales also occur, and the uppermost Glenwood bed (usually less than 5 inches thick; 12 cm) is commonly a sandstone. The Glenwood ranges in thickness from about 1 to 10 feet (0.3-3 m) in the study area. A thick sandstone, the Starved Rock Sandstone, is present in parts of Illinois, southeast Iowa, and northern Missouri, and this sandstone is stratigraphically equivalent to a portion of the Glenwood. Templeton and Willman (1963) defined the Starved Rock Sandstone as a member of the St. Peter Sandstone although in southeastern Iowa the Starved Rock overlies typical Glenwood green shales and is, therefore, more consistently included as a member of the Glenwood Formation in Iowa. The thin sandstone at the top of the Glenwood in the study area may correlate, in part, to the much thicker Starved Rock Sandstone to the south. The Glenwood Shale and St. Peter Sandstone collectively comprise the Ancell Group (Templeton and Willman, 1963). The Glenwood is Middle Ordovician (Black Riveran) in age.

Platteville Formation

The Platteville Formation in the study area is a carbonate sequence that ranges in thickness from 62 to 77 feet (19-23 m) and overlies the Glenwood possibly with minor disconformity (Templeton and Willman, 1963). In adjacent Illinois the Platteville is accorded group status and is subdivided into five formations and 24 members (ibid.). In eastern Iowa the Platteville is retained as a formation and is subdivided into three members which, in ascending order, are the Pecatonica, McGregor, and Quimbys Mill. The Pecatonica Member ranges from about 10 to 15 feet (3-4.5 m) in thickness in the study area and is characterized by dense to vuggy, partly fossiliferous dolomite and dolomitic limestone. The basal portion of the Pecatonica is commonly sandy and includes scattered to abundant phosphatic (apatite) pellets and clasts. The McGregor Member, about 40 to 60 feet (12-18 m) thick, conformably overlies the Pecatonica. The McGregor is a dense, fossiliferous, extremely finely crystalline limestone and dolomitic limestone generally occurring in wavy beds, and thin shale partings are commonly noted along bedding. The McGregor is locally sandy. The uppermost 5 feet (1.5 m) or so of the Platteville in the eastern part of the study area is assigned to the Quimbys Mill Member, a brown lithographic limestone with brown shale partings. The Quimbys Mill is, in part, fossiliferous and argillaceous and typically breaks with conchoidal fracture. The Quimbys Mill is recognized in cores from Jackson County, but is absent in areas to the north and west. The Platteville Formation is Middle Ordovician in age (possibly Rocklandian).

Galena Group

The Galena Group includes the Ordovician sequence above the Platteville Formation and below the Maquoketa Formation. It is relatively uniform in thickness throughout the study area, ranging from 235 to 265 feet (72-80 m). As previously utilized by the Iowa Geological Survey, the Galena was accorded formational status and the Decorah Formation was not included within the

Galena. Elevation of the Galena to group status and inclusion of the Decorah within the Galena Group is recommended in Iowa. This redefinition is, in part, an effort to standardize regional usage of the term Galena in order to be more consistent with present usage of the term in adjacent parts of Illinois. Inclusion of the Decorah Formation within the Galena Group properly reflects "the general lithologic similarity" of the Decorah and overlying Galena strata (Templeton and Willman, 1963, p. 95). Additionally, the thick Decorah shales in western Iowa are lithostratigraphically equivalent to the lower interval of Galena carbonates in eastern Iowa and Illinois, and the Decorah shale/Galena carbonate contact is diachronous (Witzke, 1980a, p. 8). These relationships suggest that the Decorah can reasonably be included with the Galena carbonates in a single stratigraphic package (the Galena Group). The Galena Group is subdivided into four formations in eastern Iowa; in ascending order these are the Decorah, Dunleith, Wise Lake, and Dubuque.

Decorah Formation

The Decorah Formation is divisible into three members in eastern Iowa. The lowest member, the Spechts Ferry Shale, ranges from about 5 to 12 feet (1.5-3.7 m) thick in the study area and is primarily a fossiliferous green calcareous shale with fossiliferous limestone interbeds. A widespread bentonite is commonly noted near the base of the member (Millbrig K-bentonite; Willman and Kolata, 1978). The Spechts Ferry Shale disconformably overlies the Platteville Limestone. The Guttenberg Member overlies the Spechts Ferry and is a dense fossiliferous limestone interbedded with reddish brown shale. It ranges from about 10 to 15 feet (3-4.5 m) in thickness in the study area. The Spechts Ferry and Guttenberg are both assigned formational status in adjacent Illinois where the top of the "Decorah Subgroup" is drawn at the top of the Guttenberg. In eastern Iowa an interval above the Guttenberg of interbedded argillaceous fossiliferous dolomitic limestone and green calcareous shale is assigned to the Ion Member of the Decorah Formation. The Ion is locally sandy. The Ion ranges from about 10 to 20 feet (3-6 m) in thickness in the study area but thickens northward and westward in Iowa. Ion equivalents in Illinois are included in the Dunleith Formation. The Ion is retained as a member of the Decorah Formation in Iowa, in part because the Ion shales are laterally equivalent to green calcareous shales included in the Decorah Formation north and west of the study area (Witzke, 1983a). The Ion is a transitional facies separating Decorah green shale facies in northern and western Iowa from equivalent lower Dunleith carbonates in northern Illinois. The Decorah Formation is Middle Ordovician in age (possibly Kirkfieldian).

Dunleith Formation

The Dunleith Formation conformably overlies the Decorah Formation in eastern Iowa and is characterized by argillaceous to non-argillaceous fossiliferous dolomitic limestone and dolomite. The Dunleith is dominated by dolomite along the eastern portion of the Plum River Fault Zone in Iowa, but includes in excess of 30% limestone along the western extent of the fault zone (Witzke, 1983a). Chert is scattered throughout much of the Dunleith and is especially prominent in the upper one-third of the formation. Stylolites and hardgrounds are also noted, and several persistent bentonites are present in the Dunleith sequence. Two zones of abundant receptaculitid algae are present in the formation, and a variety of invertebrate fossils are noted throughout the Dunleith. The Dunleith varies from about 110 to 135 feet (34-41 m) thick in the study area. The type Dunleith in Illinois includes equivalents of the Ion Member of the Decorah Formation in Iowa, and, hence, the Dunleith is used in Iowa in a slightly restricted sense as first suggested by Levorson and Gerk (1972). The Dunleith of eastern Iowa is divisible into eight members, although "differentiation of the Dunleith Formation into members is difficult and requires detailed study to become familiar with minor lithologic features that seem very subtle on preliminary observation" (Willman and Kolata, 1978, p. 46). Dunleith strata in eastern Iowa were formerly included within the "Prosser" Member of the Galena Formation by the Iowa Geological Survey, and the top of the "Prosser" was picked at the top of the highest cherty Galena beds with the overlying non-cherty beds assigned to the "Stewartville" Member. However, Agnew (1955, p. 1723) recognized that dividing the Galena into distinct units based on the presence or absence of chert did not afford a "consistently recognizable" stratigraphic breakdown. Furthermore, "division of the Galena into a cherty Prosser and non-cherty Stewartville has little stratigraphic utility since the top of the cherty interval climbs up and down section across Iowa" (Witzke, 1980a, p. 9). Additionally, the "Prosser" as used in Iowa excluded correlates of the upper beds of the type Prosser section in Minnesota. Therefore, suppression of the "Prosser" as a stratigraphic term in Iowa is recommended, and the Dunleith Formation, as originally defined by Templeton and Willman (1963), is adopted as a consistently recognizable stratigraphic unit in eastern Iowa. The Dunleith is Middle Ordovician in age (probably Kirkfieldian and Shermanian), although the upper portion could conceivably be Late Ordovician (Edenian).

Wise Lake Formation

The Wise Lake Formation in eastern Iowa is characterized by fossiliferous, dense to vuggy dolomite and dolomitic limestone and is about 70 to 80 feet (21-24 m) thick. It conformably overlies the Dunleith. The Wise Lake is dominated by dolomite along the trend of the Plum River Fault Zone, although some limestone is noted locally (Witzke, 1983a). In the study area the Wise Lake is differentiated from the underlying Dunleith primarily by the scarcity of argillaceous impurities and general absence of chert. However, chert is present in the Wise Lake Formation south and west of the study area, and the basal portion of the formation commonly includes chert in eastern Iowa (ibid.). The "upper *Receptaculites* zone" occurs within the Wise Lake. Gastropod fossils are often conspicuous on outcrop and in core, and additional invertebrate fossils are present. The Wise Lake is divided into two members, the Sinsinawa below and Stewartville above, based largely on bedding thickness and siliciclastic content. Both members are decidedly lower in insolubles than the underlying Dunleith Formation. Stylolites and hardgrounds are developed in the Wise Lake. A few thin shale partings are present near the top of the formation, and the contact with the overlying Dubuque Formation is drawn at the base of a widely traceable 4 to 6 inch (10-15 cm) thick carbonate "marker bed" that is set off by prominent 1 inch (2.5 cm) shale partings (Levorson et al., 1979; Willman and Kolata, 1978). All strata now included in the Wise Lake in eastern Iowa were previously referred to the "Stewartville" Member of the Galena Formation by the Iowa Geological Survey, although only the upper portion of the Wise Lake, in fact, correlates with the type Stewartville section in Minnesota. The Wise Lake is Late Ordovician in age (Edenian).

Dubuque Formation

The Dubuque Formation in the study area is characterized by well-bedded dolomite with thin shale interbeds and conformably overlies the Wise Lake Formation (Witzke, 1983a). The dolomites become increasingly argillaceous upward in the Dubuque sequence. The argillaceous content of the Dubuque Formation contrasts with the relatively pure carbonates of the underlying Wise Lake. Echinoderm debris and brachiopods are the most conspicuous fossils noted, although other invertebrates are also present. The Dubuque Formation is about 35 to 40 feet (11-12 m) thick in the study area. The top of the Dubuque is drawn at the base of the overlying Maquoketa phosphorites and brown shales. The Dubuque was recently subdivided into three widely traceable units by Levorson et al. (1979). The Dubuque is Late Ordovician in age (Maysvillian).

Maquoketa Formation

The sequence of Upper Ordovician shales and carbonates above the Galena Group is referred to as the Maquoketa Formation, an interval named after the Little Maquoketa River valley in Dubuque County, Iowa about 33 miles (53 km) north of the Plum River Fault Zone. In adjacent Illinois the Maquoketa has been elevated to group status (Templeton and Willman, 1963), and elevation of the equivalent interval in Iowa to group status may be appropriate at a later date. However, largely because of our inability to correlate Maquoketa strata along portions of the Plum River Fault Zone with the Maquoketa subdivisions utilized in northeast Iowa and Illinois, the Maquoketa is presently retained in formation rank in east-central Iowa. Only after the Maquoketa facies relations have been more clearly defined would recognition of the Maquoketa as a group be recommended. In northeastern Iowa the Maquoketa has been divided into four members, which in ascending order are: Elgin (carbonate, cherty carbonate, shale, phosphorite), Clermont (green dolomitic shale with some carbonate interbeds), Fort Atkinson (carbonate, cherty carbonate), and Brainard (green dolomitic shale, carbonate interbeds). The term Elgin was first used for a Pennsylvanian stratigraphic unit in Kansas, and it may be desirable to rename the Elgin Member of Iowa to avoid duplication. Probable correlates of the Elgin-Clermont interval in Illinois were assigned to the Scales Formation (brown and green shale, carbonate, phosphorite) by Templeton and Willman (1963), who also elevated the Fort Atkinson and Brainard to formational rank. Along most of the Plum River Fault Zone and across much of east-central and southeast Iowa, the Fort Atkinson carbonate interval is not present, and shales largely indistinguishable from Brainard and Clermont lithologies occupy the general position of the Fort Atkinson (Parker, 1970). In the absence of a Fort Atkinson carbonate interval in the Maquoketa sequence, recognition of the Clermont, Fort Atkinson, and Brainard boundaries is not possible at present. Kolata and Graese (1983) recognized similar relations in northwest Illinois. Additionally, the Fort Atkinson carbonates, where present, vary greatly in thickness, and the boundaries of the Fort Atkinson carbonate interval do not form consistent stratigraphic datums but occur at a variety of stratigraphic positions within the Maquoketa sequence (Witzke, 1980a, 1983a). Brown and Whitlow (1960, p. 23) recognized the stratigraphic problems in the Maquoketa sequence of Dubuque-Jackson counties stating that "the strata of the Maquoketa shale are discontinuous locally, and only approximate correlations can be made."

An informal stratigraphic breakdown of the Maquoketa sequence in Jackson County is utilized in this report because of the problems outlined in the previous paragraph. The lower portion of the Maquoketa in Dubuque and Jackson counties was termed the "Brown Shaly Unit" by Brown and Whitlow (1960), which ranges in thickness from about 30 to 50 feet (9-15 m) (fig. 7), and corresponds to the bulk of the Scales Formation in Illinois. The "Brown Shaly Unit" is composed primarily of non-calcareous, partly silty, laminated dark brown, brownish gray, and rarely black shale, and these shales are often graptolitic. Scattered trilobites and inarticulate brachiopods are observed. Brown argillaceous dolomites, in part phosphatic, occur within the "Brown Shaly Unit," in which trilobites, nautiloids, gastropods, bivalves, echinoderms, brachiopods, bryozoans, and graptolites have been noted. The most unusual aspect of the "Brown Shaly Unit" is the presence of phosphorite and phosphatic dolomite horizons. The base of the Maquoketa is typically marked by a 0.3 to 3 foot (10-100 cm) thick phosphorite on a corroded Dubuque Formation surface with up to 3 inches (8 cm) of relief. This surface is encrusted locally with a phosphatic (apatite) or pyritic rind. The phosphorites are composed primarily of concentrically laminated apatite pellets (less than 1 mm), irregularly shaped apatite clasts, and a diverse assemblage of diminutive ("depauperate") phosphatized fossils (usually less than 1 mm). In addition, the phosphorites contain quartz silt, clay, pyrite, and dolomite. The diminutive faunas are molluscan dominated and include bivalves, gastropods, scaphopods, monoplacophorans, polyplacophorans, nautiloids, brachiopods, trilobites, ostracodes, bryozoans, sponges, ophiuroids, pelmatozoans, conularids, scolecodonts, conodonts, and graptolites. A subaerial erosional unconformity has been interpreted at the Dubuque-Maquoketa contact by many previous workers, although in this report the contact in Iowa is regarded as essentially conformable. Both the Dubuque and basal Maquoketa are Maysvillian in age based on conodont faunas. Aside from the abundance of phosphate, the pitted corrosion surface on the top of the Dubuque closely resembles many submarine hardground surfaces in the Galena Group. Higher in the "Brown Shaly Unit" additional hardground surfaces on carbonate interbeds have been noted at several localities in eastern Iowa that are also capped by phosphorites. The "Brown Shaly Unit" correlates with the lower part of the Scales Formation in Illinois, although, in the absence of Fort Atkinson carbonates, the top of the Scales Formation cannot as yet be picked. The "Brown Shaly Unit" probably correlates with the Elgin Member in northeastern Iowa.

The "Brown Shaly Unit" in Jackson-Dubuque counties is conformably overlain by a thick green to gray dolomitic shale interval reaching a maximum thickness of about 190 feet (58 m). This interval was assigned to the Brainard Member by Brown and Whitlow (1960), although it undoubtedly includes Clermont and Fort Atkinson equivalents as well. In this report the thick shale sequence above the "Brown Shaly Unit" is informally assigned to an undifferentiated Clermont-Brainard interval. The shales are unfossiliferous throughout much of the sequence, although burrows and trilobites are noted. Argillaceous dolomite interbeds are scattered within the shale sequence, and in the upper half of the interval the dolomite interbeds include scattered to abundant invertebrate fossils. The upper portion of the Clermont-Brainard interval is exposed along the Plum River Fault Zone and correlates with the Brainard Shale of northeastern Iowa. Fossiliferous dolomite and dolomitic limestone interbeds become prominent in the upper part of the Maquoketa Formation (generally the upper 20-80 feet; 6-24 m), and this interval has been termed the "*Cornulites* zone" by Ladd (1929). A brief description of the fauna and lithology of these fossiliferous carbonates was given by Witzke (1978).

The Clermont-Brainard shale interval extends westward along the Plum River Fault Zone in the subsurface into portions of Jones and Linn counties.

The shale-dominated "Brown Shaly Unit" is replaced in Jones County by a 50 foot (15 m) thick argillaceous dolomite-dominated sequence with subordinate brown shale and is reasonably assigned to the Elgin Member (fig. 7). In portions of the Linn County subsurface a Maquoketa sequence closely similar to that described from the northeast Iowa outcrop belt is documented. There the lower 50 to 90 feet (15-27 m) of the formation is characterized by argillaceous dolomite, in part cherty and fossiliferous, with interbeds of brown shale; a basal phosphorite bed is usually noted. This lower Maquoketa interval is assigned to the Elgin Member. The Elgin is overlain by the Clermont Member, a green to gray dolomitic shale with some carbonate interbeds, which ranges in thickness from about 20 to 45 feet (6-14 m). Unlike the sequence along the Plum River Fault Zone in Jackson County, the Linn County Maquoketa sequence locally includes a carbonate interval assignable to the Fort Atkinson Member, which is characterized by cherty to very cherty argillaceous dolomites with some dolomitic shale interbeds (fig. 7). It ranges in thickness from 50 to 115 feet (15-35 m). The Brainard Member is the upper unit of the Maquoketa present in Linn County, where it is typified by green to gray dolomitic shales with interbeds of argillaceous dolomite; argillaceous dolomite locally makes up to 50% of the Brainard interval. It varies in thickness from 35 to 90 feet (11-27 m).

The uppermost member of the Maquoketa Formation, the Neda, is only locally preserved in Iowa and has not yet been noted along the trend of the Plum River Fault Zone. The Neda is known to outcrop in Dubuque County (Brown and Whitlow, 1960) and is present in the subsurface at scattered localities across much of the state (Parker, 1970). The Iowa Neda is characterized by red silty dolomitic shales and oolitic ironstones. The ooids are composed primarily of goethite with scattered laminae of apatite and chamosite. Apatite clasts are also noted. In east-central Iowa the Maquoketa Formation varies greatly in thickness, primarily because of extensive pre-Mosalem (Silurian) erosion. Documented Maquoketa thicknesses in the area, where capped by Silurian strata, range from 114 to 275 feet (35-84 m). At Bellevue State Park (Jackson Co.) the measured Maquoketa interval (capped by Silurian) is only about 115 feet (35 m) thick, and is the thinnest known complete Maquoketa exposure in the state. At the time Silurian deposition began in east-central Iowa, up to 135 feet (41 m) of vertical relief was present on the Maquoketa Shale surface (Brown and Whitlow, 1960). The Maquoketa Formation is Late Ordovician in age (Maysvillian, Richmondian).

Silurian System

The Silurian stratigraphy is considered at some length in this report for two reasons. 1) Silurian dolomite exposures form the bulk of the outcrops along the Plum River Fault Zone in east-central Iowa, and an understanding of the Silurian stratigraphy is critical to interpreting the present-day structural features along the fault zone. 2) New stratigraphic terminology and interpretations are presented which necessitate expanded discussion and definition. On first inspection the Silurian dolomites of eastern Iowa appear to consist of a monotonous sequence of generally uniform lithology, and this impression may, in part, account for the general failure of many previous investigators to understand the structural significance of outcrops along the trend of the recently discovered Plum River Fault Zone in east-central Iowa. Recent studies (Johnson, 1975, 1983; Witzke, 1978, 1981a, b) have clarified some of the Silurian stratigraphic relations in eastern Iowa, and this report expands and revises some previous interpretations of the Silurian sequence.

Mosalem Formation

The Mosalem was originally defined by Brown and Whitlow (1960) as a member of the Edgewood Formation. Willman (1973) elevated it to formational status. Witzke (1978, p. 5) stated that "the term Edgewood is inappropriate in eastern Iowa" because of lithic and age discrepencies with the type Edgewood in northeastern Missouri. The Mosalem is the most argillaceous unit noted in the Silurian of Iowa and is primarily characterized by dense, thin-bedded, argillaceous dolomite. The lower portion of the formation locally includes dolomitic shales, and generally becomes less argillaceous upward. Where the Mosalem is thick, bands of chert nodules are sometimes observed in the middle or upper portions. The basal portions of the formation are often silty and locally contain conglomeratic zones of reworked dolomite, fossils, shale, and ironstone ooids derived from underlying rocks of the Maquoketa Formation (Brown and Whitlow, 1960, p. 39). Pyrite and glauconite are also noted in the lower part of the formation. A stromatolitic horizon is locally present near the base of the formation, and halite pseudomorphs and ripple marks have been noted in the lower Mosalem at Bellevue State Park, Jackson County, Iowa.

The Mosalem rests unconformably on the Brainard Shale of the Maquoketa Formation, although the contact appears gradational with the underlying shales at some localities (Willman, 1973, p. 33). The Mosalem fills in the lows on an eroded Maquoketa surface and varies in thickness accordingly from 0 to 100 feet (30 m) (fig. 8). It is absent west of Jones County, Iowa. The Mosalem is sparsely fossiliferous, although graptolites, *Lingula cuneata*, rhynchonellid and strophomenid brachiopods, gastropods, worms, and plant filaments are noted (Johnson, 1975, 1977a; Witzke, 1983b). Graptolites collected near the base of the formation in Jackson County (Ross, 1964) indicate an Early Silurian age (early Llandoverian).

Tete des Morts Formation

The Tete des Morts was originally defined as a member of the Edgewood Formation, and Willman (1973) elevated it to formational rank. As previously noted, the Edgewood is an inappropriate stratigraphic term in Iowa. The Tete des Morts Formation is characteristically a massive, fine- to medium-grained, dense to vuggy dolomite. It is a prominent cliff-forming unit in outcrop. It is slightly glauconitic, and a zone of chert nodules is often present near the middle of the unit. Silicified corals and stromatoporoids are frequently noted. The Tete des Morts is conformable with both the underlying Mosalem Formation and the overlying Blanding Formation. The massive-bedded vuggy dolomites of the Tete des Morts contrast markedly with the underlying thin-bedded, argillaceous dolomites of the Mosalem Formation. In areas where the Mosalem is absent, the Tete des Morts rests unconformably on the Brainard shales or Neda shaly ironstones of the Maquoketa Formation. Common Tete des Morts fossils include stromatoporoids, tabulate corals (*Favosites*, *Syringopora*, *Halysites*), rugose corals, brachiopods, and crinoidal debris.

The Tete des Morts is absent west of Jones County and is absent locally in areas farther east (Willman, 1973). It reaches a maximum thickness of

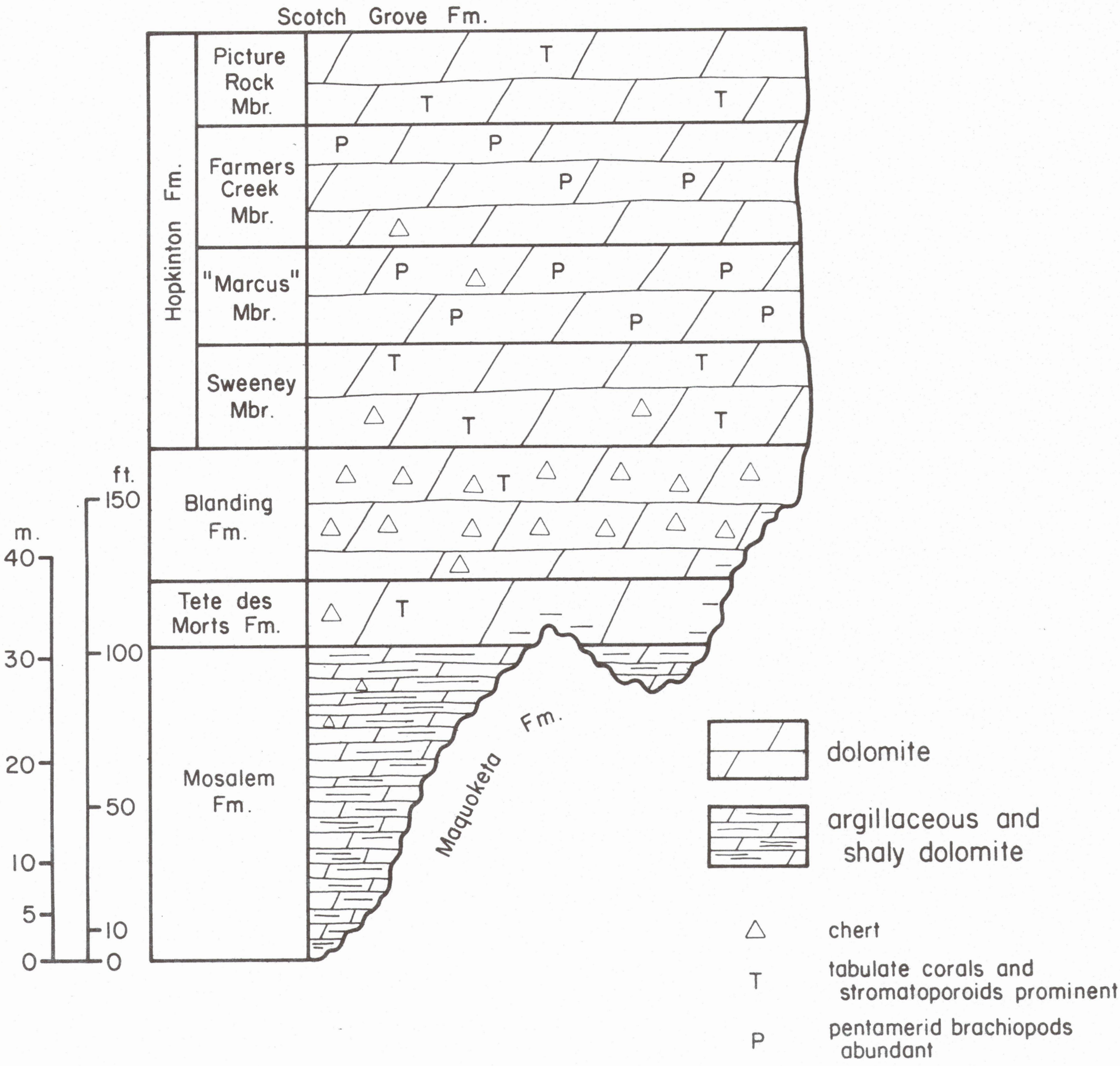


Figure 8. Generalized sub-Scotch Grove Lower Silurian stratigraphic sequence, outcrop belt of east-central Iowa.

about 24 feet (7.3 m) in eastern Iowa and northwestern Illinois (Brown and Whitlow, 1960, p. 40; Willman, 1973, p. 34). In the subsurface of Jones and Jackson counties, Iowa, the Tete des Morts is not always separable from the underlying Mosalem Formation, and the entire Mosalem-Tete des Morts interval may be characterized by dense, argillaceous, cherty dolomites. These sections are labelled Mosalem-Tete des Morts undifferentiated.

Blanding Formation

The interval now termed the Blanding Formation was previously labelled the Kankakee Formation by most previous workers and by the Iowa Geological Survey. Due to previous miscorrelations of rock-stratigraphic units, the Kankakee is not an appropriate name for the cherty beds presently included within the Blanding Formation (Willman, 1973; Witzke, 1978). In the outcrops of Jackson and Dubuque counties, Iowa, the lower 8 feet (2.4 m) of the Blanding is a dense, extremely fine to fine-grained dolomite in beds 3 to 8 inches (7.5 to 20 cm) thick with scattered chert nodules. This is the "Lower Quarry beds" of Calvin and Bain (1900). The upper portion of the Blanding in the same area is a dense to slightly vuggy, extremely fine to fine-grained, thin-to medium-bedded dolomite with abundant chert nodules and continuous beds of white chert. Chert locally makes up as much as 50 per cent of the Blanding section (Brown and Whitlow, 1960, p. 42; Willman, 1973, p. 36). West of Jones County, Iowa, the Blanding is argillaceous and pyritic with a basal zone containing abundant quartz silt. Silicified corals are prominent on outcrop. Fossils in the Blanding include stromatoporoids, tabulate corals (*Favosites*, *Syringopora*, *Halysites*), rugose corals, brachiopods (including *Cryptothyrella*, orthids), and echinoderm debris (Johnson, 1977a).

The Blanding is conformable with the underlying Tete des Morts Formation, but the contact is marked by "an abrupt lithologic change" (Brown and Whitlow, 1960, p. 42). A conspicuous bed of chert helps delineate the top of the Blanding Formation. West of Jones County, Iowa, the Blanding unconformably overlies the Brainard shales of the Maquoketa Formation as the Mosalem and Tete des Morts disappear (fig. 8). Where the Blanding overlies the Tete des Morts it varies from 28 to 60 feet (8.5-18 m) in thickness. In the subsurface west of Jones County where it rests on the Brainard Shale it varies from 13 to 33 feet (4-10 m) in thickness. The Blanding is Early Silurian (probably middle Llandoverian) in age (Johnson, 1975).

Hopkinton Formation

The Hopkinton Formation is herein restricted to include beds above the Blanding Formation and below the "upper quarry beds" of Calvin and Bain (1900) and the "building stone beds" of Wilson (1895). Thus, it is returned to the original definition of Calvin (1896, p. 49). The upper boundary of the Hopkinton, as defined, is a recognizable lithologic boundary across eastern Iowa, and beds previously included within the Hopkinton above the base of the "upper quarry beds" are herein assigned to a new formation, the Scotch Grove. A brief resume of the historical development of the Silurian stratigraphic nomenclature in Iowa is needed to explain the placement of the upper boundary of the Hopkinton as utilized in this report.

Most previous workers have assumed that the Hopkinton Dolomite underlies the Gower Dolomite, although the nature of this contact has received little attention in the literature. Unfortunately, the top of the Hopkinton has been defined by previous workers at more than one position, and the stratigraphic position of the upper boundary needs to be clarified. Wilson's (1895) report on the Silurian stratigraphy of northeastern Iowa was utilized by Calvin (1896, p. 49) to define the "Delaware Stage" of the Silurian which included "divisions one to four of Professor Wilson's paper." Division four was termed the "upper Coralline beds" lying in stratigraphic position above division three, the "*Pentamerus* beds." Utilizing the stratigraphic terminology of

Johnson (1983), the top of Wilson's fourth division would correspond to the top of the Picture Rock Member (the "*Favosites* Beds" of Johnson, 1975). Because the term Delaware was pre-occupied for a Devonian rock unit in Ohio, Calvin introduced the "Hopkinton Stage" as a replacement in 1906.

Later workers (e.g. Boucot, 1964; Berry and Boucot, 1970; Johnson, 1975, 1983; Witzke, 1978) included a thick stratigraphic interval above Wilson's fourth division within the Hopkinton Dolomite, although Calvin's original definition of the upper boundary of the Hopkinton remains the most practical. The interval above Wilson's fourth division, but below the first laminated Gower dolomites, includes *Pentameroides*-bearing cherty dolomites (*Pentameroides* is a large pentamerid brachiopod). Earlier inclusion of *Pentameroides*-bearing beds within the Hopkinton Dolomite by Calvin may have influenced the stratigraphic placement of these beds by later workers. The inclusion of *Pentameroides*-bearing beds within the Hopkinton by Calvin, however, resulted from the fact that "Calvin equated the occurrence of *Pentamerus oblongus* from his *Pentamerus* Beds with that of *Pentameroides subrectus* from younger strata" (Johnson, 1975, p. 132). Johnson (1975, p. 133) further explained: "The error caused Calvin to exclude several distinctive beds from his stratigraphy. He was perplexed that faunas succeeding what he thought to be the same *Pentamerus* beds at localities not far separated were so different. This factor Calvin attributed to the inconstancy of life on the Silurian sea bottom." Calvin's error in fossil identification had the effect of elevating the top of the Hopkinton to include rocks not originally included within his definition of the Hopkinton. The "upper Hopkinton" beds, i.e., strata above the Picture Rock Member of Johnson (1983), had previously been referred to the LeClaire Dolomite by some workers (e.g. Rowser, 1929).

The Hopkinton Formation previously was divided into a series of stratigraphic units named after distinctive fossils present (Calvin and Bain, 1900; Johnson, 1975). Witzke (1980b) outlined a westward thinning of the sub-"*Favosites* Beds" (Picture Rock Member) portion of the Hopkinton Dolomite between the outcrop belt of Jackson-Dubuque counties and the subsurface of Linn-Johnson counties, and, for the first time, the "upper Hopkinton Formation (post-*Favosites* beds)" was noted to be partly a facies equivalent of "porous, crinoidal beds" that had previously been "assigned to the Gower Formation." Witzke (1981b, 1983b) informally subdivided the Hopkinton into three rock units termed Hopkinton A, B, and C, which he correlated from the outcrop belt into the subsurface of Linn, Johnson, and Benton counties. Johnson's extensive studies (1975, 1977a, 1977b, 1979, 1980, 1983) have clarified much of the stratigraphy, paleontology, and paleoecology of the Lower Silurian in Iowa. Johnson (1983) formally named a series of members within the Hopkinton Dolomite, and these subdivisions are used in this report (fig. 8). However, the three upper members of Johnson (1983) (Johns Creek Quarry, Welton, and Buck Creek Quarry members) occupy a position above Wilson's (1895) fourth division and are excluded from the Hopkinton Dolomite in this report. Johnson (1980, 1983) placed the Hopkinton-Gower contact at the top of the *Pentameroides*-bearing Buck Creek Quarry Member; lithologic criteria for recognizing this contact were not given, although Johnson (1983, p. 13) suggested that the Hopkinton Dolomite "disconformably underlies the Gower Formation."

The placement of the upper boundary of the Hopkinton at the top of the *Pentameroides*-bearing Buck Creek Quarry Member is rejected in this study for several reasons: 1) it is not the position at which Calvin (1896) originally defined it; 2) stratigraphic cross sections of complete Silurian core penetrations in Johnson-Benton-Linn counties demonstrate that the top of the *Pentameroides*-bearing beds occurs at a variety of stratigraphic positions, and its utility as a stratigraphic datum is, therefore, seriously impaired; 3) *Pentameroides*-bearing beds are locally absent in many subsurface core sections; 4) where *Pentameroides*-bearing beds are present, they are commonly overlain by rocks of closely similar lithology. For these reasons, it seems inappropriate to define the top of the *Pentameroides*-bearing beds as the top of a formation. These objections are avoided if the top of the Hopkinton is placed at the top of the "upper Coralline beds" (= Picture Rock Member of Johnson, 1983) where Calvin (1896) defined it.

The Hopkinton Formation in the outcrop belt of eastern Iowa (Clinton, Delaware, Dubuque, Jackson, and Jones counties) varies from 115 to 160 feet (35-49 m) in thickness and averages about 130 feet (40 m) thick. In the subsurface of Linn and eastern Benton counties, the Hopkinton interval is thinner, varying from 62 to 94 feet (19-29 m) in thickness. In the subsurface of Johnson County it is thinner still, varying from 38 to 62 feet (12-19 m).

Sweeney Member. The Sweeney Formation was named by Willman (1973) for exposures near Savanna in Carroll County, Illinois. He equated the unit with the "*Syringopora* Beds" of Calvin and Bain (1900) in Iowa. Johnson (1983) recognized the Sweeney in eastern Iowa as the basal member of the Hopkinton Dolomite, and his assignment is followed in this report. The Sweeney is characterized by massive to medium-bedded, extremely fine- to medium-crystalline, often stylolitic dolomite. It ranges from dense to vuggy, and light yellow orange microporous mottled zones usually are scattered throughout the interval. The rock unit is locally cherty, especially in the middle, although scattered to abundant chert nodules may be present at any level within the interval. The dolomites are characterized by skeletal-moldic wackestone and dolomite-replaced skeletal wackestone and packstone fabrics. Green clay partings are noted in the easternmost Iowa outcrops. Silicified or moldic tabulate corals (*Favosites*, *Halysites*, *Syringopora*) and stromatoporoids are prominent. Rugose corals, brachiopods, gastropods, trilobites, and echinoderm debris are also present (Johnson, 1977a). Strata containing concentrations of the brachiopod *Stricklandia lens progressa* are recognized in the middle portion of the member (Johnson, 1980, 1983). The Sweeney Member rests conformably on the Blanding Formation, and the contact is characteristically sharp. The underlying thin- to medium-bedded, very cherty (including bedded cherts) dolomites of the Blanding contrast markedly with the medium- to massive-bedded, sporadically cherty (nodular) dolomites of the Sweeney. The top of the member is drawn at a sharp lithologic break below strata containing packed concentrations of large pentamerid brachiopods (*Pentamerus oblongus*). The Sweeney Member ranges from 33 to 45 feet (10-16.5 m) in thickness in the outcrop belt of eastern Iowa (Jackson, Dubuque, Jones, Clinton, Delaware, Bremer counties), averaging about 40 feet (12 m). In Jackson County the Sweeney Member shares complementary thickness variations with overlying *Pentamerus*-bearing strata, suggesting that where the Sweeney is thickest, the upper portion may be a partial facies equivalent of the "*Pentamerus* Beds." In the subsurface in the western part of the study area the distinction between coralline Sweeney and *Pentamerus*-bearing strata is more obscure; coralline and *Pentamerus*-bearing beds locally interfinger, and *Pentamerus* is locally sparse to absent. In that area the top of the Sweeney cannot be consistently recognized, where the member is informally included in the "Lower Hopkinton" interval (see later section). As such, the Sweeney Member is restricted to the outcrop belt of eastern Iowa and northwestern Illinois, and it loses its distinction to the west. The Sweeney Member is of Early Silurian (late Llandoverian, C_1 or C_2) age (Johnson, 1979, 1983).

"Marcus Member." The Marcus Formation was named by Willman (1973) for the 40 foot (12 m) thick interval of dolomite above the Sweeney in Carroll County, Illinois. He noted that the lower 5 to 15 feet (1.5-4.5 m) of the Marcus contains packed concentrations of *Pentamerus oblongus*, and he equated the unit with the "*Pentamerus* beds" in Iowa. Johnson (1983) formally introduced the Marcus as a member of the Hopkinton Dolomite to replace the term "*Pentamerus* Beds" (Calvin and Bain, 1900; Johnson, 1975). However, he acknowledged that "it is possible that part of the '*Cylocrinites* Beds' was included by Willman in his Marcus Formation" (Johnson 1983, p. 14). If true, Johnson's use of the term Marcus in Iowa differs from the original usage of the term in the type area of northwestern Illinois. Preliminary investigations of Marcus strata in Carroll County, Illinois, by the authors of this report suggest that strata equivalent to much or all of the "*Cyclocrinites* Beds" (Farmers Creek Member of Johnson, 1983) are included within the Marcus Formation of Illinois. Additional studies hopefully will clarify the stratigraphic relations between the "*Pentamerus*" and "*Cyclocrinites* Beds" in Iowa and type Marcus strata in Illinois. Pending these studies, the *Pentamerus*-bearing interval in the Hopkinton of Iowa is tentatively included in the "Marcus Member" following Johnson's (1983) usage.

The "Marcus Member" in eastern Iowa is characterized by medium-bedded fine-to medium-crystalline dolomite containing scattered to abundant molds of the large pentamerid brachiopod *Pentamerus oblongus*. The dense concentrations of brachiopods form a distinctive lithology that contrasts markedly with underlying Sweeney strata. The "Marcus Member" is locally cherty, and scattered chert nodules can occur at any stratigraphic position within the member. In addition to abundant *Pentamerus*, the member has also yielded specimens of tabulate and and rugose corals, additional brachiopod genera, bryozoans, gastropods, trilobites, and echinoderm debris (Johnson, 1977a). Accumulations of pentamerid brachiopods in the "Marcus" are preserved in two general ways with "a complete preservational spectrum" between the two extremes: 1) articulated shells in life position with scattered truncated "submarine erosion" surfaces (bored hardrounds), and 2) "disarticulated shells accumulated as coquinas" (Johnson, 1977b, p. 86, 92).

The "Marcus Member" in the outcrop belt of eastern Iowa (Jackson, Clinton, Dubuque, Jones, Delaware counties) ranges from 11 to 32 feet (3.4 -9.8 m) in thickness. It is conformable with overlying and underlying units. *Pentamerus*-bearing strata in the subsurface west of Jones County locally interfinger with coralline Sweeney-like strata or are locally absent. In that area the "Marcus" is not recognized as a distinct stratigraphic unit, and equivalent strata are included in the upper portion of the "Lower Hopkinton" interval. The "Marcus Member" in Iowa is an Early Silurian (late Llandoverian, C_2 or C_3) rock unit (Johnson, 1979, 1983).

"Lower Hopkinton." As previously discussed, the Sweeney and "Marcus" members are not readily separable in the subsurface west of Jones County, Iowa, and equivalent strata are informally included in the "Lower Hopkinton." Witzke (1981b, 1983b) grouped the interval including coralline and *Pentamerus*-bearing strata in the lower Hopkinton into a single informal stratigraphic unit, Hopkinton A. He (1981b, p. 45) noted that there commonly is "no lithologic break corresponding to the appearance of *Pentamerus* in the sequence" in the subsurface west of Jones County, where the Hopkinton A interval could not be consistently subdivided into two units.

The "Lower Hopkinton" interval in the western portion of the study area is characterized by extremely fine to medium-crystalline dolomite. It is commonly stylolitic, and nodular chert may occur locally at any level within the interval. The "Lower Hopkinton" ranges from dense to vuggy, and microporous mottled zones are scattered throughout. The dolomites are characterized by skeletal-moldic wackestone and dolomite-replaced wackestone and packstone fabrics. Silicified and moldic tabulate corals and stromatoporoids are common through much of the interval, and scattered to abundant molds of *Pentamerus oblongus* are commonly noted in the upper one-half to one-fourth of the "Lower Hopkinton." The combined Sweeney and "Marcus" members in the outcrop belt of east-central Iowa (Clinton, Delaware, Dubuque, Jackson, Jones counties), which is equivalent to the "Lower Hopkinton" interval farther west, varies from 54 to 70 feet (16.5-21 m) in thickness. The "Lower Hopkinton" interval thins to the west, and in the subsurface of Linn and eastern Benton counties it ranges from 40 to 61 feet (12-19 m) in thickness. In the subsurface of Johnson County the interval is thinner still, ranging from 18 to 45 feet (5.5-14 m) in thickness.

Farmers Creek Member. The Farmers Creek Member was defined by Johnson (1983) for strata previously termed the "Cerionites Beds" (Calvin and Bain, 1900), "*Cyclocrinites* Beds" (Johnson, 1975), and "Hopkinton B" (Witzke, 1981b, 1983b). The member is characterized by its "massive, unbedded appearance" (Johnson, 1975, p. 136) and is primarily a microcrystalline to finely crystalline dolomite. It locally includes medium crystalline dolomites. It is characterized typically by skeletal-moldic wackestone fabrics and also includes dolomite-replaced skeletal grains in a wackestone to packstone fabric. The Farmers Creek Member interval is generally very porous due to abundant fossil molds and solutional vugs. It is usually chert-free, although scattered chert nodules are noted in some sections. Stylolites are generally absent, although they are present locally in some subsurface sections. Abundant molds of pentamerid brachiopods are often present in the upper half of the interval in the outcrop sections, although the zone of abundant pentamerid brachiopod molds is usually developed in the lower half of the interval in the subsurface sections west of Jones County (Witzke, 1983b). The member is very fossiliferous, and well preserved fossil molds are common. The golf-ball-shaped green algae, *Cyclocrinites*, is locally prominent (Nitecki and Johnson, 1978). In addition, tabulate corals, rugose corals, gastropods, nautiloids, trilobites, bryozoans, and a variety of brachiopods (including the stricklandid *Stricklandia laevis* and the globular pentamerid *Harpidium maquoketa*) are present. Echinoderm debris is scattered throughout the sequence, and articulated cystoid, paracrinoid, and crinoid cups are noted locally, sometimes abundantly (Witzke, 1976; Witzke and Strimple, 1981; Frest et al., 1980).

The Farmers Creek Member is conformable with underlying "Marcus" and "Lower Hopkinton" strata, from which it is distinguished by its finer-grained crystalline texture, more massive bedding, general lack of chert and stylolites, and its common vuggy and porous character. The lower boundary is usually sharp, although in the western subsurface sections the boundary is gradational where the lower pentamerid brachiopod-bearing portion of the Farmers Creek resembles the pentamerid-bearing beds in the upper part of the "Lower Hopkinton." The Farmers Creek Member in the outcrop belt of eastern Iowa varies from 30 to 48 feet (9.1-14.6 m) in thickness. It is thinner in the subsurface of Linn and eastern Benton counties, where it varies from 13 to 25 feet (4-7.6 m) in thickness. It becomes thinner still in the subsurface of Johnson County, ranging from 5.3 to 13.3 feet (1.6-4.1 m).

Part or all of the Farmers Creek Member may correlate with the upper twothirds or so of the Marcus Formation in northwestern Illinois, but further study is needed. The Farmers Creek is the main cave-forming interval in the Silurian sequence (Witzke, 1978), and the interval forms a productive carbonate aquifer in eastern Iowa because of its porous and permeable character. The Farmers Creek Member is Early Silurian (late Llandoverian, probably C_4) in age (Johnson, 1979, 1983).

Picture Rock Member. The Picture Rock Member was defined by Johnson (1983) for strata previously termed the "upper Coralline beds" (Wilson, 1895), "*Favosites* Beds" (Johnson, 1975), and "Hopkinton C" (Witzke, 1981b, 1983b). The member is characterized by medium- to thick-bedded, extremely fine to coarse crystalline dolomite. It is often vuggy and microporous mottled, and stylolites are usually present. The dolomites are typified by small crinoidal debris molds in a wackestone fabric and by dolomite-replaced crinoid grains in a wackestone to packstone fabric. Silicified or moldic tabulate corals and stromatoporoids are prominent in most sections. Chert nodules are generally absent, but are rare in some sections. Tabulate corals (*Favosites*, *Halysites*, *Syringopora*) and stromatoporoids are conspicuous on outcrop (Witzke, 1983b), and rugose corals, bryozoans, brachiopods (including occasional *Pentamerus oblongus* near the base), and echinoderm debris are also observed (Johnson, 1977a). In thin section, echinoderm grains are volumetrically the most significant skeletal component of the Picture Rock Member (Witzke, 1981b). In a portion of Delaware County equivalents of the member crop out as a limestone (lower LaPorte City Formation) where the rock is a crinoidal wackestone to packstone with scattered tabulate corals and stromatoporoids.

The Picture Rock Member is sharply bounded but conformable at both upper and lower contacts. It is less porous and more coarsely crystalline than the underlying Farmers Creek Member. It is also more coarsely crystalline than the overlying dense to skeletal-moldic more finely crystalline rocks of the Scotch Grove Formation. The upper boundary of the Picture Rock Member is a distinctive and widely traceable marker in eastern Iowa. The member in the outcrop belt of eastern Iowa (Clinton, Delaware, Dubuque, Jackson, and Jones counties) varies from 17 to 42 feet (5.3-12.8 m) in thickness. It is thinner in the subsurface of Linn and eastern Benton counties, varying in thickness from 8.2 to 16.3 feet (2.5-5 m). The Picture Rock Member thins dramatically in the subsurface of Johnson County where it varies from 3.6 to 13 feet (1.1-4 m). Its general lithologic and paleontologic character is closely similar to that of the Sweeney Member, and similar depositional and diagenetic environments must have affected both rock units. The Picture Rock Member correlates with the lower portion of Willman's (1973) "Racine Formation" in northwestern Illinois. It should be noted that the base of Willman's "Racine" is not equivalent to the base of the type Racine Formation in Wisconsin. The Picture Rock Member is Early Silurian (late Llandoverian, C_4 or C_5) in age (Witzke, 1981b).

Scotch Grove Formation (new)

The Scotch Grove Formation is herein formally defined as the dolomite and cherty dolomite interval above the Picture Rock Member and below the base of the laminated and mounded dolomites of the Gower Formation. The Scotch Grove was introduced by Witzke (1981a, p. 11) as "an informal rock unit." For historical and lithostratigraphic reasons the top of the Hopkinton Formation is presently drawn at the top of Johnson's (1983) Picture Rock Member. The

Gower Formation was assumed to lie in stratigraphic position above the Hopkinton by most previous workers. The type locality of the Gower Formation is characterized by laminated dolomites (Norton, 1899), and Norton also included all rocks laterally equivalent to the laminated sequence within the Gower. Norton's definition of the Gower is maintained in this report, even though Norton and other workers later included, apparently inadvertently, a sequence of flat-lying and mounded dolomites within the Gower Formation that is not equivalent to any portion of the laminated Gower dolomite sequence but, in fact, lies stratigraphically below it. Therefore, an interval of rocks above the top of the Picture Rock Member and below the base of the first laminated Gower dolomites (or their equivalents) cannot be assigned to either formation if the original formational definitions are maintained. Two options are available to resolve the dilemma: 1) expand the original definition of Gower and/or Hopkinton formations to include this interval, or 2) erect a new formation to include the interval. As noted in the Hopkinton Formation discussion, the top of the *Pentameroides*-bearing beds does not form a consistent stratigraphic datum and cannot readily serve as a formational boundary. Instead, a complex of related carbonate facies occupies the interval in question. Therefore, no attempt has been made to define a Hopkinton/Gower contact within the interval, but rather a new formation name, the Scotch Grove, is introduced. The original definitions of both the Hopkinton and Gower are preserved with the introduction of the Scotch Grove Formation.

No single locality exposes the entire Scotch Grove Formation, and the type locality exposes only the lower portion (40 feet; 12 m) of the formation where it is in contact with the underlying Hopkinton. A quarry area (NW SE SE and SE NW SE sec. 7, T85N, R2W) north of the town of Scotch Grove, Jones County, Iowa, is designated the type locality. A measured stratigraphic section of the sequence at this locality can be referenced in Johnson (1977a, p. 130). The type section exposes a stratigraphic sequence above the Picture Rock Member of the Hopkinton Dolomite that includes, in ascending order, 10 feet (3 m) of the Johns Creek Member and 30 feet (9 m) of the Welton Member. The constituent members of the Scotch Grove Formation are described in subsequent sections of this report. Additional localities in eastern Iowa that expose portions of the Scotch Grove sequence as well as subsurface core sections are described by Johnson (1977a) and Witzke (1981a,b). A reference core section reposited at the Iowa Geological Survey contains the lower and upper contacts of the Scotch Grove Formation. This reference core was drilled near Walford in northern Johnson County (SE SW SW SW sec. 5, T81N, R8W) and was described by Witzke (1981b, p. 514-516). It contains the following sequence of members within the formation: Johns Creek Quarry Member, 14.6 feet (4.5 m); Buck Creek Quarry Member, 58.5 feet (17.8 m); Fawn Creek Member, 56.6 feet (17.3 m); and Waubeek Member 39.2 feet (12 m).

A great diversity of carbonate facies are present within the Scotch Grove, including: 1) flat-lying, dense, cherty to very cherty, sparsely fossiliferous dolomite; 2) flat-lying, dense, non-cherty, sparsely fossiliferous dolomite; 3) flat-lying, porous, skeletal-moldic (especially crinoid-moldic) and dolomite-replaced skeletal (crinoidal) dolomite (wackestone and packstone fabrics); 4) mounded (biohermal) crinoid-moldic and crinoid-replaced dolomite (mostly wackestone and packstone fabrics); and 5) mounded, dense, sparsely fossiliferous dolomite. There is a broad range of overlap and intergradation between some of these lithofacies. Seven general lithofacies, based on lithologic characteristics and general stratigraphic position within the sequence, were recognized by Witzke in the Scotch Grove Formation and were given informal stratigraphic names. Johnson (1983) described three new members in the "upper Hopkinton" that are reassigned herein to the Scotch Grove Formation. In addition, three facies described by Witzke (1981a,b) are designated formally as members in this report (fig. 9).

The Scotch Grove Formation is conformable with underlying Hopkinton and overlying Gower Formation strata. The upper and lower contacts are characteristically sharp. The upper contact may be locally disconformable (Philcox, 1972). Complete core penetrations of the Scotch Grove interval in Johnson County where the Scotch Grove/Gower contact is preserved vary in thickness from 100 to 169 feet (30-52 m). Cored intervals in portions of Linn County, where the Gower rocks are erosionally removed, document thicknesses of the Scotch Grove Formation to 240 feet (73 m). The maximum thickness of the Scotch Grove Formation is not precisely known, although considering the vertical dimensions of carbonate mounds in the upper Scotch Grove (see section on Palisades-Kepler Member) and the minimum thicknesses of up to 240 feet (73 m) in Linn County, the Scotch Grove Formation reasonably can be inferred to reach thicknesses to 300 feet (90 m). The Scotch Grove reaches its greatest thickness in portions of Cedar, eastern Linn, Jackson, Jones, Clinton, and Scott counties, where it ranges from about 150 to 300 feet (45-90 m). The Scotch Grove thins westward and southwestward into Benton and Johnson counties where maximum thicknesses range from about 100 to 170 feet (30-70 m). In Delaware, Benton, and portions of Linn and Johnson counties, Devonian stratigraphic units lie unconformably on an eroded Scotch Grove Formation surface, and the thickness of the formation is correspondingly reduced. The lower one-third or so of the Scotch Grove Formation includes conodont and brachiopod faunas indicative of an Early Silurian (late Llandoverian, C_5-C_6) age. The middle portion of the formation includes conodont (*amorphognathoides* Zone) and graptolite faunas that indicate a position near the Llandoverian-Wenlockian boundary. The upper one-third or so of the formation includes conodont and brachiopod faunas indicative of an early to middle Wenlockian age. The upper boundary of the Scotch Grove Formation is probably middle or late Wenlockian in age (Witzke, 1981b).

<u>Johns Creek Quarry Member</u>. One of four general lithologies are locally noted above the Picture Rock Member in the basal portion of the Scotch Grove Formation: 1) mounds of dense, coral-bearing dolomite; 2) flat-lying, dense, well-bedded, locally argillaceous dolomite; 3) flat-lying, porous, abundantly fossiliferous dolomite; or 4) flat-lying, sparsely fossiliferous, very cherty dolomite. Johnson (1983) included strata of the first two lithologies in a new member, the John's Creek Quarry Member. This member is recognized as the basal unit of the Scotch Grove Formation in this report. Witzke (1981b, 1983b) informally included strata of the first two lithologies in the Castle Grove Mound facies and Johns Creek facies, respectively. Carbonate mounds were first recorded at this stratigraphic position by Philcox (1970a) who noted a series of small mounds 20 to 30 feet (6-9 m) thick spaced 50 to 100 feet (15-30 m) apart at a locality in Jones County. Johnson (1975, 1977a, 1980) also described and illustrated a carbonate mound above the Picture Rock Member in Dubuque County. These mounds are, in part, buried and flanked by crinoidal dolomites (wackestone and packstone fabrics). Flat-lying coral-bearing biostromal beds also occur in the Johns Creek Quarry Member in Jones County (Johnson, 1977a, p. 146; Witzke, 1981b, p. 59).

The Johns Creek Quarry Member at most exposures in east-central Iowa is typified by dense, flat-lying, bedded, unfossiliferous to fossiliferous, extremely fine to microcrystalline, locally cherty dolomite which Johnson (1975) interpreted as an "interreef facies" stratigraphically equivalent to

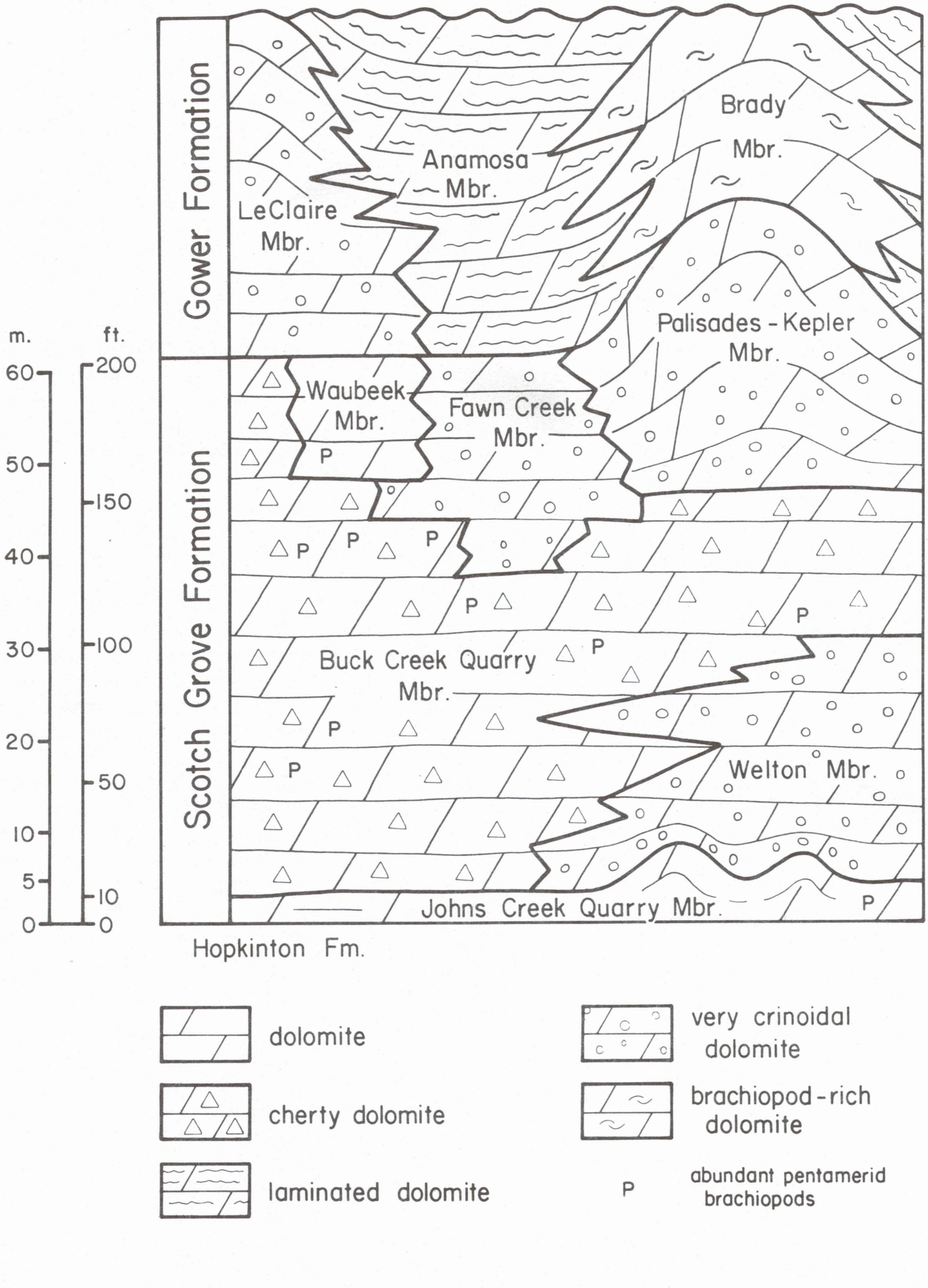


Figure 9. Generalized stratigraphic sequence of the Scotch Grove and Gower formations, Silurian of east-central Iowa.

the carbonate mounds. The basal Scotch Grove "interreef facies," where noted, varies from about 1.5 to 16 feet (0.5-5 m) in thickness in the outcrop belt. Fossils recovered from this facies include tabulate corals, solitary rugose corals, bryozoans, brachiopods (including *Pentameroides* and *Costistricklandia*), trilobites, and echinoderm debris (including *Petalocrinus*).

In the subsurface of Johnson, Linn, and eastern Benton counties the Johns Creek Quarry Member, where present, is a dense, slightly argillaceous, extremely fine to microcrystalline dolomite varying in thickness from about 5 to 15 feet (1.5-4.5 m). It commonly contains horizontal burrows, is generally sparsely fossiliferous, and locally includes argillaceous partings and streaks. Scattered chert nodules occur in some sections. Fossils are generally rare, but brachiopod, trilobite, bryozoan, and small crinoid debris molds are present locally. Lithologies of the Johns Creek Quarry Member in the subsurface in the western extent of the study area closely resemble those in the outcrop belt, although the western facies is notably more argillaceous. A reference core section of the western facies, drilled near Center Point, Linn County (SE NE SE SE sec. 24, T85N, R8W), contains an 11.6 foot (3.5 m) thick section of the member (Witzke, 1981b, p. 492).

Johnson (1983) illustrated possible interfingering of the Picture Rock and Johns Creek Quarry members in Jackson County and suggested that, where the Picture Rock is thickest, it may be a facies of the Johns Creek Quarry Member. He (1983, p. 17) observed that "the Johns Creek Quarry Member pinches out as it approaches the Plum River Fault Zone." While it is true that the Picture Rock Member is thickest in a belt, roughly 15 miles (24 km) wide, north of the Plum River Fault Zone in Jackson and eastern Jones counties, perhaps suggesting structural downwarping in that area during deposition of the member, there is no complementary thickness relationship between the Picture Rock and Johns Creek Quarry members in eastern Iowa. The thinnest known exposure of the Picture Rock Member in Iowa (NW NE sec. 17, T83N, R2E, Clinton Co.) is capped by a thin Johns Creek Quarry Member (1.5 ft; 0.5 m). However, the thickest known exposure of the Picture Rock Member in the state (type locality, SE SW sec. 32, T86N, R2W, Jones Co.) is capped by a thick interval of the Johns Creek Quarry Member. In addition, the Johns Creek Quarry Member is absent locally in areas where the Picture Rock Member is thin (e.g. Cou Falls, Johnson Co.; Covington, Linn Co.; Witzke, 1981b, p. 488, 497). These stratigraphic observations are inconsistent with the suggestion that the upper Picture Rock and Johns Creek Quarry members are facies equivalents. The two members have not been observed to interfinger at any locality, casting further doubt on the suggestion of facies relations.

The Johns Creek Quarry Member is absent locally, and the Welton or Buck Creek Quarry Member of the Scotch Grove Formation locally overlies the Hopkinton Formation. These relations suggest that the three members may be partial facies equivalents. The Johns Creek Quarry Member is cherty at some localities (subsurface of Linn and Johnson counties; Witzke, 1981b, p. 492, 526), where it is largely indistinguishable from lithologies of the Buck Creek Quarry Member. In addition, the Johns Creek Quarry Member is interbedded locally with abundantly fossiliferous dolomites indistinguishable from lithologies of the Welton Member (Witzke, 1981b, p. 59, 542). The carbonate mounds of the Johns Creek Quarry Member are buried by skeletal-rich dolomites of the Welton Member. The inter-mound facies of the Johns Creek Quarry Member is overlain by the Welton or Buck Creek Quarry Member in areas west of Jones County. Although complex facies variations characterize basal Scotch Grove strata in eastern Iowa, the Johns Creek Quarry Member can be distinguished readily as a distinct rock unit at most localities. The member is Early Silurian (Late Llandoverian, C_6) in age.

Welton Member. The Welton Member was named by Johnson (1983, p. 17) for strata above the Johns Creek Quarry Member composed of fossiliferous, "poorly bedded, very finely crystalline dolomite." These strata were termed the "*Cyrtia* Beds" by Johnson (1975), and were informally included in the "Emeline facies" by Witzke (1981b, 1983b). The Welton Member is characterized by horizontally-bedded, porous, abundantly fossiliferous (typically very crinoidal), non-cherty dolomite. Most skeletal debris is preserved as molds, but beds containing dolomite-replaced skeletal debris are also noted (crinoidal wackestone and packstone fabrics). Some beds of sparsely fossiliferous, dense dolomite occur in the member (locally containing accumulations of *Callipentamerus* in Jackson County). An abundant and diverse marine fauna is documented in the member that includes tabulate corals, rugose corals, stromatoporoids, sponge spicules, bryozoans, abundant brachiopods (including *Costistricklandia*, *Cyrtia*, *Dicoelosia*, *Eospirifer*), molluscs, trilobites, and echinoderms (Johnson, 1977a). Articulated cystoids, blastoids, eocrinoids, and crinoid cups are locally abundant (Witzke, 1976; Witzke and Strimple, 1981). The member contains the most diverse fauna known in the Silurian sequence of Iowa (Witzke, 1981b).

The Welton Member is well developed in the Scotch Grove outcrop belt of Jackson, Clinton, Dubuque, and northeast Cedar counties, where it encompasses most of the exposed Scotch Grove sequence. Welton strata locally are interbedded with the Johns Creek Quarry and Buck Creek Quarry members in the subsurface of Jones, Linn, Johnson, and Benton counties. The Welton Member is also recognized in Carroll County, Illinois (included in the "Racine Formation" by Willman, 1973). The maximum thickness of the Welton Member in the outcrop belt is unknown due to erosional truncation, although thicknesses of up to 75 feet (23 m) are exposed in portions of Jackson County.

The Welton Member conformably overlies the Johns Creek Quarry Member, and it locally overlies the Picture Rock Member of the Hopkinton in Jackson County. A coarse crinoidal dolomite facies of the Welton buries the carbonate mounds of the Johns Creek Quarry Member. The Welton Member forms the bedrock surface over large areas of east-central Iowa, obscuring stratigraphic relations with overlying units. As noted by Johnson (1983, p. 17), "a good section showing the upper contact with the overlying member has still not been discovered." However, residual chert geests occur locally above Welton strata in western Dubuque and Jones counties (ibid.), suggesting that the Buck Creek Quarry Member overlies the Welton in that area. However, farther to the east cherty Buck Creek Quarry strata or geests have not been recognized. Grabens along the Plum River Fault Zone in Jackson County, Iowa, and Carroll County, Illinois, locally contain upper Scotch Grove and Gower strata, but cherty Scotch Grove strata have not been recognized. In those areas the Welton Member probably encompasses the entire lower and middle portions of the Scotch Grove Formation, where the member is apparently overlain by the Waubeek and Palisades-Kepler members of the upper Scotch Grove. Drill cores within these grabens are needed to verify the proposed stratigraphic relations in the eastern portion of the study area. The Welton Member ranges in age from late Llandoverian (C_5 or C_6) to early Wenlockian (Witzke, 1981b).

Buck Creek Quarry Member. The Buck Creek Quarry Member was named by Johnson (1983, p. 17) for an interval of "finely crystalline dolomite with numerous white chert horizons." The member encompasses the "*Pentameroides* Beds" of Johnson (1975). The unit was informally termed the "Buck Creek Quarry facies" by Witzke (1981a,b; 1983b). The member is characterized by dense, sparsely fossil-moldic, very cherty dolomite locally containing abundant molds of the large pentamerid brachiopod, *Pentameroides*. Although nodular cherts make up to 35% of the member by volume, it characteristically contains about 5 to 15% chert, and locally contains less than 5% chert. Some beds within the member are chert-free, consisting of dense, sparsely fossiliferous dolomite. Quartz crystal-lined vugs are common in some sections. The Buck Creek Quarry Member is locally argillaceous in the lower portion of the Scotch Grove Formation, where it lithologically resembles the Johns Creek Quarry Member. The Buck Creek Quarry Member is interbedded locally with strata resembling the Welton and Fawn Creek members. The Buck Creek Quarry Member varies greatly in thickness due to complex facies relationships with other members of the Scotch Grove Formation, ranging from about 20 to 167 feet (6-51 m) in thickness in Delaware, western Jones, Linn, Johnson, and eastern Benton counties. It averages about 100 feet (30 m) in thickness in the western extent of the study area.

Johnson (1983) interpreted the Buck Creek Quarry Member to lie in stratigraphic position above the Welton Member, although Witzke (1981b) recognized that the member is locally a lateral facies equivalent of all members of the Scotch Grove Formation. In general, the Welton Member is replaced laterally (westward) by the Buck Creek Quarry Member in western Dubuque and Jones counties. In portions of Jones, Linn, Delaware, Johnson, and Benton counties the Buck Creek Quarry Member encompasses virtually the entire Scotch Grove Formation. In outcrop and in the subsurface, the member is observed to laterally replace upper Scotch Grove strata of the Palisades-Kepler, Waubeek, and Fawn Creek members (ibid.). Although Johnson (1975, 1977a, 1983) largely equated the Buck Creek Quarry Member with the *Pentameroides*-bearing interval of the Iowa Silurian above the Johns Creek Quarry Member, cherty strata of similar lithologic aspect are known to occur above and below *Pentameroides*-bearing beds in the western portion of the study area (Witzke, 1981b). In addition, thick intervals of the Buck Creek Quarry Member are known in Johnson, Linn, and Benton counties that lack *Pentameroides*. The cherty dolomites of the Buck Creek Quarry Member are replaced westward in central Benton County by cherty limestones of the LaPorte City Formation (ibid.).

Pentameroides-bearing beds are locally developed in the Buck Creek Quarry Member in the middle porton of the Scotch Grove Formation and reach thicknesses to 75 feet (23 m). Additional brachiopods (including *Costistricklandia*), corals, crinoids, (especially *Petalocrinus*), and other fossils occur in these beds, and diverse coral faunas are associated locally with *Pentameroides* (Witzke, 1981b). Where the Buck Creek Quarry Member is developed in the lower portion of the Scotch Grove Formation, a sparse fauna of similar taxonomic composition to the Welton Member is noted. Buck Creek Quarry strata in the upper portion of Scotch Grove Formation contain faunas dominated by small solitary rugose corals and several species of brachiopods (ibid.). Johnson (1983, p. 17) suggested that the large pentamerid brachiopod, *Callipentamerus*, is "apparantly restricted to the top horizon of the member." *Callipentamerus* locally occurs associated with *Pentameroides* at localities in Jones and Delaware counties in the middle portion of the Scotch Grove Formation, but uppermost Buck Creek Quarry strata lack these taxa. Johnson (ibid.) assigned a *Callipentamerus*-bearing bed at a locality in Jackson County to the Buck Creek Quarry Member. However, this bed is not cherty and occurs within a sequence, both above and below, displaying lithologies characteristic of the Welton Member; this occurrence of *Callipentamerus* is more consistently included within the Welton Member. The Buck Creek Quarry Member collectively spans the entire Scotch Grove interval and, therefore, ranges in age from late Llandoverian (C_5 or C_6) to middle Wenlockian (Witzke, 1981b).

Fawn Creek Member (new). The Fawn Creek Member is designated formally in this report as a stratigraphic unit in the middle and upper portions of the Scotch Grove Formation. It was described originally as an informal stratigraphic unit, the "Fawn Creek facies," by Witzke (1981a,b; 1983b). The Fawn Creek Member is characterized by horizontally-bedded, porous, abundantly fossiliferous, non-cherty dolomite. The unit is prominently crinoid-moldic (disarticulated debris), and beds containing dolomitized crinoid debris also occur; dolomitic skeletal wackestone and packstone fabrics predominate. Some beds of sparsely fossiliferous dolomite, commonly vuggy, occur in the member. The type locality is designated in Wapsipinicon State Park, where a series of natural exposures along both sides of the Wapsipinicon River upstream from the mouth of Fawn Creek are accessible (N 1/2 SE 1/4 sec. 11, T84N, R4W, Jones Co.). The type locality exposes a 50 foot (15 m) thick section of the member. Although no locality exposes the upper and lower contacts of the member, the type area displays the lateral facies relationships of the Fawn Creek Member. The member is replaced laterally by cherty strata of the Buck Creek Quarry Member immediately east and west of the park boundaries. In addition, the Fawn Creek Member is replaced to the north by the Palisades-Kepler Member; the transition between these two members can be seen along the northwestern edge of the type locality (Witzke, 1981b, p. 67). A reference core section of the Fawn Creek Member from Fairfax (NW NW SW NE sec. 20, T82N, R8W, Linn Co.) is reposited at the Iowa Geological Survey and contains a 62.7 foot (19.1 m) thick interval of the member in stratigraphic position above *Pentameroides*-bearing Buck Creek Quarry strata and below the Waubeek Member.

The Fawn Creek Member lithologically resembles the Welton Member, but it occupies a different stratigraphic position within the Scotch Grove Formation. The Welton Member overlies the Johns Creek Quarry or Picture Rock members, whereas the Fawn Creek Member overlies the Buck Creek Quarry Member. Like the Welton Member, the Fawn Creek shares lateral facies relationships with the Buck Creek Quarry Member. At some subsurface localities in Linn County the basal portion of the Fawn Creek Member includes interbeds of cherty dolomite, illustrating interfingering of the member with the Buck Creek Quarry Member. The base of the Fawn Creek Member does not occur at the identical stratigraphic position within the formation at all localities, resulting from complex facies variations between Buck Creek Quarry and Fawn Creek strata in the middle portion of the Scotch Grove Formation. The Fawn Creek Member is overlain by the Waubeek Member at localities in Linn, Johnson, and Benton counties. At other localities in the same counties the Fawn Creek Member extends higher in the Scotch Grove sequence, where it is apparently a lateral facies equivalent of the Waubeek Member. In those areas, the Fawn Creek is overlain by the Anamosa Member of the Gower Formation. The Fawn Creek Member is overlain by Middle Devonian strata in areas of Johnson and Linn counties. The Fawn Creek Member ranges in thickness up to 97 feet (29.5 m), and is commonly about 60 feet (18 m) thick. Fawn Creek intervals as thin as 5 to 10 feet (1.5-3 m) occur locally as interbedded strata in the Buck Creek Quarry Member.

The Fawn Creek Member is abundantly fossiliferous, although taxonomically indeterminate echinoderm debris molds generally dominate. However, articulated cups of camerate crinoids are locally common. The member is coralline at many localities, and tabulate corals, solitary and colonial rugose corals, and stromatoporoids have been identified. A variety of brachiopods also occur in the member (Witzke, 1981b). *Pentameroides* occurs locally in the basal portion of the Fawn Creek where the member is thickest. *Costistricklandia* is noted at some localities above occurrences of *Pentameroides* in the Buck Creek

Quarry Member; the genus is absent in the upper portion of the Fawn Creek Member. In addition, green algae, bryozoans, molluscs, and trilobites are noted in the member. The stratigraphic position of the Fawn Creek Member in the middle and upper portions of the Scotch Grove Formation indicates an early to middle Wenlockian age for the member (ibid.).

Waubeek Member (new). The Waubeek Member is designated in this report as a formal member in the upper part of the Scotch Grove Formation. It was termed informally the "Waubeek facies" by Witzke (1981a,b; 1983b). The Waubeek Member is characterized by sparsely fossiliferous, dense to vuggy, chert-free, finely crystalline to microcrystalline dolomite. Fossil molds are locally common in some beds. The type locality is located along the Wapsipinicon River 1 mile (1.6 km) northeast of Waubeek in Linn County (SE SW NE sec. 17, T85N, R5W), where a 42 foot (12.8 m) thick section of the member is exposed in a series of roadcuts and natural exposures. The Waubeek Member at the type locality is characteristically chert-free, thick-bedded (1.5 to 3 ft.; 0.5 to 1 m), partly vuggy, and contains quartz-lined vugs in the lower beds. The generally unfossiliferous to sparse skeletal-moldic nature of the member is similar to dolomite fabrics in the Buck Creek Quarry Member, and the two members are distinguished from each other primarily on the presence or absence of chert.

A reference core section from near Walford in Johnson County (SE SW SW SW sec. 5, T81N, R8W), reposited at the Iowa Geological Survey, contains a 39.2 foot (12 m) thick interval of the Waubeek Member in contact with the Anamosa Member of the Gower Formation above and the Fawn Creek Member of the Scotch Grove Formation below. The sharp contact of the Waubeek and Anamosa members is well exposed at Stone City, Jones County (SE SW NE sec. 6, T84N, R4W). The Waubeek Member averages about 40 feet (12 m) in thickness, and locally reaches thicknesses to 56 feet (17 m). The member is thinner where erosionally bevelled beneath Middle Devonian strata. The Waubeek Member is replaced laterally by strata assignable to the Buck Creek Quarry, Fawn Creek, and Palisades-Kepler members (Witzke, 1981b, p. 91-92). The Waubeek Member overlies the Buck Creek Quarry or Fawn Creek members in the western portion of the study area (Linn, Johnson, Benton, western Jones counties). The stratigraphic relations of the Waubeek Member in grabens along the Plum River Fault Zone in the eastern part of the study area (Jackson County) are not apparent, although the absence of the Buck Creek Quarry Member in that area suggests that it may overlie a thick Welton Member.

Although the Waubeek Member is only sparingly fossiliferous, the unit has produced a variety of fossils, and some beds are moderately fossiliferous. Small echinoderm debris molds are scattered (in places abundantly) throughout much of the member, but are absent locally in the upper part. Tabulate and solitary rugose corals are common in some beds. Bryozoan, sponge spicules, and gastropods also occur. Trilobites (especially *Encrinurus*) are present in the lower and middle portions of the member. Brachiopods occur throughout much of the Waubeek Member. Atrypids, spiriferids, gypidulinids, strophomenids, and orthids are noteworthy (Witzke, 1981b). Dense accumulations of the large pentamerid brachiopod, *Rhipidium*, occur within the member along the Plum River Fault Zone in Jackson County (SW NE NE SW sec. 33, T84N, R3E). The uppermost beds of the Waubeek Member contain a unique association of fossils that differs from that in underlying beds. The uppermost beds are typically brachiopod-dominated, commonly including rhynchonellids (*Stegerhynchus*), athyrids, and *Meristina*. These beds locally contain small tabulate and solitary rugose corals, but trilobites, bryozoans, and echinoderm debris characteristically are lacking (Witzke, 1981b, 1983b). The uppermost Waubeek faunas share their closest similarity to faunas in the overlying Gower Formation. The Waubeek Member is probably of early to middle Wenlockian age (ibid.).

Palisades-Kepler Member (new). The most complex rock unit in the upper Scotch Grove Formation is a mounded sequence of very crinoidal skeletal-moldic and skeletal-replaced dolomite with wackestone, packstone, and rare grainstone fabrics. This unit is here named the Palisades-Kepler Member after the Iowa state park by the same name (type locality E 1/2 SW and SW NE sec. 4, T82N, R6W, Linn Co.). Natural exposures extend along the banks of the Cedar River for 1.6 miles (2.6 km) within Palisades-Kepler State Park, and up to 85 feet (26 m) of vertical section is accessible. Witzke (1981a, b; 1983b) informally termed the unit the "Palisades-Kepler Mound facies." The member is characterized by mounded carbonate buildups ("reefs"), including isolated single mounds or coalesced complexes of mounds at various localities. The type locality of the Palisades-Kepler Member is within a coalesced complex of six or more carbonate mounds that extend laterally for about 1.5 miles (2.4 km). Single isolated mounds in the member range from about 200 to 300 feet (60 to 900 m) in lateral extent, and vertical dimensions range from about 30 to 200 feet (9 to 60 m) (Witzke, 1981b). During deposition the carbonate buildups of the Palisades-Kepler Member were expressed as topographic highs on the sea floor. The topographic relief on the upper surface of the Scotch Grove Formation varies locally from a few feet up to 125 feet (38 m), and the thickest Scotch Grove sections are present where the Palisades-Kepler Member is developed. Beds within the mounded facies dip at angles ranging from 0 to 50°, generally in a radial fashion away from the mound centers. However, within mounded complexes the configuration of dipping beds is very complicated. No skeletal framework is generally developed within the mounds, and the mounds became rigid topographic features on the sea floor primarily through submarine cementation processes.

Mounded, skeletal-rich strata of the Palisades-Kepler Member are replaced laterally by horizontally-bedded ("interreef") strata in the upper Scotch Grove Formation, as displayed at several localities in eastern Iowa. At the type locality, the mounded Palisades-Kepler Member can be traced laterally into cherty, horizontally-bedded, sparsely fossiliferous strata of the Buck Creek Quarry Member (Philcox, 1970b; Witzke, 1981a,b). A single mound of the Palisades-Kepler Member near the type locality of the Fawn Creek Member in Jones County can be traced laterally into strata of the Fawn Creek and Buck Creek Quarry members (Witzke, 1981a,b). At other localities in Jones and Clinton counties the Palisades-Kepler Member is replaced laterally by horizontally-bedded strata of the Waubeek Member (ibid.). The base of the Palisades-Kepler Member has not been observed at any locality in Iowa, although subsurface relationships suggest that it overlies the Buck Creek Quarry Member in the central and western portions of the study area in eastern Iowa. Subsurface relationships are unclear in Jackson and Clinton counties, but the member may overlie the Welton Member in that area. Where covered by younger Silurian strata, the Palisades-Kepler Member is sharply overlain by Gower strata of the Anamosa or Brady members. Philcox (1972, p. 704) noted "a possible erosion surface" beneath thin-bedded Gower rocks in Jones County, and he inferrred that the tops of the mounds within the Palisades-Kepler Member may have been subaerially exposed during basal Gower deposition.

All previous workers have included the stratigraphic unit here termed the Palisades-Kepler Member within the "LeClaire Facies" of the Gower Formation.

In this report, the LeClaire is still considered a valid term, but only for Gower deposits in the type area of Scott County. With the exception of Philcox (1970b, 1972) and Henry (1972), all previous workers have incorrectly assumed that rocks now included in the Palisades-Kepler Member (i.e. "LeClaire") are laterally equivalent to laminated dolomites of the Anamosa Member (Gower Fm.). Although in the type area the LeClaire Member is demonstrably laterally equivalent to laminated dolomites, many mounded sequences elsewhere in Iowa have been mistakenly correlated with the LeClaire. The Palisades-Kepler Member is nowhere laterally equivalent to laminated Gower beds or their equivalents. Instead, the Palisades-Kepler Member is equivalent to horizontally-bedded, non-laminated strata in the upper Scotch Grove Formation (Witzke, 1981a).

Echinoderm debris forms the bulk of the fossils observed in the Palisades-Kepler Member, but a diverse assemblage of other fossils are present including green algae, sponge spicules, stromatoporoids, tabulate corals, colonial and solitary rugose corals, bryozoans, brachiopods (including the large pentamerids *Conchidium* and ?*Lissocoelina*), nautiloids, bivalves, gastropods, and trilobites (Witzke, 1981b). Although skeletal debris is an important constituent of the Palisades-Kepler Member, the bulk of the mounds is composed of an extremely fine to coarse crystalline dolomite that forms the matrix of the crinoidal wackestones and packstones. The fauna and stratigraphic position of the Palisades-Kepler Member suggest an early to middle Wenlockian age (Witzke, 1981b).

Gower Formation

The Gower Formation, as originally defined and as utilized in this study, includes laminated dolomites and their lateral facies equivalents (in part, mounded). The type locality of the Gower Formation is at the Cedar Valley (Bealer's) Quarry in Gower Township, Cedar County (SE NE NW sec. 19, T80N, R2W), where 108 feet (33 m) of laminated Gower rocks were exposed during quarry operations (Norton, 1901). As discussed earlier, not all mounded sequences in the Iowa Silurian are laterally equivalent to laminated Gower dolomites, and, therefore, not all mounded features should be included within the Gower Formation. The Gower Formation has been conventionally divided by previous workers into two facies: a laminated Anamosa facies and a mounded or "reefal" LeClaire facies. Most previous workers included the Palisades-Kepler Member of this study within the "LeClaire facies" and incorrectly inferred lateral equivalency of these mounds with laminated Anamosa rocks. This procedure has promulgated serious stratigraphic errors and led to an oversimplification of the complex facies relationships in the Scotch Grove and Gower formations. In this study the LeClaire is given member rank and applied to mounded and flat-lying sequences within the Gower Formation of Scott County. The definition of the LeClaire Member is restricted to include only those rocks that are known to be laterally equivalent to laminated Anamosa sequences. Three general carbonate lithofacies are given member rank in the Gower Formation: Anamosa, Brady, and LeClaire.

The Gower Formation is erosionally truncated beneath Devonian strata in western Linn County and varies from about 0 to 100 feet (30 m) in thickness in eastern Linn and western Jones counties. Gower sections up to 100 to 150 feet (30-45 m) thick in Jones, Cedar, and Scott counties represent the maximum known thickness of the formation. The thickness of the Gower is complementary to that of the Scotch Grove Formation; where the Scotch Grove is thickest (i.e., where the Palisades-Kepler Member is present) the Gower is thinnest, and vice versa. Structurally-preserved blocks of Gower Formation rocks are noted along the Plum River Fault Zone in southern Jackson (Baik, 1980; Chao, 1980) and southern Jones (Saribudak, 1980) counties. The Gower conformably overlies the Scotch Grove Formation at most localities, but the contact may be locally disconformable above the Palisades-Kepler Member. The contact is characteristically sharp, but interbedding of laminated Gower rocks and Waubeek-like strata occurs at some localities in the basal Gower (Witzke, 1981a). Where such interbedding occurs, the base of the Gower Formation is drawn at the base of the lowest laminated dolomite bed. Up to 150 feet (45 m) of relief is developed locally on the Scotch Grove-Gower contact in the vicinity of carbonate mounds of the Palisades-Kepler Member, indicating that the upper Scotch Grove mounds were topographic highs at the onset of Gower deposition (Witzke, 1981b). The Gower Formation is overlain unconformably by Devonian strata of the Bertram, Otis, or Wapsipinicon formations in the western portion of the study area. The Otis Formation locally overlies mounded Gower strata with angular unconformity in portions of Linn and Cedar counties. The Gower Formation is overlain unconformably by Pennsylvanian strata (Caseyville Formation) in portions of Scott County. The exact age of the Gower Formation is not known, although its position above Wenlockian carbonates suggests an upper Wenlockian to Ludlovian age, in part, for the Gower Formation. The Gower brachiopod fauna generally supports such an assignment (Witzke, 1981b).

Anamosa Member. The "Anamosa stage" was named by Calvin (1896, p. 56) for a sequence of laminated dolomite quarrystone and dense dolomite at Stone City, Jones County, Iowa. This rock unit is given formal status as a member of the Gower Formation in this report; the type of locality is designated at Stone City (SW SE NE and NE NE SE sec. 6, T84N, R4W), about 2.5 miles (4 km) west of Anamosa. An 87 foot (26.5 m) thick sequence of the Anamosa Member is exposed at the type locality, where it sharply overlies the Waubeek Member of the Scotch Grove Formation. The member was termed the "Anamosa facies" by previous workers (Hinman, 1968; Philcox, 1970b, 1972; Henry, 1972; Witzke, 1981a, b). The laminated dolomites of the Anamosa Member include both wavy- (or crinkly-) laminated and planar-laminated rock types; the laminae range from less than a millimeter to about 1 cm in thickness. The laminae are usually uninterrupted and can be traced laterally along quarry walls for hundreds of feet. The laminated dolomites are interbedded at some localities with dense, microcrystalline to fine crystalline dolomite (ranging from less than 1 cm to more than 1 m in thickness); these dense dolomite interbeds are termed "flints" by quarrymen. Individual "flint" bands are traceable for distances of up to 2 miles (3.3 km) in the Stone City area (Henry, 1972). Skeletal-moldic and skeletal-replaced dolomites are also present in the Anamosa Member at different localities to varying degrees. At Stone City the Member is largely unfossiliferous, although a few thin bands of brachiopod-moldic and bivalve/ostracode-bearing rocks are noted. At other localities, such as the type Gower section at Cedar Valley (Cedar County), flat-lying beds of skeletal-rich dolomite (skeletal-moldic and skeletal-replaced wackestones and packstones) in excess of 5 feet (1.5 m) thick are noted in the laminated dolomite sequence; these beds contain abundant brachiopods and small tabulate corals. Skeletal-rich beds are most abundant in the Anamosa Member only in the proximity of Brady Member mounds. Additional rock types are also noted in the Anamosa Member at some localities and include: 1) beds of vuggy, sparsely fossiliferous to fossiliferous dolomite, 2) brecciated and intraclastic laminated dolomite, 3) obscurely laminated dolomite, and 4) evaporite-crystal-moldic dolomite. Smooth chert nodules are noted within the Anamosa Member at several localities.

Skeletal fossils are absent within the laminated dolomites of the Anamosa Member at most localities, although "rods" (enigmatic rod-shaped bodies about 1 cm x 2 mm) are locally abundant along bedding surfaces. The "rods" have been variably interpreted as fecal pellets or gelatinous dwelling tubes (Henry, 1972). The preservation of thin continuous laminae in the Anamosa Member indicates that burrowers were generally absent during deposition of the laminated muds. Non-laminated fossil-bearing rocks within the laminated sequence are most frequently characterized by low-diversity brachiopod-rich faunas (*Protathyris*, *Hyattidina*, atrypaceans, rhynchonellids), although corals (small favositids, small cup corals) are sometimes prominent. In addition, bivalves (*Pterinea*), small gastropods, and ostracodes (*Leperditia*) are present. Crinoid debris molds are sparsely represented in non-laminated dolomites of the Anamosa Member at some localities, although the general scarcity or absence of echinoderm debris in the Anamosa Member contrasts markedly with the general abundance of crinoid molds in the underlying beds of the upper Scotch Grove Formation. Trilobites, stromatoporoids, nautiloids, and bryozoans have not been noted in the Anamosa Member.

The Anamosa Member is known to occur in Linn, Jones, Johnson, Cedar, southeastern Clinton, and Scott counties. Laminated Anamosa rocks are preserved in grabens along the Plum River Fault Zone in Jackson County, Iowa, and Carroll County, Illinois. Laminated dolomites reasonably assigned to the Anamosa Member are also recognized in Rock Island County, Illinois. The Anamosa Member is known to overlie the Waubeek, Fawn Creek, Buck Creek Quarry, and Palisades-Kepler members of the upper Scotch Grove Formation. The Anamosa Member reaches thicknesses to 150 feet (45 m) in Scott County. The member is laterally replaced by and locally interfingers with the Brady and LeClaire members of the Gower Formation.

Brady Member. Philcox (1970b) introduced the term "Brady facies" for the Brady (McGuire) Quarry in Cedar County (SE SE sec. 14, T80N, R3W). The rock unit is accorded member status, the Brady Member, in the Gower Formation in this report. The Brady Member is characterized by mounded sequences of fossil-moldic and fossil-replaced dolomites (wackestone, packstone, and boundstone fabrics) with abundant brachiopods and/or rugose corals (*Fletcheria*). Dense non-laminated to laminated dolomites, in part with prominent domal stromatolites, are interbedded with the fossil-rich beds. General rock types present in the Brady Member resemble those of the Anamosa Member, although the Brady is markedly more skeletal-rich. The Brady Member is laterally equivalent to flat-lying laminated dolomites of the Anamosa Member and is known to overlie upper Scotch Grove Formation strata, including the Palisades-Kepler Member and its inter-mound equivalents. Beds generally dip radially outward from the central area of the Brady mounds, although smaller "satellite" mounds locally alter this general picture. Dips average between about 10 and 50°, but S-shaped "slump folds" in the Brady Member locally achieve dips up to 90° (Hinman, 1963; Smith, 1967; Philcox, 1972).

At Palisades-Kepler State Park (Linn Co.) the Brady Member occurs in stratigraphic position above the crinoidal mounds of the Palisades-Kepler Member, and Philcox (1970b) defined the base of the "Brady facies" at the lowest brachiopod-rich bed (which is laterally equivalent to the basal laminated Anamosa dolomites). "The contact appears to represent a time-surface . . . there is no evidence for interfingering between the Brady and crinoidcoelenterate facies [i.e., Palisades-Kepler Member] . . . the Brady facies at the Palisades characteristically takes the form of large-scale wedge-beds with dips up to about 45° which thin and flatten out down dip" (Philcox, 1970b, p. 176). These exposures, and others in Linn, Cedar, and Jones counties, demonstrate that the mounded Brady Member is, in part, developed on the tops and flanks of the older crinoid-rich carbonate buildups of the Palisades-Kepler Member. Additionally, the Brady Member carbonate mounds probably acted as loci for the distribution of skeletal debris that spread laterally away from the mounded areas and interfingered with the inter-mound laminated carbonates of the Anamosa Member. The lateral equivalence of the Brady and Anamosa members is clearly displayed at several localities in eastern Iowa, and if individual bedding surfaces are traced along exposures displaying these relations, the Brady Member is noted to be two to more than five times greater in thickness than the equivalent beds in the Anamosa Member (Witzke, 1981a). This observation suggests that the Brady Member mounds were developed as topographic highs above the surrounding inter-mound area where laminated carbonate mudstone deposition was taking place.

The Brady Member represents a second stage of carbonate mound building above and around the older buildups of the Palisades-Kepler Member. In addition, co-extensive Brady mounds expanded laterally away from the site of older Scotch Grove buildups into areas previously occupied by upper Scotch Grove Formation inter-mound facies. The Brady Member is interpreted to both bury and spread laterally away from the older Palisades-Kepler mound developments (Witzke, 1981a). The fossil fauna of the Brady Member is primarily characterized by an abundance of brachiopods, and literally millions of brachiopods are exposed at some Brady Member outcrops. Most brachiopods are small (less than 1 cm). Large numbers of a few species characterize most Brady Member brachiopod faunas (*Protathyris*, *Hyattidina*, rhynchonellids, cyrtiids, ?*Fardenia*). Preservation of delicate brachiopod spiralia and spondylia is noted in some brachiopod-rich beds of the Brady Member. An additional brachiopod association is noted only in the central portion of the Brady Member mounds which includes *Atrypa*, *Leptaena*, ?*Sphaerirhynchia*, orthids, strophomenids, *Trimerella*, and *Harpidium* (a large pentamerid reaching lengths to 10 cm). Clusters of rugose corals (*Fletcheria*) are prominent in many Brady Member mounds, and the rugose corals can form boundstones of cemented "reef" rock (remnant fibrous cement fabrics are noted, probably originally submarine aragonite cements). A small favositid tabulate coral is locally prominent in portions of some Brady Member mounds, and small zaphrentid cup corals, *Favosites*, and *Halysites* are also noted. Additionally, ostracodes, small gastropods, bivalves (*Pterinea*), and rare nautiloids are locally present. Domal stromatolites are developed in the Brady Member at some localities, and green algae (*Ischadites*) are also noted. While echinoderm debris has been noted in the Brady Member at some localities, it is characteristically rare to absent. Trilobites, bryozoans, and stromatoporoids are absent in the Brady Member.

The Brady Member is restricted geographically to areas of Jones, Cedar, and Linn counties. Brady lithologies have also been noted in grabens along the Plum River Fault Zone in Jackson County. The member represents a discontinuous mounded, brachiopod-rich carbonate facies surrounded by laminated dolomites of the Anamosa Member. Strata assigned to the LeClaire Member of the Gower Formation occupy the same stratigraphic position in Scott County, Iowa, and adjacent parts of Illinois.

LeClaire Member. From the town of LeClaire, Scott County, Iowa, Hall (*in* Hall and Whitney, 1858) described a horizontally-bedded and "folded" sequence of hard dolomite, in part with abundant brachiopod and rugose coral molds, as the "limestone of LeClaire rapids." Worthen (1862, p. 47) noted that the "LeClaire Limestone" is intercalated with laminated rocks (i.e. Anamosa Member) in the area of LeClaire, Iowa. In general, subsequent studies included the non-laminated, fossiliferous strata at LeClaire in the "LeClaire facies" of the Gower Formation. The LeClaire is defined as a member of the Gower in this report. The LeClaire Member includes both mounded and flat-lying dolomite sequences in Scott County. Although no specific type locality has been designated, previously described exposures at the town of LeClaire presumably would include the type locality. However, many of the localities described by Hall and others in the vicinity of LeClaire are no longer accessible, and the best Silurian exposure presently available in town (quarry NW NW sec. 35, T79N, R5E) does not expose the LeClaire Member but is developed in the laminated Anamosa Member.

The primary reference section of the LeClaire Member is designated along Bud Creek 3 miles (5 km) north of LeClaire (SE SW sec. 11, T79N, R5E), where natural exposures display up to 45 feet (14 m) of the member. The horizontally-bedded sequence includes: 1) dense, unfossiliferous dolomite, 2) sparse to abundant brachiopod-moldic dolomite, 3) tabulate and rugose coral-moldic rock, and 4) dolomite with sparse crinoid debris molds. An additional reference section, 40 feet (12 m) thick, is accessible at Princeton, Scott County (SW SW SE sec. 35, T80N, R5E). The horizontally-bedded sequence at this locality includes several types of dense, hard dolomite: 1) horn coral-rich (*Fletcheria*) beds, 2) brachiopod-rich beds, and 3) tabulate coral-rich (favositid) beds. A thin zone (4 inches; 10 cm) of laminated dolomite of Anamosa aspect is interbedded with the brachiopod-bearing beds.

The LeClaire Member in the type area of eastern Scott County, Iowa, includes both mounded and flat-lying sequences of dense, crystalline dolomite varying from unfossiliferous to very fossiliferous. Like the Brady Member, small brachiopod and horn coral (*Fletcheria*) molds are prominent in many exposures of the LeClaire Member. Additionally, at several Scott County exposures the LeClaire Member is developed as flat-lying beds, unlike the Brady Member which is only known to occur as a mounded facies (which interfingers with flat-lying Anamosa beds). The LeClaire Member is interbedded with laminated Anamosa rocks at several Scott County exposures, and, therefore, the member can be logically included within the Gower Formation. Although the LeClaire Member resembles the Brady Member in several respects, important differences serve to differentiate the two units. Some features are noted only in the Brady Member but not in the LeClaire Member: 1) great profusion of abundant brachiopod-moldic and -replaced dolomites with wackestone and packstone textures, 2) large domal stromatolites, and 3) high-angle dips and "slump folds." Features noted only in the LeClaire Member that are not observed in typical Brady Member exposures include: 1) abundant crinoid-moldic wackestones, 2) nautiloid-rich "pockets" and beds, and 3) presence of trilobites. The presence of crinoidal debris in the LeClaire Member (both mounded and flat-lying sequences) is especially noteworthy, since the Brady Member is characterized by the absence or scarcity of echinoderm remains.

Many studies greatly modified and expanded the original definition of the LeClaire in Iowa to include rock types and strata not exposed in the type area of eastern Scott County. Older strata, presently assigned to the Scotch Grove Formation, were included in the "LeClaire facies" by Rowser (1929), Smith (1967), Hinman (1968), and Philcox (1972). In this report the LeClaire Member in Iowa is recognized only in the outcrop belt of Scott County. Strata elsewhere in the state that were assigned to the "LeClaire facies" by earlier workers are now included in the Palisades-Kepler Member of the Scotch Grove Formation and the Brady Member of the Gower Formation. The presence of crinoidal wackestone textures, tabulate coral-bearing beds, and nautiloid-rich "pockets" in the LeClaire Member of Scott County is reminiscent of the Palisades-Kepler Member, and, in this regard, it is understandable how the term "LeClaire" was extended by many previous workers to include strata now assigned to the Palisades-Kepler Member. The most important stratigraphic distinction between these two members is the lateral equivalence of the LeClaire Member and the laminated Anamosa Member, whereas the Palisades-Kepler Member is laterally equivalent to non-laminated strata in the upper Scotch Grove Formation. The full thickness of the LeClaire Member in Scott County is not known with certainty, but its lateral stratigraphic relationship with the Anamosa Member suggests comparable thickness to 150 feet (45 m).

The LeClaire Member in Scott County includes a variety of marine invertebrate fossils in both flat-lying and mounded sequences. Echinoderm-rich strata contain an abundance of disarticulated crinoid debris as well as camerate and inadunate crinoid cups (Witzke, 1981b). Rhynchonellid and atrypacean brachiopods and trilobites are associated with the crinoidal strata. Coralline strata, including tabulate (*Halysites*, favositids, syringoporids) or rugose (*Fletcheria*) corals, are developed locally. An assemblage characterized by large brachiopods (*Conchidium, Trimerella, Dinobolus*) with gastropods, bivalves (*"Megalomus"*), and corals also occurs in the member. Accumulations of abundant nautiloids are present locally within the LeClaire mounds. The diversity of fossils in the LeClaire Member contrasts with the Brady and Anamosa members and indicates that the LeClaire was deposited in more open-marine environments than the other two members of the Gower (Witzke, 1983b).

Savage (1926) named the dolomite unit across the river from LeClaire at Port Byron, Illinois, the "Port Byron Limestone." Savage subsequently "decided that these beds are the same as the LeClaire of Iowa and has suggested that the name Port Byron be abandoned in favor of the older LeClaire" (Sutton, 1935, p. 274). However, Sutton (1935, p. 274, 277) used the term Port Byron for uppermost Silurian strata in northwestern Illinois, and restricted the term LeClaire to the "'reef' phase of the Iowa Gower," noting that the LeClaire "reefs" merge "into the Anamosa type of sediment in almost horizontal beds." Strata equivalent to the LeClaire Member in northwestern Illinois (Rock Island and Henry counties) were included in the Racine Formation by Willman (1973). Carbonate mounds in Rock Island County, Illinois, locally interfinger with laminated dolomite, in a manner identical to that observed in the Iowa Gower. It is suggested in this report that the terms LeClaire and Anamosa members have utility in northwestern Illinois, where the "Racine Formation" remains poorly defined. The LeClaire Member is also recognized in grabens along the Plum River Fault Zone in Carroll County, Illinois.

Devonian System

Middle and Upper Devonian rocks are widely distributed throughout Iowa. Exposures of Middle Devonian rocks along the Plum River Fault Zone are limited to small structurally preserved outliers down-dropped within the fault zone. Upper Devonian rocks have been erosionally removed across extreme eastern and northeastern Iowa, but stratigraphic leaks occurring within Middle Devonian and Silurian carbonates are recorded across the area. Devonian rocks are unconformable to and regionally truncate older Paleozoic rock units (Collinson et al., 1967, p. 948), recording a major epeirogenic tectonic episode in the midcontinent region (Sloss, 1963).

Bertram Formation

The Bertram was originally defined by Norton (1895b, p. 135) and assigned to the Silurian. In 1920 Norton "provisionally" reassigned the Bertram to the Wapsipinicon Formation based upon its texture and extent of brecciation. Differences of opinion have existed since that time as to the exact age relationships of the Bertram to overlying Wapsipinicon and underlying Silurian strata. The lack of fossils has hampered the ability to establish time stratigraphic relationships, necessitating the use of such criteria as textural and lithologic similarities, and stratigraphic relationships to the overlying and underlying units. Redefinition and clarification of the underlying Silurian rock units has shown the Bertram to be unconformable to this system, variably overlying either the Scotch Grove or Gower formations in its area of occurrence. The upper beds of the Bertram are extensively fractured, and the upper contact of the Bertram below the Coggon marks a sharp lithologic break (Sammis, 1978, p. 30), suggesting a possible unconformable relationship between the two rock units. Sammis (1978, p. 196) interpreted the Bertram as having been formed in "a very restricted nearshore to terrestrial environment with rapidly fluctuating conditions," suggesting that the Bertram may represent in part a paleocaliche, or soil horizon, which accreted in a topographic depression developed on the Silurian surface. Subsurface mapping of the thickness and distribution of the Bertram shows that it occupies a small elliptical-shaped area in east-central Iowa. Comparison of the mapped area of the Bertram to the area of occurrence of the other overlying members of the Wapsipinicon Formation (Church, 1967) shows it to have a very limited geographic distribution.

In light of the preceding discussion in which the Bertram: 1) was originally assigned to the Silurian and then "provisionally" assigned to the Devonian; 2) apparently is unconformably bounded above and below; and 3) is of limited geographic extent in comparison to overlying and underlying units, the Iowa Geological Survey is elevating the Bertram from member status in the Wapsipinicon Formation to formational rank.

No specific type section was designated for the Bertram. The type region was originally considered as the area along Big Creek in southeastern Linn County (Norton, 1895b, p. 135-136) in the vicinity of the town of Bertram (fig. 6) extending to Paralta. At least part of Nortron's original reference area is still exposed and accessible. A single quarry section (SE NE sec. 34, T83N, R6W) exposing 14.6 feet (4.5 m) was described by Sammis (1978), and can be used as a surface reference section for the Bertram. A reference core section (NW NW SW NE sec. 20, T82N, R8W) containing the Bertram Formation (17 ft; 5.2 m) is reposited at the Iowa Geological Survey and was described by Sammis (1978).

The Bertram Formation is a brown to gray, unfossiliferous, vuggy, usually laminated, sometimes intraclastic, fine-grained to sublithographic dolomite. Rounded quartz sand grains and fine pyrite crystals are disseminated throughout; sandy gray-green shales occur locally near the base. The Bertram ranges from 0 to greater than 75 feet (23 m) in thickness in east-central Iowa.

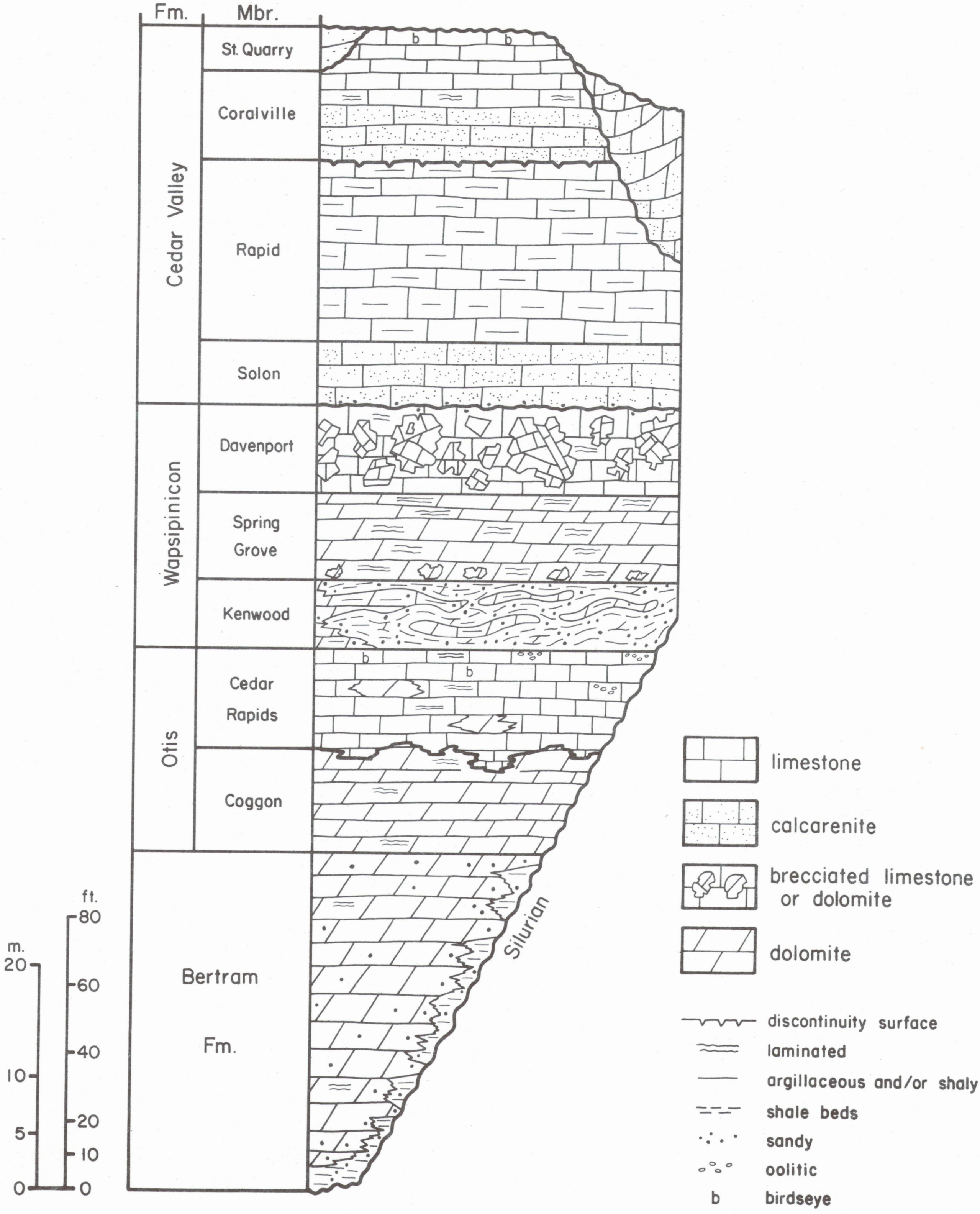


Figure 10. Generalized stratigraphic sequence of the Middle Devonian strata in east-central Iowa.

Otis Formation

Norton (1894) first recognized the existence of a Devonian unit subjacent to the Kenwood beds in Linn County, based upon the occurrence of the brachiopod *"Spirifer subumbonus"* Hall (i.e. *Emanuella* sp.), and termed this unit the Otis. He (1895b) later assigned these rocks to the Wapsipinicon Formation. The underlying Coggon at that time was "provisionally" assigned along with the Bertram to the Silurian. The later identification of *"Spirifer subumbonus"* in the Coggon (Norton, 1901, p. 320) necessitated the reassignment of the Coggon to the Devonian (ibid., p. 321). The similarity of the brachiopods found in the Coggon to those found in the Otis suggested to Norton (1901, p. 323) the possible uniting of these two members into a single unit. In 1920, Norton subdivided the Otis into four "phases" --Cedar Rapids, Coggon, Westfield, and Vinton --recognizing the Coggon as the lowermost "phase." Except for Stainbrook's (1935) reassignment of the Coggon to member status below the Otis, subsequent workers have eschewed these other subdivisions.

Sammis (1978) noted that the Otis and Coggon members are similar "in original overall composition with most present-day differences probably generated by diagenetic effects in the Otis." Sammis (1978, p. 199) further noted that "the Coggon in both core and outcrop . . . is mineralogically and texturally homogeneous," whereas the Otis ". . . exhibits significant variability in extent and nature of authigenic calcite and silica, thus giving rise to macroscopic variations that may have prompted initial delineation of separate members by previous workers." The contact between the Otis and Coggon is also a transitional interval which has led to arbitrary selection of the contact in the subsurface, and the utilization of the thicker-bedded nature of the Coggon in outcrop. Subsurface mapping (Church, 1967) also showed distinct similarities in the geographic distribution of the Otis and Coggon, and suggested similarities in the overall depositional framework (Sammis, 1978, p. 199).

It is apparent from the historical perspective of interpretations of the Otis and Coggon interval, and the latest petrographic work by Sammis (1978), that Norton's 1920 classification of the Coggon as the lower "phase" of the Otis was probably a correct interpretation, and as such the Coggon is now included as part of the Otis.

The Otis is regionally restricted to east-central Iowa and adjacent northwestern Illinois. Prior to the 1980s, Otis or Otis-equivalent rocks had not been identified in northeastern Iowa. Original correlations of the basal Devonian strata in northeastern Iowa were based primarily on supposed biostratigraphic similarities to the east-central Iowa Cedar Valley Limestone sections (Calvin, 1903, 1906). These correlations allowed for an interpretation that suggested that Otis or Otis equivalent rocks were absent in northern Iowa by regional overlap of the upper part of the Wapsipinicon Formation in Fayette County (Scobey, 1940, Stainbrook, 1944), which in turn was overlapped by the Cedar Valley Limestone (Stainbrook, 1935). However, recent lithostratigraphic and conodont biostratigraphic studies (Klapper and Barrick, 1983; Bunker et al., 1983; Witzke and Bunker, 1984) of the Devonian rocks across northern Iowa, has shown that the basal Devonian strata in this area do not correlate with any part of the Cedar Valley Limestone, but instead apparently correlate with the Otis. Based upon these new lithostratigraphic and biostratigraphic studies, a reevaluation of the stratigraphic classification of the Otis appears to be in order.

The basal Devonian rocks in northeastern Iowa have been formally designated the Spillville Formation (Klapper and Barrick, 1983), and shown to be physically separated from the Otis in east-central Iowa by a northeastward trending pre-Devonian Silurian escarpment (Bunker et al., 1983). Lithostratigraphic relationships also show the Spillville and Otis to be directly overlain with an abrupt lithologic break, and regionally overlapped by the Kenwood Member of the Wapsipinicon Formation (Bunker et al., 1983). Because the Otis is a distinct, mappable rock unit restricted to east-central Iowa, and is apparently the stratigraphic equivalent of the Spillville Formation of northeastern Iowa, it is elevated to formational rank in this report.

Norton's (1920) four "phase" classification of the Otis has not been utilized by subsequent workers. Since the Otis is now elevated to formational rank, and the Coggon is included as a distinct part of the Otis, it is appropriate to consider the Coggon as the lower member of the Otis Formation (fig. 10). The other three "phase" subdivisions that Norton (1920) defined were either not utilized or discarded in later discussions of the Otis. The "Vinton phase" was discarded in 1944 by Stainbrook when it was shown that this interval represented fossiliferous Cedar Valley strata. The "Westfield phase" was never utilized and is now known to be equivalent to the Kenwood Member of the Wapsipinicon Formation, which has priority. The "Cedar Rapids phase," as originally proposed, replaced the originally defined Otis interval (Norton, 1895b, 1901), and when Stainbrook (1935) removed the Coggon from the Otis, the "Cedar Rapids" was dropped in favor of Otis. Based upon the elevation of the Otis to formational rank with the Coggon as its lower member, the Cedar Rapids is now considered the upper member of the Otis Formation (fig. 10), replacing the previously defined "Otis Member."

Norton (1895b) designated several quarries in the vicinity of Otis, a station to the southeast of Cedar Rapids on the Chicago and Northwestern Railroad, as the type locality, but never formally designated a type section for the Otis. However, later workers (Stainbrook, 1935; Church, 1967) have referred to the first section listed (Norton, 1895b, p. 138) as the type section (SW NW NW sec. 31, T83N, R6W). This section is still partially exposed and was used by Sammis (1978) as an "Otis Member" surface reference section. Exposures of the contact with the overlying Kenwood are common within the type area, but the underlying contact with the Bertram Formation is not exposed. A reference core section reposited at the Iowa Geological Survey contains the lower and upper contacts of the Otis Formation. This reference core was drilled by IGS near Fairfax, Iowa (NW NW SW NE sec. 20, T82N, R8W), and was described by Sammis (1978). It contains both members of the Otis Formation --Coggon (20.1 ft.; 6.1 m) and Cedar Rapids (23.2 ft.; 7.2 m) --within the described interval.

Lithologically, the Otis consists of thin to medium-bedded, interbedded limestone and dolomite in the upper part (Cedar Rapids Member), to predominantly thick-bedded dolomite in the lower half (Coggon Member). The Otis ranges in thickness from 0 to 50 feet (15.2 m) across east-central Iowa. However, a thicker section may occur at the Pleasant Hill Outlier in southern Jones County, but this hypothesis needs further testing.

Scattered throughout the Otis are external molds of a small brachiopod of the genus *Emanuella* (formerly referred to as "*Spirifer subumbonus*"). Gastropods, bryozoans, spirobid worm tubes (Scobey, 1938, p. 32; Sammis, 1978, p. 104), rare corals, and trilobites *(Dechenella)* have also been noted. Conodonts have been recovered from the Coggon Member, which represents an impoverished fauna, consisting only of *Ozarkodina raaschi* (Klapper and Barrick, 1983); this species is well represented in the Spillville Formation of northern Iowa (ibid.). Conodonts recovered from the Spillville Formation allow for a zonal assignment of the Spillville to the *ensensis* and *kockelianus* Zones (ibid.) of the standard conodont zonation of the Middle Devonian. The

Eifelian-Givetian stage boundary is placed within the *ensensis* Zone, thus indicating a late Eifelian to early Givetian age for the Spillville Formation. Probable equivalency of the Otis and Spillville Formations based upon the earlier discussion allows for a tentative age assignment of the Otis Formation to the Late Eifelian-early Givetian.

Coggon Member. Norton (1895b, p. 147) first referred to the section exposed at Ashby's Quarry near the Illinois Central Railroad Station at Coggon as illustrative of the character of the Coggon beds. However, when Norton (1901) formally designated the Coggon as a member of the Wapsipinicon Formation he made no further reference to Ashby's Quarry as a type section. In 1920, Norton designated Ashby's Quarry as typical of the Coggon "phase." The Ashby Quarry section is no longer accessible, but the Coggon is exposed in the nearby Betenbender Quarry (SW NE SW sec. 3, T86N, R6W) and is utilized as a primary reference section.

Lithologically the Coggon Member consists of slightly laminated, porous, ferroan, microcrystalline, light buff to brown dolomite. It is strikingly homogeneous throughout the region, varying megascopically in extent and type of porosity, amount of iron, and degree of lamination. The contact with the overlying Cedar Rapids Member is transitional being marked in the Otis reference core section by a change from ferroan, slightly laminated dolomite to slightly calcitic, locally non-ferroan dolomite. In outcrop the contact with the Cedar Rapids Member is likewise transitional, but unlike the reference core, a contact is somewhat easier to delineate due to differences in observed bedding and weathering characteristics. Molds of the spiriferid brachiopod *Emanuella* sp. are locally abundant.

Cedar Rapids Member. In 1920, when Norton first proposed the term Cedar Rapids, no type section was designated as had been done with the Coggon. It is clear from Norton's (1920) discussion, however, that he was referring to those beds previously restricted to the Otis superjacent to the Coggon, and exposed in the quarries of the type area of the Otis to the southeast of Cedar Rapids. In lieu of the utilization of the Cedar Rapids terminology in substitution for the previously defined "Otis Member," the type section of the Cedar Rapids Member is considered to be the same as the type section of the Otis Formation (SW NW NW sec. 31, T83N, R6W). This section is still partially exposed. The reference core of the Otis also contains the member and was described by Sammis (1978; under the heading Otis Member).

Lithologically, the Cedar Rapids Member consists of fine-grained to sublithographic, locally laminated, stylolitic, thin-to medium-bedded, interbedded limestone and dolomite. Varying degrees of calcitization and silicification occur throughout the member. Birdseye structures are noted locally, occurring near the top of the unit. Pelletal (Sammis, 1978) and oolitic (Dille, 1928) limestones have also been noted. The Cedar Rapids Member ranges in thickness from 0 to 30 feet (9.1 m) across east-central Iowa. It is a fossiliferous unit of low diversity; the brachiopod *Emanuella* sp. is common throughout the member.

Wapsipinicon Formation

The Wapsipinicon Formation was named by Norton (1895b) for exposures along the Wapsipinicon River in northeastern Linn County, Iowa. According to the original definition, the Wapsipinicon consisted of four members which included, in ascending order: the Otis, Kenwood (including the Independence Shale), Lower Davenport, and the Upper Davenport. Later workers (Norton, 1920; Stainbrook, 1935; Scobey, 1938, 1940; and Church, 1967) modified the terminology and recognized up to six members, which in ascending order were: the Bertram, Coggon, Otis, Kenwood, Spring Grove, and Davenport. Based upon the preceding discussions and reclassifications of the Bertram, Coggon, and Otis, the Wapsipinicon Formation is now restricted to three members, which include, in ascending order: the Kenwood, Spring Grove, and Davenport.

The Wapsipinicon Formation has long been recognized as one of the most enigmatic rock units of the Iowa stratigraphic column. The stratigraphic position of the Wapsipinicon Formation and its contained members has been a matter of great confusion because of the poorly to unfossiliferous nature of the section, repeating lithologies, extensive brecciation, and the regionally persistent occurrence of Upper Devonian stratigraphic leaks within and adjacent to the unit. Sammis (1978) provided an exhaustive petrographic analysis of the members of the Wapsipinicon Formation, and described a core (NW NW SW SE sec. 20, T82N, R8W) drilled by IGS, which is used as a subsurface reference section.

Kenwood Member. Norton (1895b, p. 156) originally proposed the term Kenwood for shaley beds immediately overlying the Otis. Norton (1985b) utilized the Kenwood as a local synonym for non-fossiliferous shales he believed were equivalent to the Independence Shale. In 1920, Norton dropped the term Kenwood in favor of Independence. However, Stainbrook (1935, p. 251) revived the term Kenwood, because he believed from faunal and lithologic evidence that the Independence occupied a higher stratigraphic position than the Kenwood. Later studies indicated that the Independence represents a stratigraphic leak of Upper Devonian shales into the underlying Middle Devonian carbonates.

The type section of the Kenwood (fig. 6) is in the right bank of Indian Creek in northeastern Cedar Rapids, Linn County (NW SW NW NE sec. 14, T83N, R7W). As originally described, Norton (1895b) included within the Kenwood at this exposure the Spring Grove, a unit later given member status by Stainbrook (1935).

The Kenwood is the most lithologically variable of all the members of the Wapsipinicon Formation. Sammis (1978, p. 108) described the Kenwood as "a sequence of bluish-gray, locally somewhat folded or disturbed, sandy, ferroan, argillaceous, unfossiliferous limestones and dolomites with intercalated calcitic and/or dolomitic sandy shales." In the subsurface of southeastern and central Iowa, the Kenwood locally contains abundant gypsum and anhydrite. Across east-central Iowa the Kenwood ranges in thickness from 0 to 20 feet (6.1 m), and lies with an abrupt lithologic break over the Otis carbonates.

Spring Grove Member. The Spring Grove was named by Stainbrook (1935, p. 251) for exposures of distinct laminated dolomite in Spring Grove Township (NW SW sec. 24, T86N, R7W) along the bank of the Wapsipinicon River in northern Linn County, Iowa (fig. 6). Norton (1895b) originally considered that these beds belonged to the upper portion of the Kenwood.

The Spring Grove is predominantly a porous, laminated, nonferroan, fine- to medium-crystalline dolomite that is tan to rust in color. In east-central Iowa it ranges in thickness from 20 to 25 feet (6.1-7.6 m). The member is essentially unfossiliferous. However, Church (1967, p. 72) noted ostracodes in this interval from extreme eastern and northeastern Iowa, and stromatolites have been reported in northeastern Iowa (Bunker et al., 1983). The laminations are probably the most striking megascopic characteristic of the Spring

Grove, and the entire interval is laminated to some degree. The laminations are horizontal to slightly crenulated, dark brown in color and probably organic in nature. When freshly broken the Spring Grove characteristically exhibits a strong petroliferous odor.

Davenport Member. Norton (1895b) named an Upper and Lower Davenport for exposures in the vicinity of Davenport, Scott County, Iowa, and assigned them both to the Wapsipinicon Formation. Stainbrook (1935, p. 252) rejected the term Upper Davenport, and assigned these beds to the Cedar Valley Limestone, because of their close faunal relationships. He further simplified the name Lower Davenport to Davenport, and retained these beds as the uppermost member of the Wapsipinicon Formation.

No specific type section was ever designated for the Davenport, other than a type area in the vicinity of Davenport, Iowa. A surface reference section at the Linwood Quarry, near Buffalo, Iowa, in Scott County, was described and petrographically examined by Sammis (1978).

The Davenport is a light gray to dark brown sublithographic to lithographic limestone ranging in thickness from 20 to 25 feet (6.1-7.6 m) across eastern Iowa. It is generally unfossiliferous; although, algal stromalite-like structures have been noted in the Cedar County, Iowa, and Rock Island, Illinois areas (Don Koch, Iowa Geological Survey, pers. comm.). It is thin-to medium-bedded containing numerous stylolites and commonly is highly brecciated. The breccias ("Fayette Breccia," Norton, 1895b, p. 157) are composed of angular chips and blocks of lithographic limestone in a matrix of sandy, argillaceous limestone or dolomite. In his treatise on the "Wapsipinicon Breccias of Iowa," Norton (1920, p. 355-547) provided detailed descriptions of these breccias which crop out in east-central Iowa. The origin of the brecciation found in the Wapsipinicon has long been a controversial subject. The intimate association of gypsum-anhydrite with the Wapsipinicon in the subsurface (Dorheim and Campbell, 1958; Church, 1967; Sendlein, 1964, 1968, 1972) has led to the suggestion that evaporite solution-collapse was largely responsible for the widespread brecciation.

The Davenport is disconformably overlain by the Solon Member of the Cedar Valley Limestone. The contact is difficult to identify with any consistency where the Davenport has been brecciated, because of the stratigraphic leakage of Solon materials downward into the brecciated interval. Davenport breccia fragments and/or blocks are surrounded by a matrix of basal Solon sand ("Hoing Sandstone") and/or carbonate mud, intraclasts, and fossils in the upper part of the member. Generally, the contact is chosen at the highest occurrence of Davenport clasts within the brecciated interval.

Cedar Valley Formation

The Cedar Valley Formation generally has been considered a widespread fossiliferous Middle Devonian rock unit, which records a major marine transgression (Taghanic Onlap) into the Upper Mississippi Valley region. However, this assumption has never been proven within a lithostratigraphic and/or biostratigraphic framework outside of the Johnson County, Iowa area, especially where this interval has been encountered in the subsurface of central Iowa. Three members have been defined in the Cedar Valley sequence of Johnson County: the Solon, Rapid, and Coralville. Recent studies of Devonian rocks in northern Iowa (Klapper and Barrick, 1983; Bunker et al., 1983; Witzke and Bunker, 1984), and central-northwestern Iowa (Klug, 1982; Tynan, 1982, unpublished IGS manuscript) call into question the usage of the term Cedar Valley, especially the present member subdivision, in those parts of the state. The recognition of a series of repetitive shallowing-upward marine-carbonate cycles in northern Iowa (Witzke and Bunker, 1984) indicates that greater caution should be utilized in the application of Solon, Rapid, and Coralville terminology outside of Johnson County, Iowa, until further bio- and lithostratigraphic studies are completed. Consideration of these problems is presently beyond the scope of this paper.

The Cedar Valley Formation was originally named for the series of exposures of Devonian carbonate rocks along the valley of the Cedar River, although no type locality was designated (McGee, 1891). As originally defined, it included the Devonian sequence above the Silurian unconformity and below the "Hackberry Shale" (Lime Creek Formation). Subsequent recognition of the Wapsipinicon and Shell Rock Formations restricted the Cedar Valley Formation to a position between those two formations. Although no specific type section was ever designated for the Cedar Valley Formation, a surface reference section (Conklin Quarry, secs. 32 and 33, T80N, R6W) has been described (*in* Bunker and Hallberg, 1984) in the Johnson County, Iowa area. This reference section has been used by many authors in the past as representative of the Cedar Valley Formation in east-central Iowa.

Previous usage by the Iowa Geological Survey recognized the Cedar Valley Formation as consisting of three members. These members were originally defined in terms of faunal zones established by Stainbrook over a period of 20 years (1938, 1942, 1943, 1945a, b). The lithostratigraphic investigations by Kettenbrink (1973) further defined the same member boundaries. These members in ascending order are: the Solon, Rapid, and Coralville, all named for exposures near Iowa City, Johnson County, Iowa.

Solon Member. The Solon was originally named by Keyes (1912), and Stainbrook (1941) later designated a type section (secs. 23 and 24, T81N, R6W; now covered) in the vicinity of Solon, Johnson County, Iowa (fig. 6). Lithologically the Solon consists predominantly of dense, thick-bedded, fine-grained skeletal calcarenites, which vary in color from yellowish olive gray to very dark yellow brown. The Solon ranges in thickness from 6 feet (1.9 m) near Davenport, Scott County, Iowa, to 19 feet (5.8 m) at Conklin Quarry, Johnson County Iowa, to approximately 35 feet (10.7 m) at Brook's Quarry (NW 1/4, sec. 2, T88N, R9W), Buchanan County, Iowa. The base of the Solon is characterized by thin discontinuous transgressive sandstones and/or arenaceous carbonates, which often extend downward into the brecciated interval of the upper part of the Wapsipinicon. These thin sandstones and/or sandy carbonates are genetically related to the Hoing Sandstone, which occurs at or near the base of the Cedar Valley Formation in west-central Illinois (Hinds, 1914, p. 12; Howard, 1961).

The Solon has been subdivided biostratigraphically by Stainbrook (1941) into two macrofaunal "zones." Brachiopods are dominant in the lower portion of the Solon and this interval is referred to as the "'*Atrypa*' *(Desquamatia) independensis* zone." The upper Solon is characterized by colonial and solitary rugose corals that are abundant enough to locally form biostromes and rare bioherms (Mitchell, 1977). This interval is referred to as the "*profunda* zone" (*Hexagonaria*, *Asterobillingsa*, *Cystiphylloides* and tabular stromatoporids are common), which derives its name from the colonial rugose coral *Hexagonaria profunda*. In Benton, Buchanan, and northern Linn counties shell accumulations comprised of the terebratulid *Rensselandia* occur in the upper part of the "*profunda* zone." At most localities this brachiopod is not abundant, but local accumulations of disarticulated and broken valves result in a coquina 1 to 5 feet (0.3 -1.5 m) thick (Kettenbrink, 1973).

The Solon is assigned to the Middle Devonian-Givetian (Klapper, 1975a; Klapper and Johnson, 1980, Tables 10, 11; Bunker and Klapper, 1984). The upper Middle and Upper *varcus* Subzones are identified within the Solon (except for the uppermost beds) chiefly on the presence of the nominal species, *Icriodus brevis*, *Polygnathus ovantinodosus*, and *Ozarkodina semialternans*. The uppermost beds are assigned to the Lower *hermanni-cristatus* Subzone.

Rapid Member. The Rapid was named by Stainbrook (1941) for exposures north of Iowa City (fig. 6) along Rapid Creek, Johnson County, Iowa. Stainbrook designated an exposure near the mouth of Rapid Creek (S 1/2 sec. 27, T79N, R6W) as the type section.

The Rapid has generally been considered the most widespread and lithologically homogeneous member of the Cedar Valley Formation. It consists predominantly of light bluish gray, barren to extremely fossiliferous, argillaceous, medium bedded calcilutites with thin shaley intervals in the lower part. Local concentrations of white chert nodules and glauconite occur in the upper part. The Rapid Member appears to average approximately 50 feet (15.2 m) in thickness across east-central Iowa.

Following Stainbrook (1941), the Rapid has been subdivided into three macrofaunal "zones," which in ascending order include: 1) the "*bellula* zone," dominated by atrypid brachiopods, but *Strophodonta*, *Elita*, and *Spinocyrtia* commonly occur; 2) the "*Pentamerella* zone," which contains two thin calcarenitic coral-stromatoporoid biostromes (Zawistowski, 1971) that average from 2 to 5 feet (.6-1.5 m) in thickness and occur above the middle of the member in the Johnson County area; and 3) the "*waterlooensis* zone," characterized by atrypid brachiopods, with occurrences of *Strophodonta* and *Schizophoria* common.

The Rapid (below the biostromes of Zawistowski, 1971) has been assigned to the upper Middle Devonian-Givetain *hermanni-cristatus* zone (Klapper, 1975a; Klapper and Johnson, 1980, Table 12; Bunker and Klapper, 1984). Klapper and Johnson (1980) formally subdivided the *hermanni-cristatus* Zone into Upper and Lower Subzones. The lower boundary of the Lower Subzone is defined by the first occurrence of *Schmidtognathus hermanni*, and the lower boundary of the Upper Subzone by the first occurrence of *Polygnathus cristatus*. Characteristic species in the Cedar Valley that first occur in the Lower Subzone are *Schmidtognathus wittekindti* and *Polygnathus limitaris;* (Klapper and Ziegler, 1967; Ziegler et al., 1976); *Schmidtognathus peracutus* first occurs in the Upper Subzone. This faunal association is regarded as diagnostic of the zonal assignment indicated, notwithstanding that neither name bearer has yet been found in the Cedar Valley Formation of Iowa.

The upper part of the Rapid (above the biostromes) and the lower part of the Coralville in the Johnson County, Iowa, area has yielded a conodont fauna dominated by *Icridodus subterminus* and *Polygnathus xylus* (Klapper, 1975a, p. 8). This interval in Johnson County may correlate with some part of the *disparilis* Zone, however, diagnostic species of the *disparilis* Zone, which is developed in an offshore conodont biofacies, have not been noted in association with the nearshore *I. subterminus* Fauna (Bunker and Klapper, 1984).

Coralville Member. The Coralville Member of the Cedar Valley Formation was named by Keyes (1912) for exposures near the town of Coralville (fig. 6), Johnson County, Iowa. Stainbrook (1941) designated the River Products Conklin Quarry (N 1/2 sec. 33, T80N, R6W) as the type section.

The Coralville in Johnson County can be divided into two lithologically distinct rock units. The lower unit consists of brown, massively bedded, coral- or stromatoporoid-rich, skeletal calcarenite, which ranges in thickness from 16 to 23 feet (4.9-7 m). Stainbrook (1941) subdivided this interval into two macrofaunal "zones," which in ascending order are: 1) the coral-rich "*Cranaena* zone"; and 2) the "*Idiostroma* zone," a stromatoporoid-rich interval which is locally developed into a biostromal unit. A sharp, burrowed discontinuity surface marks the boundary between the Rapid calcilutites and the lower Coralville calcarenites in Johnson County, and the "*waterlooensis* zone" of the upper Rapid extends into the lower basal portion of the Coralville.

The upper Coralville in Johnson County is characterized by laminated and pelleted calcilutites and was designated the "*Straparollus* zone" by Stainbrook (1941). Kettenbrink (1973) described the upper Coralville fauna as one of low diversity with diversity and numbers decreasing upwards. He recognized five distinct lithologic subdivisions of the upper Coralville at its type section, which in ascending order are: 1) "gastropod-oncolite calcilutite"; 2) "laminated 'lithographic' limestone"; 3) "*Amphipora*-rich pelleted calcilutite"; 4) "intraclastic calcilutite to calcirudite"; and 5) "'birdseye' calcilutite." Above the "*Amphipora* beds" the Coralville is essentially devoid of macrofossils; the only biota occurring are ostracodes, algae, and foraminifera. Maximum observed thickness of the Coralville in Johnson County is 45 feet (13.7 m).

The Coralville in Johnson County has been assigned to the uppermost Middle Devonian-Givetian (Klapper, 1975a; Bunker and Klapper, 1984). The lower part of the Coralville has yielded a conodont fauna dominated by *Icriodus subterminus* and *Polygnathus xylus*, and as commented upon previously in the Rapid Member discussion, may correlate with some part of the *disparilis* Zone. However, the remainder of the Coralville has proven barren.

State Quarry Limestone Member. The State Quarry Limestone has not been observed along the Plum River Fault Zone, but is exposed in portions of Johnson County. These beds are geographically restricted within the county, and form a linear northeast to southwest trending pattern. The State Quarry Limestone occupies erosional channels that cut across Coralville and Rapid strata. The skeletal calcarenites and calcilutites of the State Quarry represent a channel-filling sequence, orginally interpreted to be of post-Cedar Valley age. However, Watson (1974) suggested that the State Quarry Limestone grades laterally and intertongues with strata in the middle and upper Coralville Member. In Watson's (1974) interpretation, the State Quarry Limestone represents a marine tidal channel facies of the Coralville Member. Watson's critical exposures that purportedly show lateral gradation of Coralville and State Quarry strata are, in our opinion, equivocal, and determination of Coralville/State Quarry relationships requires additional study. At most exposures the Coralville/State Quarry contact clearly is unconformable.

The State Quarry Limestone is dominated by skeletal calcarenites; packstones and grainstones contain varying proportions of echinoderm, brachiopod, branching stromatoporoid, and coral grains and intraclasts. Pelletal calcarenites and stromatoporoidal calcirudites also occur. Fish teeth and plates are common in the calcarenites in the lower and marginal parts of the State Quarry channels ('bone bed'). Grain size generally decreases upward in the sequence and from channel-center to channel-margin (Watson, 1974). Skeletal calcilutites are interbedded with calcarenites near the channel margins. The State Quarry Limestone reaches thicknesses to about 40 feet (12 m). The old State Quarry northeast of North Liberty, Johnson County, is designated the type locality (SW SW NW SE sec. 8, T80N, R6W).

The State Quarry Limestone contains a conodont fauna characterized by *Pandorinellina insita* and species of *Polygnathus* and *Icriodus*. The *insita* Fauna encompasses the interval containing *P. insita* below the first occurrence of *Ancyrodella rotundiloba*, and is an apparent equivalent of the Lowermost *asymmetricus* Zone of latest Givetian (Middle Devonian) age (Bunker and Klapper, 1984). The occurrences of the *insita* Fauna in the State Quarry Limestone post-dates the "*subterminus* Fauna" in the upper Rapid and Coralville members of the Cedar Valley Formation in Johnson County (ibid.). The State Quarry Limestone correlates with strata included in the Cedar Valley Formation in central and northern Iowa (Witzke and Bunker, 1984) and in the Davenport area of eastern Iowa that contain the *insita* Fauna. The *insita* Fauna in central and northern Iowa occurs within the third transgressive-regressive cycle (cycle C) of the Cedar Valley Formation (ibid.); the possibility that the State Quarry Limestone was deposited in erosional channels during the marine transgression of this cycle and post-dates the Coralville regressive phase deserves additional study. Regardless, since the State Quarry Limestone correlates to a portion of the Cedar Valley Formation elsewhere in Iowa, it seems reasonable that the State Quarry should be included as a part of the Cedar Valley Formation. Watson's (1974) interpretation of possible Coralville/State Quarry facies relationships, although regarded as equivocal in this report, further suggests that the State Quarry Limestone should be included as a part of the Cedar Valley Formation. Although the State Quarry Limestone was assigned formational rank in some previous studies, its relationship to Cedar Valley strata elsewhere in the state dictates that it be included as a unit within the Cedar Valley Formation. The State Quarry Limestone is recognized in this report as a member of the Cedar Valley Formation. It is geographically restricted to portions of Johnson County, where it occurs as a marine channel facies incised into the Rapid and Coralville members.

Upper Devonian/Mississippian

Present day occurrences of Upper Devonian rocks along or near the Plum River Fault Zone are limited to small isolated stratigraphic leaks in karst-derived void or fracture fills. Multiple generation karstification and reoccupation by later stage infilling of younger age sediments has been observed at several locations.

The Independence Shale was one of the most famous stratigraphic problems in the midcontinent region during the first half of the 20th century. Controversies regarding its stratigraphic position have existed ever since Calvin (1878) first reported a dark shale below the Devonian (Cedar Valley) Limestone in a quarry at Independence (fig. 6), Buchanan County, Iowa. Urban (1972, p. 48) and Klapper (1975b, p. 8) have summarized the historical controversy surrounding the Independence Shale problem.

The Independence is now considered to be a stratigraphic leak occupying sinkholes and caverns developed during post-Cedar Valley to pre-Lime Creek regional erosion and karstification. Klapper (1975b, p. 9) noted that exposures identified as the Independence Shale display a jumbled mixture of two different shales: a grayish-green clay shale and a dark gray fissile shale. The grayish-green shale has yielded delicately preserved Upper Devonian brachiopods and conodonts of marine origin. The unnamed dark gray fissile shale has yielded Late Mississippian (Chesterian) age spores (Urban, 1971, 1972) of continental derivation. (R. Ravn, 1983, personal communication, suggested that the contained spore assemblage may be of Morrowan age.) The comingling of Upper Devonian marine sediments with Mississippian or Lower Pennsylvanian continental sediments has led to the interpretation of multiple-generation karstification and stratigraphic leaks in the Cedar Valley and Wapsipinicon formations (Urban, 1972, p. 50; Klapper, 1975b, p. 9).

Similar Paleozoic stratigraphic leaks have also been identified in the Silurian of east-central Iowa by numerous geologists. A brief summary of some of these occurrences follow:

1) Davis (1963, p. 50) noted in a quarry in the northeastern corner of Cedar County (center sec. 4, T. 81N., R. 1W.) a gray stratified shale occurring in a pocket within the Silurian dolomite. Several conodonts, one identified as "*Icriodus expansus*" (probably Upper Devonian), were recovered.

2) Smith (1967, p. 31) noted a green clay shale filling fractures, pores, and cavities in a Silurian "reef core" in southern Cedar County (Brady Quarry). The green clay is mainly illite. Devonian conodonts (*Icriodus*) and also Pennsylvanian pollen and spores have been recovered, indicating multiple-generations of stratigraphic leakage.

3) Dorheim (1968, p. 142) described a stratigraphic leak of grayish-green Upper Devonian shale near the Bertram-Silurian contact in eastern Linn County. Upper Devonian conodonts (*Palmatolepis sp.* and *P. schindewolfe* Mueller) have been recovered from this shale.

4) C. B. Rexroad (Indiana Geol. Surv., pers. comm., 1977) noted the occurrence of a green shale irregularly present in a Silurian "reef core" in southern Jones County (Wyoming Quarry). Upper Devonian conodonts were recovered, including several species of *Palmatolepis*.

Pennsylvanian System

Outliers of Pennsylvanian sandstones, shales, and conglomerates are scattered along the trend of the Plum River Fault Zone and across portions of east-central Iowa (see Appendix I). These Pennsylvanian outliers directly overlie Upper Ordovician, Silurian, and Middle Devonian rock units. Sandstone is the dominant lithology at most outliers. Petrographic analyses of the sandstones, characteristically very fine to medium grained, remain to be completed. However, two general compositional varieties of sandstone appear to be present at different localities: 1) quartzarenite, and 2) feldspathic sandstones (sublitharenite to subarkose) containing varying amounts of feldspar, mica (muscovite), and lithic grains. Cross-bedding has been noted in the Pennsylvanian sandstones at some localities (Norton, 1895a; Savage, 1906, p. 626; Udden, 1905, p. 405; Witzke and Kay, 1984; Witzke, 1984). The sandstones vary from friable to cemented (silica, iron oxide, and calcite cements); they are locally conglomeratic containing clasts of quartz, Precambrian crystalline rocks, coalified wood, and Silurian chert, dolomite, and silicified fossils. Green, gray, and carbonaceous black shales (variably plastic to fissile) and siltstones are present at some outliers. Thin coal seams have been noted at Pennsylvanian outliers in Linn, Johnson, and Clinton counties (Norton, 1895b, p. 196; Calvin, 1897, p. 82; Udden, 1905, p. 406). Most Pennsylvanian outliers of east-central Iowa are apparently remnants of channel-filling fluvial deposits. The quartz clasts and crystalline rock fragments noted at some outliers were probably derived from igneous/metamorphic terranes to the north in Wisconsin or Minnesota. In addition to the Pennsylvanian channel sandstones of east-central Iowa, fracture- and cavern-filling sequences of "clays, silts, and sands" (Udden, 1905, p. 406) have been noted within exposures of Silurian and Devonian rocks that are, in part, Pennsylvanian in age.

Channel- and cavern-filling sandstone/shale sequences in Jones, Johnson, Linn, Cedar, Clinton, and Jackson counties were assigned to the Middle Pennsylvanian Desmoinesian Series by some previous workers (Calvin, 1896, 1897; Norton, 1901; Udden, 1905; Savage, 1906) based on rather limited fossil evidence. Norton (1895a, p. 131) found *Lepidodendron*, a "calamite," and a "trigonocarp-like nut" in a Pennsylvanian sandstone outlier in Jackson County. A shale outlier in Linn County yielded a variety of Pennsylvanian fern fossils that Norton (1895a, p. 127) correlated with the Lower Coal Measures of Illinois (apparently Atokan or Desmoinesian). He also noted a gastropod and spirorbid worm tube at the same locality, and, therefore, Pennsylvanian marine or marginal marine deposits may be represented in some outliers. Calvin (1897, p. 83) reported specimens of *Lepidodendron* and *Calamites* from an outlier in Johnson County, and Udden (1905, p. 406) noted that cavern-filling carbonaceous shales in Clinton County "often contain leaves of Carboniferous ferns." Channel-filling sandstones and mudstones along the Iowa River in Johnson County have yielded specimens of ferns *(Neuropteris)*, lycopsids *(Sigillaria, Lepidodendron, Lepidostrobus)*, seeds *(Trigonocarpon)*, *Cordaites*, and *Calamites* (Adams, 1926, Witzke and Kay, 1984; Witzke, 1984). A cavern-filling sandstone at Robins, Linn County, yielded a diverse assemblage of Pennsylvanian plant fossils, including a variety of fern and fern-like taxa as well as *Calamites*, *Lepidodendron*, and *Cordaites* (Wilson and Cross, 1939). Smith (1967, p. 31) reported Pennsylvanian spores from a clay fill in a Silurian dolomite quarry in Cedar County. Sandstones near the Nolting Farm section along the Plum River Fault Zone in Jackson County have produced plant macro-fossils, including *Calamites*, that "strongly suggest a Pennsylvanian age" (Ludvigson and Bunker, 1978, p. 44).

Pennsylvanian exposures in Muscatine and Scott counties include a sequence of fluvial sandstones, conglomerates, siltstones, shales, and coal assigned to the Caseyville Formation and fluvial sandstones assigned to the "Spoon" Formation (Fitzgerald, 1977; Ravn et al., 1984). The Caseyville of Scott-Muscatine counties rests on an eroded surface of Silurian and Middle Devonian carbonates. The Caseyville Formation of Scott-Muscatine counties includes plant megafossils and spores of Morrowan (Early Pennsylvanian) age (ibid.), and the overlying sandstones included in the "Spoon" Formation may be Desmoinesian (Middle Pennsylvanian) in age. The Caseyville sandstones are mature quartzarenites, whereas the "Spoon" sandstones are immature feldspathic litharenites (Fitzgerald, 1977). It seems reasonable to suggest that the Pennsylvanian outliers of east-central Iowa are closely related to the Caseyville-Spoon fluvial deposits, especially since the Scott-Muscatine County exposures lie only a few miles south of outliers in Clinton, Cedar, and northern Scott counties and about 35 miles (55 km) south of those along the Plum River Fault Zone. This suggestion is partly corroborated by the identification of Morrowan (i.e. Caseyville) spores (R. Ravn, 1980, pers. comm.) in a cavern-filling sandstone-carbonaceous shale sequence within the Middle

Devonian sequence at the Atalissa limestone quarry (sec. 18, T78N, R2W) in northern Muscatine County. Carbonaceous shales from a karst filling in the Middle Devonian Wapsipinicon Formation at Hiawatha (NE NW NW NW sec. 4, T83N, R7W), Linn County, also have yielded Morrowan spores (R. Baker, 1983, pers. comm.). By analogy with the sandstones of Scott-Muscatine counties, compositional differences between various Pennsylvanian sandstones from outliers in eastern Iowa may prove potentially useful for contrasting Lower and Middle Pennsylvanian strata. Lower Pennsylvanian sandstones of the Caseyville Formation are compositionally mature and are dominated by quartzarenites. By contrast, the overlying Middle Pennsylvanian "Spoon" sandstones are more immature and include feldspathic litharenites and subarkoses. Middle Pennsylvanian sandstones in Iowa are characteristically micaceous to varying degrees. Further petrographic and paleontologic investigations may permit recognition of two general suites across eastern Iowa. The Pennsylvanian outliers along the Plum River Fault Zone are therefore tentatively included in the Caseyville and "Spoon" formations of Early and Middle Pennsylvanian age.

Quaternary System

Pleistocene and Holocene deposits cover most of the area around the Plum River Fault Zone. These deposits consist of Illinoian and older tills, glaciofluvial deposits as young as Wisconsinan, eolian silts and sands, and Holocene alluvium and colluvium. The Illinoian tills occur in the extreme eastern part of the state, and are bounded to the west by the Goose Lake Channel in southeastern Clinton County (George Hallberg, 1978, pers. comm.).

The pre-Illinoian Pleistocene stratigraphy of eastern Iowa has been defined and described by Hallberg (1980). This interval is divided into two formations with multiple tills, paleosols, and other sediments. The Alburnett Formation and the younger Wolf Creek Formation are primarily distinguished by clay mineralogy, with the former generally containing less than 45% expandable clays, and the latter containing greater than 60%. Hallberg (1980) showed that both of these formations were deposited over immense areas, and that their present distribution is primarily controlled by erosional truncation by the modern land surface.

The relationship between the Quaternary deposits and the Plum River Fault Zone is best known along the Goose Lake Channel in southeastern Jackson County, Iowa. A former drainageway of the Mississippi River, the channel crosses the fault zone perpendicularly with dramatic topographic effect. The channel morphology, and thus the distribution of channel-filling sediments, have been controlled by the relative erosive rates of bedrock formations on opposite sides of the fault zone (Ludvigson et al., 1978, p. 44-45). Areal studies of the Quaternary deposits in and around the Goose Lake Channel by Updegraff (1981) indicate that Illinoian deposits do not impinge on the Plum River Fault Zone in Iowa. Drillhole data from Updegraff (1981) indicate that tills of the Wolf Creek Formation are the youngest present along the Goose Lake Channel in Jackson County, Iowa. Radiocarbon ages and regional landscape relations bracket the loess-mantled terrace surface of the Goose Lake Channel between 17,000 to 20,000 R.C.Y.B.P. (Updegraff, 1981, p. 42-53).

STRUCTURAL GEOLOGY

The gross structural pattern in the area of the Plum River Fault Zone is revealed by structure contouring (fig. 11). The ability to map near-surface geologic units in the area of interest has been acquired only recently, with new refinements in the understanding of the Silurian stratigraphy of eastern Iowa (Johnson, 1975; Witzke, 1976, 1978, 1981a and b). The top of the Silurian Blanding Formation was chosen as a datum because that surface is widely distributed along the fault zone and can most easily be recognized in the subsurface. The data used in constructing figure 11 were obtained from logs of municipal and private wells on file at the Iowa Geological Survey, logs from IGS-USGS research test drilling, and Blanding outcrops in the eastern part of the area.

Kolata and Buschbach (1976) provided the first description of the structural geology of the Plum River Fault Zone and adjacent areas. They defined the fault zone as a "narrow zone of fractures that is less than half a mile wide" (Kolata and Buschbach, 1976, p. 9). Block faulting within the zone was inferred from outcrop distribution (ibid., p. 9), and from well data in Savanna, Illinois (ibid., p. 14). Silurian carbonate rock units within the fault zone along the Plum River in western Carroll County, Illinois were described as being highly brecciated, to the extent that bedding is no longer distinguishable (ibid., p. 9). They also reported the presence of stylolitic striations within these breccias, features that have also been recorded by Chao (1980, p. 41).

Ludvigson et al. (1978, p. 1) applied the term "cataclastic rocks" to refer to the brecciated carbonate rocks along the Plum River Fault Zone. This term was applied in the sense of Higgins (1971, p. 2) in referring to "rock deformation accomplished by fracture and rotation of mineral grains or aggregates; granulation." This textural variation was recognized in the Silurian rocks along the fault zone in Jackson County, Iowa, by Savage (1906, p. 612), who referred to a "granular phase of the Niagaran" which he interpreted as stratigraphic in origin.

Following a broadened usage implied by Ludvigson et al. (1978, p. 1, p. 34, p. 41-42, p. 46-47) subsequent workers (Chao, 1980, p. 27-32; Baik, 1980, p. 129-150) have used the term to refer to a variety of unusual textural and compositional alterations that occur in carbonate strata along the Plum River Fault Zone. The brittle mechanical behavior of carbonate rocks during faulting in the shallow subsurface, coupled with their chemical reactivity make them particularly susceptible to alteration within the fault zone. Thus the term "cataclastic rocks" has been used to refer to cemented breccias (Baik, 1980, p. 138), non-cemented breccias (Ludvigson et al. 1978, p. 41), quartz-replaced dolomite breccias (ibid., p. 34; Baik, 1980, p. 132), and most recently, recrystallized sacchararoidal dolomites (Baik, 1980, p. 137; Chao, 1980, p. 38-41).

The most extensive petrographic descriptions of cataclastic fabrics in carbonate rocks of the Plum River Fault Zone to date are provided by Baik (1980, p. 129-150). His investigations indicated that many of these rocks had experienced complex diagenetic histories and suggested that some may have originated from penecontemporaneous deformation during Silurian deposition (ibid., p. 139).

Preliminary field investigations in an area of former lead mine prospecting (Heyl et al., 1959, p. 296; after Owen, 1844, p. 91) along the Plum River Fault Zone in Jackson County (section 20, T.84N., R.5E.), suggested that

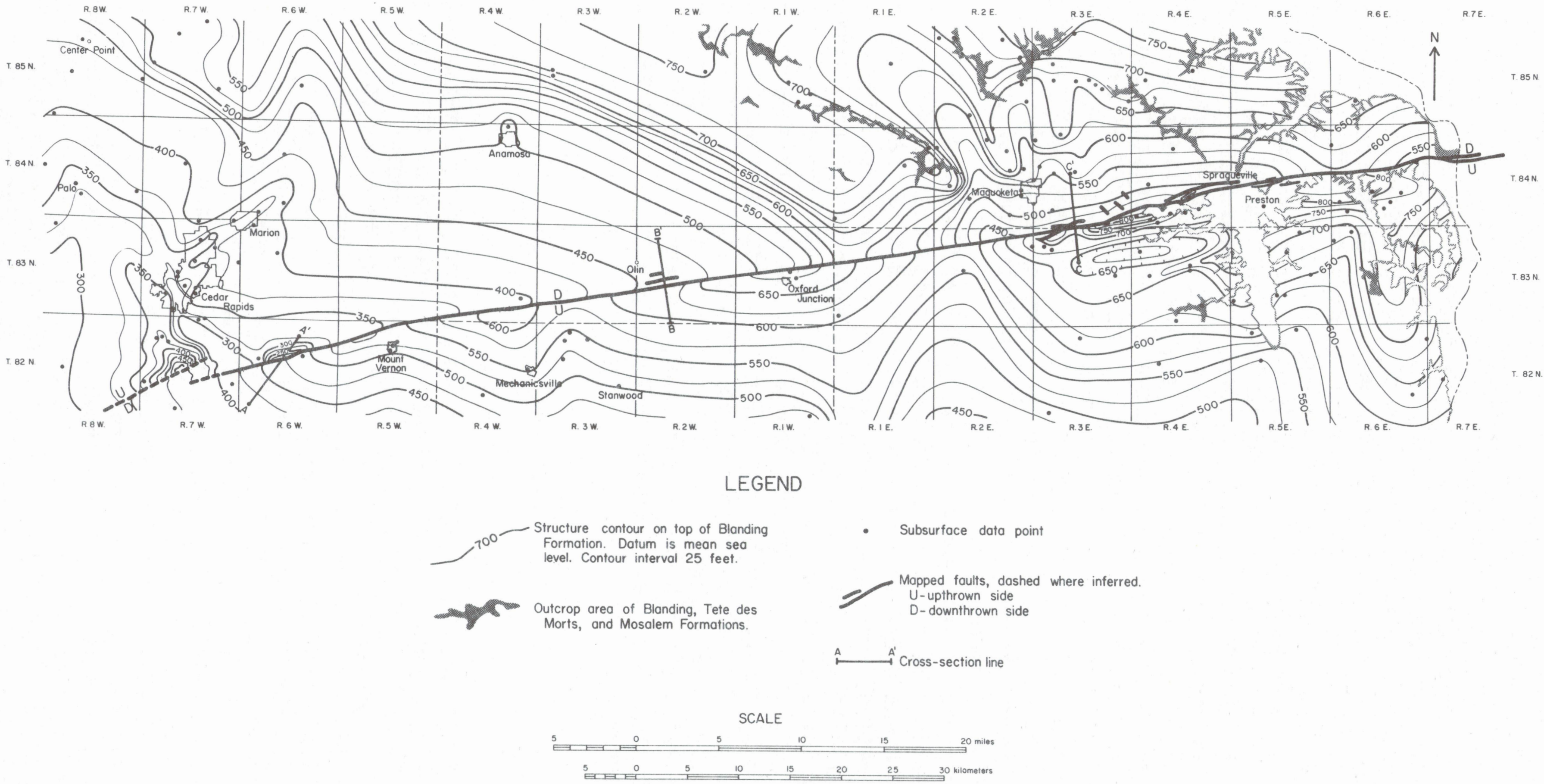


Figure 11. Structure contour map on top of the Blanding Formation (Lower Silurian of east-central Iowa, with lines of cross-sections from figures 15, 18, and 22).

fault-controlled epigenetic sulfide mineralization has occurred within and adjacent to zones of cataclastic rocks in strata currently assigned to the Scotch Grove Formation (Bunker and Ludvigson, 1977; Ludvigson et al., 1978, p. 46-47). Petrographic examination of altered Scotch Grove Formation rocks from this location has supported this conclusion (Ludvigson, 1980). More recently, Garvin (1982, 1984) has recognized structurally controlled epigenetic sulfide mineralization in Silurian strata at the Cedar Rapids Martin-Marietta Quarry, which borders the Plum River Fault Zone in southern Linn County.

Adjacent to the fault zone in Illinois, Kolata and Buschbach (1976) recognized the existence of an east-west trending synclinal axis bordering the fault on the northern, downthrown side (Upton's Cave Syncline), and anticlinal structures bordering the fault on the southern side (Forreston and Brookville Domes, Leaf River Anticline). A similar structural pattern is evident in Iowa (fig. 11). Although the precise configuration of the structures cannot be outlined from the existing data density, it is apparent that the Plum River Fault Zone is bordered on the south by a group of anticlinal structures collectively referred to as the Savanna-Sabula Anticlinal System, and on the north by synclinal structures.

Since the pioneer field mapping work by Dow and Mettler (1962), a number of geologic field mapping projects have been completed along the Plum River Fault Zone in areas where bedrock is well exposed (fig. 12). These studies have been of critical importance in recognizing the internal structure of the Plum River Fault Zone. The field mapping was aided by the acquisition of low altitude color infrared imagery (May 1977; RF 1:24,000; on file at IGS), the completion of U.S.G.S. 7 1/2 minute topographic quadrangle mapping along the fault zone, and by the regional compilation of bedrock-derived soils from published soils maps. Areas of bedrock exposure and limits of completed mapping projects are shown in figure 12. The internal structural geometry of the fault zone is revealed with the least ambiguity where rocks of the Devonian and Silurian systems are well exposed in close proximity. Three such occurrences are described here:

Silver Creek Devonian Outlier

Located in section 33 of T84N, R3E in Jackson County, this outlier was first described by Dorheim (1953), who noted an exposure of the Devonian Wapsipinicon and Cedar Valley formations. These Middle Devonian carbonate strata are structurally preserved in a narrow graben some 40 miles (65 km) east of their present erosional edge. Bunker and Ludvigson discovered deformed cataclastic Silurian rocks immediately bounding the south side of the outlier during reconnaissance mapping of the Plum River Fault Zone in 1977. This relationship led Wahl et al. (1978, p. 35) to depict the outlier as occurring on the northern, downthrown side of the Plum River Fault Zone. The possibility of structural preservation of these Devonian rocks in a graben fault block was suggested by Ludvigson et al. (1978, p. 42).

The immediate area around the Silver Creek Outlier was mapped in detail during the early fall of 1979. The results of this mapping, integrated with auxiliary data from Baik (1980), are shown in figure 13. This mapping clearly shows that the Devonian outlier is enclosed in an east-west trending graben, and is itself cut by at least one smaller fault. Dorheim (1953) noted the narrow range of elevations of several different stratigraphic units in adjacent exposures, and suggested that the Middle Devonian carbonate sequence is

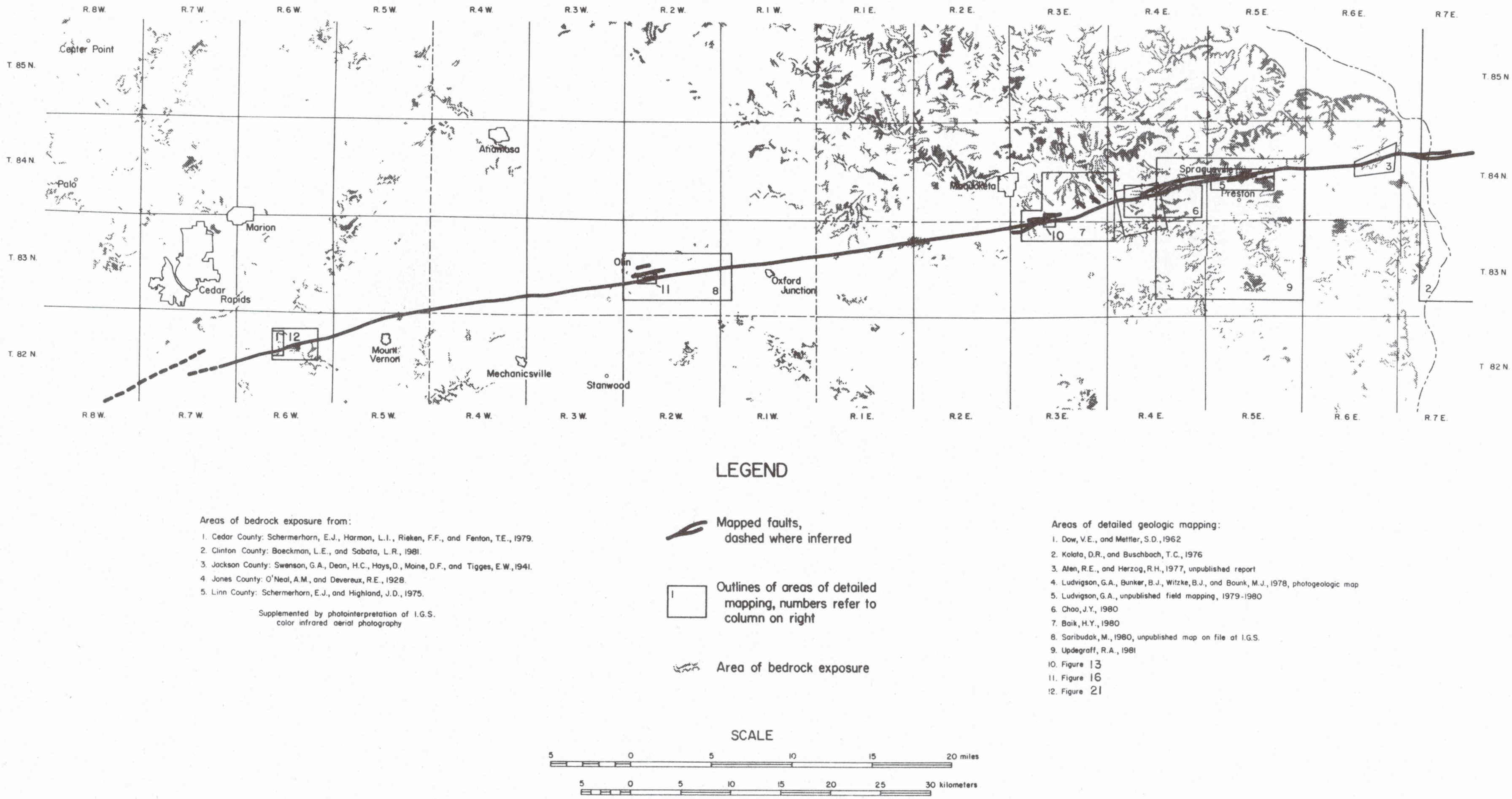


Figure 12. Areas of bedrock exposure and detailed geologic mapping in east-central Iowa.

greatly thinned in this area. The tectonic setting of this outlier, and the dip of beds measured in adjacent exposures (fig. 13) strongly suggests that the various units have been juxtaposed, and it is likely that further mapping at an even larger scale would reveal a complex pattern of minor faulting within the graben.

Several interesting fault-related alterations of Silurian carbonate units can be found in cataclastic zones around the Silver Creek Outlier. In area (A) calcite-cemented breccias (see figs. 13 and 14) are exposed immediately to the south of exposures of Devonian rocks. Petrographic examination of these rocks by Baik (1980, p. 145) suggested that cementation was accomplished by successive stages of iron oxide and calcite deposition. Examination of the calcite cement under cathodoluminescent illumination revealed an alternating sequence of brightly luminescent and non-luminescent crystal growth zones, which Baik (1980, p. 145) attributed to variations in Mn^{++} or Fe^{++} trace element concentrations. Petrographic examination of these cements using the staining method of Dickson (1965) reveals alternating growth of non-ferroan and ferroan calcites, suggesting that the cathodoluminescent banding reported by Baik (1980, p. 145) probably results from changes in the Fe^{++}/Mn^{++} ratio of the crystal growth zones (Frank et al., 1982, p. 636). In area (B), immediately to the south of the cataclastic zone containing the calcite-cemented breccias, an entire exposure of Scotch Grove Formation has been altered by pervasive silicification. At location (C) is an exposure of intensely fractured, coarse-grained, saccharoidal dolomite. This unusual rock type has been observed elsewhere along the Plum River Fault Zone (Chao, 1980, p. 38-41) and in the area around the Silver Creek Outlier (Baik, 1980, p. 77, p. 86-94). At location (D) is an exposure of iron oxide-cemented dolomite breccia.

The area around the Silver Creek Outlier was mapped by Baik (1980, p. 131), who also made supplemental gravity traverses across the Plum River Fault Zone to determine magnitudes of vertical displacement of the Precambrian basement surface. Baik (1980, p. 131, p. 170) interpreted the internal structure of the graben at Silver Creek as a rotated fault block dipping to the south. Subtracting density effects of near surface rocks, Baik (1980, p. 170) interpreted a gravity traverse across the graben to indicate 300 feet (91.5 m) of throw on the south bounding fault, and 100 feet (30.5 m) on the north bounding fault. Thus he interpreted 200 feet (61 m) of net vertical displacement of the basement surface with the north side downthrown at Silver Creek.

The interpreted magnitudes of faulting of the basement surface by Baik contrast with conservative estimates of fault throw, based on the configuration of near surface rocks. Cross section C-C' (figs. 11 and 15) depicts a simplistic structural interpretation of the graben at Silver Creek, showing no internal faulting. Assuming no dramatic thickening or thinning of Devonian or Silurian units, and using the top of the Blanding Formation as a structural datum, the throw on the south bounding fault is interpreted as approximately 460 feet (140 m), the north bounding fault as approximately 230 feet (70 m), yielding a net vertical displacement across the structure of 230 feet (70 m), downthrown to the north. Unless reversals in fault motion have occurred in the past, the net displacements on the basement surface are unlikely to be less than those observed in the near-surface rocks.

Pleasant Hill Devonian Outlier

The Pleasant Hill Devonian Outlier is a structurally preserved remnant of

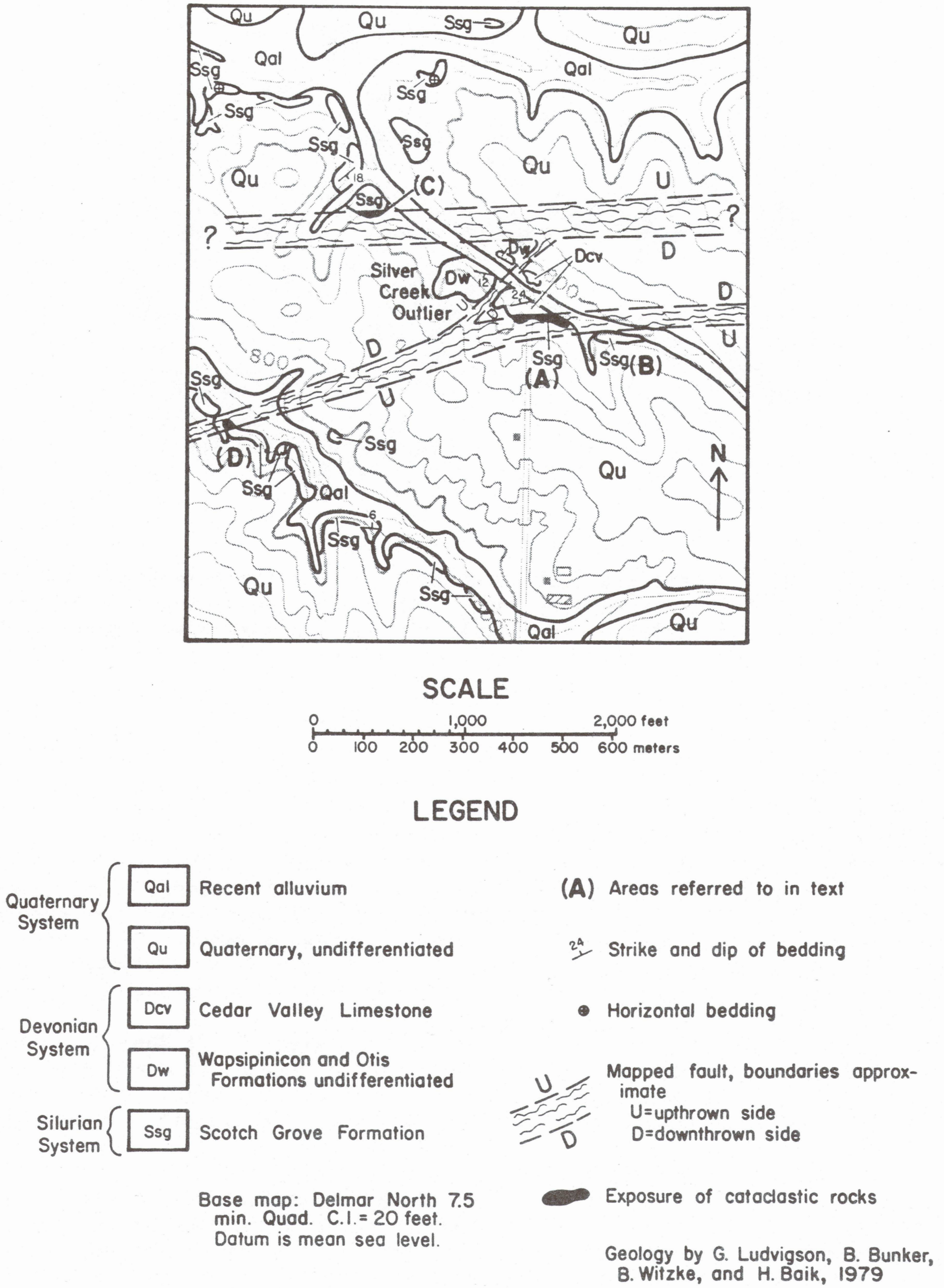


Figure 13. Geologic map of the Silver Creek Devonian Outlier. Location is shown in figure 12.

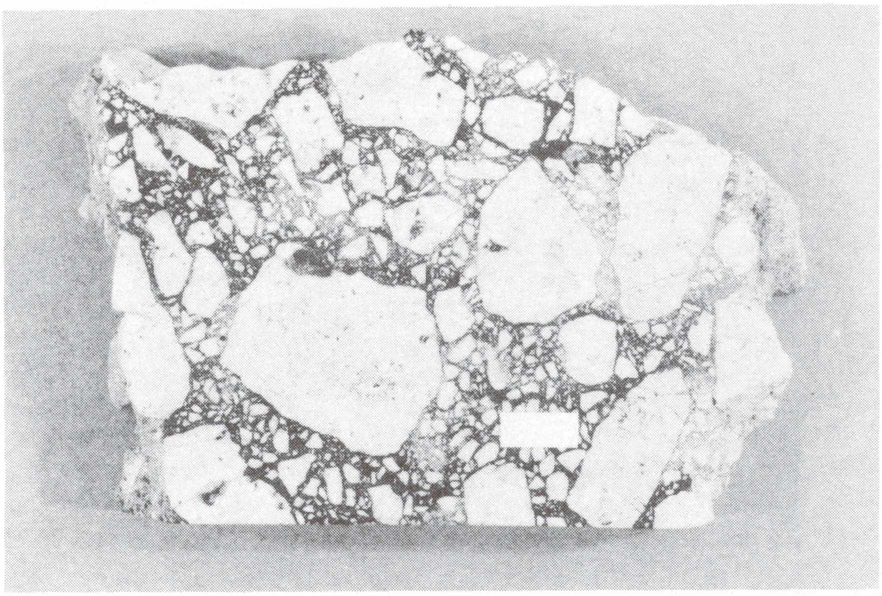


Figure 14. Calcite-cemented fault breccia from location A in figure 13. Matrix is ferruginous gouge. This rock type represents one of several varieties of cataclastic fabrics that are frequently observed along the Plum River Fault Zone.

Middle Devonian rocks that is contained within a structurally complex graben. The outlier occurs some 16 miles (27 km) east of the erosional edge of Devonian strata in eastern Iowa, isolated within the Silurian outcrop belt. Located in section 20 of T83N, R2W, it was discovered in 1977 by Bunker and Ludvigson during reconnaissance mapping of the Plum River Fault Zone. The first geologic description of the outlier appeared in Ludvigson et al. (1978, p. 30-34). The geology of the area surrounding the outlier was mapped by Saribudak (1980; fig. 12).

The internal geologic structure of the Plum River Fault Zone is best documented at the Pleasant Hill Outlier (fig. 16). The fault contact relationship between the Devonian rocks and the Silurian rocks which border them on the south was discussed in Ludvigson et al. (1978, p. 30-34). The southern margin of this major fault is exposed in the Freeman Quarry (fig. 16). A transition from undisturbed flat-lying beds of the Scotch Grove Formation to pervasively silicified cataclastic breccia can be seen along the east wall of the Freeman Quarry (fig. 17). Bordering the cataclastic zone at the north end of the quarry are a series of narrow, rotated fault blocks. The displacements of individual faults are not known, because the beds preserved within these blocks are difficult to compare from block to block, as well as to the undisturbed section exposed at the south end of the quarry. The northward dips in these blocks suggest that the faults may be step faults synthetic to the master fault exposed at the north end of the quarry. If this structural interpretation is correct, then the different lithologies observed in the rotated fault blocks are structurally preserved remnants of younger Scotch

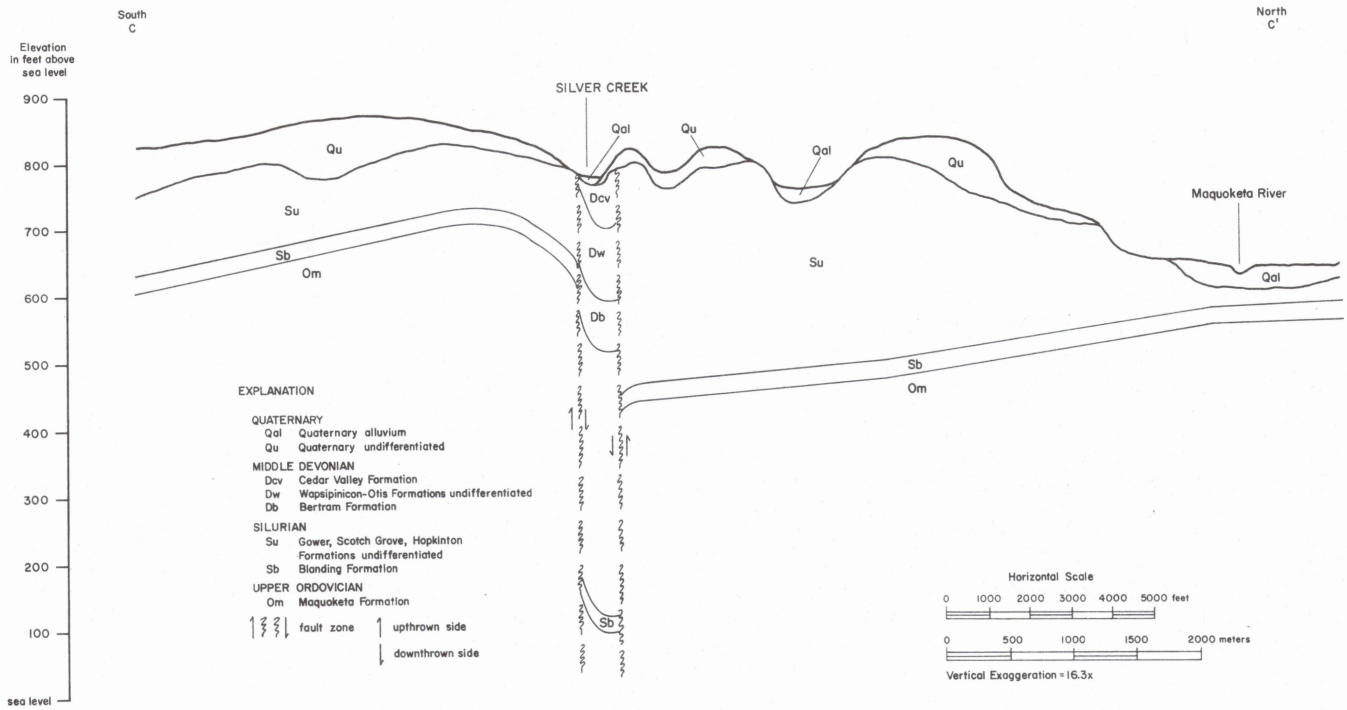


Figure 15. Cross-section showing a conservative structural interpretation of the Silver Creek Devonian Outlier along the Plum River Fault Zone. Complex faulting within the graben is indicated from field relationships. Line of cross-section is shown in figure 11.

Grove or Gower beds that have been eroded from the surrounding area. Silicified cataclastic breccia is also well exposed at the southern margin of the southern boundary fault in the westernmost part of figure 16. These rocks are bordered to the south by northward dipping rocks of the Scotch Grove Formation.

The internal structure of the graben which preserves the Devonian rocks is not known with great precision, but the distribution and structural relationships of these outcrops suggests that part of the Pleasant Hill Outlier is cut by complex block faulting. The eastern and western portions of the outlier in figure 16 differ in several important respects. The western portion of the outlier appears to be narrower, with a corresponding increase in the width of the southern boundary fault. A small outcrop of the Cedar Valley Formation occurs about 1000 feet (305 m) to the northwest of the Freeman Quarry (fig. 16) at an elevation of 795 feet (242 m) above sea level. This outcrop is bordered to the north by a large outcrop of the Bertram Formation, indicating that here the maximum downdropping occurs in the southern half of the graben. Some 800 feet (244 m) to the west of the Cedar Valley outcrop, an exposure of Scotch Grove Formation occurs at 800 feet (244 m) above sea level (fig. 16). Based on preliminary megascopic examination, this exposure appears to be a cataclastic breccia which has recrystallized to a coarse-grained saccharoidal dolomite. Immediately to the north of the Scotch Grove exposure, rocks of the Wapsipinicon and Otis formations are exposed in a south-facing roadcut. These heterogeneous rocks display a variety of lithologic characteristics which compare to several different units in the Wapsipinicon and Otis formations, and appear to have been tectonically mixed into a melange. The

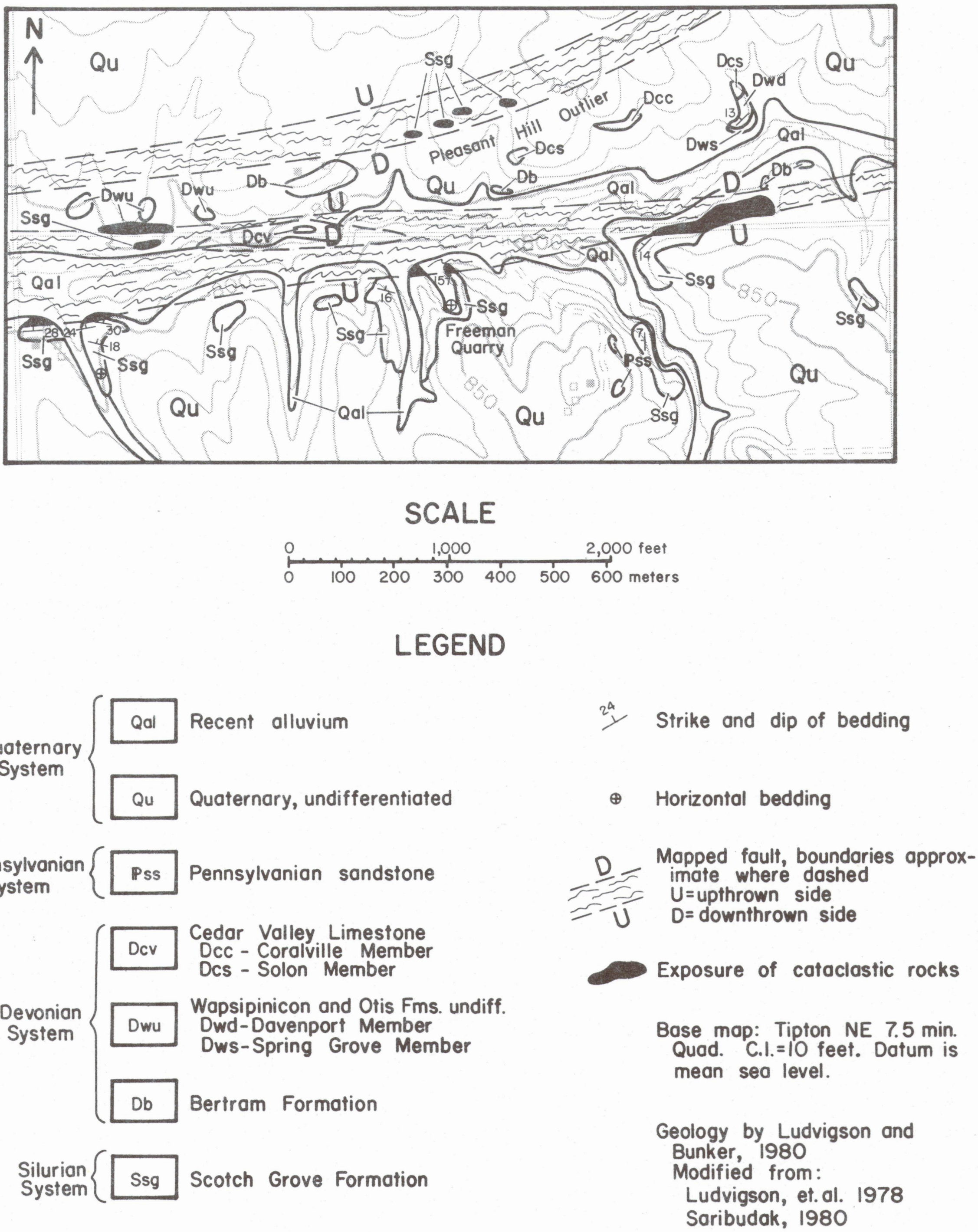


Figure 16. Geologic map of the Pleasant Hill Devonian Outlier. Location is shown in figure 12.

physical relationships between these exposures of the Cedar Valley, Scotch Grove, Wapsipinicon and Otis formations clearly require large-scale vertical faulting within the western portion of the graben structure.

In the eastern half of figure 16, the graben apparently has a greater width, and the southern boundary fault is relatively narrow. The strike and dip of bedding in the Devonian rocks are clearly revealed in an asymmetrical cuesta-like hill on which units of the Wapsipinicon and Cedar Valley formations are exposed in normal stratigraphic succession (figs. 16 and 17). Here, bedding surfaces and unit contacts dip 13° to the northwest (fig. 16). This indicates that in the eastern portion of the graben, the maximum downdropping occurs in the northern half. While zones of intense fracturing in these rocks suggest the possibility of minor faulting in this area, the position of outcrops of Cedar Valley Formation to the west and Bertram Formation to the southeast do not require faulting. In contrast to the intensely faulted structure indicated in the western portion, the eastern portion of the graben may be interpreted as a simple northwest dipping fault block.

Reconnaissance geologic mapping to the north of the Pleasant Hill Outlier by Saribudak (1980) showed that the Middle Devonian rocks are enclosed in a graben. The location of a fault which borders the northern edge of the outlier was not apparent, however. Gravity and ground magnetic profiling along the north-south road forming the eastern border of figure 16 suggested the presence of this fault, which Saribudak (1980) interpreted as an east-west trending structure. At the same time, detailed geologic mapping of the Pleasant Hill Outlier revealed the presence of a series of small, east-west trending exposures of coarse-grained saccharoidal dolomite approximately 1100 feet (335 m) north of the Freeman Quarry (fig. 17). These rocks closely resemble bodies of recrystallized Scotch Grove Formation that have been found elsewhere in intimate association with major faults in the Plum River Fault Zone. The overall east-northeast trend of these exposures, combined with the structural interpretation of Saribudak (1980), have been used to infer the approximate position of the northern boundary fault. The westward convergence of the southern and northern boundary faults implied by this interpretation suggests that these two faults may merge to the west of the mapped area in figure 16. The merging of these two boundary faults may explain the complex faulting pattern noted in the western half of figure 16. The subsurface structure of this area of complex faulting was investigated along the north-south gravel road forming the western border of figure 16 by Svoboda (1980), who utilized combined gravity and ground magnetic data. Along this traverse, Svoboda (ibid., p. 56-61) interpreted the net throw on the Precambrian basement surface to be 650-675 feet (198-206 m), with the north side downthrown. Interestingly, the gravity data suggested an internal horst structure for the fault zone (ibid., p. 58-60), with local vertical displacement of the basement surface of up to 1100 feet (335 m) (ibid., p. 61). This data, along with the surface mapping at Pleasant Hill, Silver Creek, and other localities (Chao, 1980; Baik, 1981), indicates that the internal structure of the Plum River Fault Zone changes rapidly along strike. While the present data permit only tentative conclusions, a pattern of braided, intersecting faults is hypothesized as the characteristic internal structure of the fault zone in map view. In cross-section view (see figs. 11 and 18), complex vertical block faulting is well documented. Ludvigson et al. (1978, p. 32) suggested the possibility of dextral strike-slip components of movement along the fault zone in a discussion of the geologic structure at Pleasant Hill. In recent years, Trapp and Fenster (1982), Heyl and West (1982, p. 1807) and Heyl (1983, p. 86) have all proposed that the Plum River Fault Zone is a right-lateral fault system. While the interpreted structural geometries described in this report are not incompatible with strike-slip faulting, this hypothesis is untestable with the current data set.

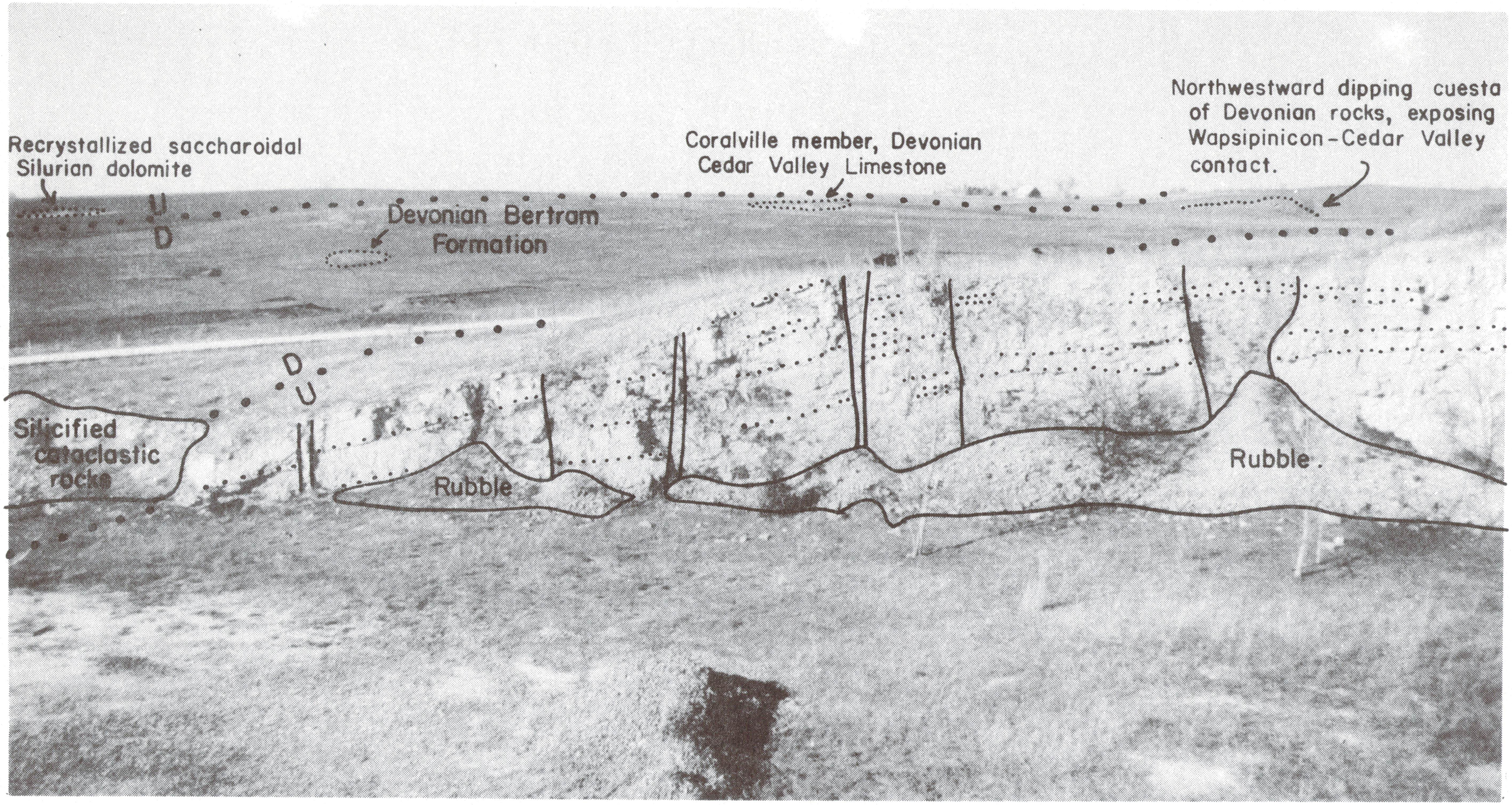


Figure 17. Annotated photograph of the east wall of the Freeman Quarry (fig. 16), looking northeast. Flat-lying carbonate strata (light dotted lines) of the Silurian Scotch Grove Formation in the southern (right) part of the wall are contrasted with northward dipping fault blocks, which are bounded by small vertical faults in the center of the figure. At the north (left) end of the wall, pervasively silicified breccias mark the southern limit of a major fault (heavy dotted line) which bounds the Pleasant Hill Devonian Outlier. Exposures of Devonian rocks are noted in the background. The approximate southern limit of a major fault (heavy dotted line) which bounds the north side of the Devonian outlier is shown near the horizon.

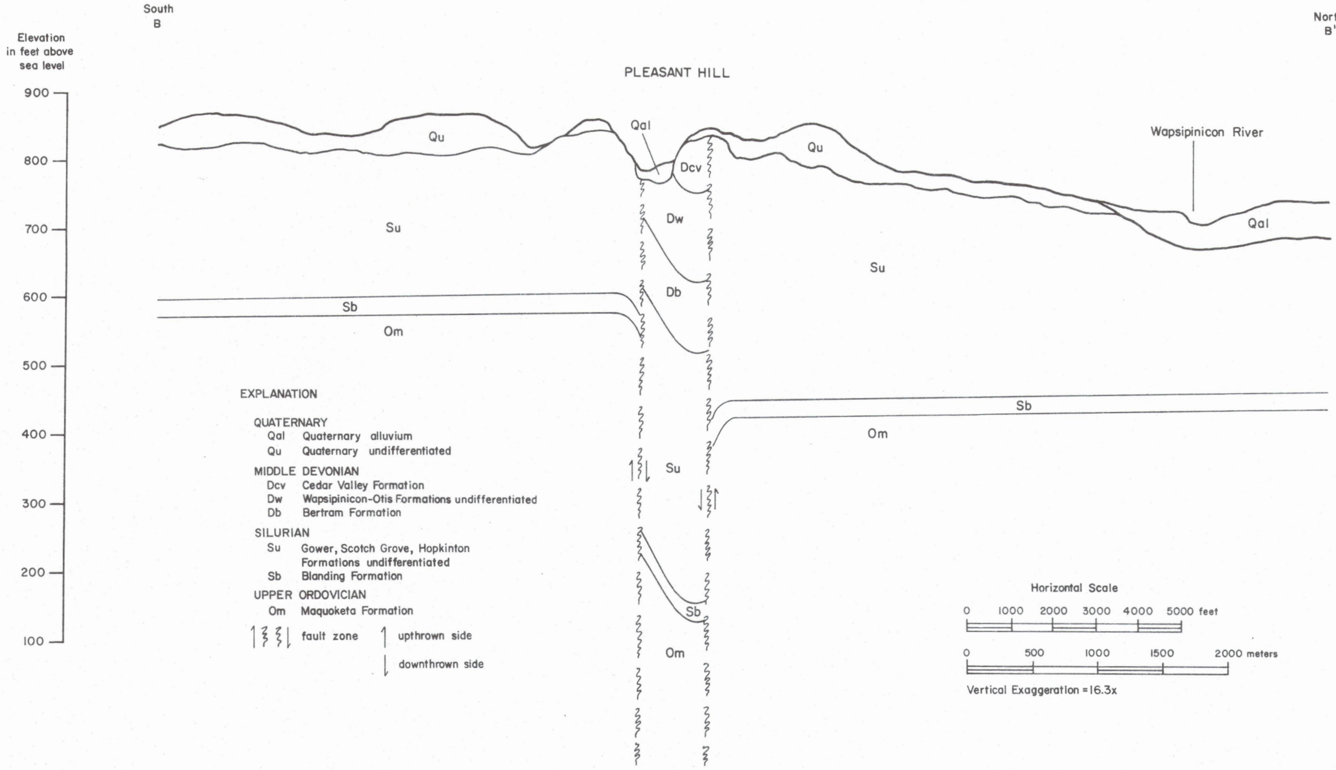


Figure 18. Cross-section showing a conservative structural interpretation of the eastern portion of the Pleasant Hill Devonian Outlier. Line of cross-section is shown in figure 11.

Saribudak (1980) mapped an east-northeast trending fault juxtaposing the Gower and Scotch Grove formations some distance to the north of the structures depicted in figure 16. This fault (see fig. 12) is downthrown to the north and is characterized by a central zone of pervasively silicified cataclastic breccia. The northern edge of this fault occurs more than 3900 feet (1200 m) to the north of the southern margin of the Plum River Fault Zone at Pleasant Hill, suggesting the possible width of deformation that might be expected along more poorly exposed sections of the fault zone.

Interpretations of the structural geology at the Pleasant Hill Outlier have been further refined by the acquisition of 600% common depth point shallow seismic reflection data across part of the Plum River Fault Zone. This project is decribed by Cumerlato (1983), who designed and supervised the field collection of seismic data (ibid., p. 18-40). Data processing procedures were performed by Seismograph Service Corporation (Data Processing Division, Denver Service Center, Denver Co., Job Number 3622) and are described in Cumerlato (1983, p. 41-63).

The seismic data were collected along a 2.3 mile (3.8 km) line between the common corner of sections 28, 29, 32, and 33 to the common corner of sections 16, 17, 20, and 21 in T83N, R2W, in southern Jones County (Cumerlato, 1983, p. 5). This seismic traverse includes the eastern boundary of figure 16, and appears to cross the entire width of the graben which encloses the Pleasant Hill Outlier (fig. 19). Note on figure 19 that the seismic traverse does not reach the projected intersection with the northern fault mapped by Saribudak (1980). Apparently, the seismic traverse of Cumerlato (1983) does not cross the entire width of the Plum River Fault Zone.

Figure 20A shows the processed record section of the traverse at Pleasant Hill, and three alternative interpretations of the seismic reflection data. Without the benefit of other geologic data to constrain the structural interpretation of the processed record section, a large number of interpretations are permissible. At present the only geologic constraints available are the results from surface mapping and the combined gravity and magnetic profiles by Svoboda (1980) one mile to the west. No subsurface drilling information is available in the immediate area.

Cumerlato (1983, p. 65-66) noted that the bed thickness resolution for these data are limited to about 40 to 80 feet (13-25 m), and discussed seismic reflection patterns that could be used to interpret the structure of the fault zone. The two most useful patterns appear to be: 1) lateral discontinuity of seismic reflection events (fault displacement of reflecting interfaces), and 2) parabolic diffraction patterns (emanating from point sources such as bed terminations by faults) (Cumerlato, 1983, p. 66).

Cumerlato (1983, p. 74-76) proposed that major reflection events in the seismic section could be correlated to the Silurian-Ordovician unconformity, the St. Peter Sandstone - Prairie du Chien Group unconformity, and the Paleozoic-Proterozoic unconformity. These reflection events are shown in figure 20B and C. In the southern half of the seismic section, laterally continuous reflection events indicate that the Paleozoic stratigraphic sequence is flat-lying and relatively undisturbed. This pattern is severely disrupted in the northern half of the seismic section, where the traverse crosses the Plum River Fault Zone. Cumerlato (1983, p. 76) observed that: "major faulting and cataclastic development associated with the Pleasant Hill Outlier are believed to be responsible for the poor data quality in the vicinity of the outlier. Detailed data interpretation within that portion of the section would be quite difficult to substantiate. However, it does serve to delineate the width of the fault zone which is seen to be 975 meters (3200 feet) here." Using a depth scale relating two-way travel times to calculated depth, Cumerlato (1983, p. 77) estimated fault displacements on the top of the Prairie du Chien Group, suggesting 230 feet (70 m) of net throw, and throws up to 550 feet (172 m) within the fault zone (fig. 20B). Figure 20C shows a second interpretation of the seismic section by Cumerlato (1983, p. 83). In the rationale for this interpretation, Cumerlato (ibid., p.79-84) discussed the possibility of down-to-the-south early Paleozoic faulting in the southern half of the seismic section.

The seismic reflection data at Pleasant Hill, and the accompanying interpretations of Cumerlato (1983) appear to corroborate earlier structural interpretations which were based primarily on geologic mapping. However, the correlation of stratigraphic units and individual seismic reflection events without supporting drillhole information or sonic log data is tenuous. This is the case even for the southern half of the seismic section, where structural complications are relatively minor (figure 20B and C). It is clear, however, that the attempt to identify unit contacts within the fault zone (figure 20B and C) is probably futile. Interpretations of dip angle, sense of displacement and location of individual faults based solely on the seismic data should also be viewed cautiously.

Figure 20D shows a third interpretation of the seismic section, in which perceived seismic reflection events are annotated, but no attempt is made to uniquely specify the identity or source of these events. Certain laterally

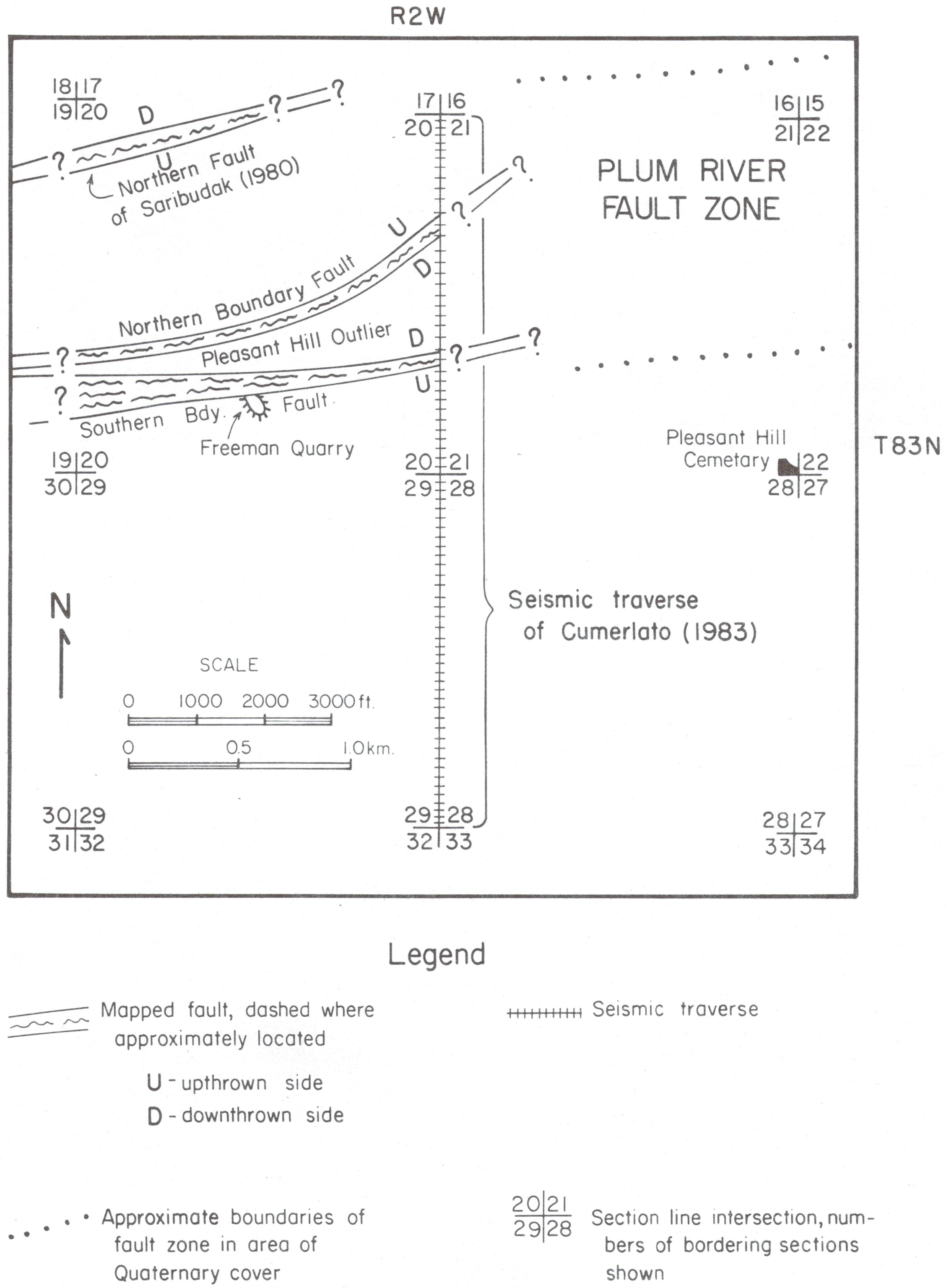
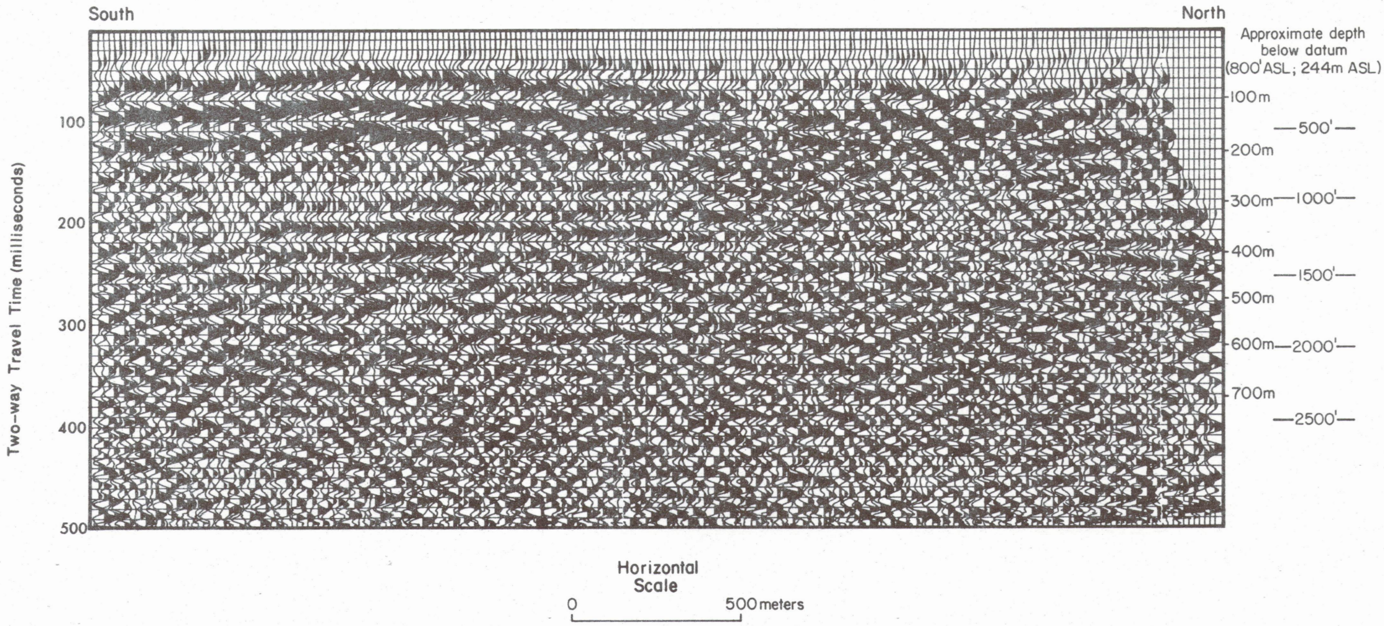
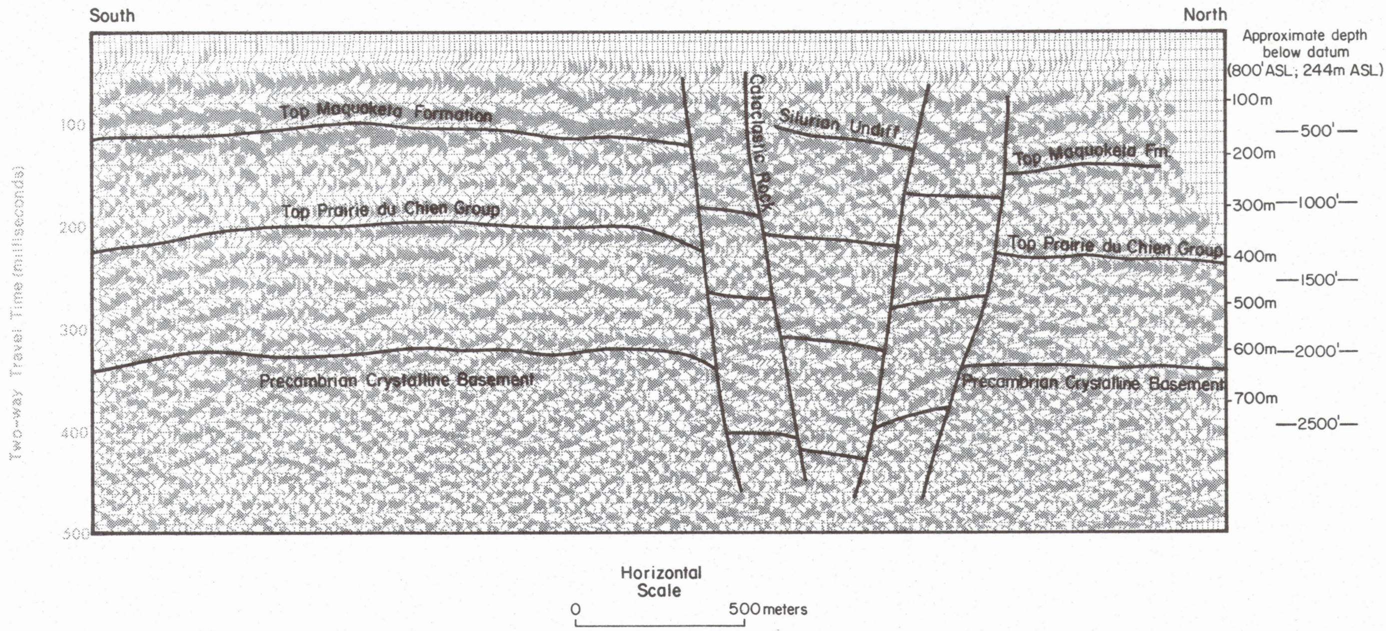


Figure 19. Location of the seismic reflection traverse of Cumerlato (1983) in relation to mapped geologic structures in the area around the Pleasant Hill Devonian Outlier.

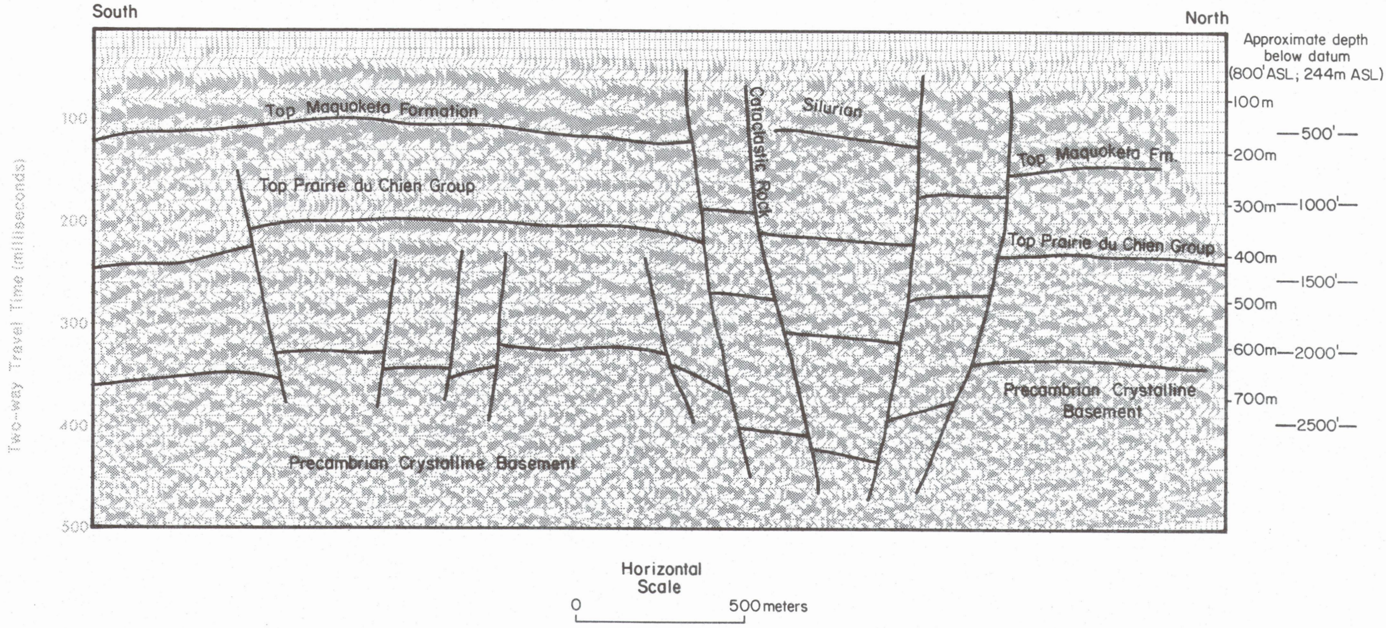


A. Processed record section.

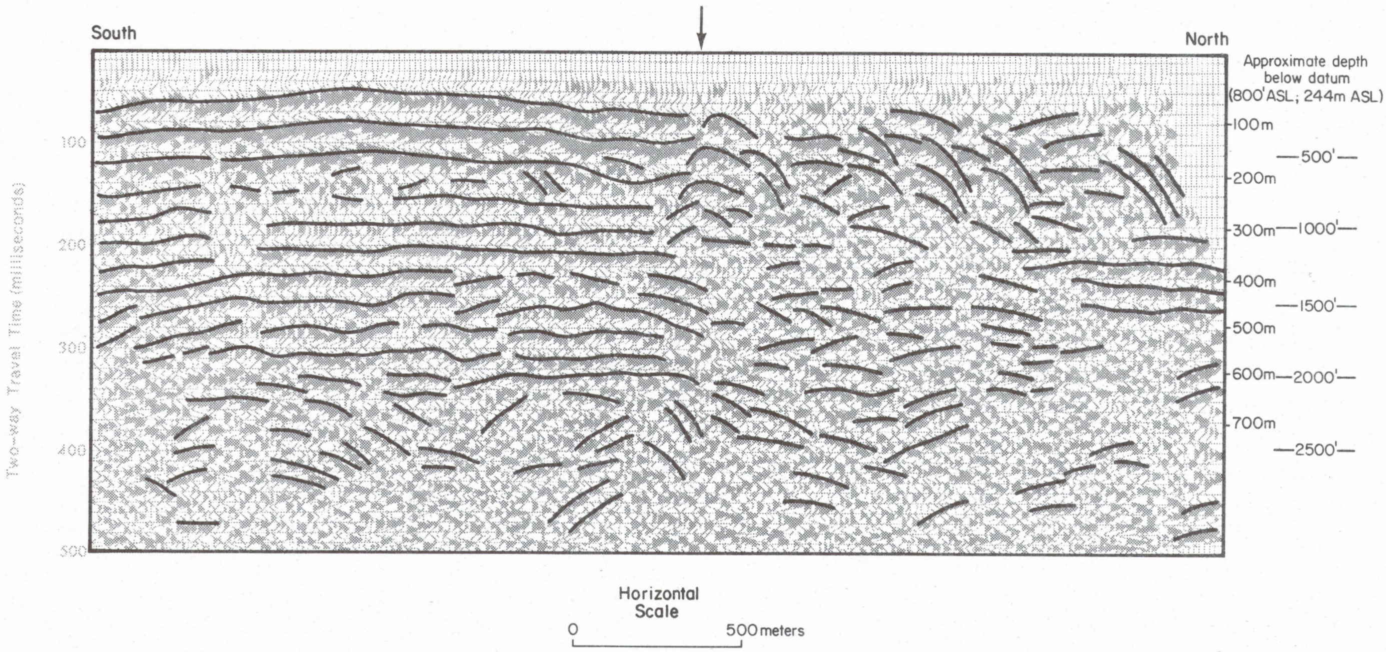


B. An interpretation of the geologic structure by Cumerlato (1983, p. 73).

Figure 20. Processed record section of the seismic reflection data of Cumerlato (1983), and three alternative structural interpretations of the data.



C. An alternative interpretation of the geologic structure by Cumerlato (1983, p. 83).



D. An alternative interpretation of the geologic structure by the authors of this report. The arrow marks the approximate southern edge of the Plum River Fault Zone.

Figure 20. Continued.

continuous reflection events like those noted in the southern half of the seismic section (fig. 20D) undoubtedly depict flat-lying Paleozoic strata. Likewise, in the northern half of the seismic section, certain convex upward reflection events are easily interpreted as parabolic diffractions (fig. 20D). The abrupt boundary between these two characteristic patterns probably marks the southern edge of the Plum River Fault Zone (arrow, fig. 20D). The occurrence of laterally continuous reflection events at the northern end of the seismic section (fig.20D) might possibly indicate flat-lying Paleozoic strata to the north of the graben which bounds the Pleasant Hill Outlier. At present it is unclear whether the poor data quality at the outlier results from pervasive minor faulting within the graben or high fracture densities as noted in outcrop. Finally, the vertical change from laterally continuous horizontal reflection events to underlying discontinuous, steeply-dipping reflection events noted to the south (upthrown side) of the Plum River Fault Zone (fig. 20D) could well record the presence of the Paleozoic-Proterozoic unconformity at an approximate depth of 2000 feet (610 m).

Cross section B-B' (figs. 11 and 18) depicts a conservative interpretation of the structure of the eastern segment of the graben at Pleasant Hill. Assuming no dramatic thickening or thinning of Devonian and Silurian units, and using the top of the Blanding Formation as a datum, the throw on the southern boundary fault is interpreted to be 320 feet (98 m), and the throw of the northern boundary fault is interpreted to be 150 feet (46 m). Thus the net throw of the fault zone is calculated to be 170 feet (52 m), although approximately 440 feet (134 m) of total structural relief occurs within the fault zone. The northern fault of Saribudak (1980) is not shown on this cross-section, because of uncertainties of configuration. It should be noted, however, that this fault increases the net displacement across the Plum River Fault Zone.

Skvor-Hartl Area

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 in Linn County, based on integrated core studies and surface geologic mapping. According to the structural interpretation of Dow and Mettler (1962, p. 329, 331), the Cedar Valley and Wapsipinicon formations are preserved along the axis of the syncline, which is bordered on the south by an anticline which exposes undifferentiated Silurian rocks. The stratigraphy and carbonate facies of the Silurian rocks of the area, particularly in the area around Palisades-Kepler State Park (fig. 21) have been described by Philcox (1970b) and Witzke (1981a). Bunker and Ludvigson (1977) related the structures of the Skvor-Hartl area to the regional pattern of deformation along the Plum River Fault Zone. Ludvigson et al. (1978, p. 27) suggested that the Plum River Fault Zone continues through the Skvor-Hartl area, bisecting the paired west-southwest plunging folds of Dow and Mettler (1962). The Skvor-Hartl area is the westernmost study area where the general structural characteristics of the Plum River Fault Zone can be deduced from surface exposures. Subsurface data suggest that the western terminus of the fault zone occurs approximately 3-5 miles (4.8-8 km) west of the western border of figure 21.

The detailed geologic structure of the Skvor-Hartl area is difficult to interpret, because the bedrock exposures are more widely spaced than in other field study areas. This problem necessitated field mapping of a larger area at a smaller scale in order to deduce the general structural pattern. No exposures of cataclastic rocks have been found at the Skvor-Hartl area. This could possibly be explained by the sparse distribution of bedrock exposures, or might possibly indicate that extensive brittle cataclastic deformation does not occur in the fault zone near its western terminus. Nevertheless, the outcrop distribution of several different stratigraphic units, and structural relationships observed in the scattered bedrock exposures permit several important generalizations to be made about the bedrock structure.

The most critical structural and stratigraphic relationships in the Skvor-Hartl area are observed along the valley walls of a northward draining creek in the southwestern part of figure 21. Exposures of the Scotch Grove, Gower, Otis, Wapsipinicon, and Cedar Valley formations occur at roughly equivalent elevations along a 2000 foot (610 m) north-south traverse. At the north end of this traverse, the Coralville Member ("*Idiostroma* beds") of the Cedar Valley Formation is exposed on the west wall of the valley, along the axis of the west-southwest plunging syncline of Dow and Mettler (1962, p. 329). The cumulative stratigraphic thickness of the Devonian units present in this area, including the Bertram Formation, is 261 feet (79.6 m) (ibid., p. 328-331), suggesting the minimum vertical displacement observed along this north-south traverse. If it is assumed that all of this vertical displacement is related to folding (i.e., Dow and Mettler, 1962), then the average northward dip along a 1200 foot (366 m) traverse from the southernmost exposure of the Cedar Valley Formation to the northernmost exposure of Silurian rocks should be 12.3°. Nowhere along this traverse have northward dips as steep as 12° been observed. Dow and Mettler (1962, p. 329) recorded northward dips of 8-11° in the Cedar Valley Formation on the south limb of the west-southwest plunging syncline. Farther south along this traverse, the exposure of Otis Formation is flat-lying, and the northernmost exposure of Silurian rocks dips 6° to the north. Therefore, the observed stratigraphic displacement must be explained by local steepening of the dip and/or faulting. The interpretation of a west-southwest plunging anticline farther south along this traverse (Dow and Mettler, 1962, p. 329, 331) has not been supported by subsequent field mapping, or by investigations of the Silurian rocks to the east in Palisades Kepler State Park, or at the Martin-Marietta Quarry (Philcox, 1970b; Witzke, 1981a).

The Wapsipinicon and Otis 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 figure 21. Here, these rocks are flat-lying. The Wapsipinicon, Otis and Bertram formations are also exposed to the north of U.S. Highway 30 in the northwest part of figure 21. These rocks are mostly flat lying, except for a local disturbance exposed in a roadcut along the entrance ramp to Iowa Highway 13.

While the field relationships at the Skvor-Hartl area are less clear than at the other two special study areas, several important similarities can be seen. The structural significance of each of these localities was originally recognized by the juxtaposition of widely separated stratigraphic units. In each case, exposures of the Cedar Valley Formation occur in a localized down-dropped structure. These down-dropped structures are internal features of the Plum River Fault Zone, a regionally continuous east-west trending zone of deformation which is downthrown on the north side. At Silver Creek and Pleasant Hill, outliers of the Cedar Valley Formation occur in grabens which are clearly bounded by faults. At the Skvor-Hartl area, near the western terminus of the Plum River Fault Zone, the structural relationships between the central

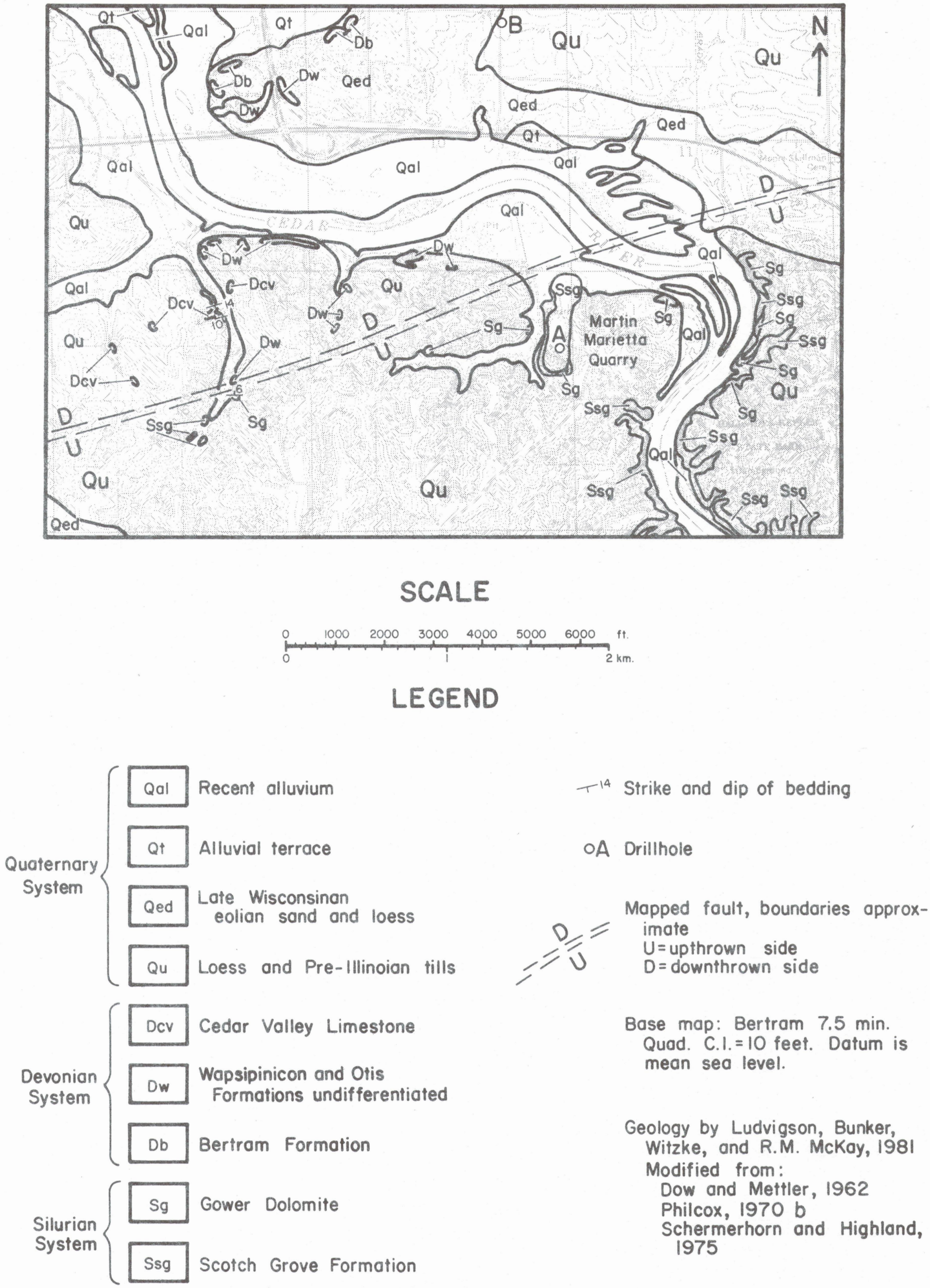


Figure 21. Geologic map of the Skvor-Hartl area. Location is shown in figure 12.

down-dropped area and the rocks to the north and south are less clear. Cataclastic rocks, which have proved to be useful in mapping faults along other portions of the fault zone, have not been found here. This may possibly be explained by lower outcrop density. Structural relationships at the Skvor-Hartl area indicate that the Cedar Valley Formation is preserved along the axis of a west-southwest plunging syncline, which is bounded on the south by a similarly trending narrow zone of faulting and/or locally steepened dip (monoclinal folding). No evidence suggestive of faulting or large scale folding has been found to the north of the synclinal axis. While the maximum vertical displacement occurs along a narrow belt between the synclinal axis and the exposed Silurian rocks to the south, a net downward displacement to the north is apparent from the outcrop distribution, and from closely-spaced subsurface data.

Locations A and B on figure 21 denote the positions of two drillholes which fully penetrated the Silurian System and reveal the net vertical displacement across the Skvor-Hartl area. Data from drillhole A is found in the Linn County outcrop file at IGS, and is a log of an exploratory core at the Martin-Marietta Cedar Rapids South Quarry (sec. 15, T82N, R6W). Drillhole B was drilled as an IGS-USGS hydrogeologic research well (W-23838). At drillhole A, the top of the Blanding Formation was penetrated at 486 feet (148 m) above sea level, and the top of the Maquoketa Formation was penetrated at 464 feet (141 m) above sea level. At drillhole B, the top of the Blanding Formation was penetrated at 325 feet (99 m) above sea level, and the top of the Maquoketa Formation was penetrated at 298 feet (91 m) above sea level. Thus, the net vertical displacement across the Plum River Fault Zone between drillholes A and B is 161-166 feet (49-51 m), with the north side downthrown.

Svoboda (1980, p. 109-111) investigated the subsurface structure of the Precambrian basement surface along the Plum River Fault Zone, utilizing combined gravity and ground magnetic profiles along a gravel road one mile (1.6 km) west of the western border of figure 21. These data suggest the presence of a zone of block faulting approximately one mile (1.6 km) wide, with a graben-horst-graben configuration (ibid., p. 65-67). The net throw on the basement surface was interpreted to be 650 feet (198 m), with the north side downthrown (ibid., p. 117). The relationship between this interpretation and the near-surface structure observed at the Skvor-Hartl area is unclear, and further studies will be required to clarify the structural relationships between the basement and Paleozoic rocks of the Plum River Fault Zone.

Cross-section A-A' (figs. 11 and 22) shows the interpreted structure of the Skvor-Hartl area. Particularly noteworthy is the absence of the Bertram Formation on the southern, upthrown side. The throw of the Plum River Fault Zone, using the top of the Blanding Formation as a datum, is 270 feet (82 m), although a total structural relief of 330 feet (101 m) is interpreted.

The geology of the Skvor-Hartl area has gained some notoriety because of the recent discovery of hydrothermal mineralization of the Mississippi-Valley-Type (MVT) in the Scotch Grove and Gower formations at the Martin-Marietta Cedar Rapids South Quarry (Garvin, 1982; Heyl and West, 1982, p. 1812; Coveney and Goebel, 1983, p. 235; Ludvigson et al., 1983, p. 500; Garvin, 1984). The proximity of these marcasite-pyrite-sphalerite-calcite deposits and other MVT deposits to the Plum River Fault Zone may indicate that the fault zone served as a regional conduit for the migration of hot mineralizing brines.

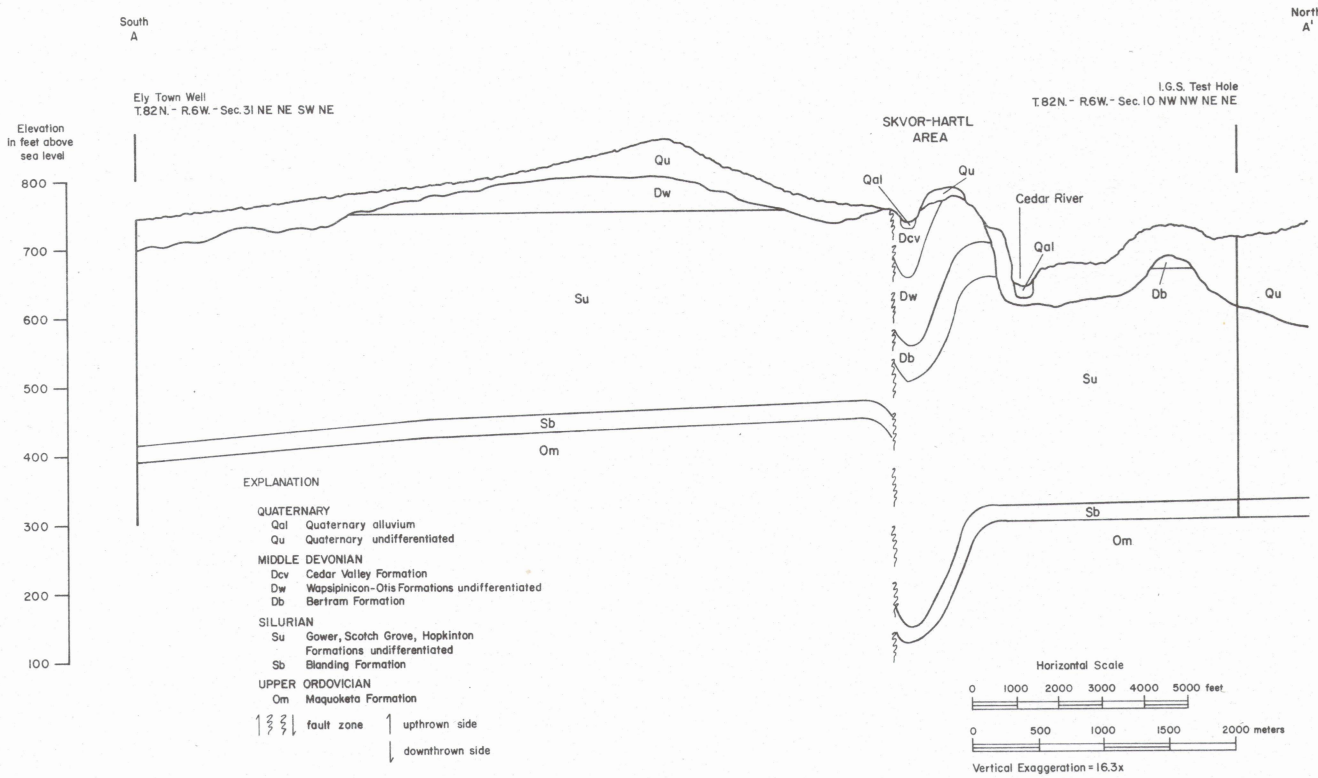


Figure 22. Cross-section showing a conservative structural interpretation of the Skvor-Hartl area. Note absence of Bertram Formation to the south of the Plum River Fault Zone. Line of cross-section is shown in figure 11.

TECTONIC HISTORY

Sloss (1963) subdivided the sedimentary record of cratonic continental interiors from very late Precambrian to the present into a series of six major sedimentary rock sequences (fig. 3), separated by major interregional unconformities. In general, each sequence is represented by a major transgressive-regressive cycle of deposition. Mapping the distribution of major stratigraphic units subcropping beneath a succeeding sequence can be important in helping to interpret the regional structural history of an area. This can also be supplemented by the use of isopach maps of key stratigraphic intervals. An isopach map shows the distribution and thickness of chosen mapping units, such as members, formations, systems, or sequences. Regional isopach maps may illustrate the size and shape of positive (arches, anticlines, etc.) and negative (basins, synclines, etc.) structural elements as they existed at a certain period in time, either at the close of a period of deposition or after an episode of regional erosional beveling marking a significant unconformity in the rock record. Isopach maps, therefore, are extremely useful in helping to unravel the tectonic history of a region under investigation. Profiles of key stratigraphic datums also provide useful insights into the geologic history of a region. A brief summary of the structural history of eastern Iowa and the development of the Plum River Fault Zone is discussed within the framework of Sloss's sequences.

Sauk Tectonic History

Sloss (1963) defined the Sauk Sequence as the sequence of strata ranging from very latest Precambrian to Early Ordovician age. In the central midcontinent region this sequence consists of Upper Cambrian (Mt. Simon Sandstone) through Lower Ordovician (Prairie du Chien Group) rocks (Table 1). Structural patterns developed during deposition of the Sauk Sequence can best be discussed through the use of an isopach map of this interval (fig. 23). The lowermost rocks (Mt. Simon), however, have not been included in the isopach because: 1) this sandstone filled many irregularities on the deeply dissected Precambrian erosional surface; and 2) because of difficulties in attempting to pick a consistent contact with underlying Keweenawan (Late Precambrian) sedimentary rocks.

The most obvious feature of the Sauk isopach map (fig. 23) is the north-south axis of thickening through eastern Iowa. Lee (1943, 1946) discussed the structural development of this structurally negative feature across southern Missouri, terming it the Ozark Basin in that area. Austin (1969, 1970) termed the northern extension of this structural depression into southern Minnesota the Hollandale Embayment. The Iowa Geological Survey has adopted use of this term for the prominent Sauk structural depression in Iowa.

Precambrian paleotopographic highs are noted in central Iowa (fig. 23), and these influenced Mt. Simon and post-Mt. Simon Sauk deposition. In a portion of east-central Iowa, isopach contours (fig. 23) show an east-west thinning of post-Mt. Simon Sauk rocks. This area of thinning corresponds to the general position of the Savanna-Sabula Anticline, which will be discussed more fully in the Absaroka tectonic history section. Thinning of Sauk rocks in this area suggests the possibility of Early Paleozoic structural movements along the trend of the Savanna-Sabula Anticline during and/or after Sauk deposition. This interpretation is speculative, and cannot be rigorously tested with the current data set.

Where the Shakopee Formation (upper Prairie du Chien Group) was erosionally removed prior to Tippecanoe deposition, thick St. Peter sections rest directly on the Oneota Formation (lower Prairie du Chien Group) or on Cambrian rock units (light stippled pattern, fig. 23). These thick St. Peter sections are generally only present northward from the general trend of the Savanna-Sabula Anticline, although a thick St. Peter channel-filling sequence is noted extending south into east-central Iowa (fig. 23). Pre-St. Peter erosion of the Prairie du Chien carbonates may have been influenced by structural movements along the Savanna-Sabula Anticline, although this remains to be demonstrated. Prior to St. Peter deposition "uplifting of the Wisconsin and Kankakee Arches" resulted in erosional truncation of Prairie du Chien and Cambrian rock units in northern Illinois, Wisconsin, and northeastern Iowa (Kolata et al., 1978, p. 24). Local pre-St. Peter, post-Prairie du Chien

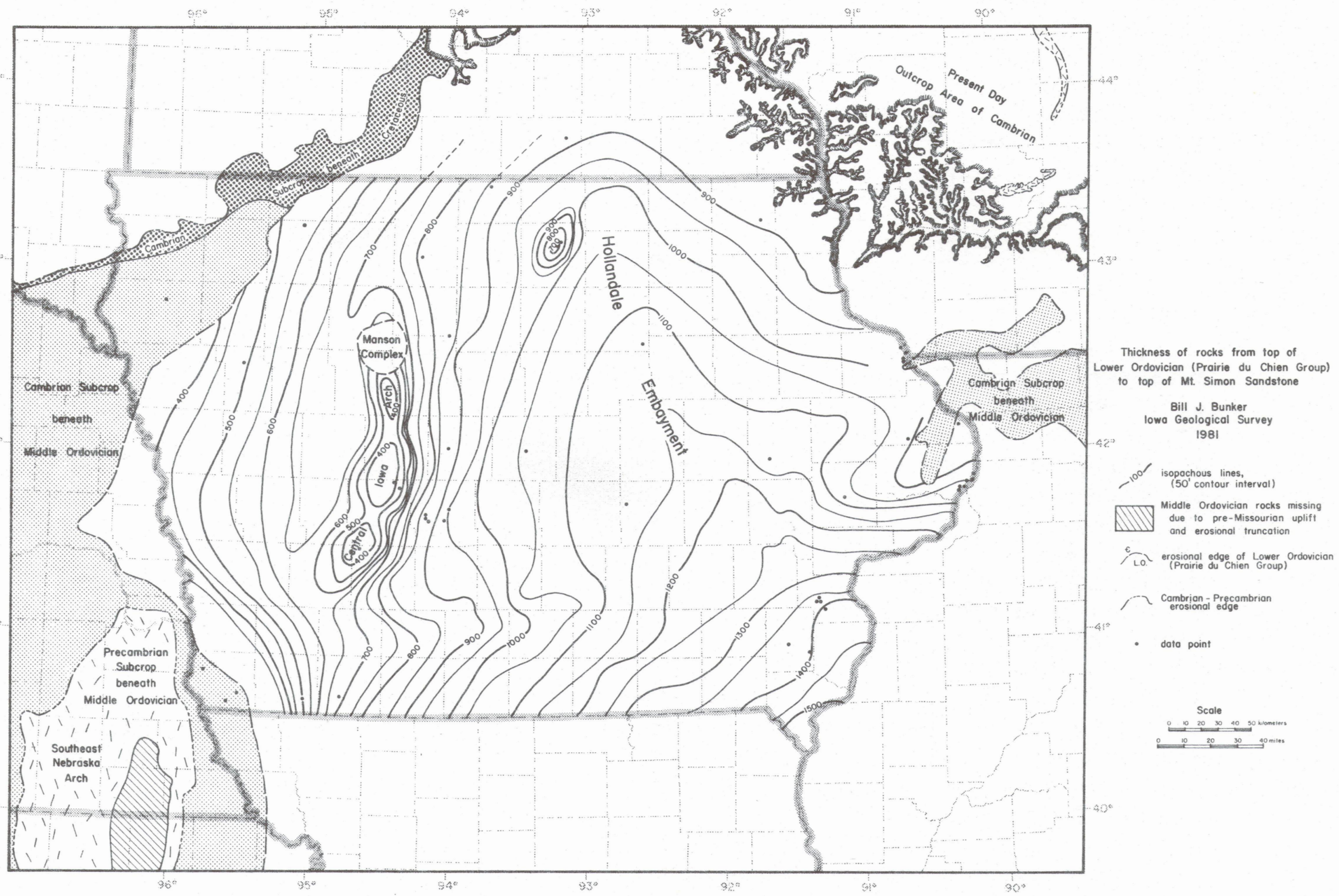


Figure 23. Isopach and paleogeologic map of the upper Sauk Sequence (adapted from Bunker, 1982).

structural movements are documented along the trend of the Sandwich Fault Zone in northern Illinois, and available "evidence suggests that unknown local structural features are concealed beneath the St. Peter Sandstone" (ibid.).

Tippecanoe Tectonic History

Sloss (1963) defined the Tippecanoe Sequence as the sequence of rocks from the base of the Middle Ordovician to the top of the Lower Devonian (fig. 3). In the central midcontinent region, this includes rocks from the base of the St. Peter Sandstone (Middle Ordovician) to the top of the Silurian (Table 1). The Early Devonian was marked by major marine regression out of the central midcontinent region, and rocks of this age are not present in the study area.

Earlier structural patterns developed during the deposition of the Sauk Sequence were greatly modified during Tippecanoe deposition (Middle Ordovician-Silurian) in eastern Iowa. The north-south oriented Hollandale Embayment (fig. 23) of eastern Iowa, south-central Minnesota, and its southward extensions into the ancestral Illinois and Ozark basins were disrupted by the uplift of the northward-trending Northeast Missouri Arch (Lee, 1943, 1946) and the east northeast-west southwest oriented Sangamon Arch (Whiting and Stevenson, 1965) across central Illinois.

The Tippecanoe structural grain in the Upper Mississippi Valley region (fig. 24) is principally oriented east-west in marked contrast to the north-south Sauk structural grain. During deposition of the Tippecanoe Sequence, subsidence of a new structural and depositional basin was initiated. This east-west oriented mid-Paleozoic feature was first noted by Lee (1946) and has been termed the East-Central Iowa Basin (Bunker, 1981). The apparent structural high in eastern Iowa (possible precursor to the Savanna-Sabula Anticlinal System) noted on the Sauk isopach map (fig. 23) became part of the East-Central Iowa Basin during Tippecanoe deposition. Further discussions regarding the structural development and destruction of the East-Central Iowa Basin area follow in this section as well as in the succeeding Kaskaskia and Absaroka tectonic history sections.

Examination of the Tippecanoe isopach map (fig. 24) illustrates the general post-Sauk structural reorganization of eastern Iowa and adjoining states that had occurred by the end of Tippecanoe deposition. The East-Central Iowa Basin is evident on this map (fig. 24), as well as the broad northward trending Northeast Missouri Arch.

The widespread distribution and relative uniformity in thickness of post-St. Peter Middle Ordovician stratigraphic units in east-central Iowa and northern Illinois suggests that no large-scale structural movements occurred along the trend of the Plum River Fault Zone during the Middle Ordovician. However, certain local stratigraphic anomalies suggest local structural activity. The local absence of the Pecatonica Member of the Platteville Formation in the area of the Forreston Dome on the south side of the Plum River Fault Zone in Ogle County, Illinois suggests "that the Forreston Dome may have been active early in Platteville time" (Kolata et al., 1978, p. 24). Templeton and Willman (1963, p. 137) suggested that broad structural movements in northeastern Iowa, southwestern Wisconsin, and northwestern Illinois, perhaps related to the Wisconsin Arch, may be responsible for the erosional truncation of Platteville strata beneath the Spechts Ferry Shale. Minor structural activity on the Oregon Anticline, a structural feature present north of the western terminus of the Sandwich Fault Zone and south of the eastern terminus of the Plum River Fault Zone in Illinois, was inferred by Kolata et al. (1978, p. 24) because the Guttenberg thins out over the anticline. The area along the Plum River Fault Zone was relatively stable tectonically during the remainder of Galena Group deposition, as evidenced by the widespread stratigraphic unity of the Galena rock package and the general uniformity of thickness of individual rock units within the Galena Group. Nevertheless, a shallow cratonic basin probably developed in northern Iowa and southern Minnesota during Galena deposition, as reflected on the Galena Group isopach map (Witzke, 1983a, p. 4). The Galena Group thickens eastward from central Iowa into east-central Iowa and northwestern Illinois, suggesting that development of the East-Central Iowa Basin was initiated during Galena deposition (Witzke, 1983a, p. 7).

The Galena Group was erosionally truncated around the Ozark Uplift in eastern Missouri and southwestern Illinois prior to Maquoketa deposition

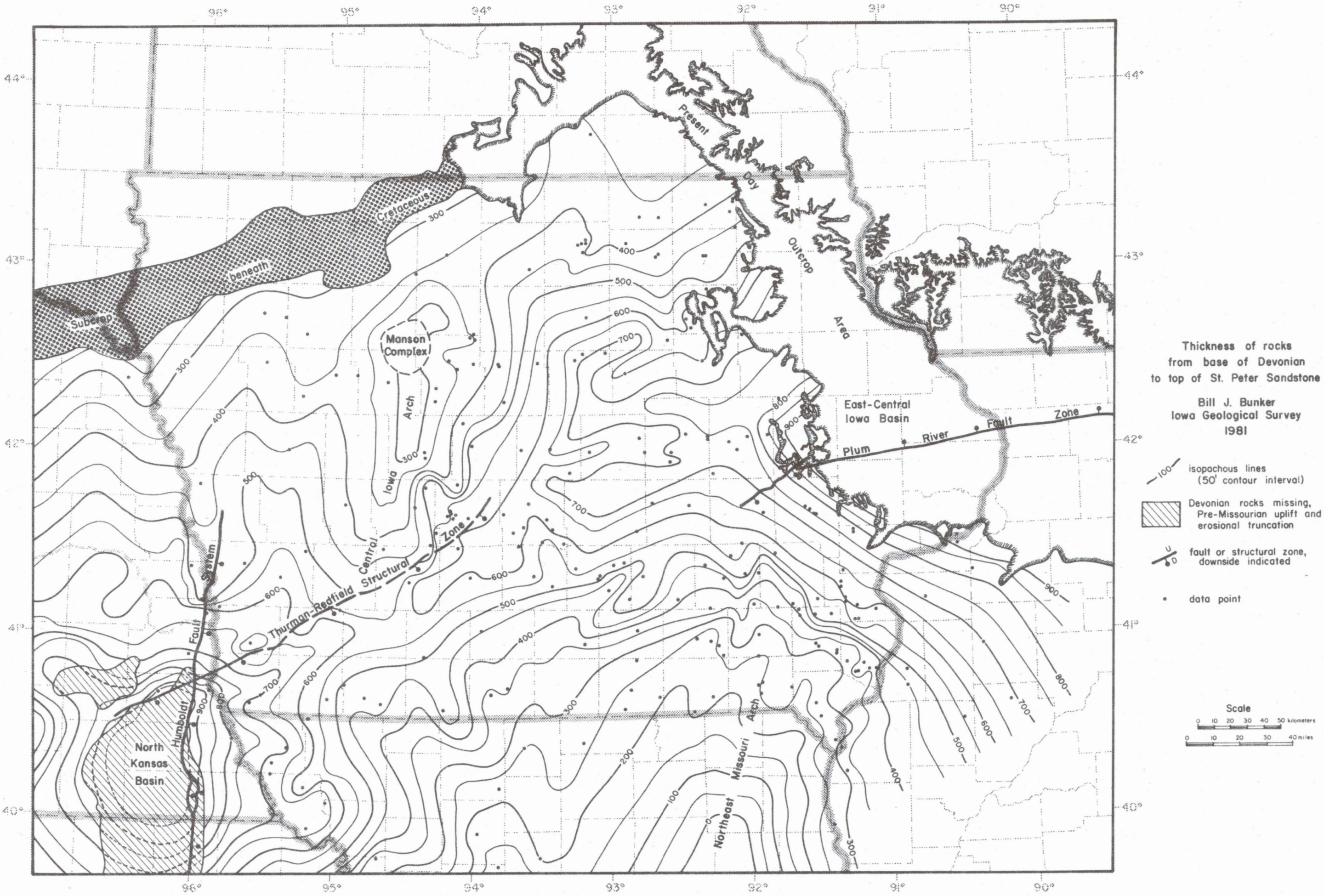


Figure 24. Isopach map of the Tippecanoe Sequence in Iowa and adjoining states (adapted from Bunker, 1982). Structural zones have been added to this map to show their general positions, although Tippecanoe faulting is not necessarily implied. However, the apparent coincidence of linear isopach trends and the position of known structural zones may indicate possible structural movements during Tippecanoe deposition.

(ibid.), but no erosional truncation of the Galena Group beneath the Maquoketa has been documented along the trend of the Plum River Fault Zone. Because of the large erosional unconformity separating Maquoketa and Silurian rock units in east-central Iowa and northern Illinois, Maquoketa isopach maps are difficult to interpret in a structural context. If the erosional relief on the Maquoketa surface is discounted, the Maquoketa Formation is relatively uniform in thickness in east-central Iowa and northwest Illinois, and the region is interpreted to have been relatively stable tectonically during Maquoketa deposition. However, Kolata and Graese (1983, p. 7) interpreted possible small-scale movements along the Wisconsin Arch and LaSalle Anticlinal Belt in northern Illinois contemporaneous with Maquoketa deposition. It is not known if the distribution of the various carbonate/shale facies in the Maquoketa was influenced structurally. The general restriction in Iowa of the "Brown Shaly Unit" to the east-central and southeastern portions of the state and the absence of Ft. Atkinson carbonates across much of east-central and central Iowa may possibly relate to the development of a subtle structural sag across east-central and central Iowa. Lower Maquoketa deposition in northern Iowa and Minnesota probably occurred in shallower water environments than in east-central Iowa and northwestern Illinois (Witzke, 1980a, p. 12), suggesting that the early stages of subsidence in the East-Central Iowa Basin influenced depositional patterns.

The isopach map of the total Silurian interval in the study area (fig. 25) clearly delineates an eastward-trending basin in east-central Iowa, the East-Central Iowa Basin. The thickest Silurian intervals in this basin presently occur within the Silurian outcrop belt. This pre-Middle Devonian basin was first recognized by Lee (1946) who noted "an eastward-trending basin in east-central Iowa." An extensive period of erosion following Silurian deposition removed Silurian rocks from much of the North American Midcontinent prior to the deposition of Middle Devonian carbonates, and the most complete sequences of Silurian rocks are found only in areas where they were structurally preserved (e.g., East-Central Iowa Basin). The preserved Silurian rocks were then covered and overlapped by Middle Devonian carbonates.

An east-west stratigraphic cross-section within the East-Central Iowa Basin (fig. 26) demonstrates a basinward thickening of individual Silurian stratigraphic units. The most dramatic thickening is an eastward doubling in thickness of the Hopkinton Formation. Therefore, the East-Central Iowa Basin can be considered a Silurian depositional basin. This observation runs counter to Johnson (1980, p. 200) who stated that the eastern Iowa post-Mosalem Silurian "formations in the Llandovery and basal Wenlock [i.e. Blanding, Hopkinton, and part of the Scotch Grove formations of this report] exhibit a relatively uniform thickness over their full range." The eastward thickening of the Blanding and Hopkinton formations into the central area of the East-Central Iowa Basin indicates that basinal subsidence was contemporaneous with Early Silurian deposition.

The maximum known thickness of Silurian rocks in Iowa, 480 feet (146 m), occurs adjacent to the Plum River Fault Zone in the Silurian outcrop belt of Jones County (Ludvigson and Bunker, 1978, p. 23). Since this thick Silurian sequence is not overlain by Devonian strata but was subjected to post-Devonian erosional truncation, Silurian thicknesses in excess of 480 feet (146 m) can be inferred in the central portion of the East-Central Iowa Basin prior to Middle Devonian deposition. Extrapolation of west-to-east thickening trends in the Silurian interval of eastern Iowa (see fig. 26) suggests that about 600 feet (180 m) or more of Silurian strata may have been present in portions of easternmost Iowa prior to the onset of Middle Devonian deposition. Unfortunately, extensive post-Devonian erosion has greatly reduced the thickness of the Silurian sequence in easternmost Iowa and removed all Middle Devonian rocks from the region except where preserved within grabens along the Plum River Fault Zone. The maximum thickness of the Silurian interval in Jackson-Jones counties, Iowa, can be more accurately established once holes are drilled and/or seismic profiles are run within the grabens containing Middle Devonian strata.

The distribution and thickness of Silurian rocks in the study area was strongly influenced by post-Silurian, pre-Middle Devonian structural development and post-Silurian erosion. Therefore, the Silurian isopach map does not necessarily reflect the precise position of the East-Central Iowa Basin center

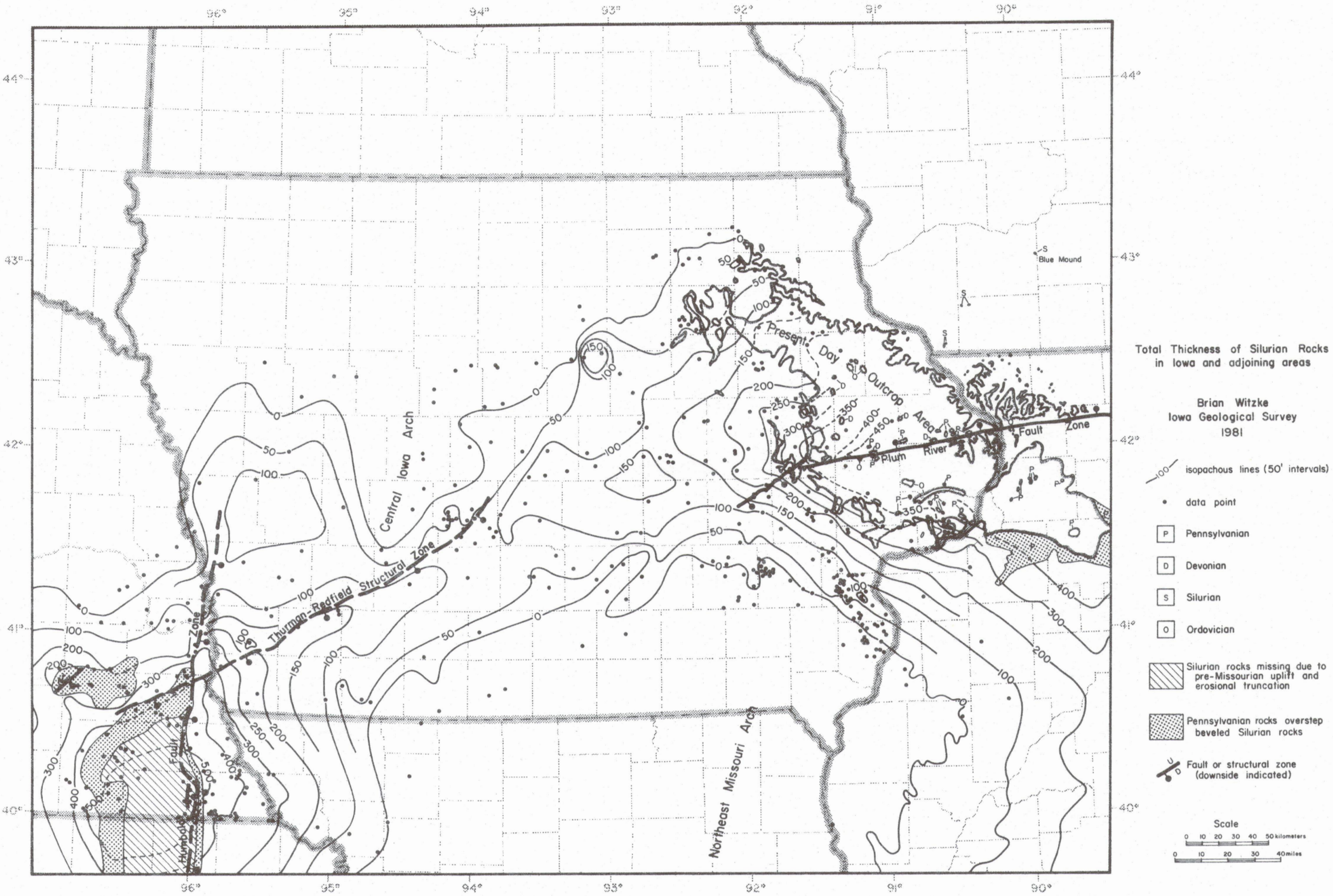


Figure 25. Isopach of the total Silurian System of Iowa and adjoining states (modified from Witzke, 1981a, b).

during various phases of Silurian deposition. Delimiting the eastern margins of the Silurian depositional basin is seriously impaired by extensive erosional truncation of Silurian rock units in northeastern Iowa, northwestern Illinois, and southwestern Wisconsin. The lowermost portion of the Silurian sequence (Mosalem, Tete des Morts, Blanding fms.) is recognizable throughout much of eastern Iowa and adjacent Illinois, and erosional outliers of this portion of the Silurian sequence are identified in southwestern Wisconsin, northwestern Illinois, and along the Silurian escarpment in northeastern Iowa. An isopach map of the Blanding Formation interval in the study area was constructed utilizing subsurface and surface control points (fig. 27), and the erosional outliers in Wisconsin and Illinois provided greater regional control outside of the main Silurian outcrop belt. The Silurian outlier at West Blue Mound (fig. 27) is 85 feet (26 m) thick (D. Mikulic, 1982, pers. comm.); beds containing *Pentamerus oblongus* occur at the top of the mound suggesting correlation with the Marcus Member of the Hopkinton Dolomite in Iowa. By

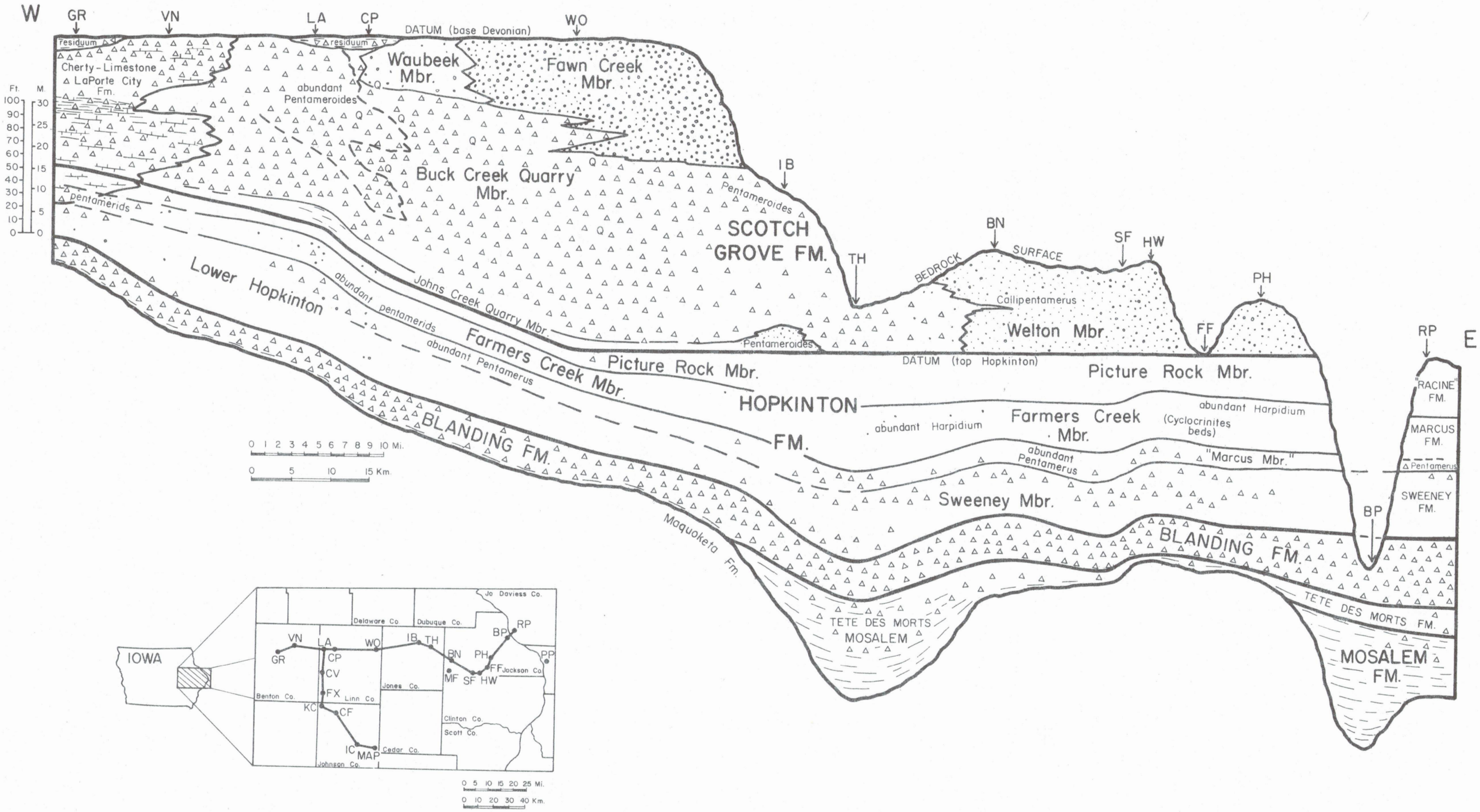


Figure 26. East-west stratigraphic cross-section of Silurian strata in east-central Iowa. Silurian section unconformably overlies Ordovician Maquoketa shale. Datum is base of Devonian in west; datum is lowered to top of Hopkinton Formation in east where Devonian rocks have been erosionally removed. Cross-section displays eastward thickening of Lower Silurian rock units into the East-Central Iowa Basin. Symbols: triangles (chert), small circles (abundant crinoid-moldic and replaced dolomitized wackestone and packstone fabrics), Q (quartz-lined vugs), discontinuous thin lines (argillaceous carbonates), hatchured lines (Silurian limestones of the LaPorte City Fm.).

analogy with the Iowa sequence, the sub-Hopkinton Silurian interval at West Blue Mound is probably 35 to 45 feet (11-14 m) thick. This suggests that the Blanding interval thins towards the northeast in southwestern Wisconsin. Closure of the 35-foot Blanding contour is tentatively drawn near West Blue Mound to reflect this thinning (fig. 28B).

The reconstructed Blanding isopach map (fig. 28B) delineates maximum thicknesses in northwesternmost Illinois, southwesternmost Wisconsin, and adjacent portions of eastern and northeastern Iowa. The thickest Blanding sections generally occupy a shallow basinal depression, the East-Central Iowa Basin, north of the Plum River Fault Zone. Therefore, during Blanding deposition the axis of the East-Central Iowa Basin occupied a position slightly north of the basin axis as delineated on the total Silurian isopach map (fig. 25). Thickening of the Blanding Formation north of the Fayette Structural Zone (figs. 28B, 4) suggests the presence of a structural sag during Blanding deposition in that area. The Blanding Formation thins westward in Iowa, although the formation is not readily distinguished in portions of central and northeastern Iowa where it is locally replaced by a non-cherty facies or by a portion of the Waucoma Limestone facies (fig. 28B; see also Witzke, 1981b).

Pre-Blanding Silurian deposition in the study area (i.e. Mosalem and Tete des Morts formations) was restricted to the central and northern portions of the East-Central Iowa Basin, and over most of Iowa the Blanding Formation rests directly on the Maquoketa surface (fig. 28A; see also Witzke, 1981b). Deposition of the Mosalem Formation, an argillaceous carbonate unit that filled in valleys and depressions on the Maquoketa Shale surface, and the overlying Tete des Morts Formation occurred only within the general confines of the East-Central Iowa Basin as Early Silurian marine transgression spread westward into Iowa from Illinois and/or Wisconsin. Although the Mosalem-Tete des Morts interval varies locally in thickness in response to variations in paleotopography on the Maquoketa surface (Brown and Whitlow, 1960; Willman, 1973), the thickest portions of this interval (50-135 ft; 15-41 m) generally occupy an area coincident with maximum Blanding thicknesses (fig. 28A).

The Middle Silurian (late Llandoverian-Wenlockian) development of the East-Central Iowa Basin is characterized by maximum subsidence parallel to the general trend of the present-day Plum River Fault Zone. A north-south stratigraphic cross-section cutting perpendicular to the axis of the western portion of the basin (fig. 29) illustrates a pronounced thickening of the Silurian interval along the northern edge of the Plum River Fault Zone. Of special note is the relative uniformity of thickness of the Blanding-Hopkinton interval along the cross-section line, and the Silurian thickening is primarily a function of thickening of the Scotch Grove Formation. In fact, the Scotch Grove interval doubles in thickness over a distance of only 10 miles (16 km) as one moves from northwestern Johnson into southwestern Linn County. These observations suggest that maximum subsidence in the western portion of the East-Central Iowa Basin occurred during Scotch Grove deposition. Unfortunately, post-Silurian erosion removed most evidence of the upper Scotch Grove and Gower sequence in the central area of the East-Central Iowa Basin, except in structurally preserved blocks along the Plum River Fault Zone, and it is not presently known if the Scotch Grove-Gower interval thickens into the basin center. The zone of abundant *Costistricklandia* within the Scotch Grove Formation, as illustrated on figure 29, can be interpreted either as a diachronous brachiopod association or as an essentially isochronous interval. If the *Costistricklandia* association is essentially isochronous, maximum Silurian subsidence can be identified as occurring in western Linn County during depo-

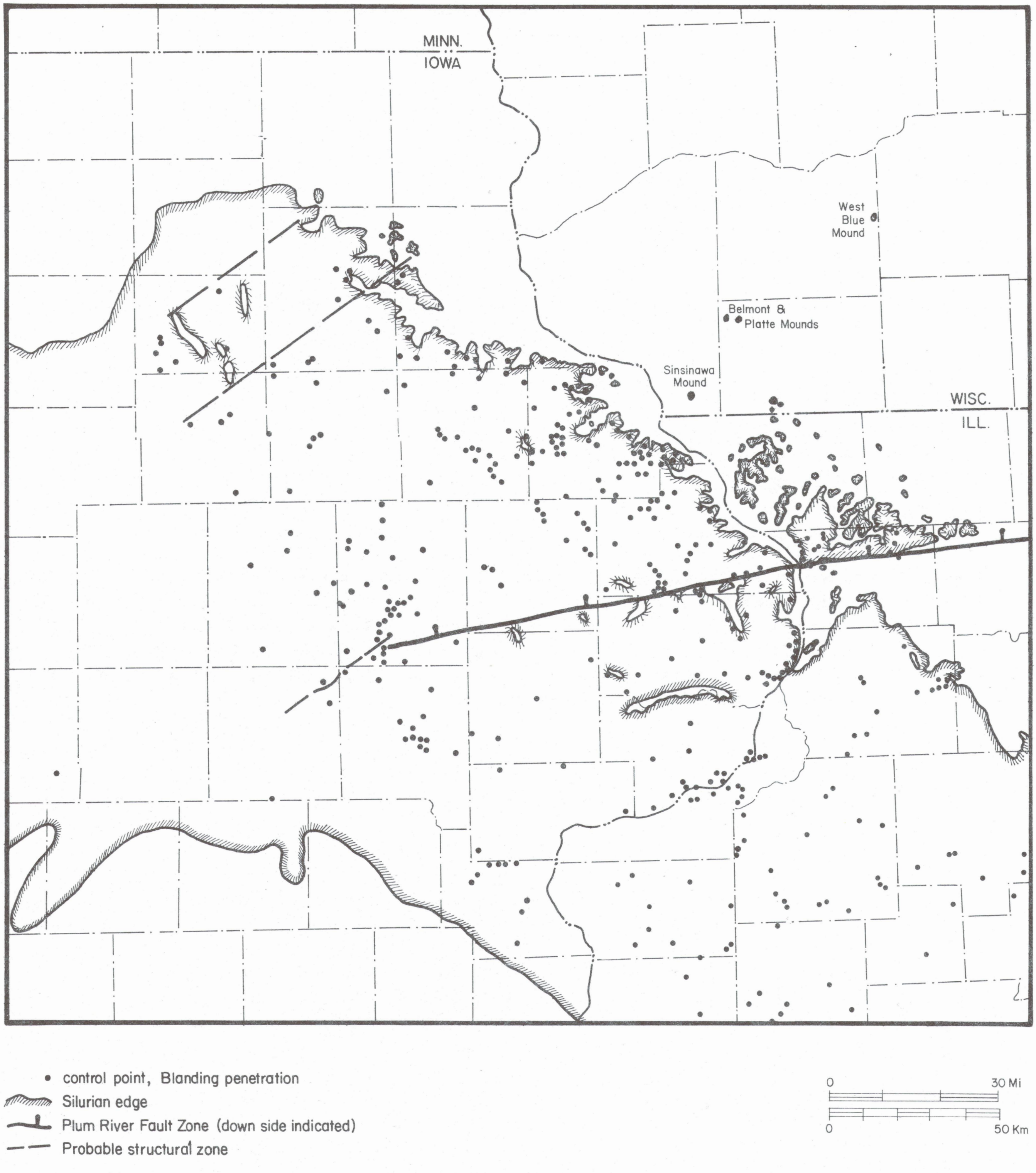


Figure 27. Lower Silurian data control map, eastern Iowa and adjacent Illinois and Wisconsin. Well records for individual control points on file at the Iowa and Illinois State Geological Surveys. Outcrop control data in part from Johnson (1977a), unpublished Iowa field studies, and Willman (1973).

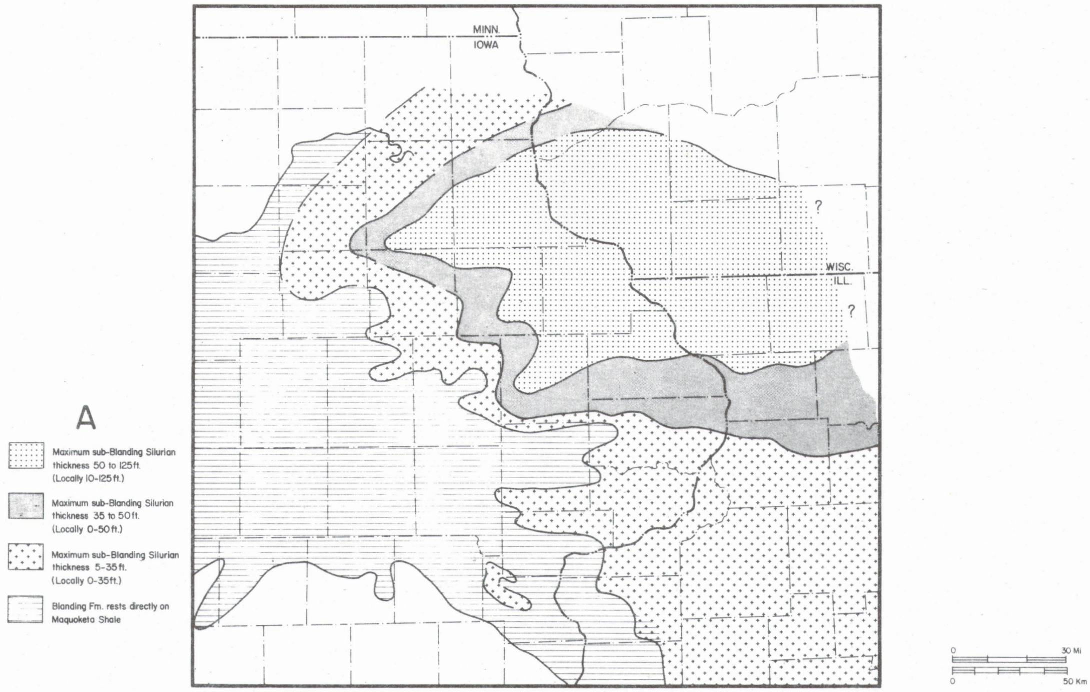


Figure 28. A. Generalized maximum sub-Blanding Silurian thicknesses (Mosalem and Tete des Morts fms.). Due to significant local variations in sub-Blanding Silurian thicknesses, largely a function of sub-Silurian relief on the Maquoketa shale surface, only the locally thickest intervals were utilized in constructing the map. Base map and data control shown in figure 27.

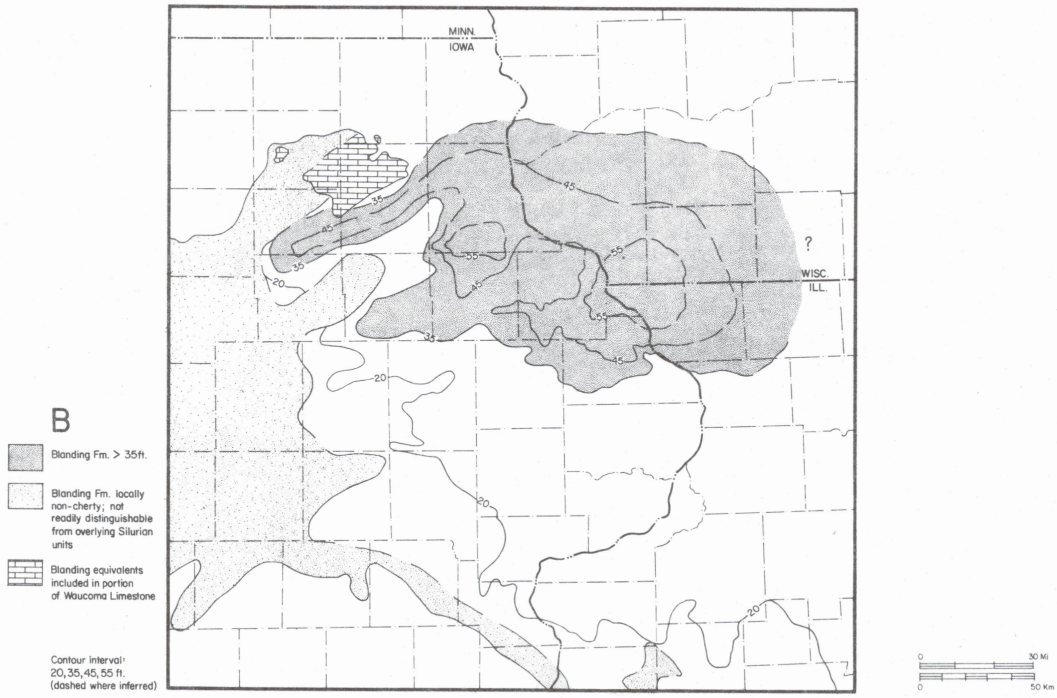


Figure 28. B. Generalized Blanding Formation isopach map (in feet). Thickest Blanding intervals delineate central area of East-Central Iowa Basin. Base map and data control shown in figure 27.

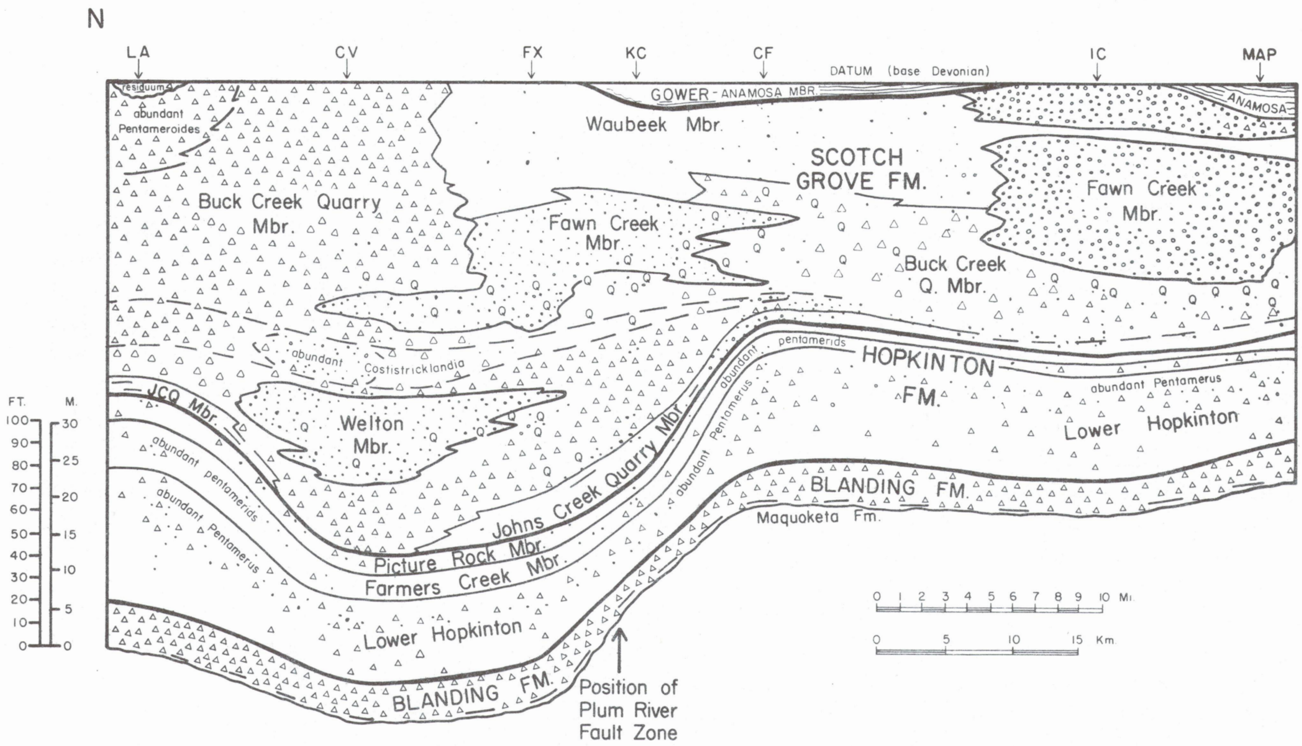


Figure 29. North-south stratigraphic cross-section in western Linn and Johnson counties, Iowa. Cross-section lines shown on figure 26. Symbols same as in figure 26; squiggly lines are laminated dolomites of the Gower Formation (Anamosa Facies). Thickening of Scotch Grove Formation north of the Plum River Fault Zone suggests that maximum Silurian subsidence in the western portion of the East-Central Iowa Basin may have occurred during Scotch Grove deposition (adapted from Witzke 1981a, b).

sition of the lower Scotch Grove Formation. An eight-fold increase in thickness in the Scotch Grove interval beneath the *Costistricklandia*-bearing zone is noted between northwestern Johnson and southwestern Linn counties. It may be coincidental in this interpretation, but maximum structural subsidence in western Linn County apparently occurred coeval with maximum marine transgression, as interregional interpretations of Johnson and Campbell (1980) suggest peak transgression in latest Llandoverian, which is coincident with lower Scotch Grove deposition.

The Gower Formation, where preserved, is restricted to the general confines of the East-Central Iowa Basin in Jackson, Jones, Linn, Clinton, Cedar, Johnson, and Scott counties. The deposition of laminated unfossiliferous Gower carbonates in eastern Iowa and the presence of evaporite crystal molds in the Gower (Henry, 1972) contrasts markedly with the underlying Scotch Grove-Hopkinton carbonates with an abundant marine fauna. Henry (1972, p. 78) suggested that the laminated Gower carbonate sequence "was deposited under unusual, possibly high saline, conditions in which a normal marine fauna could not develop." Philcox (1972, p. 701) interpreted the depositional environment of the laminated Gower carbonates as one of "low energy," "hostile to most organisms," perhaps in a situation of "restricted circulation" that "led to high salinities." What conditions in eastern Iowa might be responsible for the change from "normal" marine deposition in the upper Scotch Grove Formation to possible hypersaline deposition in the Gower? A change in circulation patterns needs to be invoked, and a barrier to open circulation is undoubtedly necessary. Witzke (1981a, p. 17) proposed "that the progressive regression of the seas during the Middle and Late Silurian left central Iowa emergent at the beginning of Gower deposition, and open circulation across the carbonate shelf was thereby cut off" leaving "east-central Iowa as a restricted embayment of the Silurian sea." Additionally, "carbonate buildups and skeletal/mud banks of the LeClaire Facies [i.e. LeClaire Member] in the eastern portion of the East-Central Iowa Basin may have served to attenuate open marine circulation between Illinois and eastern Iowa" (Witzke, 1981a, p. 21). These interpretations point out the influence of the East-Central Iowa Basin on Gower deposition.

Silurian structural history over the entire extent of the East-Central Iowa Basin has not been worked out in detail, although several pertinent points need to be stressed. The eastward thickening of individual stratigraphic units in the Blanding-Hopkinton interval of eastern Iowa delineates the central region of a Silurian stratigraphic basin. As evidenced by the Blanding isopach map (fig. 28B) and the distribution of sub-Blanding Silurian units (fig. 28A) in the study area, the axis of the East-Central Iowa Basin occupied a position 30 to 45 miles (48-72 km) north of the general trend of the Plum River Fault Zone during the Early Silurian (early to mid Llandoverian). However, during the Middle and Late Silurian (late Llandoverian-Ludlovian) the axis of thickening in the East-Central Iowa Basin apparently shifted southward to occupy a position adjacent to or coincident with the trend of the Plum River Fault Zone. The eastern portion of the basin in east-central Iowa represents the area of maximum pre-Middle Devonian subsidence, as reflected by the total Silurian isopach map. Maximum Silurian subsidence in the eastern half of the basin is coincident with the present-day Plum River Fault Zone. Maximum vertical displacements along the Plum River Fault Zone presently occur in the same general region where maximum Silurian subsidence is noted. Moving westward in Linn County, vertical displacement along the Plum River Fault Zone decreases until the fault zone ultimately disappears in northeastern Iowa County. In Linn County maximum Silurian subsidence is noted sub-parallel to the Plum River Fault Zone on the north side of the fault. Although there is no evidence of actual faulting along the Plum River Fault Zone during deposition of the Silurian carbonates, there is close correlation between trends of basinal subsidence, as shown by isopach and facies variations in the Silurian and later development of the Plum River Fault Zone. This suggests that deep-seated crustal features may have controlled recurrent structural movements along the trend of the Plum River Fault Zone, and these movements have been variably expressed as epeirogenic downwarping (basinal subsidence and faulting).

Kaskaskia Tectonic History

The Kaskaskia Sequence in the central midcontinent region consists of strata that rest upon the interregional unconformity developed on Tippecanoe and older rocks, and that underlie an interregional unconformity at the base of the overlying Absaroka Sequence (fig. 3). Sloss (1963) defined the age of this sequence of rocks as ranging from late Early Devonian to latest Mississippian (Chesterian). Lower Devonian rocks are not recognized in Iowa, and Chesterian(?) continental sediments are present only as stratigraphic leaks into the older Kaskaskia rocks (Table 1).

Prior to the basal Kaskaskia (Middle Devonian) transgression into Iowa, an extensive period of erosion stripped several hundred feet of Sauk and Tippecanoe rocks from the north-central midcontinent region. Figure 30 is a paleogeologic map showing the distribution of Tippecanoe rocks subcropping beneath the Kaskaskia Sequence. The Northeast Missouri Arch has expression in the southern part of the map area as noted by the regional truncation of Tippecanoe rocks. Silurian rocks occupy a northeast to southwest trending trough-like depression across the central part of Iowa. The Silurian isopach map (fig. 25) as discussed earlier subdivides this depressed area into two structural/depositional basins (Witzke, 1981b): the North Kansas Basin (Rich, 1933) and the East-Central Iowa Basin (Bunker, 1981, p. 6; Witzke, 1981a, b). The youngest Silurian rocks in Iowa are preserved within these basins, marking the areas of maximum pre-Middle Devonian subsidence.

Collinson and James (1969) have referred to the Middle Devonian rocks of eastern Iowa and northwestern Illinois as the southeasternmost transgressive deposits of a vast seaway that extended northwestward into western Canada. McCammon (1960, p. 23) also concluded that much of the Dawson Bay macrofauna of the Williston Basin area is comparable to that of the lower Cedar Valley Formation in eastern Iowa, suggesting partial or direct seaway connections across the Transcontinental Arch. Faunal similarity with age equivalent rocks of the Traverse Group in the Michigan Basin area is also suggestive of partial or direct seaway connections to the east across the Wisconsin Arch. Norris et al. (1982) recognized a series of "distinct depositional cycles" in the Devonian sequence of Manitoba. Witzke and Bunker (1984) also noted a series of depositional cycles in the Devonian of northern Iowa that appear to relate in timing and development to those described in Manitoba. These major cycles of deposition can be traced as far south as central Benton County, Iowa, but extrapolation of these cycles into the Johnson County, Iowa, sequence of the Cedar Valley Formation must await further study. However, a general discussion of the Middle Devonian depositional framework in eastern Iowa follows.

Areas of pre-Kaskaskia erosional topographic relief developed on the Tippecanoe surface as a result of lithologic variations between the Silurian carbonates and the Upper Ordovician Maquoketa shales of eastern Iowa. Erosional escarpments, similar to the present day Niagaran Escarpment (Prior, 1976, p. 30) of northeastern Iowa, developed at the pre-Kaskaskia erosional margins of the Silurian (Bunker et al., 1983), and served as effective barriers to open-marine circulation during the initial transgression of the Middle Devonian seas. A stratigraphic profile drawn using the top of the Wapsipinicon Formation as a datum (fig. 31) illustrates this erosional surface and shows a general cross-section view of the East-Central Iowa Basin.

The present geographic distribution of the Bertram Formation (fig. 32), the lowermost Kaskaskia unit in Iowa, is limited to a small region near the axial center of the subsident area that represents the East-Central Iowa Basin. Pre-Absaroka (Pennsylvanian) uplift and erosion has limited the areal distribution of the formation and made an interpretation of its relationship to overlying and underlying geologic units difficult to resolve. Petrographic analysis by Sammis (1978) suggested that the Bertram was formed in "a very restricted nearshore to terrestrial environment with rapidly fluctuating conditions," and caliche fabrics noted in the unit suggest meteoric-vadose

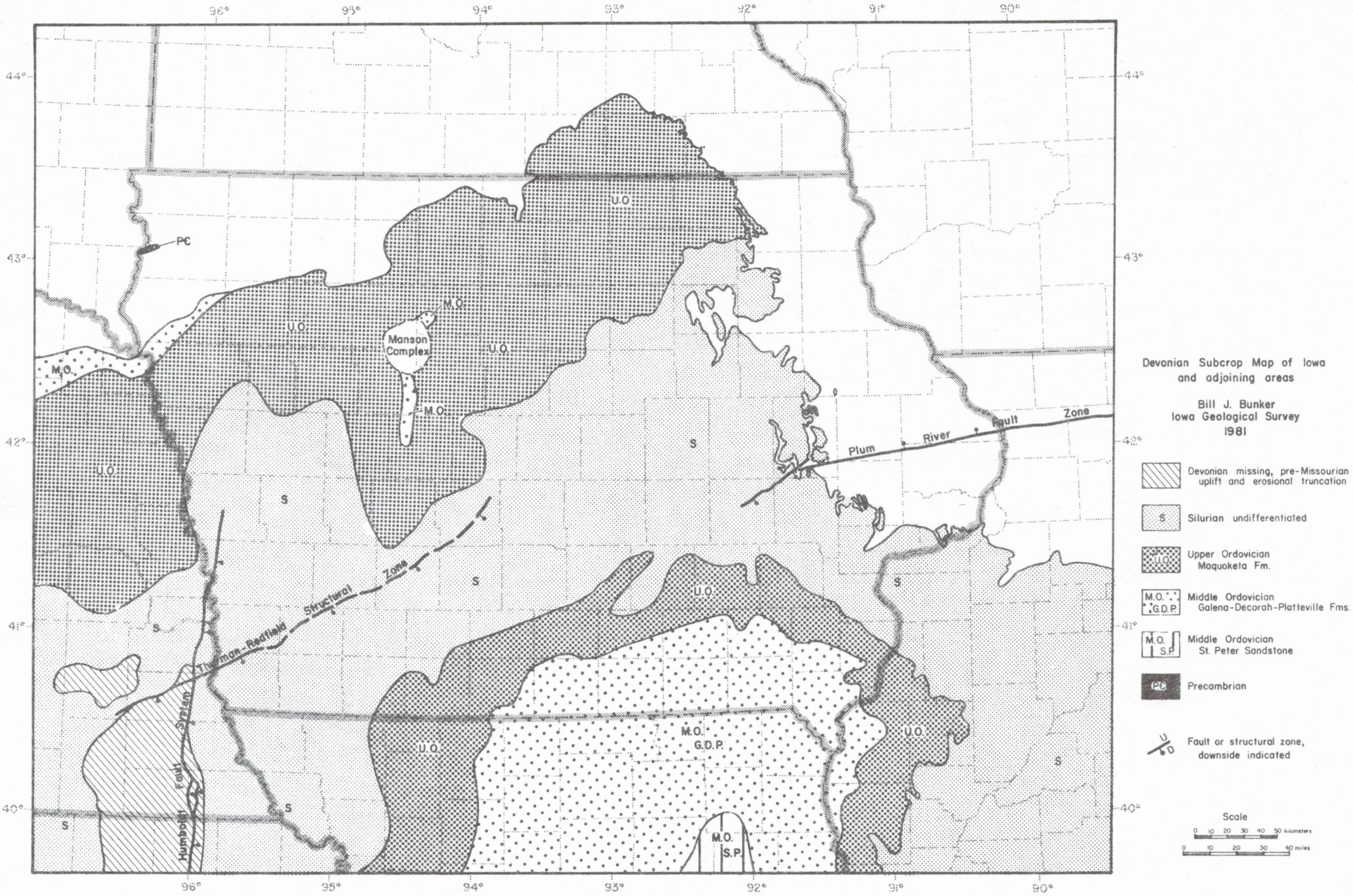


Figure 30. Paleogeologic map of the pre-Kaskaskia surface of Iowa and adjoining states (adapted from Bunker, 1982). Structural zones have been added to this map to show their general positions, although faulting is not implied at any particular point in time. However, the apparent coincidence of geologic patterns and the positions of known structural zones may indicate possible structural movements prior to Kaskaskia deposition.

calcite cementation in a deposit which accreted in a topographic depression developed on the Silurian surface. Regional thickening of the Bertram (fig. 32) along the trend of the basin axis, and the unconformable relationship with the underlying Silurian rock units supports this interpretation. The depositional basin of the Bertram Formation displays a pronounced asymmetry (fig. 32), with the greatest thickness noted immediately to the north (downthrown side) of the Plum River Fault Zone. The Bertram is absent on the south (upthrown side) of the fault (figs. 31 and 32). This lithostratigraphic re-

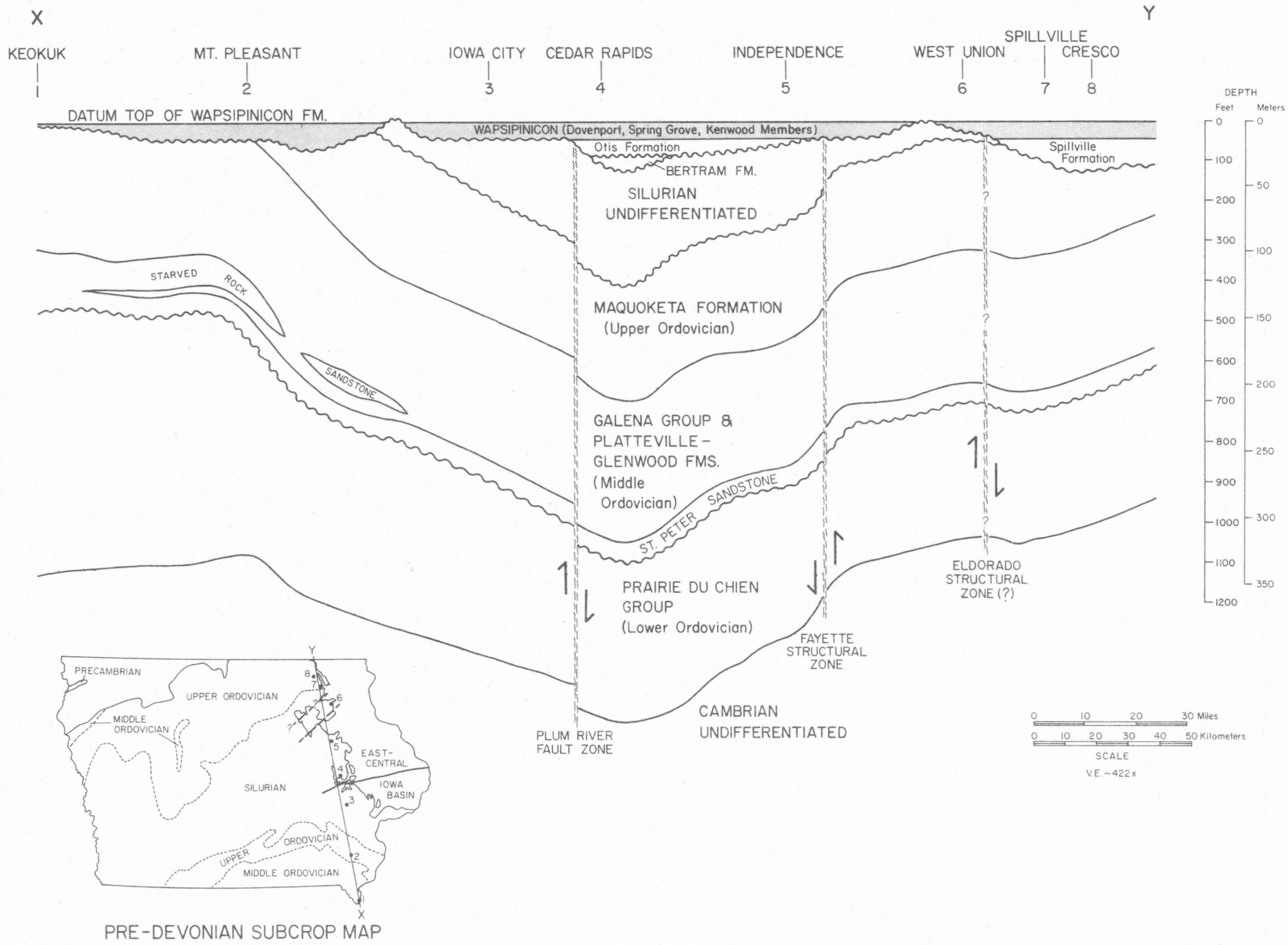


Figure 31. North-south stratigraphic cross-section, utilizing the top of the Wapsipinicon Formation (Middle Devonian) as the datum. The East-Central Iowa Basin is evident as a pre-Kaskaskia structural feature on this profile line.

lationship suggests faulting penecontemporaneous with Bertram sedimentation, and the depositional basin is postulated to have been an actively subsiding half-graben. Thus, both erosionally and tectonically controlled topography are believed to have been major controlling factors on initial Kaskaskia deposition in the East-Central Iowa Basin area.

The Otis-Spillville formations represent the initial marine deposits of the transgressing Middle Devonian seas in eastern Iowa. Examination of the stratigraphic profile (fig. 31) illustrates the general relationship of the Otis Formation to the East-Central Iowa Basin. The lateral restriction of Otis strata within the confines of the basin, can be noted along the profile line. The Otis isopach map (fig. 33) shows eastward thickening in the direction of the basin center. Examination of the profile line (fig. 31) also shows the presence of a topographic barrier (Bremer High, see fig. 33) that

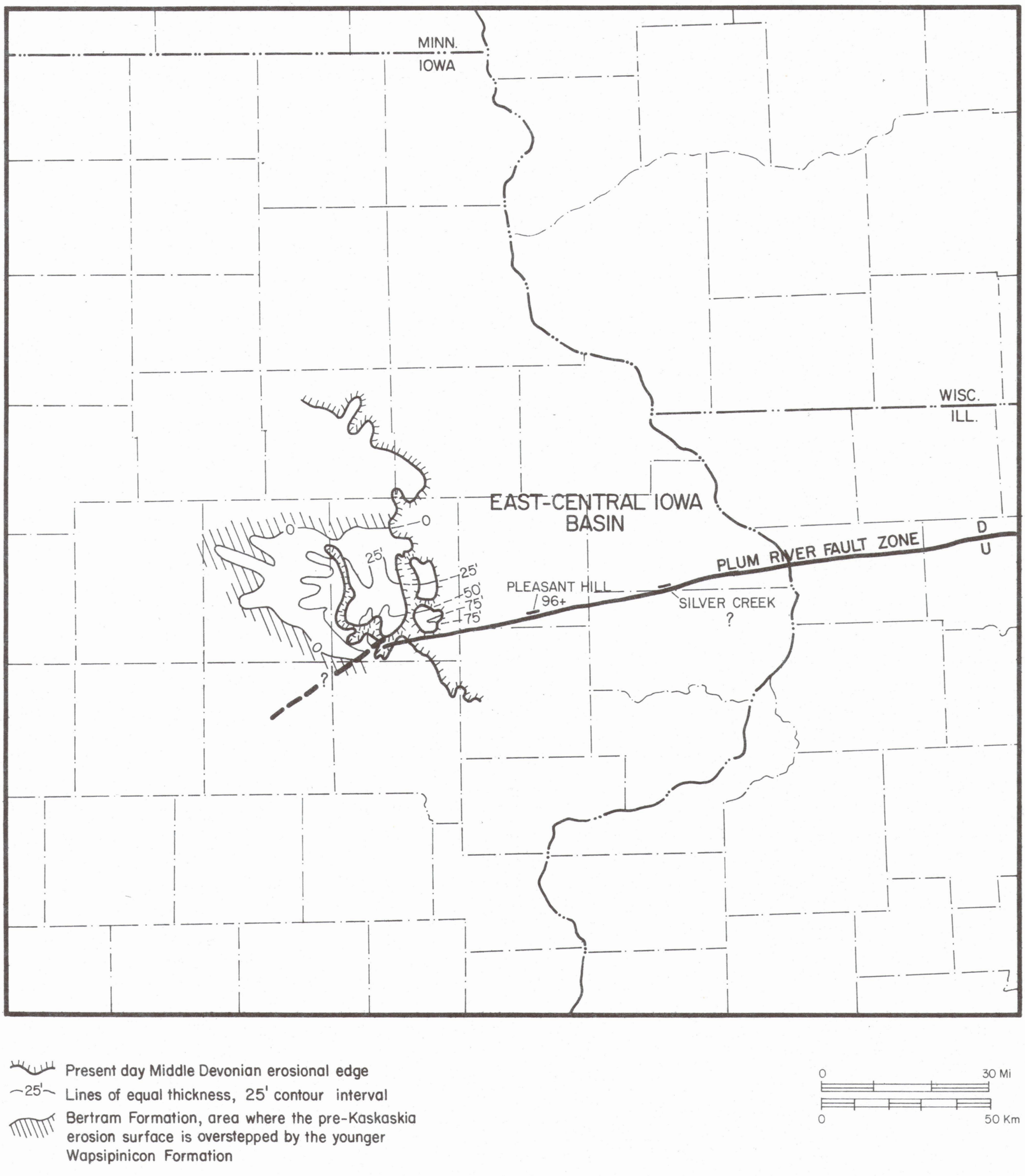


Figure 32. Isopach of the Bertram Formation in east-central Iowa (modified from Church, 1967; Ludvigson et al., 1978, p. 25).

separates the Otis of east-central Iowa from the Spillville Formation of northeastern Iowa. This topographic barrier, an erosional escarpment of Silurian carbonates, was progressively buried by the southward overlap of Middle Devonian strata. An excellent example of the Middle Devonian overlap and progressive burial of paleotopographic highs along the crest of the escarpment has been described by Dorheim and Koch (1962) in the Loomis Quarry (NW sec. 29, T91N, R13W, Bremer County, Iowa). The Bremer High served as an effective barrier to the initial transgression of the Otis-Spillville seas; open marine conditions with a diverse biota characterized the region north of the barrier (Spillville) , whereas more restricted marine conditions with a low-diversity fauna characterized Otis deposition in the East-Central Iowa Basin area.

The Wapsipinicon Formation, which overlies Otis and Spillville strata, is characterized by a sequence of unfossiliferous and evaporitic rocks. Wapsipinicon sediments were apparently deposited in shallower and/or more restricted environments than the fossiliferous carbonates of the Otis-Spillville formations. This suggests that the Wapsipinicon Formation represents the regressive phase of a major transgressive-regressive depositional cycle in eastern Iowa. Although the Wapsipinicon Formation probably represents a shallowing depositional sequence, the apparent regional overlap of Otis-Spillville strata by the Wapsipinicon, as illustrated in figure 31, is not presently well understood. Progressive burial of the northern Silurian escarpment (Bremer High) and encroachment upon a similar escarpment (Washington-Louisa High, see fig. 33) along the southern margin of the basin is illustrated. These escarpments may have played an effective role in the restriction of water circulation during deposition of the Wapsipinicon gypsum-anhydrite evaporite sequence in eastern Iowa. Prior to the onset of Cedar Valley deposition, Wapsipinicon strata may have been subjected to subaerial exposure and freshwater diagenesis as the Wapsipinicon sea regressed. Partial-to-complete dissolution of the gypsum-anhydrite evaporite interval within the Wapsipinicon (Fayette Breccia, Norton, 1920) occurred within the East-Central Iowa Basin area prior to and contemporaneous with the initial transgression of the Cedar Valley sea.

Deposition of the Solon Member of the Cedar Valley Formation marked the beginning of the next major depositional cycle in eastern Iowa. The Solon closely coincides with the initial phases of the Taghanic transgression (Johnson, 1970; Klapper & Johnson, 1980), which marked the end of provincialism among brachiopods, corals, and trilobites across the North American continent during the late Middle Devonian. Southward overlap of the Wapsipinicon Formation by the Cedar Valley Formation has been identified and mapped in central Illinois by James (1968), and Collinson and Atherton (1975), documenting regional expansion of the Cedar Valley seaway. Kettenbrink (1973, p. 162-164) summarized the following depositional framework for the lower Cedar Valley: 1) fine sand (Hoing Sandstone) and fragments of the underlying Davenport calcilutite were incorporated into the basal Cedar Valley sediments (basal Solon); 2) continued expansion of the seas during the Solon resulted in deposition of fine- to medium-grained skeletal calcarenites with local development of coral or stromatoporoid rich biostromes or shell banks; and 3) slightly deeper water conditions with an increase in the influx of argillaceous material prevailed during the transgressive phase of the lower Rapid Member.

Although the entire Cedar Valley Formation has generally been considered as a single depositional cycle in Johnson County (Kettenbrink, 1973), the recognition of a series of depositional cycles in the Cedar Valley interval of

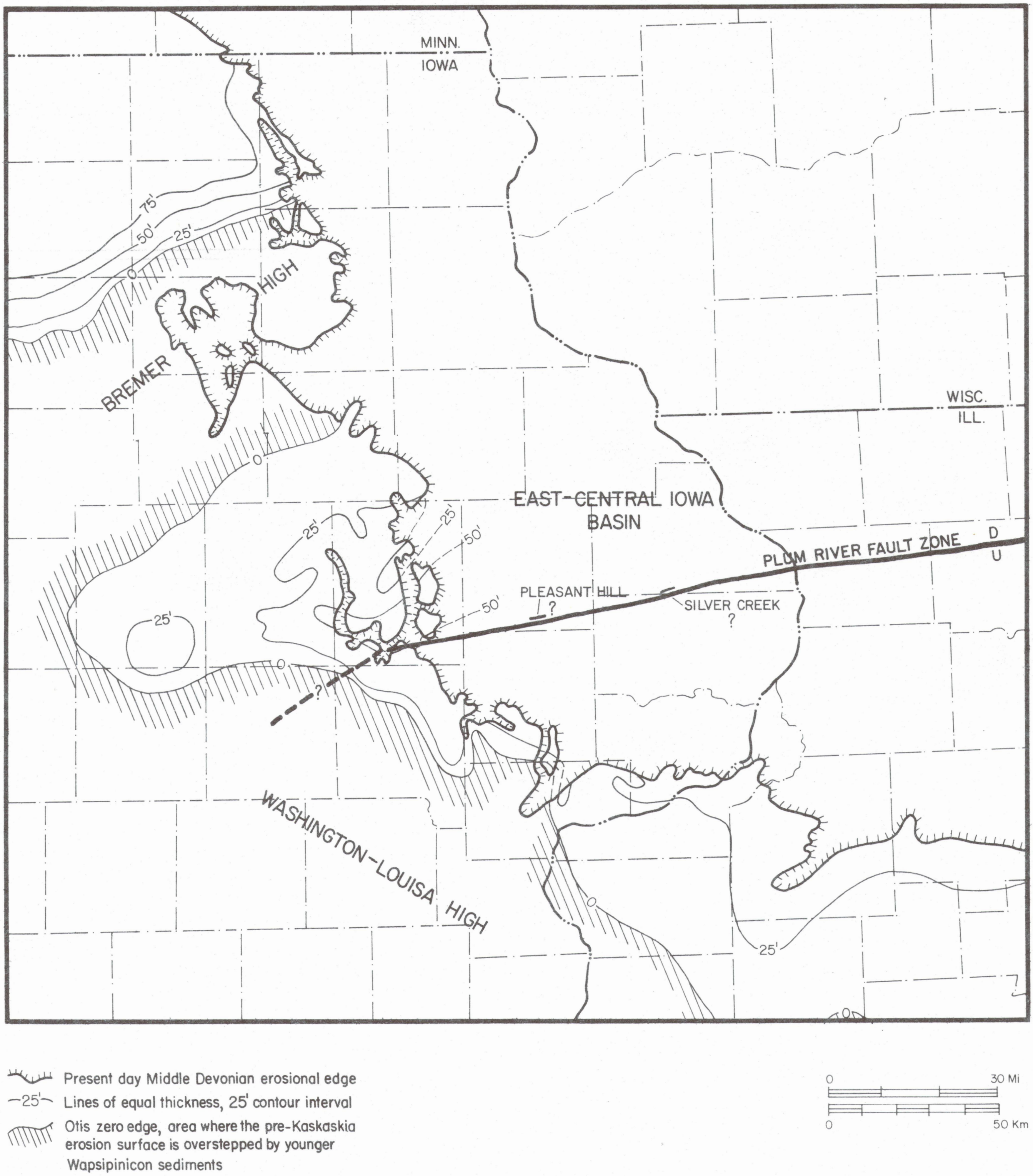


Figure 33. Isopach map of the Otis-Spillville formations in east-central and northern Iowa.

northern Iowa (Witzke and Bunker, 1984) suggests that a re-evaluation of the Coralville and upper Rapid members is in order. Unfortunately, as commented upon previously, precise correlation of the northern Iowa depositional sequence with that in Johnson County must await future studies. A prominent discontinuity surface, however, marks the Rapid-Coralville contact in Johnson County. This discontinuity surface is frequently burrowed and is characterized by a sharp change in lithology from argillaceous calcilutites to skeletal calcarenites of the Lower Coralville. It is generally agreed that the Coralville was deposited during a general regressive interval, and progressively shallower depositional environments are documented upward in the sequence.

A period of post-Cedar Valley erosion and karstification in eastern Iowa and adjacent Illinois ensued prior to the transgression of Upper Devonian shales into the midcontinent region. In places, sinkholes and caverns developed in Middle Devonian and Silurian strata, and stratigraphic leaks of the Upper Devonian Independence Shale infilled these karst features.

Interpretation of the Upper Devonian-Mississippian depositional history in the East-Central Iowa Basin area is difficult to reconstruct because of pre-Absaroka uplift and erosional beveling of these strata from the basin area. However, isopach mapping of the Upper Devonian New Albany Shale Group in Illinois (Cluff et al., 1981, p. 9) indicates gentle northwestward thickening into southeastern Iowa of these rocks. This general trend of Upper Devonian thickening in southeastern Iowa is suggestive of possible southward or southwestward shifting of the depositional center during the Middle to Late Devonian. Workman and Gillette (1956) referred to the area of maximum Upper Devonian thickening in southeast Iowa and adjacent Illinois as the Petersberg Basin. The Lincoln Fold System of northeastern Missouri-southeastern Iowa and the Wisconsin Arch of northern Illinois served as bounding structural features along the western and eastern margins of the basin respectively.

The Late Mississippian (Chesterian) records a major episode of regression from the continental interior. Erosional beveling and karstification of the earlier sedimentary sequences occurred regionally across the continental interior during this period. Stratigraphic leaks of a dark gray shale of continental derivation containing Late Mississippian (Chesterian?; Urban, 1971, 1972) or Early Pennsylvanian (Morrowan; R. Ravn, pers. comm.) spores have been noted interspersed with stratigraphic leaks of Upper Devonian (Independence Shale) marine sediments in eastern Iowa.

Absaroka Tectonic History

The Absaroka Sequence includes strata of latest Mississippian to Early Jurassic age (Sloss, 1963; fig. 3). In the continental interior, much of this sequence records a series of cyclic repetitions of marine and nonmarine sediments (cyclothems). Erosional remnants of Lower to Middle Pennsylvanian rocks are all that remain of the Absaroka Sequence in the east-central Iowa and the extreme northwestern Illinois area. These rocks consist of a succession of fluvial sandstones, conglomerates, siltstones, shales, and coal that were deposited directly on Upper Ordovician, Silurian, and Devonian rocks along the trend of the Plum River Fault Zone and regionally across the area.

The pre-Pennsylvanian uplift and erosional destruction of the East-Central Iowa Basin marked a profound change in the tectonic setting of eastern Iowa and northwestern Illinois. Movements leading to the destruction of the basin center may have been initiated during the Late Devonian and climaxed during the Mississippian. Contemporaneous periods of basinal deformation occurred in the North Kansas Basin area (Lee, 1939, 1943, 1946, 1956; Lee et al., 1948), which led eventually to the development of the Nemaha Uplift, and Forest City and Salina basins by the Middle Pennsylvanian.

Post-Kaskaskia deformation along the trend of the Plum River Fault Zone is evident by the preservation of Upper Silurian and Middle Devonian rocks along the northern downthrown side, and in grabens (Silver Creek, Pleasant Hill, and Skvor-Hartl) internal to the fault zone. The occurrence of Upper Silurian beds in extreme east-central Iowa was first commented upon by Savage (1906, p. 619-621), who described beds of the Gower Formation in west-central Jackson County. Savage originally described these beds in stratigraphic position above coralline Hopkinton beds; based upon present stratigraphic interpretations, these "Gower" outcrops should probably be assigned to the basal part of the newly defined Scotch Grove Formation. However, recent investigations (Baik, 1980; Chao, 1980) of the Plum River Fault Zone in southern Jackson County have established the existence of beds of the Gower Formation in the area. Baik (1980, p. 44-50) described outcrops of the laminated Anamosa Member (SW SW sec. 26, T84N, R3E) and the brachiopod-rich Brady Member (NW NE SW NW sec. 22, T84N, R3E) in close association with the northern downthrown side of the fault zone. Chao (1980, p. 22) also noted the occurrence of the laminated Anamosa Member associated with complex graben faulting (SE SE NE sec. 28, T84N, R4E) within the fault zone.

Numerous outliers of Pennsylvanian channel sandstone and shales have been reported across east-central Iowa (Appendix 1; fig. 34). Their scattered preservation and topographic relationships with older rocks suggests that they represent the deepest remnants of fluvial incision into the pre-Absaroka terrane. The distribution of these outliers with respect to the underlying rock units and the trend of the Plum River Fault Zone provides key information concerning pre-Pennsylvanian tectonism in the area.

Along the trend of the Plum River Fault Zone, the Pennsylvanian sandstone outliers in the southwestern corner of Fairfield Township (T84N, R4E) south-central Jackson County, are of particular interest. In this area channel sandstones are exposed in close proximity to both sides of the fault zone. On the southern upthrown side of the fault zone, these sandstones can be observed resting on Upper Ordovician (Maquoketa Shale) rocks in the NW SW corner of section 29 and the N 1/2 of section 32 (Savage, 1906; Chao, 1980). Structure contour mapping on top of the Blanding Formation (fig. 11) reveals that these Pennsylvanian sandstones occur along the crest of a doubly plunging anticline that borders the fault zone on the south. Pre-Absaroka erosional stripping of Silurian rocks along the crest of the anticline exposed the Maquoketa Shale on which the Pennsylvanian sandstones rest.

To the north a few hundred meters, on the downthrown side of the fault zone (ctr. of the N 1/2 of sec. 29), exposures of Pennsylvanian sandstone have also been noted (Chao, 1980) lying in near proximity (50 to 200 meters) to exposures of the Scotch Grove Formation. Structure contour mapping of the Blanding Formation (fig. 11) reveals that the Silurian rocks in this area occupy an east-west trending syncline bordering the north side of the fault zone. Complex block faulting involving graben development occurs locally in this area (Chao, 1980). Chao (1980, p. 24) noted a sandstone exposure along the northern boundary of the fault zone which dips 10° to the south. However, the limited areal extent of the exposure makes it difficult to determine whether the rotated block is the result of post-depositional erosional undercutting, or, possibly the result of post-depositional tectonic movement along the fault zone. Chao (1980, p. 24) also noted a few discontinuous flat-lying

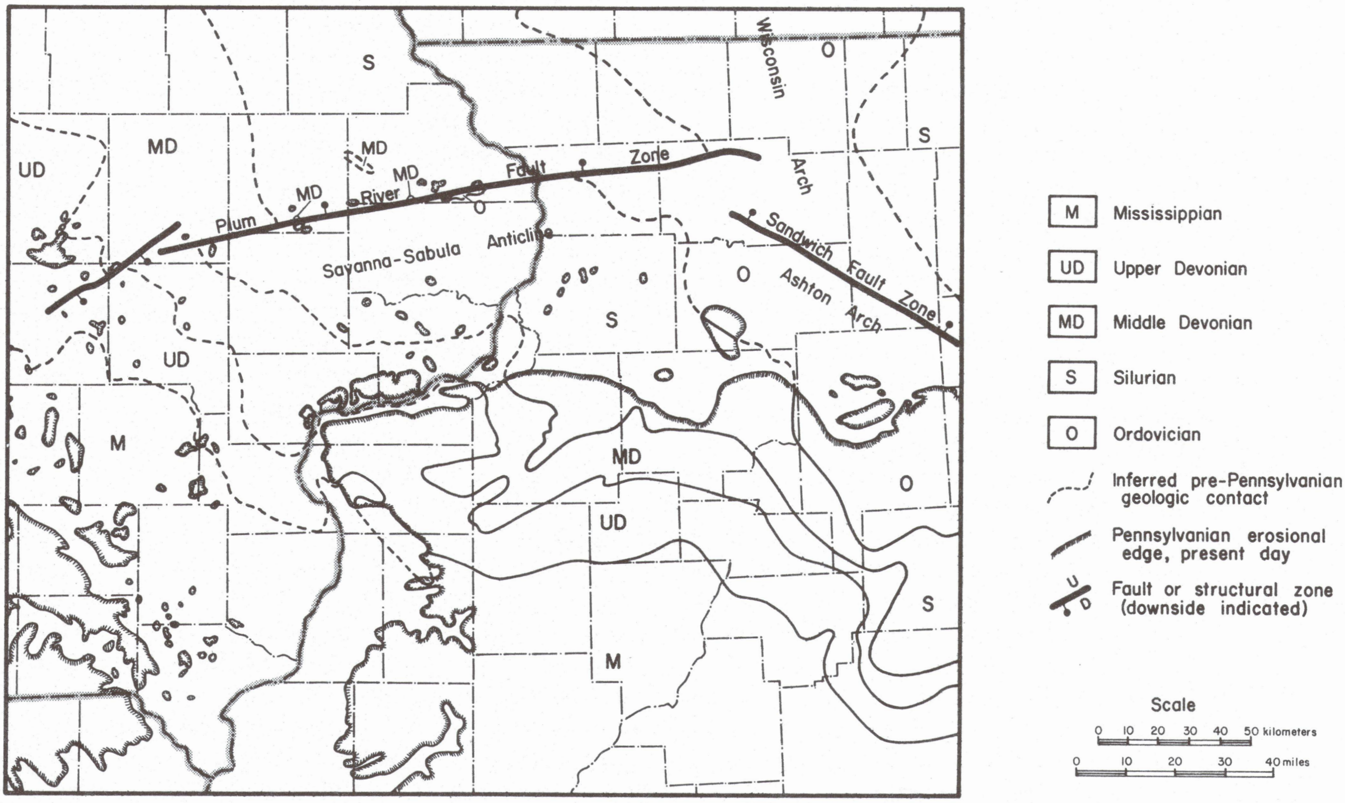


Figure 34. Paleogeologic map of the pre-Pennsylvanian surface of east-central Iowa and northwestern Illinois. Reconstruction based on the numerous Pennsylvanian outliers scattered across the study area (in part summarized in Appendix 1).

conglomeratic sandstone beds exposed near an outcrop (SE SE NE of sec. 28) of the Anamosa Member. Beds of the Anamosa Member are preserved here because of complex graben faulting and are dipping 40 to 50° to the southeast. The Maquoketa Shale is exposed about 300 meters to the southeast of the Anamosa beds at approximately the same elevation, suggesting that maximum stratigraphic throw of the Silurian-Ordovician rocks locally in this area is 150 meters (ibid., p. 32). Stratigraphic relationships of the Pennsylvanian rocks to the Silurian-Ordovician rocks of this area, indicate that most of the observed vertical displacement along the fault zone preceded Pennsylvanian deposition.

Pennsylvanian sandstone exposures in Fairfield Township occur within a limited range of elevations, which would seem to preclude any large-scale vertical displacement of these beds. Maximum and minimum observed elevations of these rocks on both sides of the fault zone range from 840 to 750 feet (256-228 meters) above sea level. This is not to suggest that there may not have been any post-Pennsylvanian movement along the fault zone, since the observed elevation differences do not preclude small-scale post-Pennsylvanian deformation, and the rocks are not continuously exposed across the fault. In addition, stratigraphic relationships between the various Pennsylvanian outliers is speculative at best. However, certain generalizations regarding the overall relationship of the outliers can be inferred. Outcrops of conglomeratic sandstone beds (ibid., p. 24) in southern Jackson County, Iowa, contain clasts of quartz, Precambrian crystalline rocks, coalified wood, and Silurian chert, dolomite, and silicified fossils. The occurrence of Precambrian crystalline clasts suggests that the source area for these fragments were probably derived from igneous/metamorphic terranes to the north and northeast in the Minnesota-Wisconsin area. This suggests that the Pennsylvanian rocks along the fault zone may possibly have been deposited as part of a broad alluvial plain complex by rivers transporting detritus from the north and east. This same depositional alluvial plain framework has been applied to the Caseyville and Spoon formations in Muscatine County, Iowa (Fitzgerald, 1977), and the Pennsylvanian outcrops along the Plum River Fault Zone are tentatively correlated with those units.

The Illinois Basin is a long-lived Paleozoic structural feature best developed in southern Illinois and adjacent areas of Indiana and Kentucky. Although the Illinois Basin has been defined by some workers to include only the deepest basinal depression in southern Illinois and adjacent areas (where it is also termed the Fairfield Basin), a more general definition delimits the Illinois Basin by the regional extent of Pennsylvanian strata in Illinois and contiguous areas (Bristol and Buschbach, 1971). The latter definition essentially makes the northern portion of the Illinois Basin a Pennsylvanian feature. Earlier structural elements in central and northern Illinois (e.g. Sangamon Arch, East-Central Iowa Basin) had disappeared before the Pennsylvanian, and Illinois Basin subsidence generally spread northward into northern Illinois and a small portion of eastern Iowa concurrent with the onset of Pennsylvanian deposition. The extension of Lower Pennsylvanian (Morrowan) rocks (Caseyville Fm.) into eastern Iowa (Scott, Muscatine cos.) has been used to mark the extreme northwestern portion of the Illinois Basin (Fitzgerald, 1977). Pennsylvanian strata in the Illinois Basin are bounded by a series of structurally positive features (fig. 2): to the north by the Savanna-Sabula Anticlinal System, to the northeast by the Kankakee Arch, to the northwest by the Mississippi River Arch, and to the west by the Lincoln Fold System and Ozark Uplift. To the east, Pennsylvanian strata thin along the trend of the LaSalle Anticline but extend southeastward to the western margin of the Cincinnati Arch. The Caseyville Formation in the northwestern portion of the Illinois Basin (Scott-Muscatine counties, Iowa and contiguous portions of Illinois) is physically separated from the main body of Caseyville strata in the southern portion of the basin by a distance of over 150 miles (275 km; Hopkins and Simon, 1975, p. 178-179). The Caseyville edge, in both the northern and southern portions of the basin, is overstepped by Middle Pennsylvanian (Atokan-Desmoinesian) strata. The absence of Morrowan strata over most of the central and northern portions of the Illinois Basin suggests that the northwestern portion of the basin occupied a structurally lower position during the Morrowan than adjacent areas of the basin. This depression may have developed in synchrony with the uplift of the "Savanna-Sabula Anticlinal System" along its northern edge. Southward flowing fluvial systems (Fitzgerald, 1977, p. 199) filled in this structural depression with sediments during the Morrowan, but by the Middle Pennsylvanian cyclic marine and nonmarine sedimentation had spread across the Illinois and Forest City basins. Interestingly, this isolated Morrowan (Caseyville) basin in the Quad Cities area occupies a position along the crest of the Mississippi River Arch.

The presence of an east-west trending structural anomaly in the Savanna, Illinois area has been noted by numerous workers over the past 100 years. Chamberlin (1882, p. 425-426; pl. 8) first noted an east-west axis of flexure in the Savanna area, but provided little detailed discussion of the anomaly. Savage (1905, p. 640-641) in summarizing the geology of Jackson County, Iowa, described a low arch extending westward from Savanna, Illinois into Iowa for a distance of 20 miles (32 km). In 1920, Cady first applied the term "Savanna-Sabula Anticline" to this structural feature.

Original interpretations of the Savanna-Sabula Anticline were hampered by the inability to perceive the regional dimensions of the structure. Stratigraphic relationships in the Silurian strata of east-central Iowa were poorly understood, thus hindering structural interpretations. However, recent stratigraphic investigations of the Silurian and Middle Devonian rocks in eastern Iowa have helped to develop a more regional perspective. Structure contour maps drawn on top of the Galena Group (fig. 4) and the Blanding Formation (fig. 11) delineate a broad east-west trending anticlinal feature corresponding to the southern upthrown side of the Plum River Fault Zone across east-central Iowa. It extends into northwestern Illinois where it merges with the northwest-southeast trending LaSalle Anticlinal System, Ashton Arch, and the Wisconsin Arch. A series of smaller domes, anticlines, and synclines are also noted superimposed on this broad anticlinal feature (i.e., Savanna-Sabula Anticlinal System).

Uplift of the Savanna-Sabula Anticlinal System initiated the final destructive erosional episode of the East-Central Iowa Basin. Scattered outliers of Pennsylvanian rocks across eastern Iowa (summarized in Appendix 1) and northwestern Illinois overstep uplifted and erosionally truncated Kaskaskia and Tippecanoe rocks (fig. 34). The Savanna-Sabula Anticlinal System is depicted by the east-west deflection of pre-Absaroka geologic contacts in extreme east-central Iowa and northwestern Illinois. The Wisconsin Arch is likewise illustrated by the southeast-northwest deflection of geologic contacts across north-central Illinois. A stratigraphic profile (fig. 35) using the base of the Pennsylvanian System as a datum illustrates uplift of the East-Central Iowa Basin center and the structural preservation of the eroded mid-Paleozoic basin margin in southeastern Iowa.

The structure contour map drawn on top of the Galena Group (fig. 4) delineates many of the present structural features known in the area. The most prominent structures are: 1) the Savanna-Sabula Anticlinal System of east-central Iowa; present-day structure along the anticline closely parallels that displayed on the pre-Absaroka paleogeologic map (fig. 34), indicating the pre-Absaroka development of that structure; 2) the Mississippi River Arch; and 3) the Lincoln Fold System of northeastern Missouri.

The Mississippi River Arch has been an ambiguous feature in the geologic literature concerning the Upper Mississippi River Valley region. Its varied placement on structure and isopach maps of the region has led to confusion as to its exact location, extent, and general regional importance. Howell (1935) originally defined the Mississippi River Arch as "a broad corrugated fold extending from south-central Wisconsin to a point north of St. Louis, Missouri." The term has generally been applied to the broad arch separating the Forest City and Illinois basins. Unfortunately, the term has been badly misused and has included other previously-named structural elements in subsequent definitions, these are: the Savanna-Sabula Anticlinal System and the Lincoln Fold System.

The Mississippi River Arch is generally considered to be Middle Pennsylvanian in age, having formed concurrently with structural movements along the Nemaha Uplift and maximum subsidence of the Forest City Basin (Lee, 1943

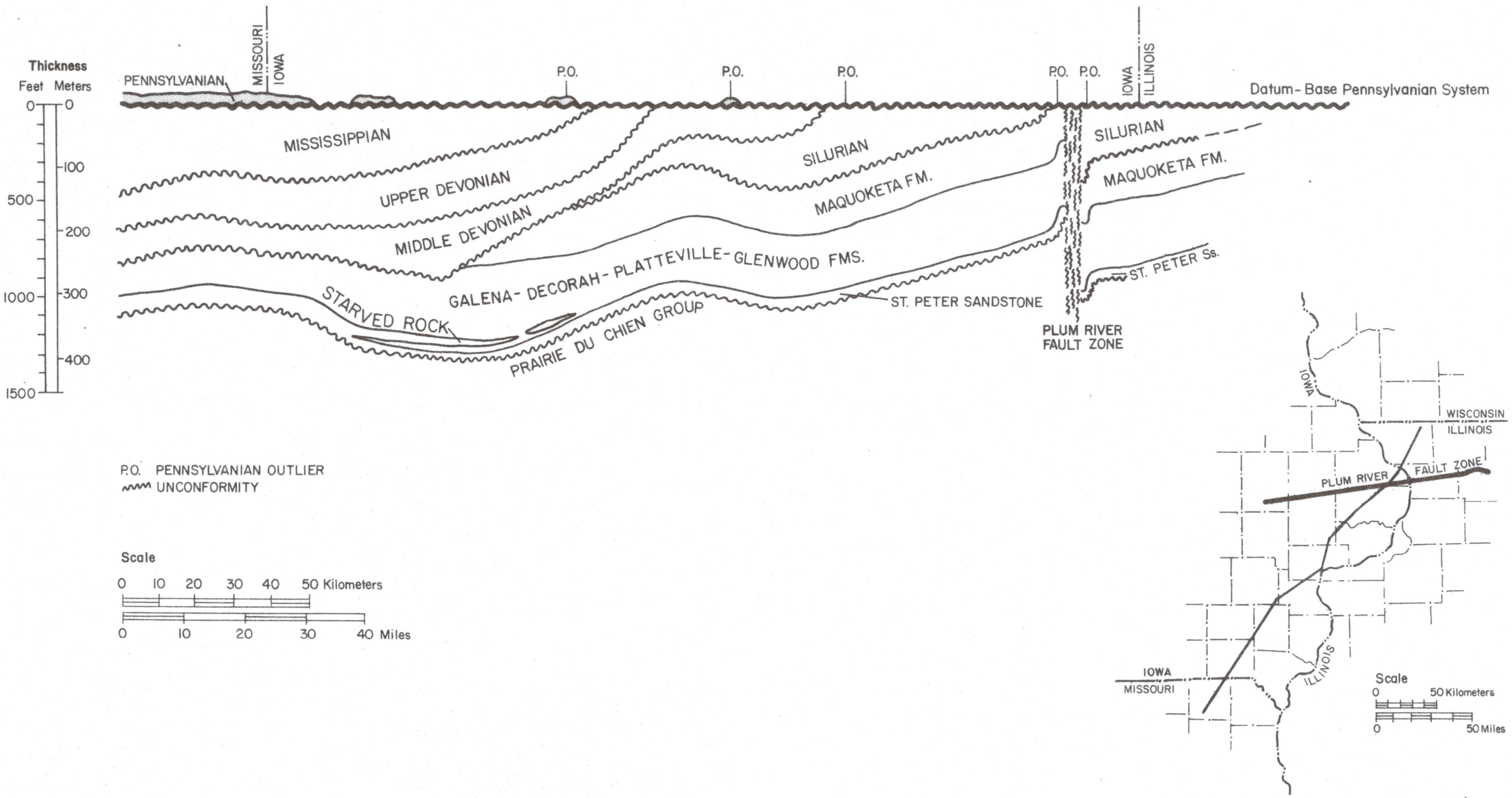


Figure 35. Northeast-southwest stratigraphic cross-section across eastern and southeastern Iowa, utilizing the base of the Pennsylvanian System as the datum. Uplift of the East-Central Iowa Basin center and preservation of the basin margin in southeastern Iowa is evident along this profile line.

1946; Anderson and Wells, 1968; Bunker et al., 1981; Bunker, 1982). Lower Pennsylvanian (Morrowan) rocks (Caseyville Formation) in the area were deposited in an isolated structural depression as evidenced by their physical separation from Caseyville strata in the Illinois Basin. The presence of Morrowan rocks along the crest of the arch runs contrary to the concept of a long lived structural arch (i.e., Mississippi River Arch) in the region.

Present-day regional structure (fig. 4) shows the Mississippi River Arch to be a northeast to southwest trending "structural saddle." It is bounded on the northeast by the Savanna-Sabula Anticline and on the southwest by the Lincoln Fold System. It also appears as a structural high between the Illinois Basin to the southeast and the Forest City Basin to the west. The crest of the Arch lies on the Illinois side of the Mississippi River along the southeastern border of Iowa. Figure 4 also shows a series of northwest-southeast oriented anticlines and synclines along the crest of the arch which parallel the axial trend of the Lincoln Fold System of northeastern Missouri. Stratigraphic studies of the Osage Series (Middle Mississippian) in southeastern Iowa (Harris and Parker, 1964) suggested that the development of these northwest trending folds pre-dated uplift of the Mississippi River Arch, and was contemporaneous with the development of the Lincoln Fold System.

The Lincoln Fold System was a prominent structural feature that influenced deposition during the Late Devonian and Early to Middle Mississippian in southeastern Iowa, west-central Illinois, and northeastern Missouri. Local doming and erosional truncation along the axis of the Lincoln Fold is suggested by the local occurrence of older Mississippian formations in contact with Middle Pennsylvanian rocks. Searight and Searight (1962) indicated that the Lincoln Fold was not a significant topographic feature during the Atokan (early Middle Pennsylvanian), but was somewhat elevated during the Desmoinesian, as evidenced by the eastward overlap onto and across the fold by successive formations during that time. The later history of the fold is obscure, because of post-depositional erosional removal of the younger Pennsylvanian beds.

In conclusion, destruction and uplift of the East-Central Iowa Basin and major faulting along the Plum River Fault Zone occurred prior to the deposition of the Absaroka Sequence in the study area. In general, the pre-Pennsylvanian structural configuration in the study area, as reflected by the sub-Pennsylvanian paleogeologic map (fig. 34), conforms closely to modern-day structural trends, indicating that post-Absaroka structural movements in eastern Iowa were relatively insignificant in comparison to earlier Paleozoic movements. However, movements along several structural features concurrent with Absaroka deposition are presently recognized in the study area: 1) structural downwarping in Scott-Muscatine counties, Iowa and adjacent portions of Illinois during the Morrowan; 2) development of the Mississippi River Arch primarily during the Middle Pennsylvanian; and 3) complex movements along the Lincoln Fold System during the Middle Pennsylvanian.

THE POSSIBILITY OF NEOTECTONISM

While it is evident that most, if not all, of the vertical displacement observed today along the Plum River Fault Zone occurred before Early to Middle Pennsylvanian deposition in eastern Iowa, the possibility of subsequent movements has not been rigorously investigated. The physical relationships between the scattered Pennsylvanian deposits and the Plum River Fault Zone are not known with sufficient precision to preclude up to 30 feet (9.1 m) of local post-Pennsylvanian vertical offset. The absence of Mesozoic and Tertiary strata in eastern Iowa leaves a large gap in the post-Paleozoic history of the area. Therefore, the physical relationships between Quaternary deposits and the Plum River Fault Zone afford the only opportunity to investigate possible post-Paleozoic movement on the fault.

Kolata and Buschbach (1976, p. 15-17) examined the near-surface structure and stratigraphy of the Plum River Fault Zone near Lanark, Illinois, utilizing a series of nine refraction seismograph profiles along a 5400 foot (1646 m) north-south traverse, supplemented by two shallow bedrock cores separated by a horizontal distance of approximately 2200 feet (671 m). These investigations led Kolata and Buschbach (1976, p. 17) to conclude that the fault zone near Lanark is mantled by several feet of the Ogle Till Member of the Illinoian Glasford Formation, in turn overlain by 10 to 15 feet (3 to 4.6 m) of loess. They (ibid., p. 17) reported that no discernable faulting displaces these Pleistocene deposits. While no evidence for large-scale Pleistocene-Holocene deformation was found, the methods employed in this investigation do not have high resolution, and the data are too inconclusive to permit meaningful evaluation of the possibility of geologically recent fault movement.

Systematic investigations of possible neotectonism along the Plum River Fault Zone have not been undertaken in Iowa, largely because no compelling evidence for Pleistocene-Holocene deformation is presently known. Studies of neotectonism in other areas of the United States, however, demonstrate that evidences are subtle, and easily missed without rigorous examination. A brief review of recent work in other areas may help to provide a comparative perspective for considering the possibility of neotectonism along the Plum River Fault Zone. Three primary lines of evidence have been used to interpret geologically recent fault movements. Geomorphic relationships between the land surface and known faults, stratigraphic relationships between recent deposits and known faults, and historic patterns of seismicity near faults have been integrated or used separately to interpret recent faulting activity.

Wallace (1977) examined the erosional degradation of young fault scarps on alluvium-colluvium mantled pediment surfaces in the arid Basin and Range region of north-central Nevada. There, it was found that recognizable fault scarps with vertical offsets around 6 meters (20 feet) can persist for more than 12,000 years, although with maximum slope angles of as little as 8° to 9°. Huntoon (1979, p. 223) argued that the age-slope relationships reported by Wallace (1977) should not be extrapolated into different, especially more humid climatic regions. This is certainly true with respect to eastern Iowa. If fault scarps on large pediment surfaces can persist for only a few tens of thousands of years in the arid Basin and Range region, then they would surely be ephemeral features in the humid, dissected till terranes of the upper midwest. Thus, in this region, the lack of observed morphological relationships between the land surface and fault traces is of little importance in the recognition of potential recent fault movements.

Kirkham (1977) reported on the structural and stratigraphic relationships between the Golden Fault and Quaternary deposits observed in a trench along the eastern margin of the Colorado Front Range. These observations were particularly significant, because "at no place along the entire length of the Golden Fault is there any known surficial evidence of Quaternary movement, except in the exploratory trench . . ." (Kirkham, 1977, p. 690). The trench exposures clearly revealed a narrow (43 feet wide, 13 m) zone of graben faulting, exhibiting at least two episodes of Quaternary reactivation (ibid., p. 692). Also important was the observation that not all of the faults exposed along the 197 foot (60 m) length of the trench displace Quaternary deposits (ibid., p. 691). Geologic mapping, drill core interpretations, and seismic reflection surveys indicate that the Golden Fault is a complex fault zone as much as 3281 feet (1000 m) wide, and that the mapped trace of the fault, based on earlier field mapping work, was 689 feet (210 m) west of the graben exposed in the trench (Kirkham, 1977, p. 692). The geomorphic, stratigraphic, and structural relationships reported by Kirkham (1977) illustrate that documentation of Quaternary faulting can be a difficult task. For the same reasons, one cannot demonstrate the lack of Quaternary faulting without similarly rigorous investigation. Considering the maximum documented width of faulting along the Plum River Fault Zone (3900 feet, 1200 m), it is evident that large scale, exhaustive studies would be required to truly demonstrate that Quaternary faulting has not occurred.

One troublesome aspect of the interpretation of neotectonics in the upper midwest is the general stratigraphic character of the Pleistocene deposits. These non-marine sediments were largely related to widespread continental glaciation. Stratigraphic markers, commonly used in the analysis of marine sedimentary sequences, are almost entirely lacking and most unit contacts are erosional surfaces. Structure contouring techniques commonly used to isolate areas of possible structural deformation are less useful in the Pleistocene, because widespread planar depositional surfaces are generally not recognized, and because of erosional complications.

Locally, however, roughly planar surfaces of ancient fluvial deposition have been preserved along the Plum River Fault Zone. Updegraff (1981) investigated the loess-mantled terraces of the Goose Lake Channel, an ancient Mississippi River channel, in southern Jackson County, Iowa. Near the north end of the channel, the terraces cross the Plum River Fault Zone. The age of these terrace remnants has been bracketed between 17,000 to 20,000 R.C.Y.B.P. (ibid., p. 42-53). Exposures of terrace sands near Spragueville, at the northern end of the Goose Lake Channel (SW sec. 19, T84N, R5E) display tabular cross-bedding units with planar foresets about one foot thick. Cross-sets dip uniformly to the south, indicating that stream drainage in the channel was from north to south. A longitudinal cross-section of the loess-mantled terrace surface along the Goose Lake Channel by Updegraff (1981, p. 55; see fig. 36) shows that over most of the length of the channel the terrace surface slopes to the south, suggesting that the surface preserves a relict stream gradient. At the northern end of the channel, where Deep Creek drains into the Maquoketa River, and where the terraces cross the Plum River Fault Zone, the terrace surface slopes to the north (fig. 36). This relationship suggests that in the vicinity of the Plum River Fault Zone, the depositional surface of the loess-mantled terrace has been modified by post-depositional erosion, tectonic deformation, or a combination of both. Figure 37 shows the location of Goose Lake Channel terrace surface remnants in relation to the approximate position of faults determined by reconnaissance field mapping. The gross distribution of the terrace surface remnants is influenced by the Plum River Fault Zone because the channel morphology is influenced by differing bedrock types. The wide channel area to the south of the Plum River Fault Zone is underlain by shales of the Maquoketa Formation, which were more easily eroded than the various Silurian dolomites exposed within and to the north of the Plum River Fault Zone. Thus, the Goose Lake Channel constricts abruptly toward the north, where it crosses the Plum River Fault Zone. In this narrow gorge-like segment of the channel, a terrace remnant is preserved across the full width of the Plum River Fault Zone, near the town of Spragueville (fig. 37). Detailed investigations of the Quaternary and Paleozoic stratigraphy and

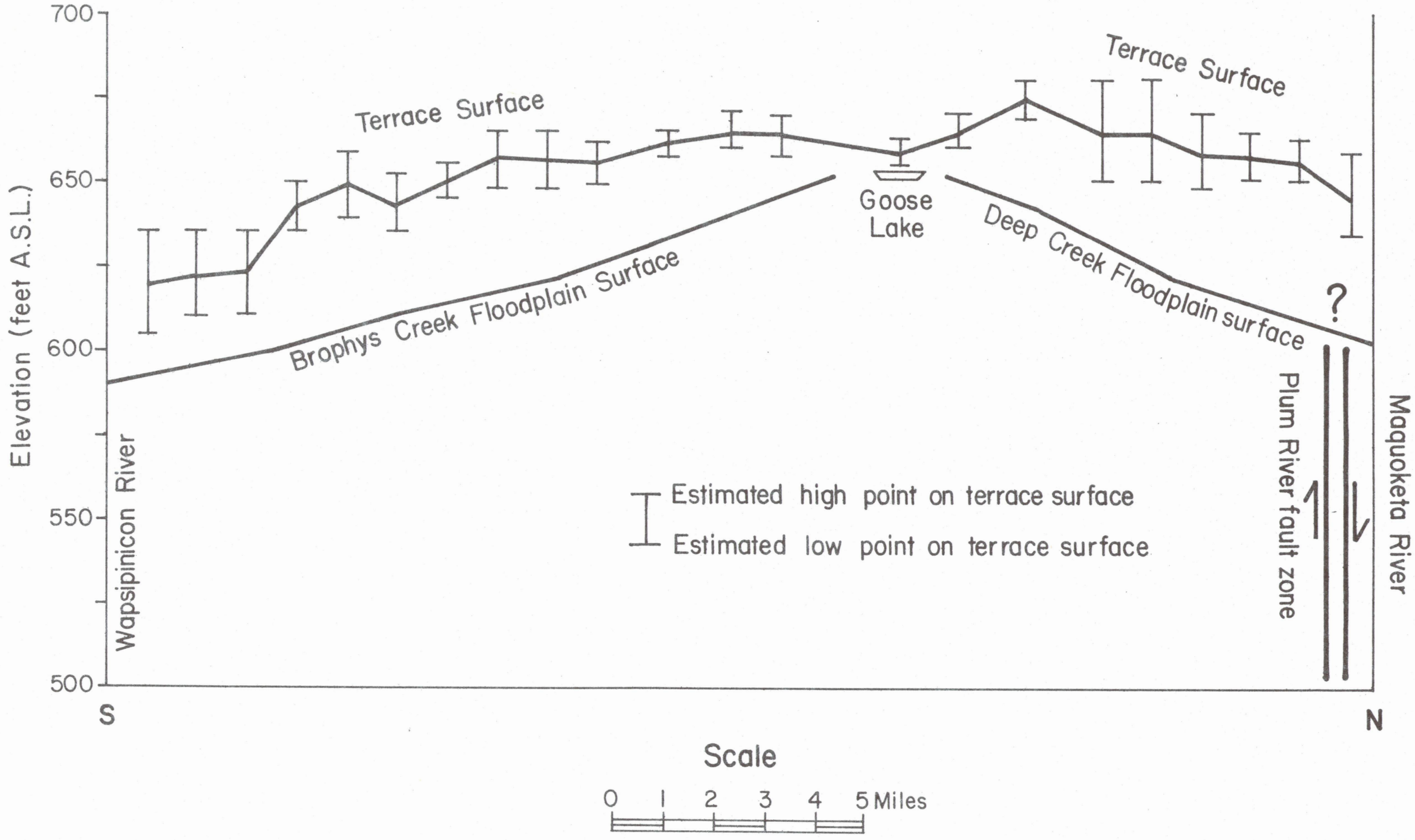


Figure 36. Elevations of modern and ancient surfaces of fluvial deposition in the Goose Lake Channel, modified from Updegraff (1981, p. 55). The gradient of the loess-mantled terrace surface in the vicinity of the Plum River Fault Zone has been modified by undetermined post-depositional processes.

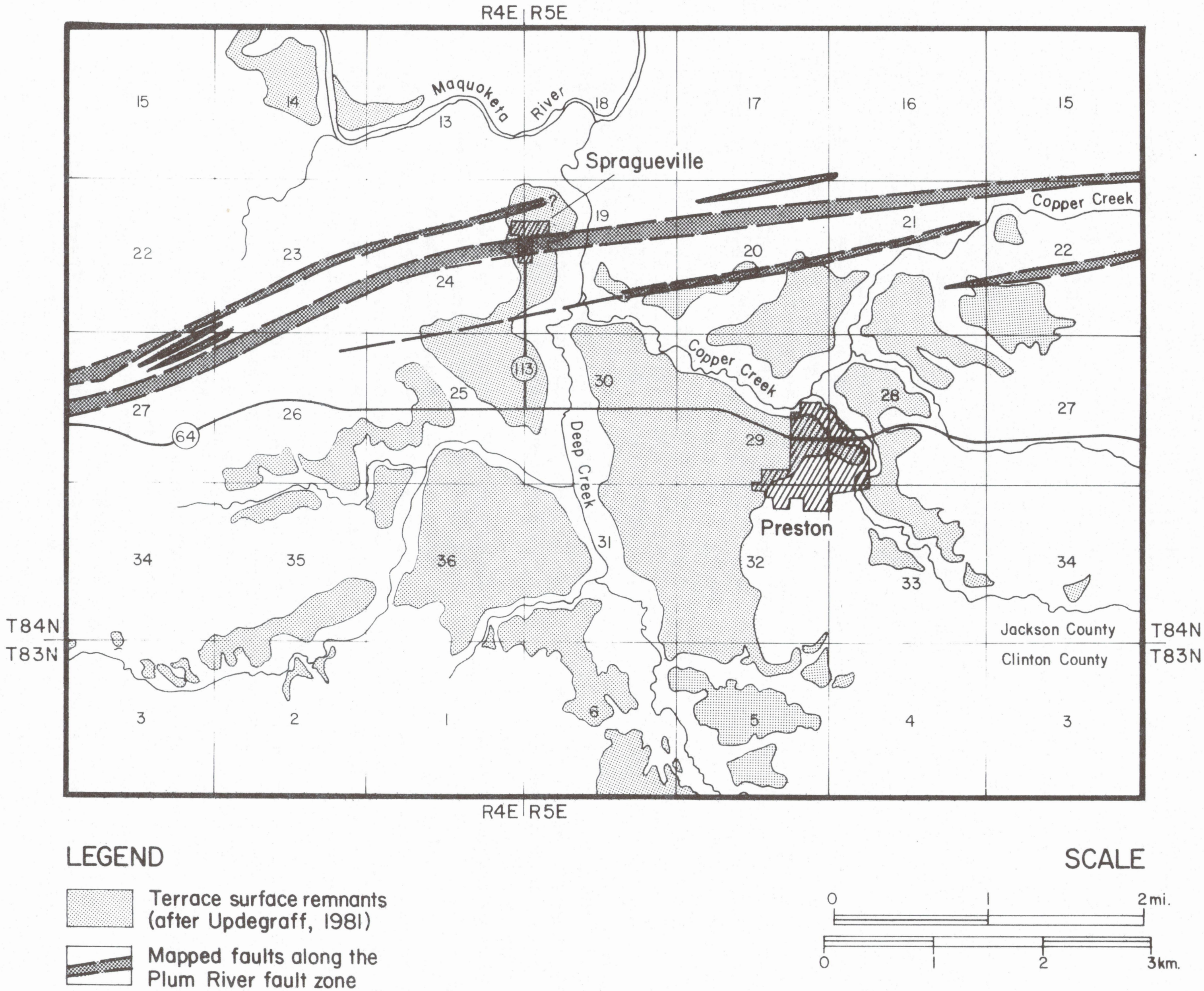


Figure 37. Location of loess-mantled terrace surface remnants in the Goose Lake Channel, in relation to mapped faults along the Plum River Fault Zone. The terrace surface remnant at Spragueville appears to be continuous across the entire width of the Plum River Fault Zone, and appears to be the most promising area for future study. Modified from Updegraff (1981).

structure of this area could determine whether or not faulting has occurred along this portion of the Plum River Fault Zone during the last 20,000 years.

Figure 38 shows the record of historic seismicity in the upper midwest (after Barstow et al., 1981). Several of the historic seismic events depicted in figure 38 occurred in relatively close proximity to the Plum River Fault Zone. It is not presently known whether or not this spatial relationship is significant. The relationships between historic seismicity and known structures in the upper Mississippi Valley are obscure because of the inherent limitations of the extant data. Nearly all of the recorded seismicity is reported using the Modified Mercali Intensity, because instrumental records from seismic stations are not available for these events. The locations of these epicenters have been inferred from written accounts and are not accurately known. Nuttli and Dwyer (1978) have shown that because of low crustal attenuation of high frequency seismic waves in the midcontinent region, local seismic events are felt over larger areas than comparable shocks in more tectonically active areas. Additionally, non-uniform population distribution and differing mechanical properties of surficial materials may also distort human perceptions of epicentral locations. One of the greatest limitations of these data is that written records in this region span only about 200 years. The midcontinent region is certainly not aseismic, but seismic events of even moderate magnitudes have long recurrence intervals, and for this reason, regional patterns of seismicity are not yet apparent. These problems have led Zoback and Zoback (1981) to assert that in evaluating seismic risk in the central and eastern United States, "attempting to identify potentially hazardous areas solely on the basis of historic seismicity is clearly inadequate" (ibid., p. 104). Thus, with the absence of long-term instrumental records on patterns of microseismicity in the region, no conclusions can be safely drawn concerning the possibility of modern seismic activity along the Plum River Fault Zone.

Perhaps the best example of the importance of instrumental data in evaluating seismic risk is the Ramapo Fault System of the Metropolitan New York City area. The Ramapo Fault System forms the northwestern boundary of the Newark Triassic Basin, but also experienced a history of Precambrian and early Paleozoic movements (Ratcliffe, 1971). Aggarwal and Sykes (1978) noted that the fault was long presumed to be inactive, a hypothesis that "now appears to have been tenable only in the near absence of local instrumental earthquake data" (ibid., p. 425). Significantly, before the availability of this instrumental data, modern seismic activity along the Ramapo Fault System had escaped detection in the most densely populated region in the United States! While the tectonic setting and magnitudes of displacement of the Ramapo Fault System differ greatly from that of the Plum River Fault Zone, both faults share the similarity of long histories of recurrent movement, spanning intervals on the order of 10^8 years (earlier discussion on tectonic history in this report; Ratcliffe, 1971, p. 138). Mechanisms of long-term fault reactivation are not clearly understood, though Zoback and Zoback (1981) have reviewed possible tectonic scenarios which may explain these phenomenon. Nevertheless, in evaluating the seismic risk potential of faults which have experienced long histories of recurring episodic movement, the terms "active" or "inactive" are probably inappropriate. The existence of and/or extent of seismic hazard associated with the Plum River Fault Zone will necessarily remain unknown until further investigations have been conducted.

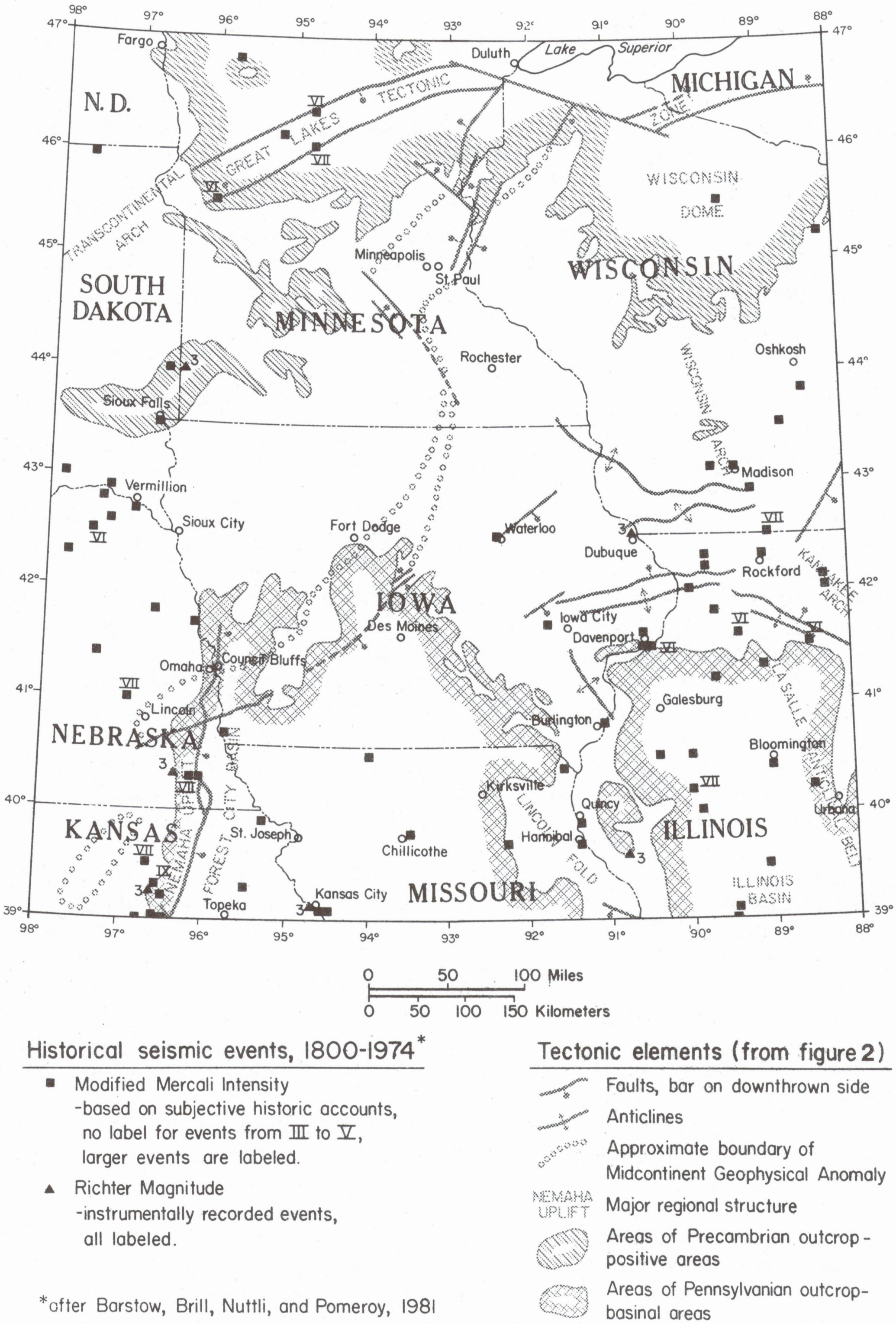


Figure 38. Historic seismicity in the upper midwest in relation to the regional structural geology. Actual epicenters of events reported in Modified Mercali Intensity are not known with precision.

SUMMARY AND CONCLUSIONS

Complementary investigations of the Plum River Fault Zone and the Paleozoic stratigraphy of eastern Iowa have demonstrated that the patterns of Paleozoic deposition and erosion in the area were influenced by tectonic activity over a long span of geologic time. While this observation echoes the earlier conclusions of many other geologic investigators who have worked in the "stable" interior region of North America, it still seems to counter conventional wisdom about tectonic stability in the craton. The central purpose of this report is to desribe the geologic structure and tectonic history of a large fault zone in the continental interior of North America. Many of the characteristics described in this report probably are not unique to the Plum River Fault Zone, but await future discovery and documentation along other structural features in the midcontinent region.

The major conclusions of this report are that:

1) The Plum River Fault Zone is a 112 mile (180 km) long, east-west trending zone of high-angle faulting in east-central Iowa and northwest Illinois.
2) The north side of the fault zone is downthrown with respect to the south side, with documented net vertical displacement of Silurian strata up to 270 feet (70 m).
3) The fault zone consists of many faults which intersect in a complex pattern, and graben and horst fault blocks occur within the structure.
4) Vertical displacements of Paleozoic strata up to 500 feet (150 m) are well documented between adjacent fault blocks, and vertical displacements of up to 1100 feet (335 m) on the Precambrian basement surface may occur within the fault zone.
5) Major faults within the Plum River Fault Zone are recognized by the occurrence of zones of brittle cataclastic deformation.
6) The maximum known width of the Plum River Fault Zone in Iowa is 3900 feet (1.2 km).
7) The Plum River Fault Zone played a role in the subsidence of the mid-Paleozoic East-Central Iowa Basin.
8) Active faulting along the Plum River Fault Zone probably occurred simultaneously with initial Middle Devonian deposition in the area.
9) Major faulting along the Plum River Fault Zone, and the development of the present regional structural configuation in east-central Iowa, occurred prior to initial Pennsylvanian deposition in the area.
10) Structural relationships between Pennsylvanian strata and the Plum River Fault Zone are not known with sufficient precision to preclude up to 30 feet (10 m) of post-Pennsylvanian fault displacement.
11) The potential for seismic hazard associated with the Plum River Fault Zone is not known.

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APPENDIX 1. Locations of Pennsylvanian outliers in east-central Iowa.

County	Location	Source of Information	Comments
Jackson	NE NE NW sec. 14, T84N, R5E	Ludvigson, 1984, unpub. field mapping	Sandstone
	NW NE NW sec. 20, T84N, R5E	Ludvigson, 1979, unpub. field mapping (see fig. 12)	NE-SW trending sandstone ridge.
	NW SE NW sec. 20, T84N, R5E	Ludvigson et al., 1978, p. 47	Sandstone float in field.
	W 1/2 sec. 17 & E 1/2 sec. 18 T84N, R5E	Savage, 1906, p. 628	Sandstone and shale.
	NW SW sec. 19, T84N, R4E & NE SE sec. 24, T84N, R3E	Savage, 1906, p. 628	Sandstone and shale.
	SE SE NE sec. 28, T84N, R4E	Chao, 1980, p. 24	Conglomeratic sandstone resting in near proximity to exposures of Gower Fm.
	N 1/2 sec. 32, T84N, R4E	Savage, 1906, p. 628	Sandstone resting on Maquoketa Shale.
		Chao, 1980, p. 24	A few discontinuous conglomeratic beds noted.
		Ludvigson et al., 1978, p. 44	Plant macrofossils (*Calamites*) recovered from this site.
	NW SW sec. 29, T84N, R4E	Chao, 1980, p. 24	Sandstone ledge resting on the Maquoketa Shale. In the N 1/2 sec. 29, Chao also noted sandstone along the northern boundary of the fault dipping 10° to the south. However, the exposure is such that it is difficult to tell whether the rotated block is the result of post-depositional erosional undercutting or tectonic in origin.
	SE sec. 4, T84N, R3E	Savage, 1906, p. 629	Reported but not observed.
	SE sec. 13, T84N, R3E	Savage, 1906, p. 627	Interbedded sandstone and shale.
	NW sec. 15, T84N, R3E	Savage, 1906, p. 629	Sandstone.
	NW SW NW sec. 25, T84N, R3E	Baik, 1980, p. 46-51	Sandstone
	NE SW SE sec. 26, T84N, R3E	Baik, 1980, p. 46-51	Sandstone
	NE NE SE sec. 21, T84N, R2E	Terry Frest (U.I.[1], pers. comm.)	Sandstone.
	SW SE sec. 31, T84N, R1E SW S 1/2 sec, 32, T84N, R1E SW SE sec. 32, T84N, R1E NW NE sec. 33, T84N, R1E extending into SE SE sec. 36 T84N, R1W (Jones County)	Savage, 1906, p. 625-627	Savage described a sandstone "ledge" extending intermittently for a distance of more than two miles, in a northeast-southwest direction. "This depression has a width of fifteen to twenty rods and a depth below the tops of adjacent Niagara ledges of fifty feet or more."
		Ludvigson et al., 1978	Noted that the sandstone channel is incised through the *Cyclocrinites* and *Pentamerus* bearing beds (Hopkinton Fm.)
		Osborn, 1892, p. 115	Specimens of *Lepidodendron* and *Calamites* were reported.
	sec. 3, T85N, R3E	Norton, 1895a, p. 131	Approximate location--sandstone ledges occur: in the base of 3 ravines trending north-northeast.
	SE sec. 9, T85N, R1E	Savage, 1906, p. 627	Sandstone.
	NW sec. 17, T85N, R1E	Savage, 1906, p. 627	Sandstone overlain by 2 1/2 to 3 feet of shale.
	SW NE sec. 18, T85N, R1E	Norton, 1895a, p. 122	Sandstone and shale, age uncertain; contains silicified Devonian fossils, lithologic similarities suggest Pennsylvanian. Norton noted some badly weathered boulders of Wapsipinicon type breccias (?); relationship unclear.
Clinton	sec. 7 & 18, T83N, R1E	McGee, 1891, p. 305	Sandstone.
		Udden, 1905, p. 404-407	Unable to locate, but noted several blocks of sandstone float in a ravine about 1 mile to the east.
	center sec. 2, T84N, R3E	Udden, 1905, p. 404-407	A thin sandstone and shale exposed in the base of the south bluff of Sugar Creek.
	1/3 mile east of NW corner sec. 1, T83N, R3E	Udden, 1905, p. 404-407	15 feet of sandstone of a few acres extent; extends across road into Jackson county on the north.
	SE NW sec. 22, T83N, R4E	Norton, 1895a, p. 131	Sandstone incised into Hopkinton Fm.
	SW NW NE sec. 12, T82N, R2E	*IGS Files	Private well, W-11503; sandstone.

County	Location	Source of Information	Comments
	SE sec. 24, T84N, R4W	Calvin, 1896, p. 60	Sandstone.
	NE NW NE sec. 9, T85N, R2W	*IGS Files	Private well, W-13925; shale.
Cedar	SE NW NW sec. 8, T82N, R1W	*IGS Files	Private well, W-05348; shale.
	sec. 28 & 29, T82N, R4W	Norton, 1895a, p. 121	Sandstone exposure along a line from east to west for nearly 2 miles along the banks of Clear Creek.
Linn	SW SE sec. 12, T82N, R5W	Norton, 1895a, p. 120	Sandstone, containing silicified Devonian fossils.
	NW sec. 35, T82N, R5W	Norton, 1895a, p. 120	Sandstone.
	SW NE SE sec. 34, T83N, R6W	Norton, 1895a, p. 118	Sandstone and shale, containing silicified Devonian fossils.
	NE NW NW NW sec. 4, T83N, R7W	*IGS Files	Carbonaceous shale and sandstone karst fill; Morrowan spores recovered.
	SE sec. 12, T83N, R7W	Norton, 1895a, p. 127	Private well, 23 feet deep; penetrated a bed of dark shale that contained characteristic "Coal Measure" plants.
	NE sec. 22, T84N, R7W	Wilson and Cross, 1939	Cavern-filling sandstone with a variety of Pennsylvanian plant fossils.
Johnson	c SW sec. 12, T78N, R8W	*IGS Files	Private well, W-5728; shale.
	SE SE NW sec. 3, T79N, R6W NE NE SW sec. 3, T79N, R6W	Calvin, 1897, p. 80; Witzke and Kay, 1984	Sandstone and shale; certain sandstone beds have yielded specimens of *Lepidodendron*, *Calamites*, and other plant fossils.
	NE NW NW sec. 3, T79N, R6W	*IGS Files	Private well, W-01980; shale and sandstone.
	SE NE NE SW sec. 3, T79N, R6W	*IGS Files	Private well, W-02972; shale and sandstone.
	NW SE NE sec. 2, T79N, R7W	*IGS Files	Rock core; shale and sandstone.
	SE NW SW sec. 30, T80N, R5W	*IGS Files	Private well, W-11240; shale and sandstone.
	NW NW sec. 4, T80N, R6W NE sec. 8, T80N, R6W	*IGS Files	Coralville Reservoir exposures of channel-filling sandstone, and shale; *Cordaites*, *Lepidodendron*.
	SW NW SW sec. 21, T80N, R6W	*IGS Files	Private well, W-26652; shale and sandstone.
Jones	SE SW SE sec. 20, T83N, R2W	Ludvigson et al., 1978, p. 131	Sandstone.
	NW NW NW sec. 20, T83N, R2W	M. Saribudak, 1980, unpublished map, *IGS files	Sandstone ledge with a 4° dip to the south.
	NE SW SW sec. 35, T83N, R2W	*IGS Files	Private well, W-08830; shale and sandstone.
	SE NE SW sec. 31, T84N, R3W	*IGS Files	Private well, W-13487; shale and sandstone.
	N 1/2 NW and W 1/4 SW sec. 33 and E 1/2 SE sec. 32, T80N, R6W	Witzke, 1984	Conklin Quarry; channel- and karst-filling sandstone and shale; Pennsylvanian plant fossils.
	NW NW SW sec. 34, T80N, R6W	Adams, 1926	Butlers Landing; channel-filling sandstone, shale, conglomerate; Pennsylvanian plant fossil.
	SE NE SE sec. 3, T80N, R8W	*IGS Files	Private well, W-10756; shale.
	SW SE SE sec. 17, T80N, R8W	*IGS Files	Private well, W-21165; shale and sandstone.
	SW SW SE sec. 28, T80N, R8W	*IGS Files	Private well, W-00383; sandstone.
	SE SE NE sec. 32, T80N, R8W	*IGS Files	Private well, W-19330; shale.
	SE SW SW SW sec. 5, T81N, R8W	*IGS Files	IGS test well, W-23166; sandstone and shale.
	sec. 27, T81N, R8W	Calvin, 1897, p. 82	Sandstone exposures near mouth of Knapp Creek.

Karst-filled voids of a dark gray silty shale are also occasionally encountered in quarry operations across east-central Iowa.

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*IGS - Iowa Geological Survey