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Dickeyville Roadcut

Highway 151
Dickeyville, WI

Dr. Bruce Brown

We will stop briefly at the junction with Hillclimb Road (old 151) to look back at the largest and highest road cut in Wisconsin. The cut immediately across from where the bus will stop is St. Peter Sandstone, with a thin Glenwood Member and the lower beds of the Platteville Formation at the top (Figure 1).

Figure 1, View east from old 151. Lower third of cut is St. Peter Sandstone, upper tw0 thirds Platteville. Dark zone at lower bench is Glenwood Member.

The main cut (Figure 2) exposes most of the Platteville Formation, the Guttenburg and Spechts Ferry Members of the Decorah Formation, and much of the Galena Formation. The Spechts Ferry is very shaly and is marked by water seeps after a rain or during spring thaw. In winter, large icicles form from these seeps, posing a potential hazard. The cut was designed with near vertical sides to minimize excavation and right of way. So far there have been few problems with rock fall, but wide ditches and a retaining wall along the pavement edge were built to prevent any falling material from landing on the roadway.

Figure 2, Dickeyville bypass cut. Cuts in distance at bottom of hill are St. Peter capped by lower Platteville. Rock at road level on left is Galena dolomite, dark band visible downhill by cars is Spechts Ferry and Guttenburg members of Decorah Formation.

Historical sedimentation on Little Platte River floodplain near Hwy 151 roadcut northeast of Dickeyville, Wisconsin.

Dr. James Knox
University of Wisconsin, Madison 53706l-1491

The Driftless Area

The Driftless Area is a hilly, highly dissected, unglaciated region located in southwestern Wisconsin and a small area of adjacent northwestern Illinois (Fig. 1 and Fig. 2). Various lobes of continental glaciers have abutted all sides of the Driftless Area, but the region was never completely encircled by glacial ice as shown on some early maps (Chamberlin, 1883) (Plate No. IX). Adjacent areas in southeastern Minnesota and northeastern Iowa frequently are referenced as part of the Driftless Area because these areas west of the Mississippi River were glaciated earlier during the Quaternary Period. Three principal factors explain the absence of glaciation in the Driftless Area. These include: (a) the southwestward diversion of ice flow into Minnesota by the deep trough associated with the structural syncline of Lake Superior, (b) southward diversion of ice flow by the deep valley now occupied by Lake Michigan, and (c) ice flow resistance against the relatively higher and erosionally resistant crystalline rocks of northern Wisconsin. The correct explanation for the origin of the Driftless Area is attributed to geologist R. D. Irving in 1877, but geologists N. H. Winchell and T. C. Chamberlin independently reached similar conclusions at about the same time (Martin, 1932)(p. 113-118). A recent new hypothesis suggests that permeable sandstone and carbonate bedrock of the Paleozoic Plateau in the Upper Mississippi Valley contributed to dewatering of glacial ice and to increased basal flow resistance for the advancing ice sheets which, in turn, then prevented ice from over-running the Driftless Area (Hobbs, 1999). Dewatering may have played some influence, but the original three factors described above seem to be the dominant factors accounting for lack of glaciation in the Driftless Area.

The "driftless" status of the Driftless Area has been challenged from time to time, but no unequivocal evidence for glaciation has ever been presented. Geologic deposits that have been presented as evidence for glaciation have been shown to result from other causes such as eolian, fluvial, or mass wasting processes (Knox et al., 1982). It is possible that a pre-Illinoian glacier(s) covered the Driftless Area, but this seems unlikely based on known evidence. Similarly eroded topography in adjacent northeastern Iowa and southeastern Minnesota contains abundant evidence of pre-Illinoian glaciation. No evidence of this type is found in southwestern Wisconsin or northwestern Illinois except on and along the upland bluff tops at the western margin of the Driftless Area, and at the mouth of the lower Wisconsin River on the Bridgeport Terrace (Knox and Attig, 1988).

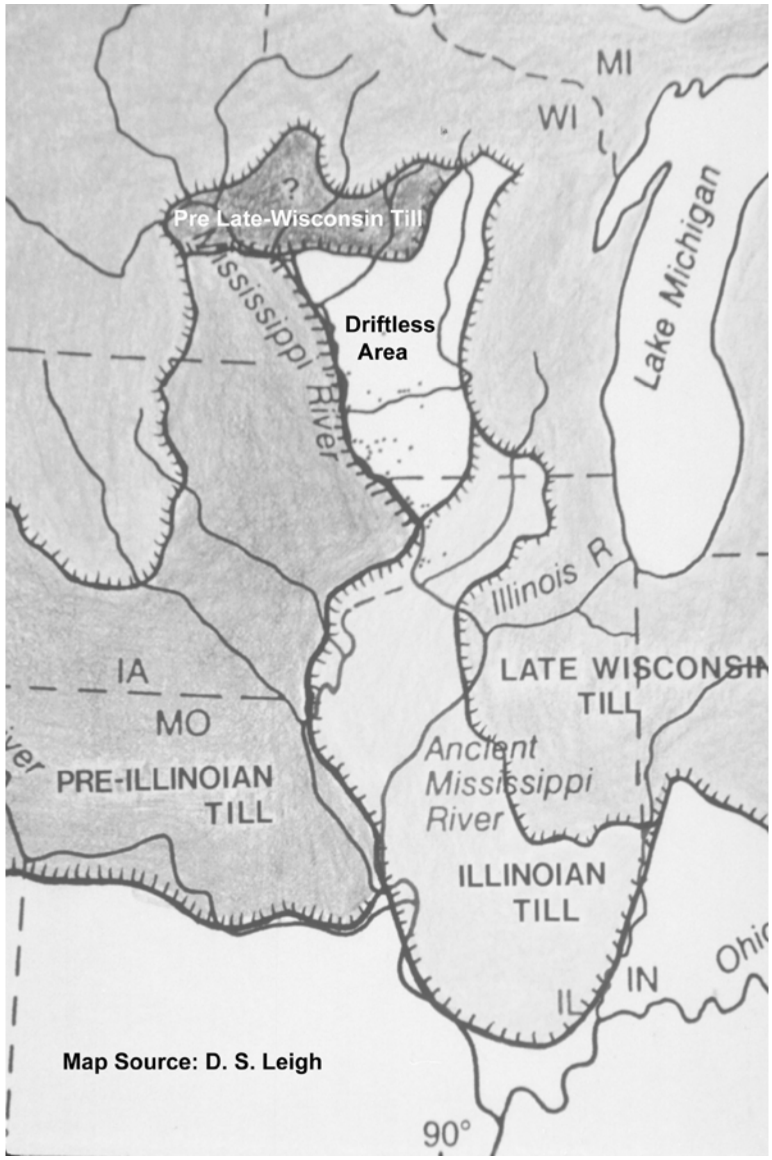

Figure 1. Glacial deposits of the Upper Mississippi Valley and the location of the Driftless Area.

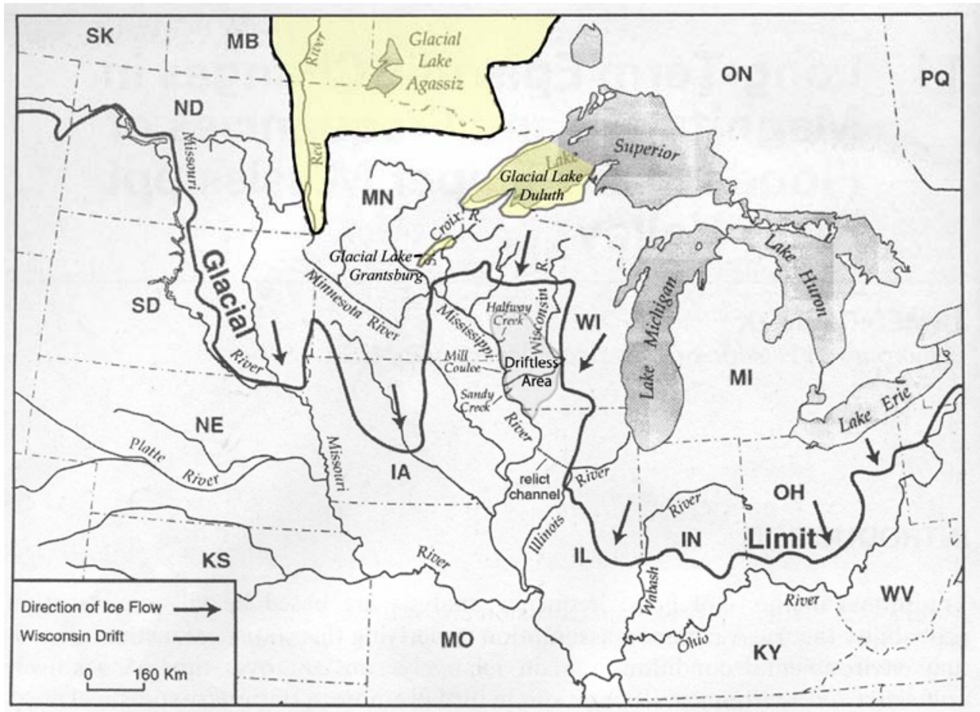

Figure 2. (Source: modified from (Knox, 1996)). After about 14,000 ^{14}C yr B.P. (16,700 cal yr B.P.) a general northward retreat of the southern margin of the Late Wisconsin continental glacier prevailed and eventually led to the modern post-glacial (Holocene) climate. The upper Mississippi River experienced massive aggradation with sand and gravel during full glacial times when meltwater and large sediment loads were input to the river system. When the continental glacier began its general retreat after about 14,000 ^{14}C yr B.P. (16,700 cal yr B.P.), lakes formed between the retreating ice front and former drainage divides to the south of the ice sheet. Glacial Lakes Agassiz, Duluth, and Grantsburg served as sediment traps and the discharges from these lakes carried relatively low sediment loads and produced large erosive discharges to the upper Mississippi River. Discharges from Glacial Lakes Duluth and Agassiz, because of their large sizes, appear to have been mainly responsible for deep incision of the alluvial fill that had accumulated in the Mississippi River between about 25,000 and 14,000 ^{14}C yr B.P. The incision produced a series of terraces in the Upper Mississippi Valley. The highest terrace, which represents the former late-glacial floodplain that existed until about 14,000 ^{14}C yr B.P. (16,800 cal yr B.P), is known as the Savanna Terrace (Flock, 1983). A sequence of terraces, inset below the Savanna Terrace, are know as the Bagley Terrace complex in the Upper Mississippi Valley between western Wisconsin and Minnesota and Iowa (Knox, 1996). The Bagley Terrace probably is equivalent to the Kingston Terrace in the middle reaches of the central Upper Mississippi Valley along western Illinois (Hajic et al., 1991). Major erosive discharges from proglacial lakes ceased to flow into the upper Mississippi River headwaters by at least 9400 ^{14}C yr B.P. (10,600 cal yr B.P.) (Fisher, 2003; Teller, 1987).

The alluviation of the upper Mississippi River by Late Wisconsin sand and gravel led to a damming and backfilling of the lower reaches of tributaries. Although the post-glacial downcutting described in the Figure 2 caption removed much of the Late-Wisconsin backfill sediment, the gradients of lower valley floodplains are anomalously flat as they approach the base level of the Mississippi River. Lower reaches of Mississippi River tributary streams that have been affected by this flattening have been susceptible to exceptionally high magnitudes of historical overbank flooding and sedimentation. An excellent example occurs along the floodplain of the Little Platte River north of Dickeyville near the first stop (Stop B – Hwy 151

Roadcut) of the field trip on Saturday. Figure 3 shows components of the alluvial stratigraphy underlying the Little Platte River floodplain about 0.6 miles upstream of Stop B.

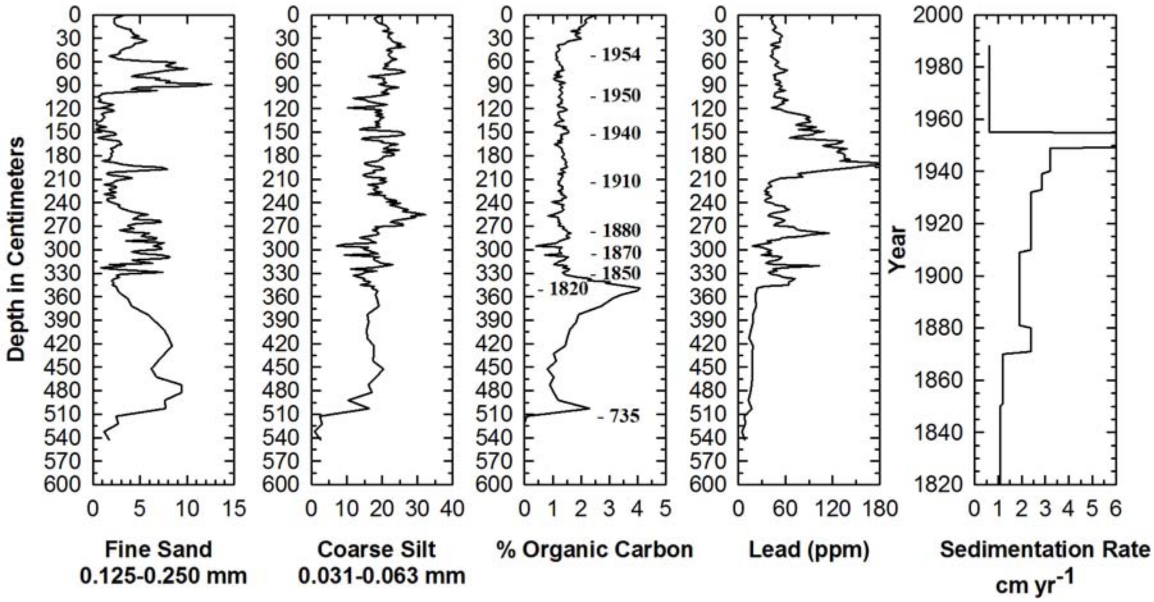

Figure 3. (source: (Knox, 2001)). Fine sand and coarse silt sediment fractions obtained from a 5.4 m drill core in the Little Platte River floodplain (SW ¼ Sec 11 T2N R2W) document variability of flooding and sedimentation here since about A.D. 735. Anomalous high fractions of fine sand are indicative of large floods as occurred in 1950, 1951, and 1954. Surveys and benchmarks established by the Corps of Engineers show the river bed in 1939 was still at the same level it had been since A.D. 735. The bed and bank tops have aggraded between 1.5 and 2 m since closure of Lock and Dam 11 at Dubuque about 25 km downstream. This shows that the Mississippi River "backwater curve effect" extends well upstream into these low gradient tributaries and contributes significantly to increased flooding and sedimentation on these lower valley floodplains. The high level of organic carbon at about 345 cm depth corresponds with the floodplain surface prior to Euro-American agricultural settlement in the upstream watershed. Variations in lead concentrations correspond with changes in mining intensity in the upstream watershed. Since mining and milling histories are documented in written records and maps, temporal variations in concentrations of lead in overbank sediments provide a very useful indicator of the approximate age and extent of vertical accretion in the historical past. Note that the Little Platte River overbank sedimentation began to accelerate significantly after most of the watershed was converted to agricultural land use by the 1860s and 1870s, and that overbank sedimentation rates peaked with the large floods at the beginning of the 1950s. There is some evidence that overbank sedimentation may have increased slightly in recent years since about 1990 at this site in response to changes in agricultural cropping practices. Nevertheless, the relatively lower sedimentation rates since the early 1950s are mostly a consequence of improved land conservation practices and a dramatic change to higher density plant cover following the introduction of heavy use of commercial fertilizer and herbicides in the 1950s.

The historical pattern of overbank sedimentation shown for the lower Little Platte River in Figure 3 is characteristic of large stream, lower valley, reaches of low gradient tributaries that directly join the Mississippi River, although the beginning of accelerated sedimentation occurs slightly later in the northern Driftless Area where agricultural settlement occurred later. Agriculturally accelerated sedimentation on floodplains and valley floors continues to occur on lower valley reaches, but farther upstream former historical floodplains of small and intermediate size tributaries have become terraces and now experience overbank sedimentation. Historical development of large capacity meander belts contain runoff associated with most floods. The effectiveness of the meander belt and modern channel to restrict overbank flooding in the upstream reaches is especially pronounced for average valley floor sedimentation rates for the period dating from the mid-20th century (Fig. 4).

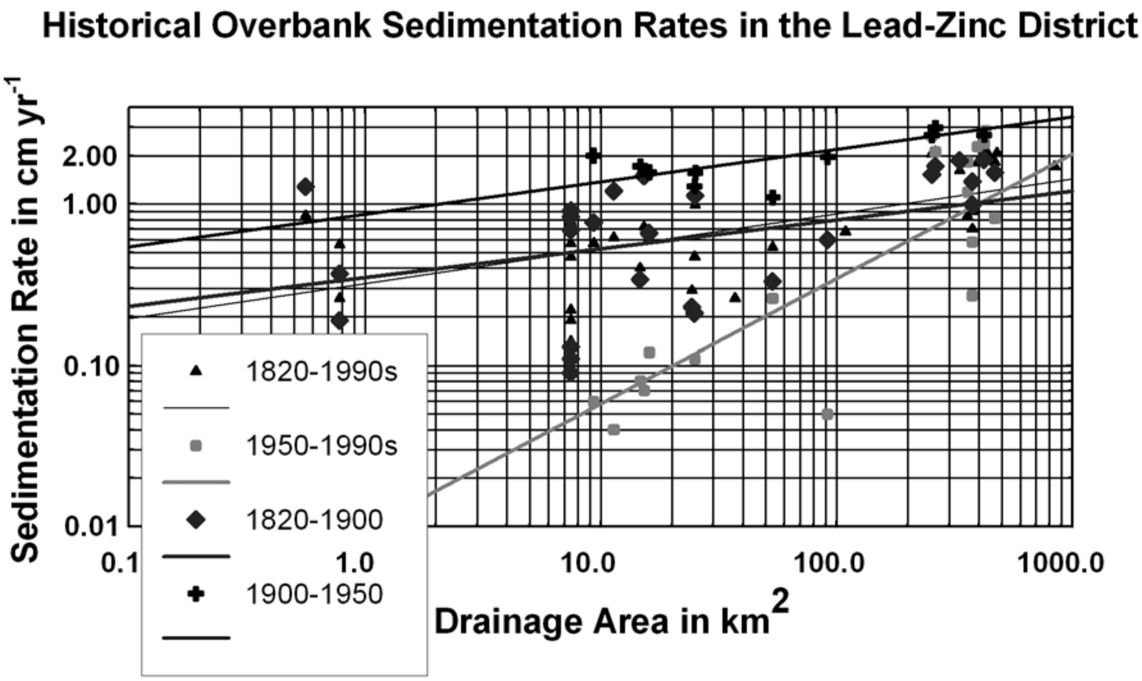

Figure 4. (Data source: (Knox, 2006)). Rates of overbank sedimentation associated with floods on the floodplains and adjacent valley floors generally increase systematically downstream with increasing drainage area. However, as flood erosional power in channels increased with increasing bank heights due to overbank sedimentation, bank erosion accelerated and this led to the historical development of a new floodplain inset within the former floodplain that became a terrace. The new historical floodplain began to form in the small, higher energy tributaries by the late 19th century and it subsequently expanded downstream. The combined capacities of the historical meander plains and modern channels effectively function like a flume to efficiently route water and sediment downstream and prevent most floods from being able to experience the effects of roughness, storage capacity, and other attenuation effects produced by the former valley floor and floodplain surfaces.

The data of Figure 4 show that average overbank sedimentation rates varied from about 0.5 to 2 cm yr^{-1} throughout much of the region before the mid-20th century. These rates are significantly higher than the long-term average Holocene rate of floodplain overbank sedimentation which averaged about 0.02 cm yr^{-1} for tributary drainages ranging between about 5 km^2 and 700 km^2 and average about 0.09 cm yr^{-1} on the adjacent Mississippi River floodplain where drainage area increases to 160,000-180,000 km^2 (Fig. 5).

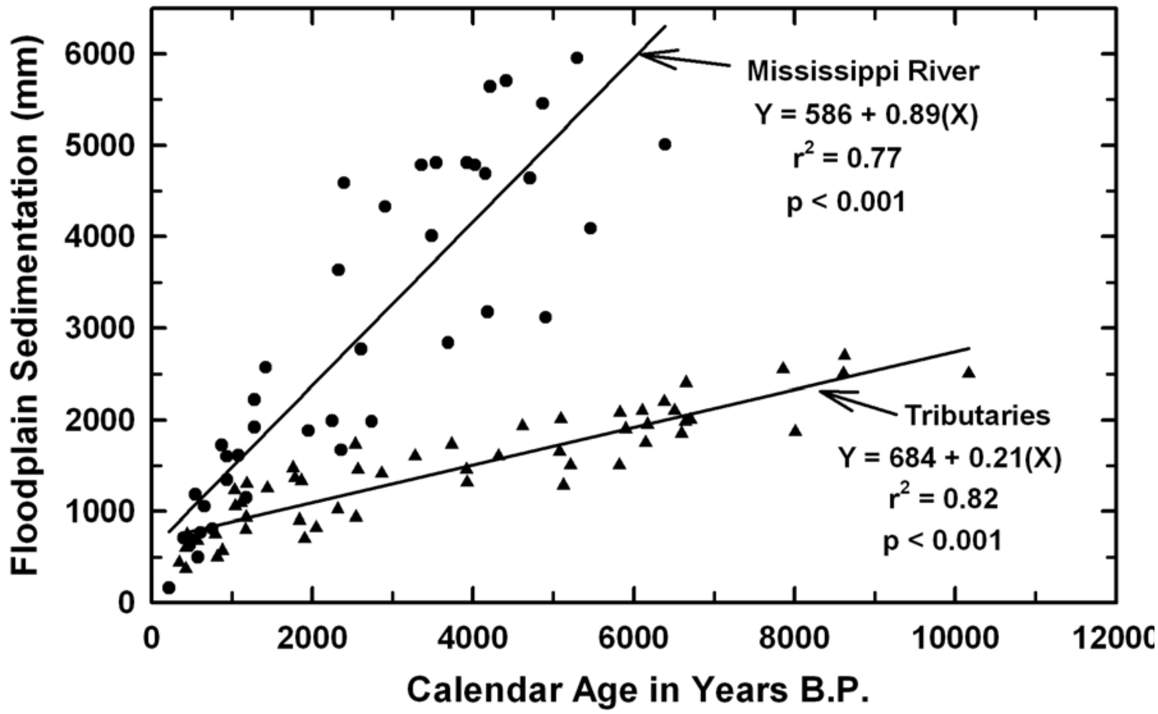

Figure 5. (Source: (Knox, 2006)). Long-term natural rates of floodplain vertical accretion prior to human influence.

Hoadley Hill Roadcut

Collapse structures in St Peter Sandstone
Highway 151 and Airport Road
Platteville, WI
Dr. Bruce A. Brown

The accompanying published outcrop description (Appendix A) provides a detailed account of the geology in this section, which is still useful in spite of the fact that the exposure has been greatly enlarged as the curve was straightened for the new highway. Geologists often lament the loss of a favorite exposure, particularly one that has been written up for field trips, but more often than not new construction exposes surprising new information.

The original cut only exposed a small section of the St. Peter, but the new cut exposes the full St. Peter and just touches the upper Prairie du Chien. The big story at this stop is illustrated in Figure 3. Note the bedding offsets marked by iron staining in the sandstone. At the base of the cut there is deformed shale of the Readstown Member. The offsets, which are really minor faults, drop the sandstone down relative to the shale to either side. There appears to be several tens of feet of downward displacement across the face, but at the top of the sandstone the upper contact with the Platteville formation is continuous and unbroken.

Figure 3 Collapse structure in St. Peter sandstone at Hoadley Hill cut. Dark band to right of center is Readstown shale cut off by minor fault in sandstone. Note sharp contact of overlying Platteville, suggesting deformation predates Platteville deposition. Prairie du Chien dolomite is exposed in ditch at far left out of view.

The interpretation of this outcrop is that collapse of paleokarst in the Prairie du Chien was occurring as the sandstone was being deposited. Collapse ceased during sand deposition before the transition to carbonate in Platteville time.

There is evidence that collapse occurred locally as late as Galena deposition. An exposure near the Iowa –Lafayette County line south of Mineral Point illustrates similar disturbance truncated above by continuous carbonate beds (Figure 4). Previously features such as these were attributed to collapse due to solution during lead-zinc mineralization which we now know occurred much later in Pennsylvanian –Permian time.

Figure 4, Collapse structures in Galena dolomite. Note continuous carbonate beds above disturbed bedding of collapse zone.

The remainder of the trip will rely on the published stop descriptions attached in Appendix A. The Bridgeport west section and the Wyalusing section are supplemental stops that we will probably not have time for, but are worth a stop when in the area. The Wyalusing Quarry, which we will probably drive through on our way from the park to the next quaternary stop at Bagley illustrates a nearly complete section near the type area for the Prairie du Chien. The Pigeon Creek stop illustrates the Platteville/St Peter transition, similar to Bridgeport. The Potosi Hill section is one of the best and most complete accessible sections in Southwest Wisconsin, and has been visited by student trips for many years.

Majestic View Dairy Parlor

5532 Commercial Road
Lancaster, WI
Don Abby, Owner

Majestic View Dairy Farm is one of the largest dairy farms in southwest Wisconsin. It is on the DNR list of Combined Animal Feedlot Operations (CAFOs). We will be given a brief tour of the facility and we ask that nobody acts rudely to the management or staff. The juxtoposition of this huge waste producing complex and the nearby trout streams is certainly worthy of critical inspection. However, the management is proud of their efficient waste handling techniques and ability to keep the nearby streams safe.

Schreiner Park

999 Schreiner Park Road
Lancaster, WI
Lee Trotta

Schneider Park in the northwest part of Lancaster is more than just a pastoral setting. The creek that runs through the park eventually feeds a very productive trout stream, the Grant River. The rocks exposed in the hillside include a layer that is seldom seen because of its fragile nature. As we have only a short time to visit this site due to impending lunchtime, a stratigraphic cross section (excerpted from Kammerer, Trotta, et al, 1998) will be used to explain the geology here. This section intersects Lancaster and runs between Prairie du Chien and Blue Mound.

Reference:

Kammerer, P.A., Trotta, L.C., Krabbenhoft, D.P., and Lidwin, R.A., 1998, Geology, ground-water flow, and dissolved-solids concentrations in ground water along hydrogeologic sections through Wisconsin aquifers: U.S. Geological Survey Hydrologic Investigations Atlas HA 731, 4 plates.

Wyalusing State Park

Bagley, WI

Quaternary History of the Lower Wisconsin River Valley

Dr. James Knox

Wyalusing Park overlooks the confluence of the Wisconsin River with the Mississippi River. Because early transportation in the region emphasized water routes, the confluence has a long history of settlement by both Native Americans and Euro-Americans. Nicholas Perrot, a fur trader, built Fort Saint-Nicholas at the mouth of the Wisconsin River between 1686 and 1688 (Scanlan, 1937). The Americans built a fort at Prairie du Chien at the beginning of the war of 1812 but it was soon captured by the British and burned. After the war, the Americans built Fort Crawford and maintained it until the mid-19th century. France surrendered Canada to Great Britain in 1763, but the village of Prairie du Chien remained dominantly a French community into the first few decades of the 19th century (Trewartha, 1938).

The lower Wisconsin River valley has special significance pertaining to the natural history of the Driftless Area because it preserves some of the oldest Quaternary sediments in the Driftless Area. These sediments are located on a rock bench (strath terrace), and it is the presence of the rock bench that prevented the overlying Quaternary deposits from being destroyed from lateral erosion by the Wisconsin River (Fig. 6). The rock strath represents a former valley bottom of the Wisconsin River. The gradient of the strath is so nearly flat, and the strath is so highly dissected, it is not possible with present stratigraphic information to determine whether its longitudinal gradient slopes westward or eastward (Fig. 7). The bedrock floor of the valley

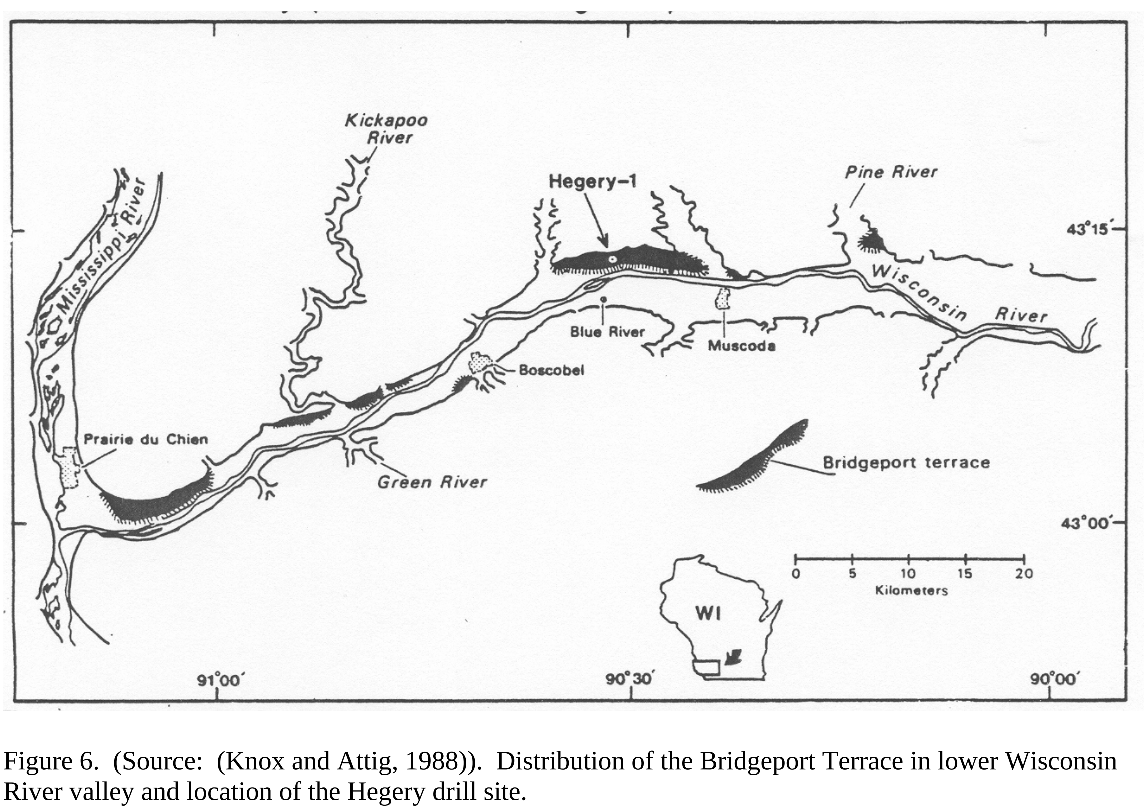

Figure 6. (Source: (Knox and Attig, 1988)). Distribution of the Bridgeport Terrace in lower Wisconsin River valley and location of the Hegery drill site.

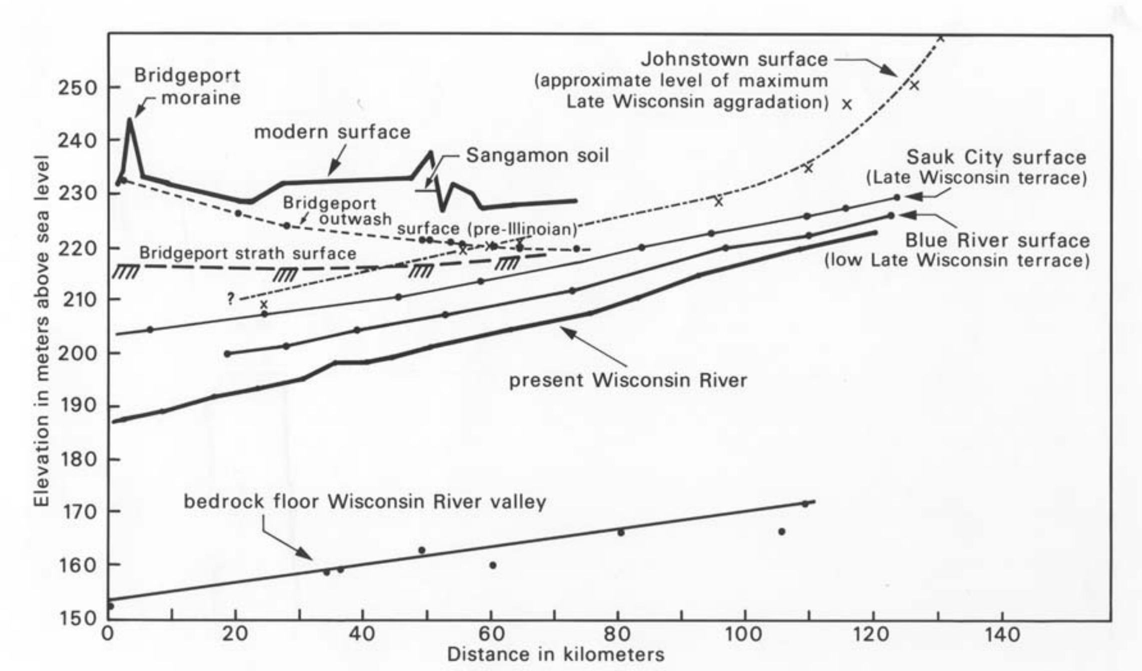

Figure 7. (Source: (Knox and Attig, 1988)). Longitudinal profiles of the Bridgeport terrace and other fluvial and bedrock surfaces in the lower Wisconsin River valley.

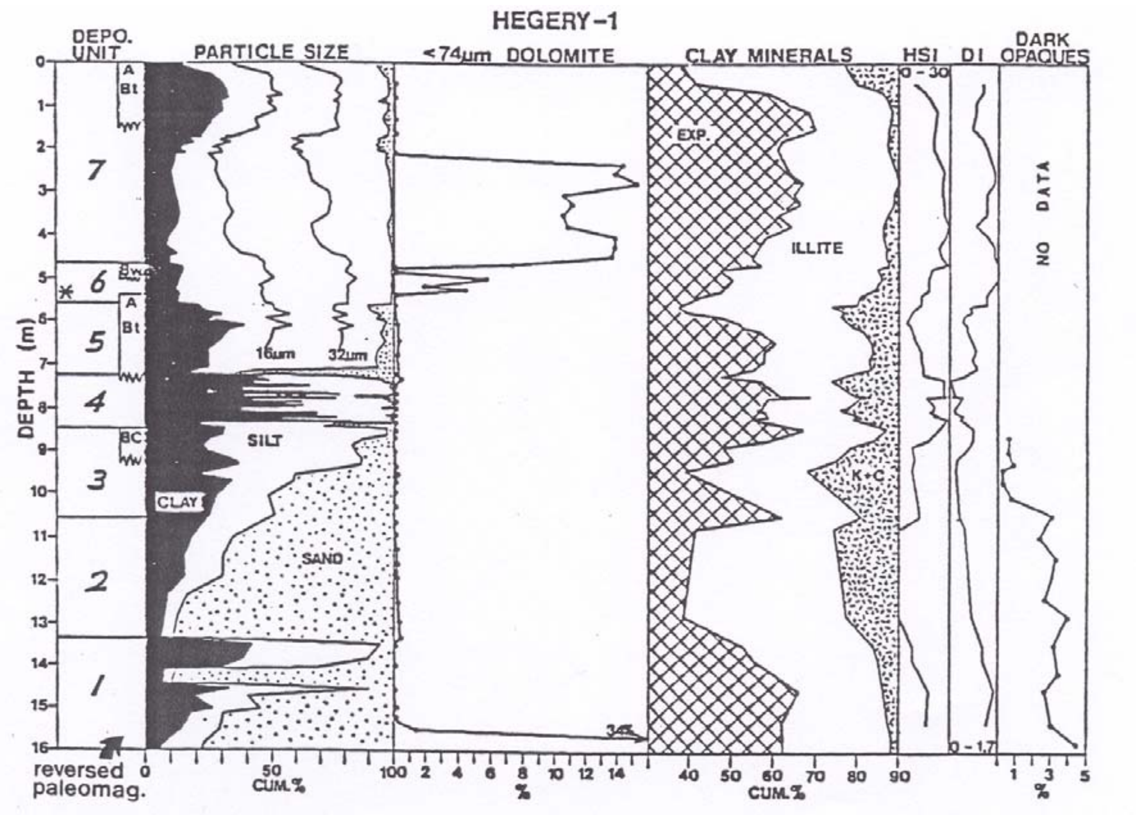

Figure 8. (Source: (Leigh and Knox, 1990)). The unconsolidated sediments of the Bridgeport Terrace are represented by seven depositional units involving differential fluvial, lacustrine, and eolian processes. Unit

1 is dominated by sandy gravel outwash from an ice advance of 2-3 km into the mouth of the lower Wisconsin River valley. Silty (lacustrine?) beds interstratified with the gravel of unit 1 were associated with reversed remanent magnetism (Knox and Attig, 1988). The reversed aspect of the magnetism implies that unit 1 probably is at least 790,000 years old. Units 2 and 3 show a fining upward trend, but their ages are unknown. The high clay and silt composition of rhythmically bedded unit 4 is indicative of lacustrine or ponding conditions. Such ponding may have occurred when the Mississippi and/or lower Wisconsin River was flooded by an ice-dammed lake, perhaps when nearby northeast Iowa was glaciated about 500,000 years ago (Hallberg, 1980), or possibly later when Illinoian ice blocked the Mississippi River (Fig. 1). The relatively higher sand content of unit 5 compared to overlying loess units 6 and 7 suggest that its deposition probably involved both fluvial and eolian processes. Unit 5 incorporates an extremely well-developed soil that is correlated with the Sangamon age paleosol elsewhere in the Upper Mississippi Valley. Units 6 and 7 respectively represent the Roxana Silt and Peoria Loess, comprised of sediments that originated from eolian processes. A ^{14}C radiocarbon age on spruce charcoal from unit 6 was 31,310 ± 470 yr BP (Leigh and Knox, 1993). The Roxana Silt was deposited between about 55,000 and 25,000 ^{14}C years ago in the Upper Mississippi Valley, whereas the Peoria Loess was mainly deposited between about 25,000 and 12,000 ^{14}C years ago in the same region.

underlying the modern Wisconsin River slopes westward at a slightly lower gradient than the present river. The bedrock floor under the present river is approximately 60 m below the elevation of the Bridgeport strath and about 35 m below the present Wisconsin River bed in the Wyalusing Park area. The strath is 145-150 m below upland ridgetops that are associated with Windrow Formation gravel. The Windrow gravel represents former stream bed deposits and are thought to be of Cretaceous age (Andrews, 1958). Assuming a Cretaceous age as reasonable for the Windrow gravel, then approximately 220 m of total downcutting have occurred over approximately that last 60-65 million years.

The most noteworthy aspect of the Bridgeport Terrace is the presence of pre-Illinoian till overlying the bedrock strath. The till extends into the valley mouth a distance of 2-3 km where its terminous is denoted by a highly eroded end moraine (Fig. 7). Proglacial outwash gravel slopes eastward from the Bridgeport moraine indicating that the lower Wisconsin River was flowing eastward at this time (Fig. 7). Coring of the Bridgeport Terrace about 50 km up-valley at the Hegery Site, coupled with study of sediment exposures in a gravel pit located near the Hegery Site, showed that the Bridgeport outwash consisted of sandy gravel that was interstratified with silty (lacustrine?) beds (Fig. 6 and unit 1 of Fig. 8). As noted in the caption of Figure 8, the silty beds are characterized by reversed remanent magnetism which indicates an age of at least about 790,000 years for this unit. This deposit represents the oldest dated Quaternary age deposit in the Driftless Area. The principal reason why there are few old Quaternary deposits in the Driftless Area is explained by the severe tundra and permafrost conditions that have occurred in the region when glacial ice abutted its margins. During these times mass wasting stripped much or most of the unconsolidated deposits from uplands and valley sides.

Don Orr Gravel Pit

Bagley, WI
Cenozoic Drainage Evolution
(This section on Cenozoic History reprinted from Knox, 2004, Groundwater Conference Field Trip)

Most of the topography and drainage of the upper Mississippi River system has developed within relatively recent geologic time, and this includes the location of the upper Mississippi River along southwestern Wisconsin and northwestern Illinois. Ages of valley fills and their relations with regional stratigraphy indicate that many deeply incised bedrock valleys in mid-continent North America probably were not present as recently as the late Pliocene (2-3 million years ago)(Anderson, 1988; Hallberg et al., 1984; Hallberg et al., 1985; Knox, 1982; Trowbridge, 1921).

The exact pre-Quaternary (pre-glacial) northern drainage divide is unknown, but regional geologic structures suggest that it followed an east-west trend along the Niagara cuesta in northern Illinois, thence northwesterly along the same cuesta in northeast Iowa, followed by a westerly course across southern Minnesota to the area of the Sioux Arch in southern South Dakota, then west-southwesterly to the Rocky Mountain Front (Fig. 3). The drainage represented by the upper Missouri, Yellowstone, and other tributaries in the Dakotas flowed north to Hudson Bay prior to glacial modification of the region (Thornbury, 1965). The effects of Quaternary glaciers and their associated deposits also account for most of the anomalous relations between geologic structures and geographic locations of the upper Mississippi River main channel. The Mississippi River along the western borders of northwestern Illinois and southwestern Wisconsin is incised across dolomite cuestas rather than being deflected by them. It is therefore said to be "out of accordance with structure" (Anderson, 1985; Anderson, 1988) (Fig 3). As recently as late Cretaceous time (60-65 million years ago), drainage in the present northern upper Mississippi River valley along western Wisconsin was apparently southwestward from the Wisconsin Arch toward the Cretaceous Ocean then located in a north-south alignment in the North American mid-continent (Molenaar and Rice, 1988) (Hallberg et al., 1985). This early drainage pattern is suggested by regional gradients for Cretaceous age stream gravels associated with former southwestward flowing channels whose deposits now cap various upland divides in southwestern Wisconsin, southeastern Minnesota, and northeastern Iowa. Mineralogical properties of the gravels indicate an origin in Precambrian crystalline rock highlands associated with the Wisconsin Arch in northern Wisconsin (Fig. 3). The origin and distribution of these upland fluvial gravels is evidence that a southward flowing Mississippi River, as occurs today between the states of Wisconsin to the east and Minnesota and Iowa to the west, was not present then.

The incision of the Mississippi River and the development of a relatively narrow gorge across the cuestas along southwestern Wisconsin and northwestern Illinois apparently occurred since the first glaciers invaded the region between 2.5 - 3.0 million years ago. Deep weathering profiles of apparent Miocene age, as young as about 15-5 million years, are associated with a former low relief erosion surface that pre-dates the deep incision of the Upper Mississippi Valley (Dury and Knox, 1971). Drainage of Quaternary proglacial lakes across preglacial drainage divides apparently was a principal cause of river incision through the cuestas. Glacial outwash gravel from a northwesterly source direction occurs on bluff tops along the eastern side of the Mississippi River in southwestern Wisconsin and northwestern Illinois and probably was deposited prior to incision of the river gorge (Knox, 1985; Willman and Frye, 1969; Willman et al., 1989). Glacial sediment in buried, incised valleys of northeastern Iowa is estimated to be about 500,000 years old (Hallberg et al., 1985), while glacial related sediment at the mouth of the Wisconsin River on a strath 60 m above the Mississippi River bedrock valley floor and 135 m below the adjacent upland ridge is at least about 790,000 years old (Knox and Attig, 1988). These ages support the idea that major valley incision here probably had occurred within the early Quaternary, and at least before about 800,000 years ago.

The Missouri, upper Mississippi, and Ohio tributary basins have experienced continental glaciations several times during the last 2-3 million years. Because details of glaciations older than about 300,000 years are not well understood, the classic four-stage model of Quaternary glaciations has been abandoned (Boellstorff, 1978; Hallberg, 1986; Roy et al., 2004). The early Quaternary stage names of Nebraskan and Kansan have been replaced with a simple Pre-Illinoian (pre-isotope stage 8) designation to reflect the occurrence of at least six glaciations prior to 300,000 years ago in western Iowa and Nebraska. Very little is known about the effects of Pre-Illinoian glaciations in influencing drainage in the Mississippi River drainage basin. The Illinoian glaciation (isotope stages 6-8), which occurred between about 300,000 and 130,000 years ago (Johnson, 1986), temporarily diverted the Mississippi River out of its existing course in northern Illinois (Fig 1, relict channel).

Glaciers of Wisconsin age (during isotope stage 3) began to influence the Mississippi River by about 55,000 ^{14}C yr B.P. (Leigh and Knox, 1994). During late Wisconsin glaciation (isotope stage 2), about 20,350 ^{14}C yrs B.P. (~ 24,000 cal yr B.P.), westward expansion of the glacier in northern Illinois again blocked the Mississippi River and resulted in a large northwestern Illinois lake (Curry, 1998). This proglacial lake drained over the uplands on its southwest margin and subsequently incised a new course for the Mississippi River along the present boundary between Illinois and Iowa from the lake's former location to approximately the mouth of the Iowa River system where flow joined an ancestral valley of the ancient Iowa River system (Curry, 1998) (Fig. 3). Ages of alluvial fill underlying terraces in the Mississippi Valley near the confluence with the Missouri and Illinois Rivers indicate that the middle reach, upper Mississippi River continued to aggrade after the diversion until at least about 18,800 ^{14}C yr B.P. (22,300 cal yr B.P.), but soon after began a series of cuttings and fillings (Hajic et al., 1991). Local reaches of the upper Mississippi River and its tributaries northward of the late Wisconsin terminal moraine in Wisconsin and Minnesota have morphologies and topography that reflect processes associated with movement and subsequent stagnation of glacial ice. These reaches also include floodplain and valley lakes that resulted from the post-glacial melt-out of massive ice-blocks which had been buried in proglacial outwash.

By about 14,000 ^{14}C yr B.P. (16,800 cal yr B.P.) a general northward retreat of the southern margin of the Late Wisconsin continental glacier was underway and large proglacial lakes were forming between the ice sheet and former ice front positions. The runoff regime for much of the upper Mississippi River system, where runoff was filtered through these proglacial lakes, shifted from dominance by aggradation to dominance by degradation at this time because these lakes served as sediment traps and released water of relatively low sediment concentration to the downstream river system. Furthermore, many outlets of proglacial lakes failed catastrophically and released large magnitude erosive floods into downstream valleys (Teller, 1987). One of the largest such floods occurred from Glacial Lake Agassiz, a proglacial lake that is believed to have formed about 11,700 ^{14}C yr B.P. (13,670 cal yr B.P.) in the eastern Dakotas, western Minnesota, and southern Canada between the northern drainage basin divide of the Mississippi River and the retreating ice front (Teller, 1987; Teller and Clayton, 1983) (Fig. 1). Glacial Lake Agassiz clearly originated after about 12,100 ^{14}C yr B.P. (14,100 cal yr B.P.) because ages of wood buried in or below glacial till in north-central Iowa indicate that active glacial ice was still present at this time in the north-central Iowa region south of the southern limit of Glacial Lake Agassiz sediments (Kemmis et al., 1981). Glacial Lake Agassiz drained into the Minnesota River and subsequently into the upper Mississippi River in southeastern Minnesota. Estimates of maximum flood magnitudes from Glacial Lake Agassiz have ranged from as high as 1,000,000 m^3s^{-1} (Matsch, 1983) to about 50,000 m^3s^{-1} (Wiele and Mooers, 1989). The last meltwater discharge through the southern outlet of Glacial Lake Agassiz to the Mississippi River occurred between about 9900 and 9400 ^{14}C yr B.P. (11,250 and 10,600 cal yr B.P.) (Fisher, 2003). A discharge of 42,000 m^3s^{-1} has been estimated for outflow during this interval (Licciardi et al., 1999). Flood magnitudes undoubtedly attenuated rapidly downstream from their proglacial lake sources. Observations of flood attenuations downstream from historical dam breaks show that after traveling about 100 km the peak discharge magnitude is commonly 10% or less of the outlet discharge (Costa, 1988). However, it is difficult to verify previous downstream attenuations of former proglacial floods on the Mississippi River because many of the large tributaries of the upper Mississippi may have contributed additional large discharges then. For example, stratigraphic and morphologic evidence of former channel dimensions input into a one-dimensional flow model indicated an estimated late-glacial paleoflood discharge as large as 30,000 m^3s^{-1} for the upper Mississippi River near its confluence with the Wisconsin River (Knox, 1996; Knox, 1999). The discharge magnitude seems at odds with expected downstream attenuation of the Agassiz flood(s), and probably reflects either combined discharges from multiple sources, or more likely, a large flood from a pro-glacial lake failure in the St. Croix River drainage in northeastern Minnesota and northwestern Wisconsin. Glacial Lake Duluth, a predecessor of modern Lake Superior, delivered floods to the upper Mississippi River via the St. Croix River, and discharges have been estimated as ranging between 30,000 and 45,000 m^3s^{-1} (Carney, 1996) (Fig. 1).

Don Orr Sand and Gravel Pit: Mississippi River Valley near Bagley, Wisconsin

The Orr Site occurs on the Bagley Terrace which is a very prominent erosional terrace complex that occurs below the Savanna Terrace in the Upper Mississippi Valley (Knox, 1996). The Savanna Terrace in the Upper Mississippi Valley is mostly preserved only in the lower reaches of tributaries due to erosional downcutting that formed the Bagley Terrace complex (Fig. 9). The Savanna Terrace represents the level of valley floor aggradation that the upper Mississippi River had achieved at the final maximum southern advance of the continental (Laurentide) ice sheet into the Mississippi River headwaters about 14,000 ^{14}C years B.P (16,800 calendar years ago). Because deep incision and downcutting of the Minnesota River valley, which drained Glacial Lake

Agassiz, resulted from rapid drainage of that lake (Wright Jr. et al., 1998), it has been commonly assumed that erosion of the alluvial fill below the level of the Savanna Terrace in the Upper Mississippi River valley also occurred mainly as a response to rapid drainage of Glacial Lake Agassiz. Valley and channel incision undoubtedly were promoted

Figure 9. (Photo source: (Brown, 1931)). Savanna Terrace and Bagley Terrace complex near Bagley, Wisconsin. Note the flood scoured landform assemblage on the Bagley Terrace.

both by catastrophic floods and by sustained high magnitude average meltwater discharges whose sediment loads were relatively low. However initial incision in the main valley upper Mississippi River possibly predates the drainage of Glacial Lake Agassiz. Flows from Glacial Lake Agassiz may have served only to further deepen the level of incision along the upper Mississippi River. Stratigraphic units exposed in the Don Orr sand and gravel pit help resolve the time of incision below the Savanna Terrace level.

Evidence suggesting that initial major erosion of the Late Wisconsin alluvial fill predates the Lake Agassiz drainage comes from a radiocarbon age of 13,545 ± 85 ^{14}C yr

B.P. (~ 16,300 cal yr BP) at the base of a channel incised 30+ m below the level of maximum late Wisconsin alluviation at a site about 75 km upstream of the Wisconsin River junction with the Mississippi River (Knox, 1996; Knox, 1999). This age predates the existence of Glacial Lake Agassiz because the Des Moines ice lobe at this time extended south into central Iowa and covered the area of Glacial Lake Agassiz. Because one radiocarbon date is not always reliable, further dating was needed to test the reliability of 16,300 cal yr BP radiocarbon age. The additional testing was done in conjunction with stratigraphic investigations at the Don Orr sand and gravel pit.

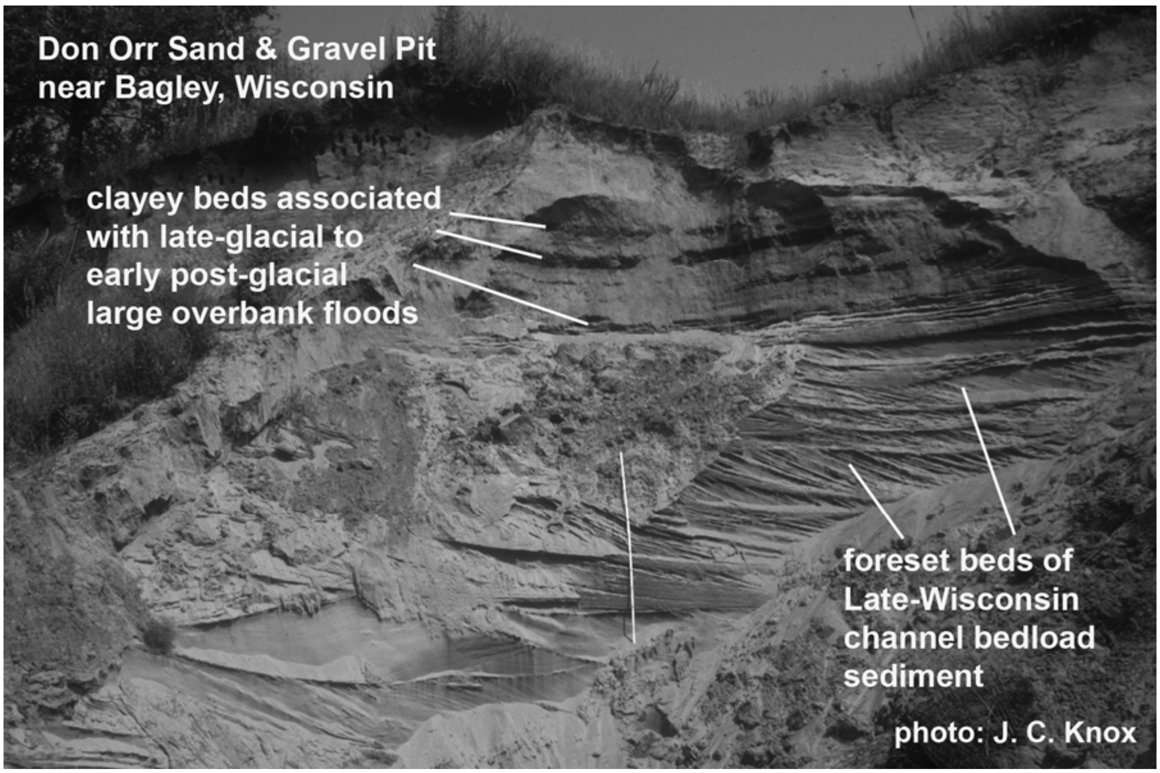

Figure 10. Clayey beds representing slackwater deposition from large overbank floods cover

Because organic materials suitable for radiocarbon dating are uncommon in the sandy deposits underlying the Bagley Terrace, the method of optically stimulated luminescence (OSL) was applied to large overbank flood deposits at the Don Orr sand and gravel pit. This technique estimates doses from ionizing radiation absorbed by minerals such as quartz and feldspars. It represents a measure of time. The method is based on the observation that electrons trapped between valence and electron bands in a crystal lattice may be freed during excitation by sunlight, and this resets the OSL clock. OSL dating therefore represents a measure of accumulated radiation dose since the last exposure of the minerals to sunlight. Seven OSL ages have been determined for the flood units overlying the Late-Wisconsin foreset beds at the Don Orr sand and gravel pit (Fig. 11). The samples were collected by J. C. Knox and L. J. Maher of the University of

Wisconsin, Madison and processed by Ron Goble at the OSL laboratory at the University of Nebraska, Lincoln. Ages determined by the OSL method typically have larger standard deviations than radiocarbon ages and are sensitive to variations in mineralogical and grain texture properties. Consequently, considerable range in ages are characteristic of the seven OSL ages representing overbank slackwater deposition from large floods that post-date deposition of Late-Wisconsin channel bed sand and gravel at the Orr sand and gravel pit (Fig. 11). Nevertheless, all of the OSL support the idea that early downcutting of 5 – 14 m occurred at the Orr location (Fig. 12). New OSL ages from the Glacial Lake Agassiz outlet at the Big Stone Moraine and oldest adjacent beaches of Glacial Lake Agassiz near the Minnesota – South Dakota border indicate the lake was present and draining between about 14,200 and 12,600 cal yr BP (Lepper et al., 2007). Therefore, given the scatter of OSL ages representing large floods at the Don Orr sand and gravel pit, it is possible that Agassiz floods contributed to the early incision of the Savanna Terrace and formation of the Bagley Terrace on the upper Mississippi River. Nevertheless, because most of

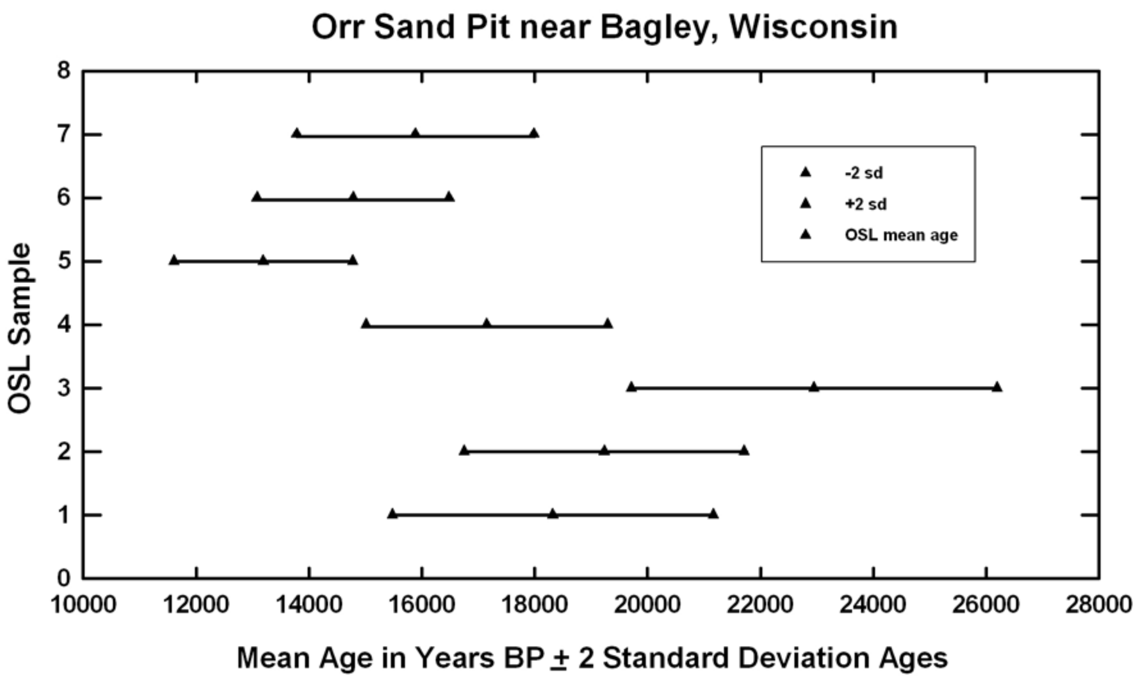

Figure 11. (Data source: J. C. Knox). Optically stimulated luminescence ages for overbank flood units at the Don Orr sand and gravel pit near Bagley, Wisconsin. The mean age is 17,360 ± 3,210 yr BP. If the anomalous sample 3 is removed, then the mean age of the remaining six samples is 16,425 ± 2,260 yr BP.

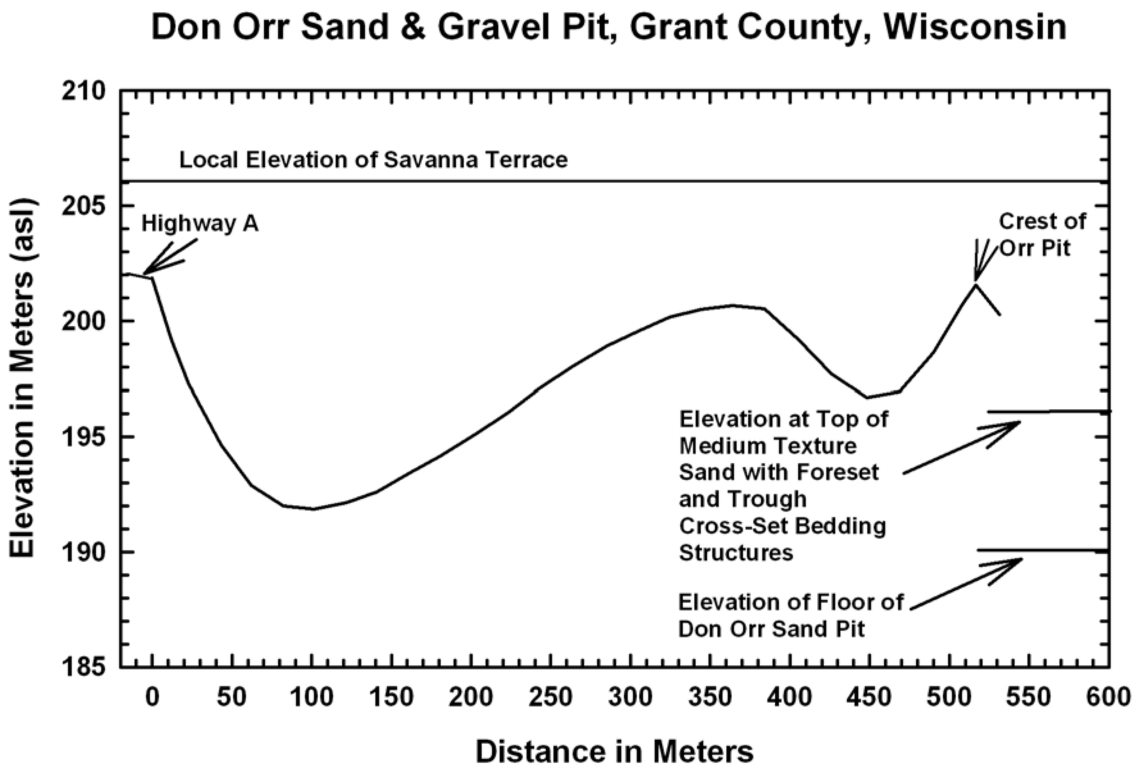

Figure 12. (Survey by J. C. Knox and C. S. Belby, May 2007). The Bagley Terrace complex between the Orr sand and gravel pit and County Highway A experienced approximately 5 to 14 m of downcutting sometime between about 16,800 and 16,400 years ago based on OSL ages shown in Figure 11. This evidence is consistent with radiocarbon evidence for rapid incision of about 30 m for the same time interval at a site 75 km up valley near Onalaska, Wisconsin.

the OSL ages shown in Figure 11 imply downcutting prior to 14,000 cal yr BP, their collective ages support the idea that formation of the Bagley Terrace here probably predates the drainage of Glacial Lake Agassiz. It also is noteworthy that when an average OSL age is computed from the six or seven available OSL ages at the Orr site, the mean age is relatively similar to the radiocarbon age of 16,300 cal yr BP obtained from the base of an alluvial fan that fills a channel incised below the Savanna Terrace and is inset into the Bagley Terrace at Halfway Creek near Onalaska, Wisconsin (Fig. 2). A probable source for early incision of the Savanna Terrace is Glacial Lake Duluth which also discharged large flows into the upper Mississippi River via the St. Croix River as noted earlier (Fig. 2). The abundance of red clay sediment in the overbank deposits of large floods recorded at the Orr site, and elsewhere in the area, have a Lake Superior source region. Therefore, these red clayey sediments provide additional support for the notion that floods and runoff responsible for the early incision of the Late Wisconsin valley fill may have entered the Mississippi River from the St. Croix River.

Boice Creek Alluvium

Boice Creek Cut-Off Valley Meander: Boice Creek near Potosi, Wisconsin
Dr. James Knox

The Boice Creek site provides an example of cut-off valley meanders that are common features in the lower reaches of valleys of upper Mississippi River tributaries in SW Wisconsin, NE Iowa, and SE Minnesota. Their origins relate to aggradation of the lower reaches of these tributaries promoted by a downstream raised base level during times when the Mississippi River was either ponded by glacial ice blockage or was experiencing massive aggradation by glacial-related sedimentation (Knox, 1982). Comparison of elevations of bedrock valley floors under cut-off valley meanders with bedrock valley floors of their adjacent modern channels indicates that at least two different ages of valley meander cut-offs are present.

The fills of the cut-off valley meanders are an important source of information about the region's natural history because much of the older stratigraphic records have been destroyed by normal fluvial and mass-wasting (gravity) forces during the long history of landscape evolution. Perhaps one of the most surprising findings is evidence that rather large magnitude valley incision into bedrock has occurred during relatively late Quaternary time as noted in the caption of Figure 13. Approximately 220 m of drainage incision has occurred since the (Cretaceous) Windrow Formation river gravels were deposited on upland interfluves. The paleo-magnetic properties of the Bridgeport sediment in the lower Wisconsin River valley show that deep incision of the Driftless Area topography had occurred by about 790,000 yrs BP. The occurrence of Pre-Illinoian sediment, with normal remanent magnetism and clay mineralogy similar to the ~500,000 yr BP Wolf Creek till of eastern Iowa, implies, since then, 15-25 m of further incision has occurred below the bedrock floor of the Boice Creek cut-off valley meander. The occurrence of a well-developed Sangamon paleosol on a low elevation valley margin 8-12 m above the bedrock floor of the nearby Little Platte River indicates only modest bedrock incision has occurred since about 125,000 years ago. During the Late-Wisconsin, the Driftless Area was characterized by tundra and extreme mass wasting on hillslopes. The Late Wisconsin periglacial climate resulted in massive alluvial accumulations of cobble and boulder gravels in the valleys. Holocene streams are mainly restricted to lateral migration on the surface of these cobble and boulder gravels because they normally due not have sufficient flow energy to incise them.

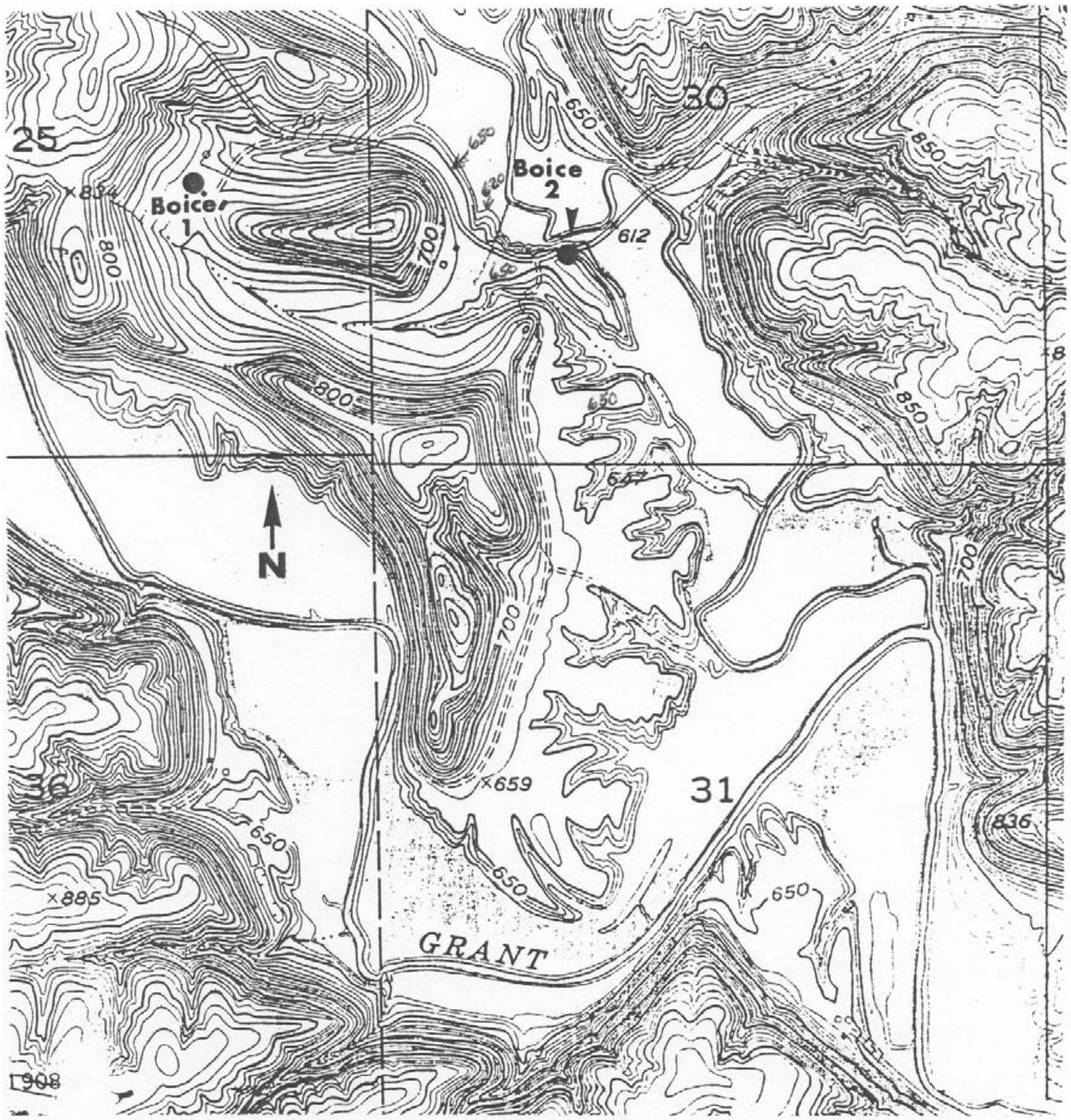

Figure 13. (Data source: J. C. Knox). Boice Creek cut-off valley meander SE1/4, Sec. 25, T3N, R4W, 4th PM, Balltown, Wisconsin, USGS 7.5 minute topographic quadrangle. The elevation of the ground surface at Boice Site 1 is 218 m asl and is 23 m above its underlying bedrock floor at 195 m asl. Boice Site 2 is located on the Savanna Terrace which averages approximately 198-200 m asl in this lower valley. The elevation of the modern floodplain of Boice Creek is 186 m asl, and the floodplain is underlain by 15 m of alluvial fill with the bedrock floor underlying present Boice Creek being at 171 m asl. The bedrock floor of the cut-off valley meander is therefore 24 higher than the bedrock floor underlying present-day Boice Creek. The sedimentary fill in the cut-off is represented by normal remanent magnetism throughout. This suggests that the 24 m of bedrock incision, since the cut-off formed, occurred since about 800,000 years ago. The fills underlying modern stream valleys in the Driftless Area appear to be primarily represented by Wisconsin age alluvium, most of which is Late Wisconsin age. Regional tectonic uplift appears to be primarily responsible for the post-800,000 yr BP incision, but climate changes and drainage modifications may also have contributed.

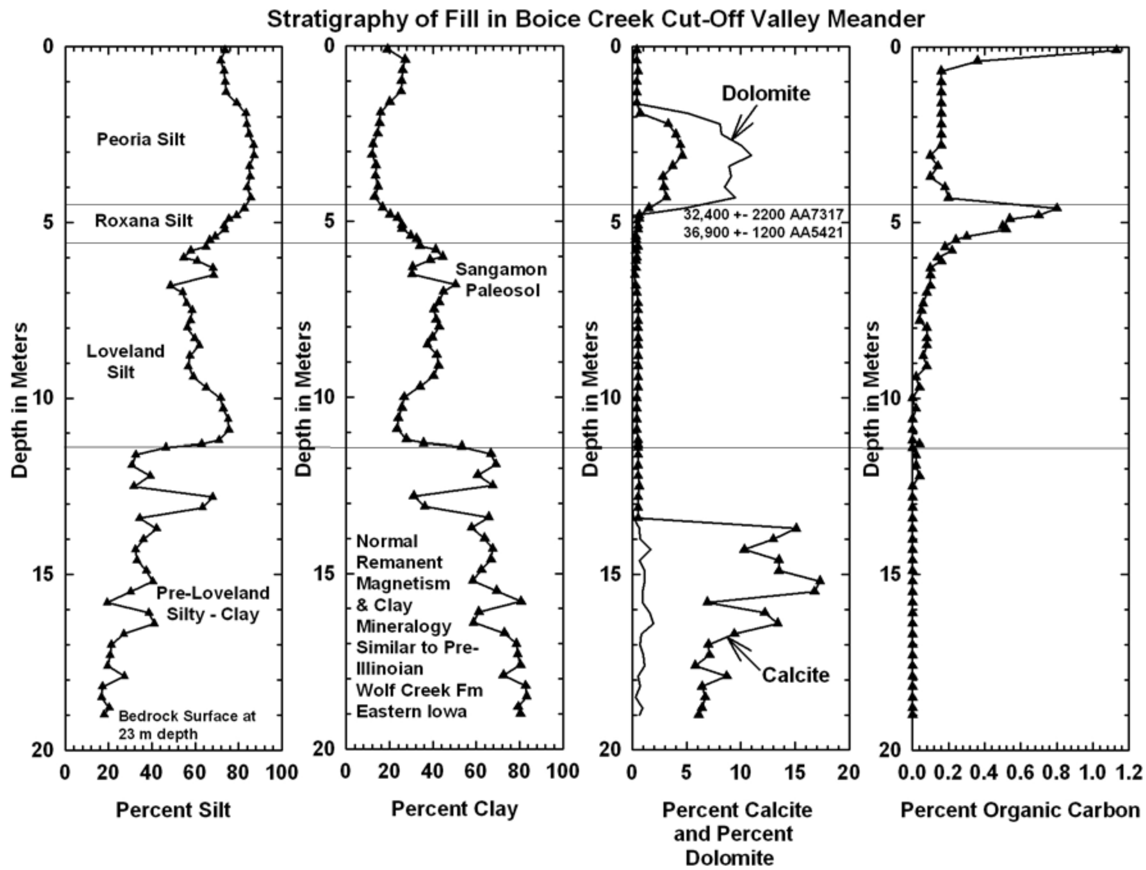

Figure 14. (Data source: J. C. Knox). The basal stratigraphic unit in Boice Creek cut-off valley meander is typically 60-80% clay composition which suggests that it accumulated in a very low energy environment such as a lake. Further evidence in support of lake deposition is the very high percentage of calcite coupled with very low percentages of dolomite. Ordinarily, in this local environment one would expect to see higher percentages of dolomite than calcite if the sediments were derived from eolian processes as illustrated by the Peoria Silt in the top unit. The anomalous high calcite percentage apparently results from precipitation in a shallow lake environment prior to burial by younger sedimentary units. Note that the top 2+ m of the basal unit has lost its carbonates due to weathering and soil formation. Clay mineralogy of the basal unit is similar to that of the ~ 500,000 yr BP Wolf Creek glacial till in eastern Iowa (Hallberg, 1980). Glacial outwash gravel from a westerly source occurs on uplands at an elevation of 250 m asl along the eastern margins of the Mississippi River about 4 km west of the Boice site. The glacial advance responsible for these upland outwash gravels probably caused the blocked drainage that produced the basal unit in the Boice cut-off valley meander. The sharp increase in silt at a depth of about 11.4 m is interpreted as the beginning of Loveland Silt deposition (Illinoian age). Between about 7.5 -10 m depth the Loveland unit is represented by thinly bedded brown and grayish brown silty clay interstratified with 3 – 5 mm thick reddish clayey beds indicative of slackwater ponding. Ponding at this time is consistent with blockage of the Mississippi River by Illinoian glacial ice in western Illinois (Fig. 2). The silt-dominated Loveland unit contains a very well-developed Sangamon paleosol in its top two meters (5.7 - ~ 7 m depth). The paleosol at the top of the Loveland Silt is overlain by mid-Wisconsin age, reddish brown, Roxana Silt that was deposited in the area between about 55,000 and 28,000 ^{14}C yr BP (Leigh and Knox, 1993). Spruce charcoal is common within the Roxana Silt and suggests that a boreal forest environment probably characterized the region then. The top unit in the Boice Creek cut-off valley meander is represented by Late-Wisconsin age Peoria Silt. Note that the depth of Holocene weathering in the Peoria Silt is very similar to that developed in the Pre-Loveland silty clay of the basal unit.

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St. John's Mine

129 South Main Street
Potosi, WI
Dr. Bruce A. Brown

See Appendix A for original outcrop description a portion is reprinted here.

Location: Opening is in valley wall on the north side of State Highway 133 about 0.2 miles south of intersection of County Highway "0" and State Highway 133 in Potosi in the SW ¼ , NW ¼ **,** SW ¼ , Sec. 34, T. 3 N., R. 2 E., Grant County (Potosi 7.5-minute topographic quadrangle, 1972).

Author: M.G. Mudrey, Jr., (Modified from St. John Mine brochure, L.C. Ihm**,** owner, and Whitlow and West, 1966).

Description: This mine is a natural cave that was extensively exploited for lead prior t o 1870. By 1843, it had yielded 250,000 pounds of lead. The Potosi sub-district produced 21,300 tons of 80 percent lead from 1862 t o 1876. Galena occurs in gash veins and openings along minor joints. The vein strikes N. **65'** W e , and is noted for its length and continuity. Host rock is Ordovician Galena Dolomite, with Maquoketa Shale on the ridge to the west. The floor of the cave is in the Dunleith Member (cherty lower unit) of the Galena Dolomite. In most outcrops, it is a pale-yellowish-brown to light-olive-gray and grayish-orange fine- to medium-grained vuggy fossiliferous dolomite containing abundant chert as nodules or as nearly continuous layers. Chert in the Dunleith Member is nodular and distributed parallel to the bedding. Near mineralized zones chert is selectively mineralized and contains microscopic grains of disseminated iron sulfide that color it bluish gray and locally very dark gray. The top of the cherty unit is marked by two discontinuous layers of chert nodules separated from the main cherty section by 6-9 feet of non-cherty dolomite.

The roof of the cave is in the Wise Lake Member (non-cherty upper unit) of the Galena Dolomite. The strata of the non-cherty unit are pale-yellowish-brown to yellowish- and grayish-orange fine-grained porous fossiliferous dolomite. The minerals of the zinc and lead deposits in the Potosi quadrangle are mostly simple sulfides, carbonates, and sulfates. The primary sulfide minerals are sphalerite, galena, pyrite, marcasite, chalcopyrite, and digenite. Galena is fairly stable and persists above the water table.; the others are commonly altered. These include smithsonite, cerussite, limonite, melanterite, malachite, azurite, and erythrite.

History: St. John Mine, originally a natural cave, was first named LaSalle Cave, after Robert Cavelier Sieur de La Salle, an early French explorer in North America, who traveled with his company on an expedition through the upper Mississippi River Valley in 1679 and again in 1687 after King Louis XIV names him Viceroy of North America. LaSalle is the man who claimed and named Louisiana Province for the French king.

St. John Mine was worked by the Indians many years before white pioneers arrived in the 1827 "lead rush". Drifts of the old mine follow the natural crevices filled with stalactites. The foxes who used it for dens are said to have uncovered the rich lead deposits near the entrance by digging and running in and out the natural cave crevice. The Indians mined galena for barter but it was left to the white men to extensively develop these diggings.

The first white man known to have worked St. ,John Mine and who gave it the name it still bears was Willis St. John, who made a small fortune from this mine between 1828 and 1870. .

In the Upper Mississippi valley, lead seems to have been discovered about 1692 by Nicholas Perrott. This metal was also noted in 1700 by LeSueur, who took lead out of a place which we believe from the description must have been Snake Hollow, now Potosi, Wisconsin. In 1766 John Carver brought to St. Louis a 500 pound hunk of lead he had received from barter with the Indians who mined a cave on the eastern Mississippi bank somewhere between the mouth of the Grant and Platte Rivers. This 500 pound piece of lead may have been taken from St. John Mine, which points to the importance St. John Mine played in bringing settlers to the lead region. and "lead rush of 1827", the convening of the first Wisconsin Territorial Legislature in 1836, Potosi and its suburbs (La Fayette, Van Buren, Dutch Hollow British Hollow, Buena Vista, and Rockville) flourished. Potosi in 1838 was hoping to become the capital of Wisconsin; first state capitol was Belmont, but Madison won out. The Mexican War of 1847; the Gold Rush of '49 and the cholera epidemic in 1854 depleted its citizens for a few years; but by 1859 when the Civil War broke out, production of lead, and with it the growth of the village of Potosi, was on an upswing.

Well over two-thirds of all lead for the North was supplied during the Civil War by the Galena, Benton, New Diggings, Shullsburg, Mineral Point and Potosi mines. The remainder was furnished by mining towns called Platteville Hardscrabble, Yuba, and Meeker's Grove, all in the southwestern Wisconsin zinc-lead region.

References: Whitlow, J. W., and West, W. S., 1966, Geology of the Potosi quadrangle, Grant County, Wisconsin, and Dubuque County, Iowa: U.S. Geol. Survey Bull. 1123-1, p. 533-571.

Heyl, A. V., Jr., Agnew, A. F., Lyons, E. J., and Behre, C. H., Jr., 1959, The Geology of the Upper Mississippi Valley Zinc-Lead District: U.S. Geol. Survey Prof. P~per 309,310 p.

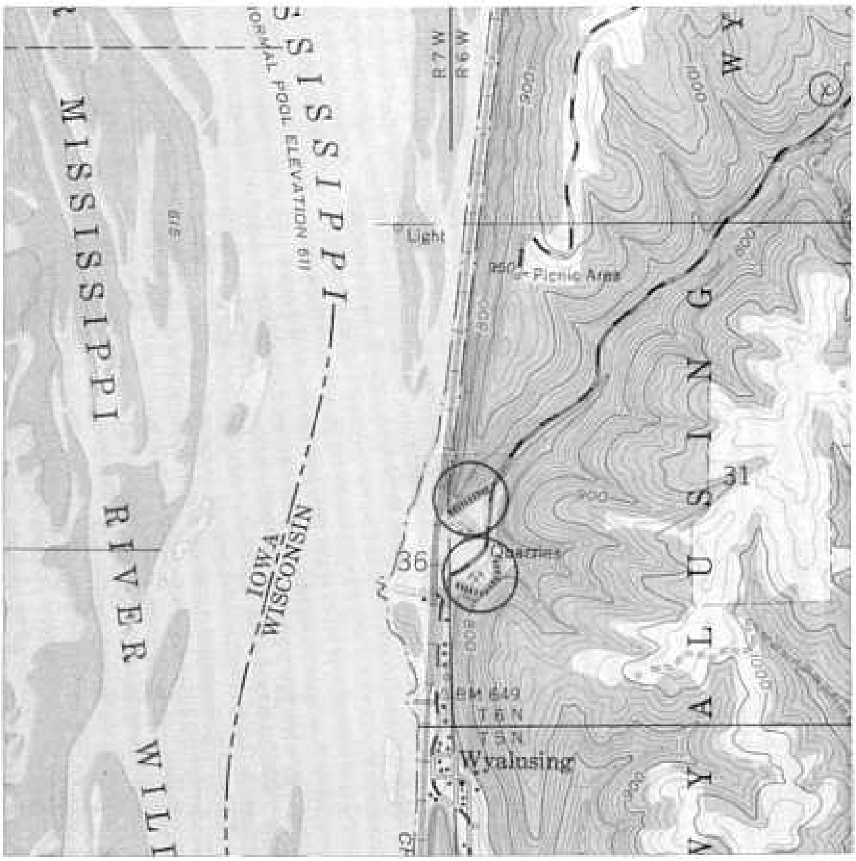
Ibm, L. C., undated, StO John Mine Brochure, Potosi, Wisconsin
39 500-3J8TO26-78

APPENDIX A

WISCONSIN GEOLOGICAL AND NATURAL HISTORY SURVEY OUTCROP DESCRIPTIONS

Title: Wyalusing

Location: Chicago, Burlington, and Quincy Railroad quarry on County Highway "X" at north edge of the community of Wyalusing in the NW$\frac{1}{4}$, SW$\frac{1}{4}$, Sec. 31, T.6N. R.6W., Grant County (Clayton 7.5-minute topographic quadrangle, 1962).



Author: M. E. Ostrom (modified from Starke, 1949, Shea, Cline 1959 and 1960, and Ostrom and Cline 1970).

Description: This quarry, exposing approximately 185 feet Prairie du Chien strata, is probably the most nearly complete exposure of the Prairie du Chien in the state of Wisconsin. George Starke (1949) described the rocks exposed in this quarry and estimated that the floor of the quarry lies some 30 feet above the base of the Oneota Formation.

The Willow River Formation at this site was studied in detail by Carozzi and Davis (1964).

Angular unconformity of beds at the Oneota/New Richmond contact is clearly shown on the high southeast quarry face. Here eastward dipping beds of Oneota Dolomite are overlain by flat-lying beds of the New Richmond Member.

There is an excellent exposure of the Oneota and Shakopee Formations in a quarry at the north village limits of Bagley. It can be reached by following

CHICAGO, BURLINGTON AND QUINCY RAILROAD QUARRY
$NW\frac{1}{4}$, $SW\frac{1}{4}$, Sec.31, T.6N., R.6W.

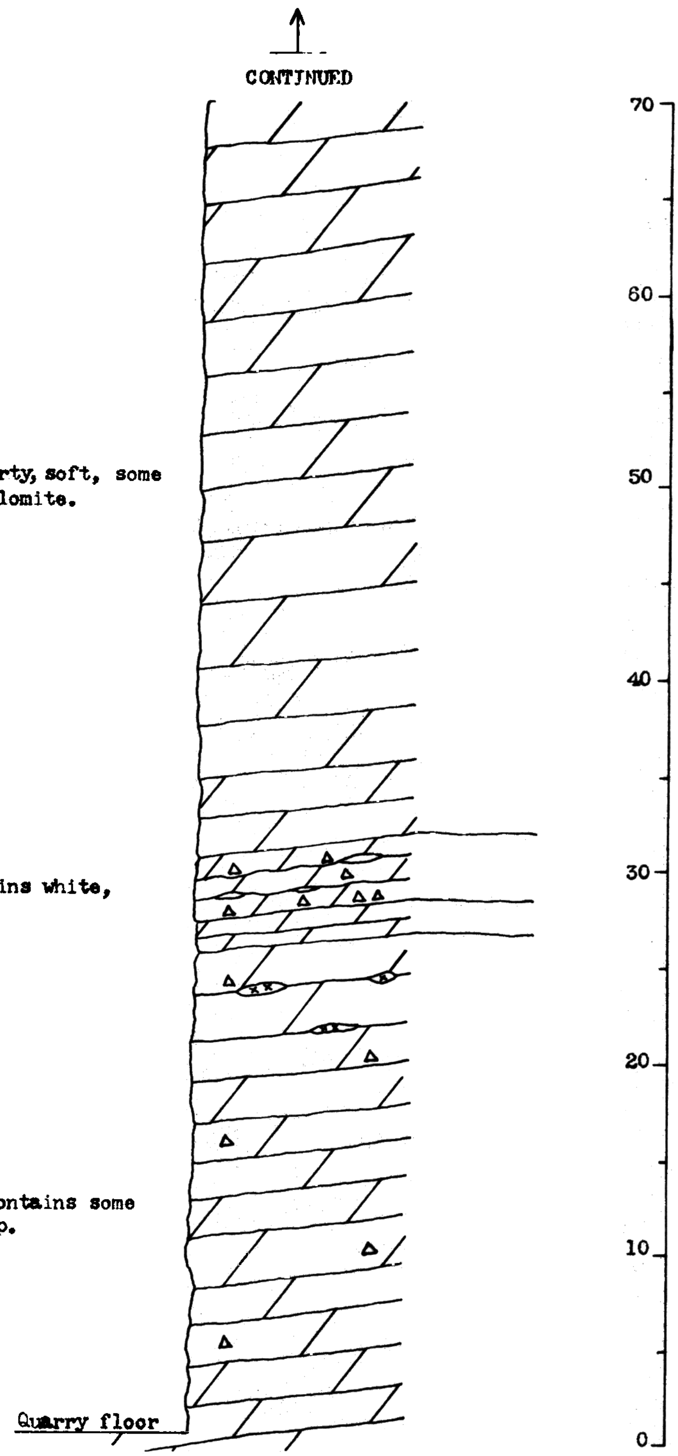
Gray, very thick-bedded, non-cherty, soft, some areas fine-grained gray buff dolomite.

ONEOTA FM.

Very fine-grained, undulating bed, contains white, gray and black chert; cavernous.

Gray to buff, coarse-grained, in 4" beds.

Gray to buff to orange, thick-bedded; contains some white chert; some clay pockets near top.



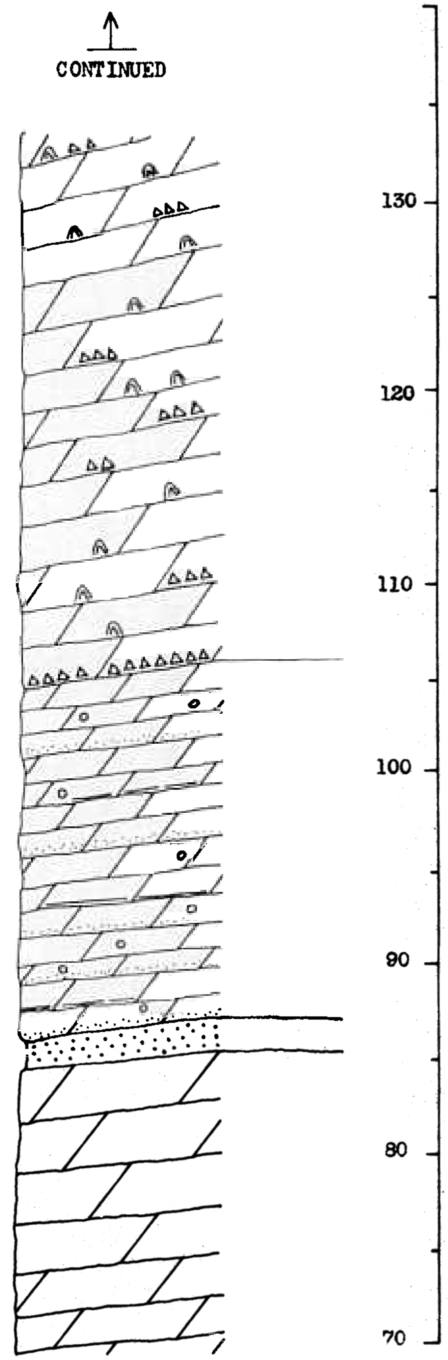
CONTINUED

Algal structures profuse; stringers of chert throughout; basal foot is weathered chert.

Conglomeratic with dark gray pebbles; well-bedded with some green shale partings and some clean white quartzose beds.

White, medium-grain; some green specks.

Gray, very thick-bedded, non-cherty, soft, some areas fine-grained gray buff dolomite.



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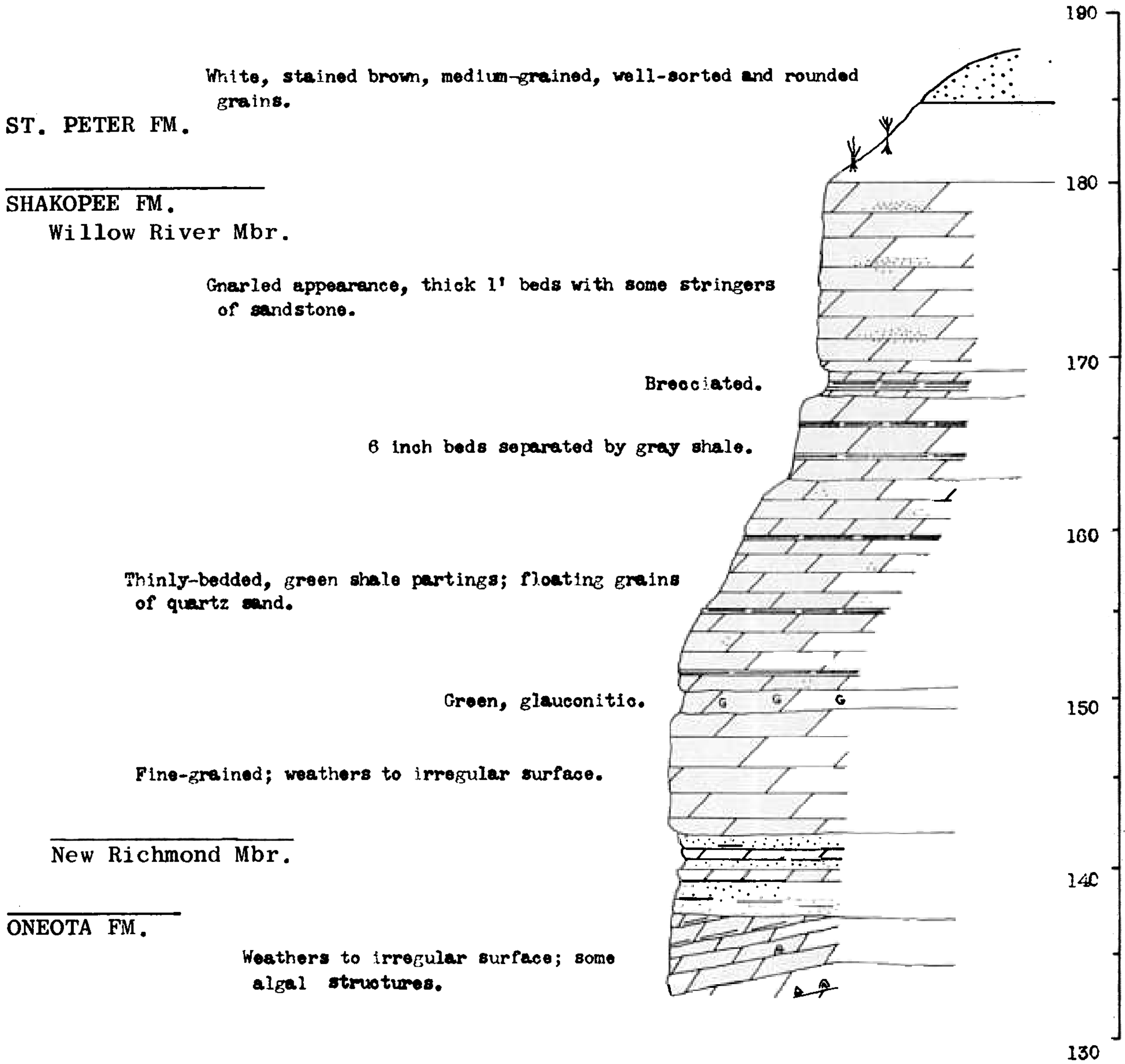


190
White, stained brown, medium-grained, well-sorted and rounded grains.
ST. PETER FM.
180
SHAKOPEE FM.
Willow River Mbr.
Gnarled appearance, thick 1' beds with some stringers of sandstone.
170
Brecciated.
6 inch beds separated by gray shale.
160
Thinly-bedded, green shale partings; floating grains of quartz sand.
150
Green, glauconitic.
G
G
G
Fine-grained; weathers to irregular surface.
New Richmond Mbr.
140
ONEOTA FM.
Weathers to irregular surface; some algal structures.
130

County Highway "X" south for a distance of 3.5 miles from the Wyalusing Quarry Stop. The Bagley quarry is an especially good location to examine fossil algae.

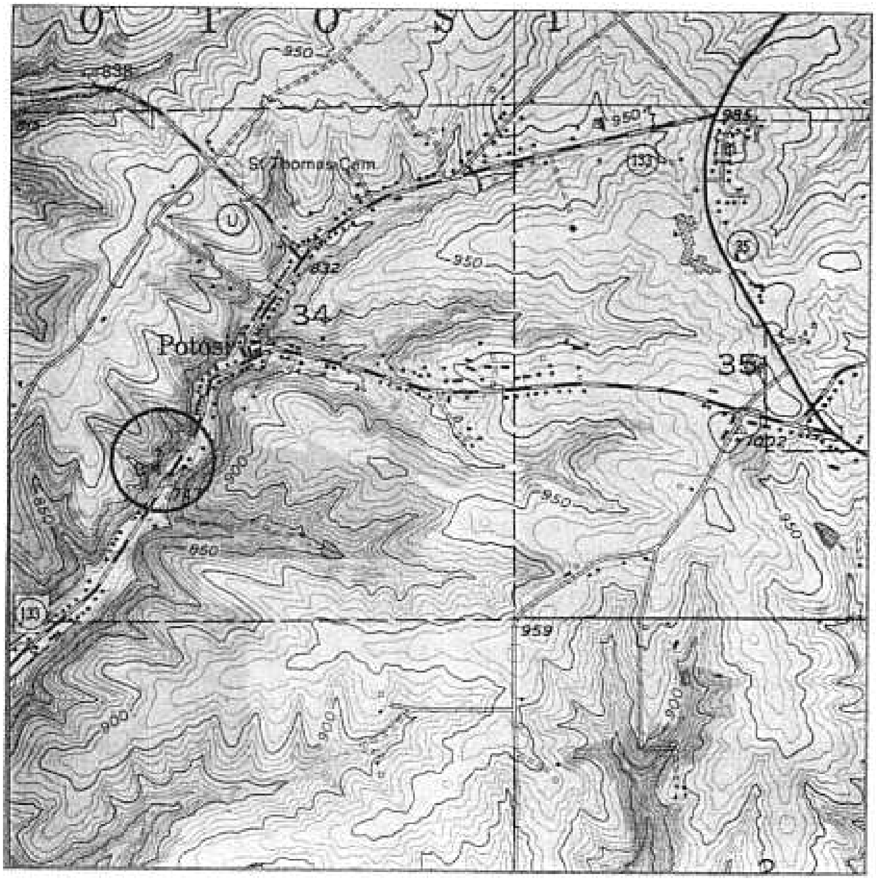
Significance: Exposure clearly illustrates the contact relationship of the Shakopee and Oneota Formations.

What is the relationship? What is your interpretation? Have you seen a similar relationship at previous stops? Closely examine the dolomite. Can you find fossils and what are they? Note the oolites. What do they signify? Note the beds of stromatolitions. What do they signify? What was the environment of deposition of the Oneota Formation?

References: Starke, 1949; Shea, 1960; Cline 1959 and 1960; Ostrom and Cline, 1970; Carozzi and Davie, 1964.

Title: St. John Mine (Snake Cave)

Location: Opening is in valley wall on the north side of State Highway 133 about 0.2 miles south of intersection of County Highway "O" and State Highway 133 in Potosi in the SW$\frac{1}{4}$, NW$\frac{1}{4}$, SW$\frac{1}{4}$, Sec. 34, T. 3 N., R. 2 E., Grant County (Potosi 7.5-minute topographic quadrangle, 1972).



Author: M. G. Mudrey, Jr., (Modified from St. John Mine brochure, L.C. Ihm, owner, and Whitlow and West, 1966).

Description: This mine is a natural cave that was extensively exploited for lead prior to 1870. By 1843, it had yielded 250,000 pounds of lead. The Potosi sub-district produced 21,300 tons of 80 percent lead from 1862 to 1876. Galena occurs in gash veins and openings along minor joints. The vein strikes N. 65° W., and is noted for its length and continuity.

Host rock is Ordovician Galena Dolomite, with Maquoketa Shale on the ridge to the west.

The floor of the cave is in the Dunleith Member (cherty lower unit) of the Galena Dolomite. In most outcrops, it is a pale-yellowish-brown to light-olive-gray and grayish-orange fine- to medium-grained vuggy fossiliferous dolomite containing abundant chert as nodules or as nearly continuous layers. Chert in the Dunleith Member is nodular and distributed parallel to the bedding. Near mineralized zones chert is selectively mineralized and contains microscopic grains of disseminated iron sulfide that color it bluish gray and locally very dark gray.

The top of the cherty unit is marked by two discontinuous layers of chert odulesn separated from the main cherty section by 6-9 feet of non-cherty dolomite.

The roof of the cave is in the Wise Lake Member (non-cherty upper unit) of the Galena Dolomite. The strata of the non-cherty unit are pale-yellowish-brown to yellowish- and grayish-orange fine-grained porous fossiliferous dolomite.

The minerals of the zinc and lead deposits in the Potosi quadrangle are mostly simple sulfides, carbonates, and sulfates. The primary sulfide minerals are sphalerite, galena, pyrite, marcasite, chalcopyrite, and digenite. Galena is fairly stable and persists above the water table; the others are commonly altered. These include smithsonite, cerussite, limonite, melanterite, malachite, azurite, and erythrite.

History: St. John Mine, originally a natural cave, was first named LaSalle Cave, after Robert Cavelier Sieur de La Salle, an early French explorer in North America, who traveled with his company on an expedition through the upper Mississippi River Valley in 1679 and again in 1687 after King Louis XIV names him Viceroy of North America. LaSalle is the man who claimed and named Louisiana Province for the French king.

St. John Mine was worked by the Indians many years before white pioneers arrived in the 1827 "lead rush". Drifts of the old mine follow the natural crevices filled with stalactites.

The foxes who used it for dens are said to have uncovered the rich lead deposits near the entrance by digging and running in and out the natural cave crevice. The Indians mined galena for barter but it was left to the white men to extensively develop these diggings.

The first white man known to have worked St. John Mine and who gave it the name it still bears was Willis St. John, who made a small fortune from this mine between 1828 and 1870.

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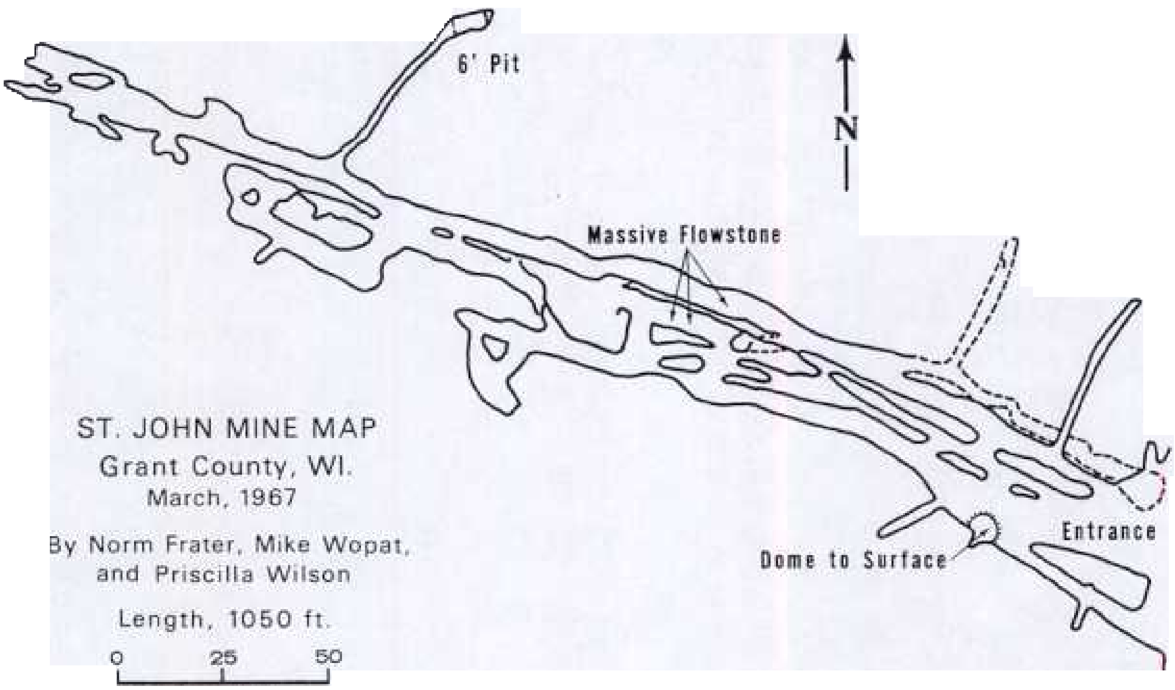
Ihm, L. C., undated, St. John Mine Brochure, Potosi, Wisconsin



Title: Pigeon Creek

Location: Exposures in roadcuts and streams cuts along County Highway "N" in the SE¼, NE¼, Sec. 19, T.4N., R.3W., Grant County (Hurricane 7.5-minute topographic quadrangle, 1962).

Author: M. E. Ostrom

Description: Description begins at top with lower massive bed of dolomite and extends downward to the St. Peter Formation.

ORDOVICIAN SYSTEM

Platteville Formation

Pecatonica Member (1.8 feet)

21.5' - 23.3'	1.8'	Dolomite, buff, dense in upper part, slightly porous in lower 10": Phosphatic nodules abundant in lower 10".

Glenwood Formation

Hennipen Member (1.8 feet)

21.4' - 21.5'	0.1'	Shale, weathers green, otherwise very dark brown and blocky. Slakes easily.

19.7' - 21.4'	1.7'	Shale, grayish brown, dolomitic, thinly laminated, blocky fracture (yeast-like). Estimated 50% carbonate.

Harmony Hill Member (2.2')

19.5' - 19.7'	0.2'	Shale, iron-rich; phosphatic grains and orange and black specks.
17.5' - 19.5'	2.0'	Shale, bright greenish gray in upper 10" changing to light gray in middle and to yellowish gray at base. Laminated and at top, and unlaminated in base part. Yellow clay slakes easily; green clay slakes.

Nokomis Member (2.5 feet)

17.0' - 17.5'	0.5'	Sandstone, limonite-centered, hard, forms ledge of irregular thickness; burrowed.
15.5' - 17.0'	1.5'	Sandstone, hard, limonite- and dolomite- silty. bed. Burrowed.
15.0' - 15.5'	0.5	Sandstone, coarse-grained to conglomerative, white to very light gray, slightly irregular base with staining below.

St. Peter Formation

0.0' - 15.0'	15'+	Sandstone, very light gray, medium and coarse-grained, well-sorted, thick-bedded.

BASE OF EXPOSURE

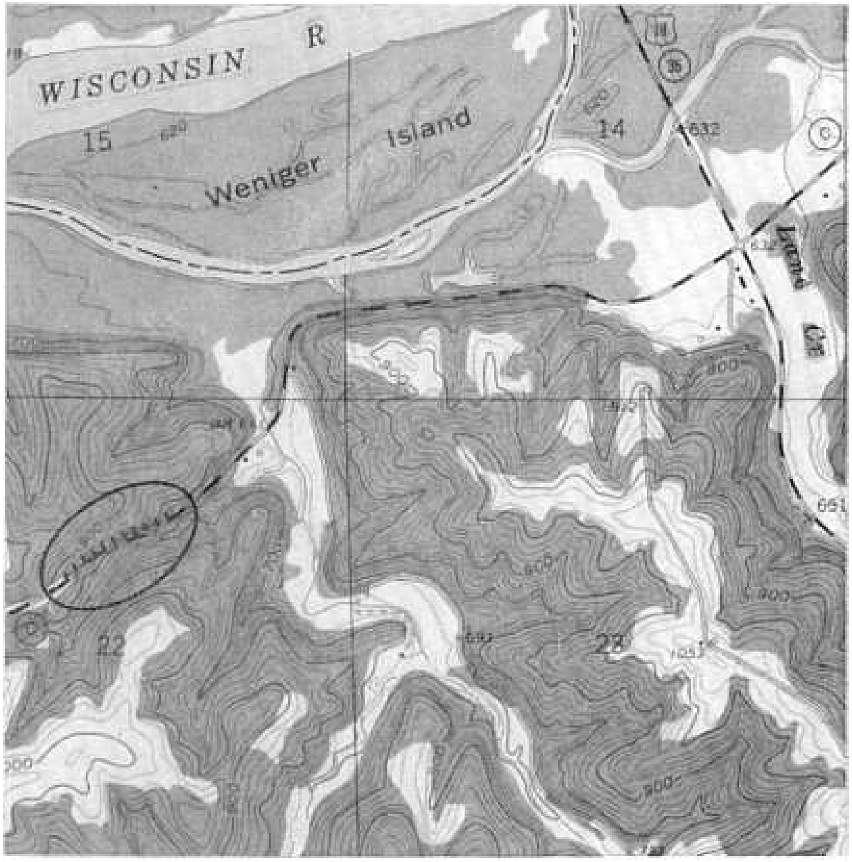
Significance: Exposure of section similar to that seen at Bridgeport West stop.

Note the individual units. Are they distinct and persistent? Recall exposures seen at Viroqua/Readstown and Bridgeport West stops. How do they compare with this exposure? Have you seen the phosphatic pebbles before? What do they signify?

References: Ostrom 1969.

Title: Bridgeport West

Location: Exposures in roadcut at north side of County Highway "C" in the SE$\frac{1}{4}$, NW$\frac{1}{4}$, NE$\frac{1}{4}$, Sec. 22, T.6N., R.6W., Grant County (Bagley 7.5-minute topographic quadrangle, 1962).



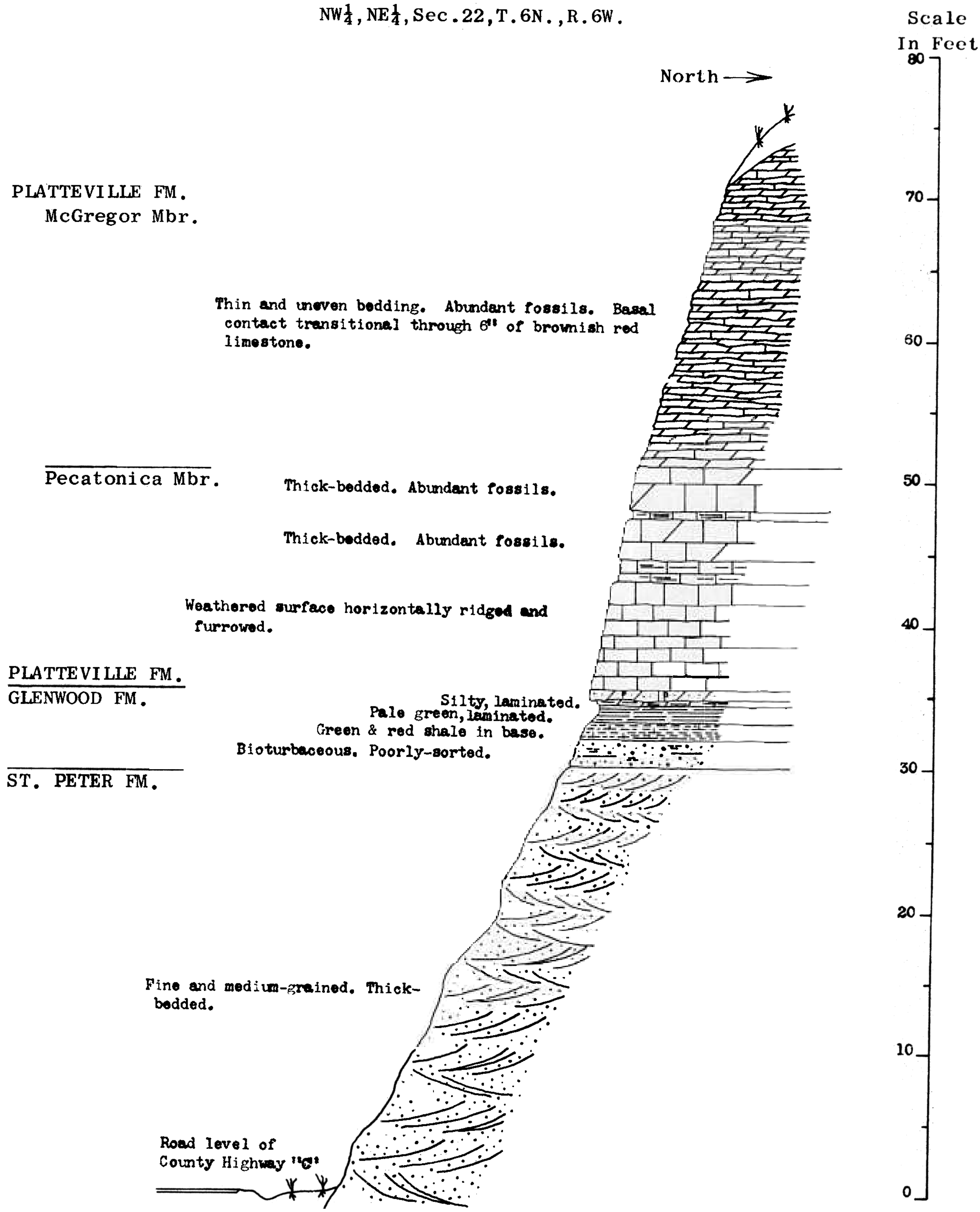
Author: M. E. Ostrom (modified from Ostrom, 1970)

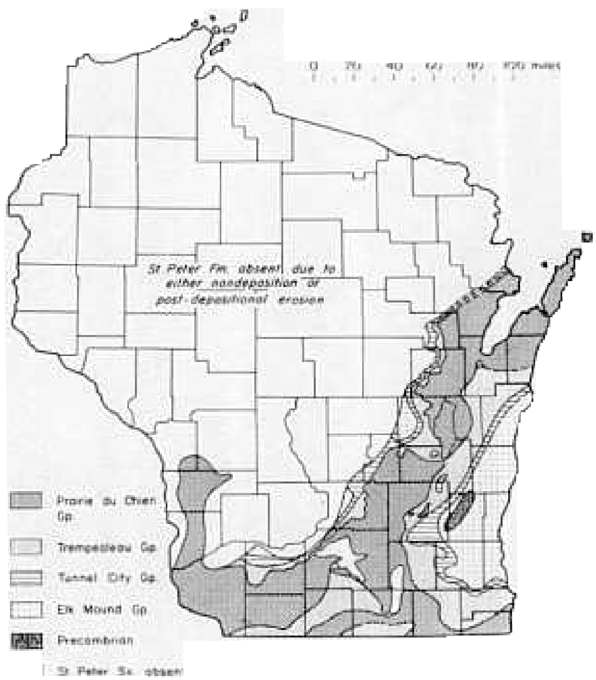
Description: At this site all four lithotopes of the uppermost cycle, which has the St. Peter Sandstone in its base, are exposed as are the contact relationships of St. Peter, Glenwood and Platteville Formations.

At the Prairie du Chien Stop the erosion surface at the base of the St. Peter was especially well shown. Dapples (1955) demonstrated that this erosion surface could be traced over a broad area extending from western Tennessee to Wisconsin. A map by Ostrom shows the geology of the pre-St. Peter surface in Wisconsin and indicates that the St. Peter rests on successively older rocks as one proceeds from western Wisconsin to the vicinity of Milwaukee. It is postulated that there was pre-St. Peter uplift in the vicinity of Milwaukee and that the uplifted surface was subsequently eroded. Although data are sparse there is very good agreement between thick sections of St. Peter which are interpreted to coincide with Pre-St. Peter surface which were likely exposed by erosion. The data suggest stream drainage to the southwest away from the Milwaukee area.

Bridgeport West

ROADCUT ON COUNTY TRUNK "C"
NW$\frac{1}{4}$, NE$\frac{1}{4}$, Sec. 22, T.6N., R.6W.





Paleogeology of the pre-St. Peter erosion surface in southern and eastern Wisconsin.

The thick-bedded quartzarenite is represented by the St. Peter Sandstone and the reworked poorly-sorted quartzarenite by the lower 2 feet of the Glenwood Formation. The reworked quartzarenite is transitional through about 1.4 feet of siltstone and shale into the shale lithotope and this is in sharp contact with the base of the overlying carbonate lithotope which is the Platteville Formation.

The Glenwood Formation in Wisconsin thins to the east toward the Wisconsin Arch from a maximum thickness of about 13 feet in the vicinity of Beetown to less than a foot near New Glarus. A study by Ostrom (1969) showed that as one proceeds eastward the upper or shale unit thins to disappearance and the underlying poorly-sorted reworked quartzarenite is in direct contact with the overlying carbonate lithotope. Traced further to the east, in the vicinity of New Glarus south of Madison, the reworked quartzarenite thins to less than 1 foot.

The contact of the Glenwood Formation with the Platteville Formation is one of apparent unconformity. This relationship is attributed to lateral variations in environmental conditions at the time of formation rather than to post-Glenwood erosion because there is no evidence except the apparent regional truncation described above that would indicate the Glenwood was eroded.

This limit of the New Richmond Sandstone is along a line trending southwestward from about Danville, in east-central Illinois, toward Cape Girardeau, Missouri (Workman and Bell, 1949).

The New Richmond developed in the littoral zone during the succeeding cycle. It is succeeded by poorly known or defined deposits of the nondepositional and depositional shelf zones similar to those of the preceding Jordan cycle, and is overlain by deposits of the carbonate zone, the Shakopee Dolomite, developed further seaward. In more seaward areas to the south the Shakopee Dolomite is continuous with the Oneota Dolomite of the preceding Jordan cycle and consists almost entirely of carbonate. Northward, as for example, near Utica in north-central Illinois, the Shakopee overlies the New Richmond and has a variable lithology which consists of dolomite containing layers of quartzarenite, shale, and discontinuous thin beds of oolitic chert. The dolomite beds range up to 10 feet in thickness, are seldom more than 3 feet thick, and are commonly very fine-grained, and their upper surfaces may be ripple marked and mudcracked. The sandstone beds may be cross-bedded and commonly contain pebbles and cobbles derived from the underlying dolomite bed in their lower part. Beds of shale reach a known maximum thickness of 6 inches. The variable lithologic character of the Shakopee in this area is interpreted to indicate frequent environmental changes and "....fluctuation of conditions of sedimentation characteristic of shallow water deposition" (Cady, 1919). It is postulated that the Shakopee Formation accumulated in a very shallow environment situated shoreward from an area of algal headlands. This zone is considered to have been a broad, flat, and shallow lagoon or shoaling area subjected to the influence of the land on one side and the algal headlands and reefs on the other, while at the same time being influenced by other factors affecting the carbonate deposition.

The Shakopee Dolomite and older strata were eroded in northerly areas in Wisconsin during pre-St. Peter regression to an indefinite northeast-trending strandline through western Kentucky (Dapples, 1955). The geology of the eroded surface is shown in the accompanying diagram (Ostrom, 1964). The surface is one of prominent relief.

Significance: This is an opportunity to examine all four lithotapes of the St. Peter cycle and their contact relationships.

How do the lithotapes exposed here compare to those seen at the Viroque/Readstown stop? What are the differences between this cycle and the Galesville cycle seen at the Bruce School, Galesville, Duerch Peak, and Mt. Zion stops? What is the environmental and historical significance of these differences?

References: Dapples, 1955; Ostrom, 1964 and 1970.

Middle Ordovician rocks at Potosi Hill, Wisconsin

M. E. Ostrom, Wisconsin Geological and Natural History Survey, 3817 Mineral Point Road, Madison, Wisconsin 53705

LOCATION

Roadcut at east side of U.S. 61 and Wisconsin 35 about 1 mi (1.6 km) northwest of bridge over Platte River and 4 mi (6.4 km) northwest of Dickeyville in the SE¼,NW¼,Sec.7, T.2N.,R.2W., Grant County, Wisconsin in the Potosi 7½-minute Quadrangle (Fig. 1).

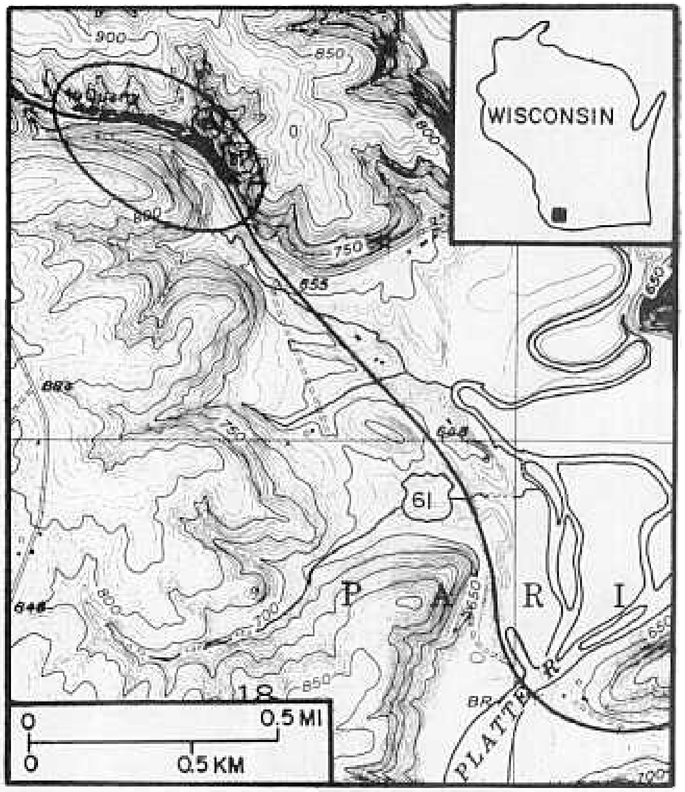


Figure 1. Location of exposure of Middle Ordovician strata in roadcut at Potosi Hill, Wisconsin.

SIGNIFICANCE

This is an excellent and easily accessible exposure of the upper few feet of the St. Peter Sandstone and the Glenwood Formation, Platteville Formation, Decorah Formation, and lower part of the Galena Formation (Fig. 2) (Ostrom, 1978). The Platteville, Decorah, and Galena Formations are the principal hosts of zinc and lead mineralization in the southwest Wisconsin zinc-lead mineralized district.

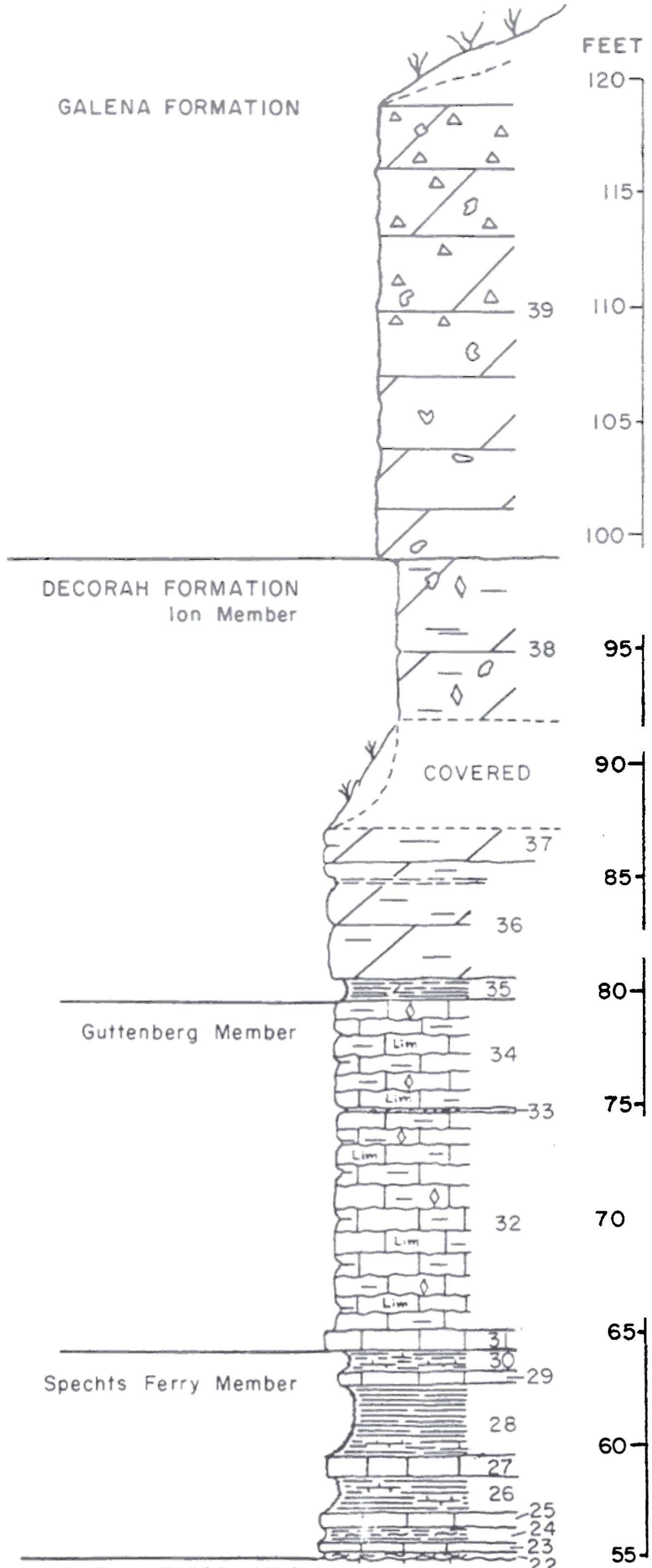
The St. Peter Sandstone was named by Owen (1847) for exposures along the St. Peter River (now the Minnesota River) near St. Paul, Minnesota. The St. Peter consists of very light yellowish gray to white, fine to coarse, subrounded to rounded quartz sand grains. It is typically very friable. It is cross-bedded, thick bedded to thin bedded, and locally massive. In the district it is from 0 to more than 300 ft (0 to 90 m) in thickness and averages about 40 ft (12 m). Variations in thickness are attributed to deposition on an erosion surface. The only fossils noted in the St. Peter are *Skolithos* (vertical straight burrows) and *Corophoides* (U-shaped burrows).

The St. Peter Sandstone is conformably overlain by the Glenwood Formation. The Glenwood was named by Calvin (1906, p. 75) from exposures in Glenwood Township (T.98N.,R.7W.) near Waukon, Iowa. Three members are recognized in the Glenwood Formation in southwest Wisconsin (Templeton and Willman, 1963; Ostrom, 1969). In the base of the Glenwood is the Nokomis Member, which consists principally of sandstone and is transitional with the St. Peter. It is distinguished from the St. Peter Sandstone by a more yellowish and greenish coloration and by a notable change in bedding character from cross-bedded, even-bedded, and uniform-textured sandstone to reworked, burrowed, and poorly sorted sandstone with more or less green clay. It is both silty and argillaceous. The Nokomis ranges from 8 ft (2.4 m) thick near Beetown (16 mi; 26 km northwest of Potosi Hill) to less than 1 ft (0.3 m) thick in the vicinity of New Glarus (about 65 mi; 105 km to the east).

The Nokomis Member is conformably overlain by the Harmony Hill Member, which consists of pale green to greenish gray shale with scattered rounded clear quartz sand grains. It decreases from 3.5 ft (1 m) thick in the western part of the district to zero in the east. The Harmony Hill is conformably overlain by the Hennepin Member. The Hennepin consists of brownish and locally calcareous shale with scattered phosphatic nodules and clear rounded quartz sand grains. It thins from 5 ft (1.5 m) thick in the western part of the district to zero in the east.

The Glenwood Formation is conformably overlain by the Platteville Formation, which is subdivided in ascending order into the Pecatonica, McGregor, and Quimbys Mill members. The Pecatonica Dolomite Member was named by Hershey (1894, p. 175) from exposures in the Pecatonica River valley in southwestern Wisconsin near the Illinois border. The Pecatonica is predominantly medium-grained, granular, thick- to thin-bedded dolomite. The lowermost bed, the Chana Member of Templeton and Willman (1963), contains phosphatic pellets and rounded clear quartz sand grains. The Pecatonica ranges from 20 to 25 ft (6 to 7.6 m) in thickness in the district.

The McGregor Limestone Member was named by Kay (1935, p. 286) from an exposure near McGregor, Iowa. It is from 25 to 30 ft (8 to 9 m) thick and consists of irregularly bedded, thin- to medium-bedded, light gray to buff argillaceous dolomite



Ordovician System
Champlainian Series
Sinnipee Group

Galena Dolomite Formation
Cherty Unit

39. Dolomite, yellowish-buff, medium- to coarse-grained, vuggy, abundant white chert in upper 10 ft (3 m).

Decorah Formation
Ion Dolomite Member
(Gray Unit)

38. Dolomite, buff, medium- to coarse-grained, thick- to massive-bedded, vuggy, green shale partings throughout, sparry calcite present. Covered interval.
37. Dolomite, buff, medium-grained, medium-bedded, with green shale partings.

(Blue Unit)

36. Dolomite, purplish-gray, medium-grained, slightly fossiliferous. Green shale present as partings, and as a 0.5 ft (15 cm) bed 0.8 ft (25 cm) below the top of the interval.
35. Shale, green. 0.3 ft (9 cm) green dolomitic shale in middle of interval.

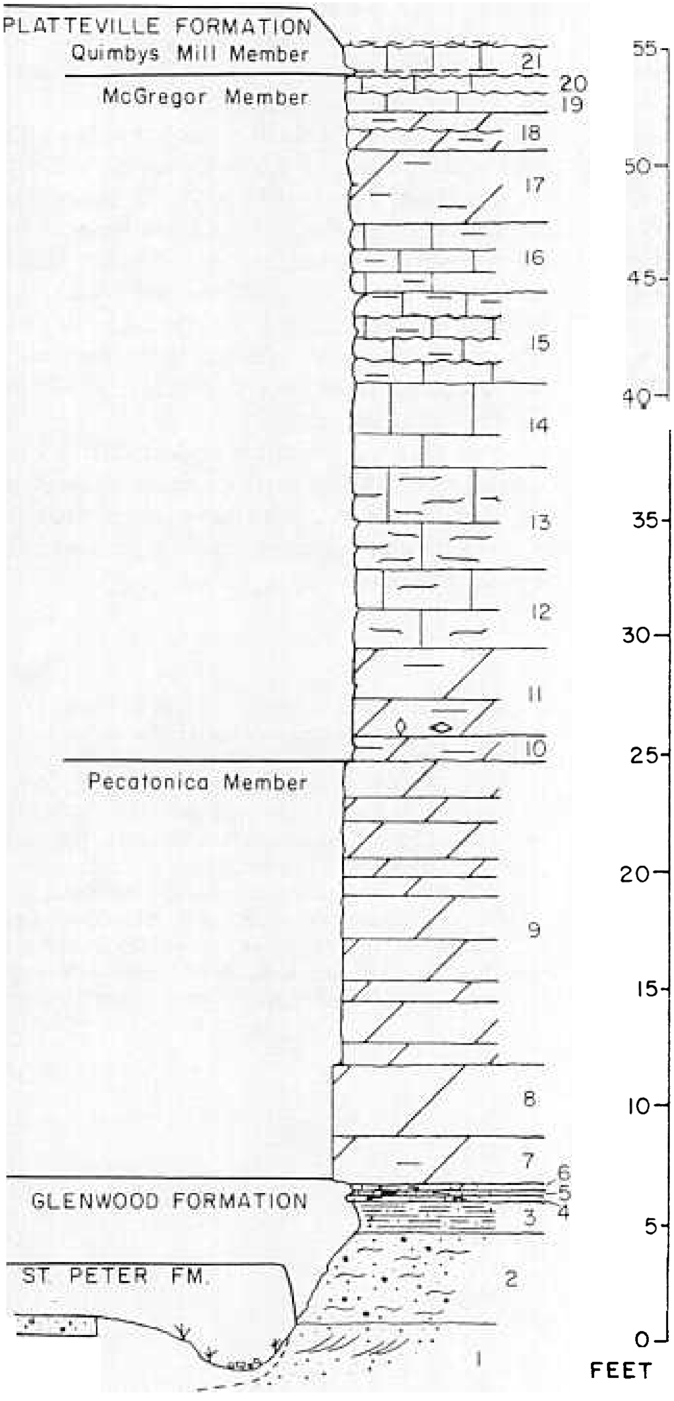
Guttenberg Limestone Member

34. Limestone, purplish-brown, fine-grained to sub-lithographic, fossiliferous, upper 1 ft (30 cm) fine- to medium-grained, brown shale present as partings, calcite and limonite after iron sulfide present in small amounts.
33. Metabentonite, brownish orange, crumbly, sticky when wet.
32. Limestone, purplish-brown, sublithographic, thin wavey bedding, fossiliferous, brown carbonaceous shale present as thin beds and partings, calcite and limonite present. Thin metabentonite bed at base.
31. Limestone brown-gray, fine-grained, thick-bedded.

Spechts Ferry Shale Member

30. Shale, orange-gray, calcareous, and limestone, tan-gray, fine-grained; limestone 0.4 to 0.7 ft (12 to 20 cm) above base.
29. Limestone, gray, fine-grained, thin-bedded.
28. Shale, gray, green, brown, fissle, some beds fossiliferous, limestone present as thin lenses near middle of the interval.
27. Limestone, tan, with iron oxide mottling, fine-grained, thin-bedded.
26. Shale, gray-green-brown. Fissle, with thin lenses of gray fine-grained limestone.
25. Limestone, dark to light gray, thin-bedded fossiliferous.
24. Shale, brown-green-orange-gray, brown carbonaceous shale parting at top, metabentonite near middle.
23. Limestone, purplish-brown, fine-grained, thin-bedded, very fossiliferous, fucoids at base.
22. Metabentonite, orange, sticky when wet, with brown shale partings.

Figure 2 (this and facing page). Section exposed in roadcut at Potosi Hill, Wisconsin.



Platteville Formation
Quimbys Mill Member

21. Limestone, purplish-ray-brown, sublithographic, thick-bedded, conchoidal fracture, irregular upper surface, shale at base. Metabentonite in top of shale.

McGregor Limestone Member

20. Limestone, purplish-gray-brown, fine to medium crystalline, thick-bedded; wavy upper surface.
19. Limestone, light gray, very fine crystalline, thin-bedded, fossiliferous; wavy upper surface.
18. Dolomite, yellowish-brown, fine crystalline, sugary texture, argillaceous, thin-bedded; discontinuous beds/nodular appearance.
17. Dolomite, light olive-brown, fine crystalline, sugary texture, argillaceous, thick bedded.
16. Limestone, light brown to greenish-brown, medium to fine crystalline, with some argillaceous partings; some fossil shells and fucoids on bedding planes.
15. Limestone, light brown to light greenish-brown, fine to medium crystalline, thin-bedded and uneven beds with nodular appearance, shale partings. Argillaceous in upper 0.6 ft (18 cm) with abundant fossils.
14. Limestone, light brownish-gray, medium to fine crystalline in medium to thick beds.
13. Limestone, light greenish-gray, fine and medium crystalline, thick-bedded with abundant discontinuous wavy shale partings. Shale partings are greenish-gray, mottled and very fossiliferous.
12. Limestone, same as above but fewer shale partings.
11. Dolomite, light gray, fine crystalline, slightly argillaceous, very fossiliferous, in medium beds, discontinuous faint shale partings. Some clear calcite and dolomite in lower 0.5 ft (15 cm).
10. Dolomite, light gray to light brownish-gray, fine crystalline with thin shale partings up to 1 in (2 cm) thick between beds. Shale is bluish-green and brown. Fossiliferous.

Pecatonica Member

9. Dolomite, brownish-gray, fine to medium crystalline, sugary texture, thin- to medium-bedded, even-bedded, beds 0.1 to 18 in (2 mm to 45 cm) thick. Weathered surface shows distinct but discontinuous thinner beds.
8. Dolomite, brownish-gray, fine crystalline, in single bed, upper 0.5 in (1 cm) stained brown. Scattered dark brown fossil molds and traces. Weathered surface shows wavy horizontal bedding features.
7. Dolomite, brownish-gray, fine crystalline, faint horizontal shale traces, dark brown phosphatic pebbles up to 2 mm, abundant dark brown fossil hash.

Glenwood Formation
Hennepin Member

6. Dolomite, very silty, yellowish-brown, very fine crystalline, abundant phosphatic pellets up to 2 mm, scattered round medium quartz sand grains.
5. Sandstone, brown, very fine- and fine-grained with little medium-grained, abundant grayish-green shale.
4. Sandstone, dark brown, fine- and medium-grained, argillaceous, poorly sorted, iron-oxide cemeted, abundant phosphate pellets up to 2 mm.

Harmony Hill Member

3. Shale, brown and bluish-green in upper 2 in (5 cm) grading downward to bluish-green with some reddish-brown; little rounded medium grained quartz sand.
2. Sandstone, mottled light yellowish-green, light greenish-yellow, and reddish-brown, medium- and fine-grained with some very fine and very coarse grains, poorly sorted, abundant pale green clay in matrix, reworked/bioturbated texture.

St. Peter Formation
Tonti Member

1. Sandstone, light yellowish-gray, very fine- to medium-grained, some light brown stains cross bedded.

Base of exposure in drainage ditch.

and limestone. The overlying Quimbys Mill Member was named by Agnew and Heyl (1946, p. 1585) from a quarry exposure about 5 mi (8 km) west of Shullsburg, Wisconsin. Its thickness varies in the district from less than 1 ft to more than 18 ft (0.3 to 5.5 m) thick and consists of purplish gray-brown, sublithographic, thick-bedded, conchoidally fractured limestone with an uneven upper surface and with shale at its base.

The Platteville Formation is overlain disconformably by the Spechts Ferry Shale Member of the Decorah Formation named from exposures in the city of Decorah, Iowa (Calvin, 1906, p. 61). It thins eastward from 8 ft thick to less than 1 ft (2.4 to 0.3 m). The Spechts Ferry Member consists of fossiliferous, gray-brown limestone with green shale interbeds. At this exposure two thin beds of "metabentonite" occur near its base. The metabentonites are orange to light reddish brown and about 2 in (5 cm) thick. Phosphatic nodules occur locally in the upper 1 ft (0.3 m).

The Spechts Ferry Member is conformably overlain and transitional with the Guttenberg Limestone Member, which consists of hard, fine crystalline, thin-bedded, fossiliferous, light brown limestone with brown petroliferous shale partings and interbeds. The presence of these interbeds has led to the member being referred to as the "Oil Rock." In the district the Guttenberg Member thins eastward from more than 14 ft to less than 7 ft (4 to 2 m).

The Spechts Ferry is conformably overlain by the Ion Member, which is a gray to blue dolomite, medium-crystalline, and medium to thick bedded with green shale interbeds. The amount of shale decreases to the east. The Ion maintains a thickness of about 20 ft (6 m) across the district.

The Decorah Formation is conformable with and transitional with the overlying Galena Dolomite Formation. The Galena was named (Owen, 1840, p. 19, 24) from exposures in the vicinity of the city of Galena in northwest Illinois. It is a light buff to drab, cherty, thick-bedded, vuggy dolomite with medium to coarse sugary grains. A zone of *Prasopora insularis* Ulrich marks the top of the Ion Member in some areas, but is absent here. In most of the district the Galena is dolomitized and the sparse fossils are poorly preserved. It is from 220 to 230 ft (67 to 70 m) thick throughout the district.

Near the north end of the roadcut, there is a quarry, now occupied by a junkyard, in which, on the southeast wall, can be seen an example of "pitch-and-flat" structure, which is the principal site of zinc and lead mineralization in the district. Here there is no mineralization.

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Middle Ordovician Platteville Formation, Hoadley Hill, Wisconsin

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LOCATION

Exposure in roadcut at north side of U.S. 151 about 6.5 mi (10.8 km) southwest of Platteville in the W½,NE¼,Sec.12, and the W½,SE¼,Sec.1,T.2N.,R.2W., Grant County, Wisconsin on the Dickeyville 7½-minute Quadrangle (Fig. 1).

SIGNIFICANCE

This is the reference section for the Platteville Formation in the southwest Wisconsin zinc-lead district (Agnew and others, 1956; Ostrom, 1978). The strata exposed here are the upper part of the St. Peter Sandstone, the Glenwood Formation, a complete section of the Platteville Formation, and the lower part of the Decorah Formation (Fig. 2).

The Hoadley Hill exposure shows the interrelationship and lithologic characteristics of the St. Peter, Glenwood, and Platteville Formations in the classic southwest Wisconsin zinc-lead district. The St. Peter Sandstone was named by Owen (1847, p. 170) for exposures in bluffs of the St. Peter River valley (now Minnesota River), near St. Paul, Minnesota. It consists of clear, fine to coarse rounded to subangular quartz grains and is generally poorly cemented. It is white to very light gray and very light buff, thin to thick bedded, locally massive and cross-bedded, and is variable in thickness in the district ranging from 0 to more than 350 ft (0 to 100 m) thick and averaging about 50 ft (15 m) thick. Thickness variations are attributed to deposition of the sand on a deeply dissected erosion surface.

The Glenwood Formation conformably overlies the St. Peter Formation. It was named by Calvin (1906, p. 75) from exposures in Glenwood Township (T.98N.,R.7W.) a short distance northwest of Waukon, Iowa. The classification used here is that of the Illinois Geological Survey (Templeton and Willman, 1963) as modified by Ostrom (1969). Three members are recognized in the Glenwood Formation in southwest Wisconsin. In ascending order these are the Nokomis, Harmony Hill, and Hennepin. The Nokomis Member consists principally of sandstone and is transitional with the underlying St. Peter Sandstone. It is distinguished from the St. Peter Sandstone by light yellowish and greenish coloration and by a notable change in bedding character from cross-bedded, even-bedded and even-textured sandstone to reworked, burrowed, and poorly sorted sandstone with more or less green clay. It is both silty and argillaceous. In the district the Nokomis ranges from 8 ft thick (2.5 m) near Beetown, about 25 mi (40 km) west-northwest of Hoadley Hill, to less than 1 ft (0.3 m) thick in the vicinity of New Glarus, located about 55 mi (88 km) to the east. The Nokomis Member is conformable with the overlying Harmony Hill Member.

The Harmony Hill Member is a pale green to greenish gray shale with scattered rounded clear quartz sand grains. It is up to

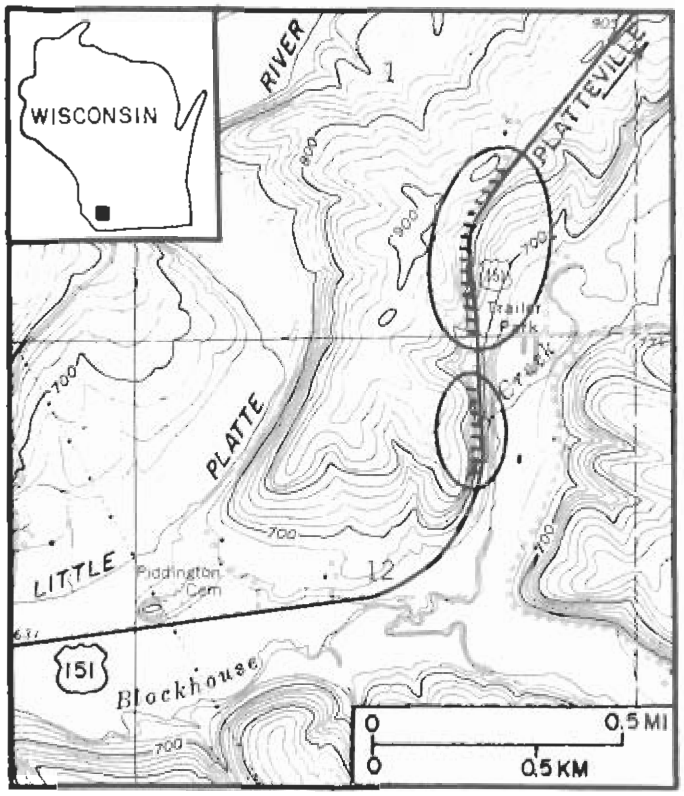
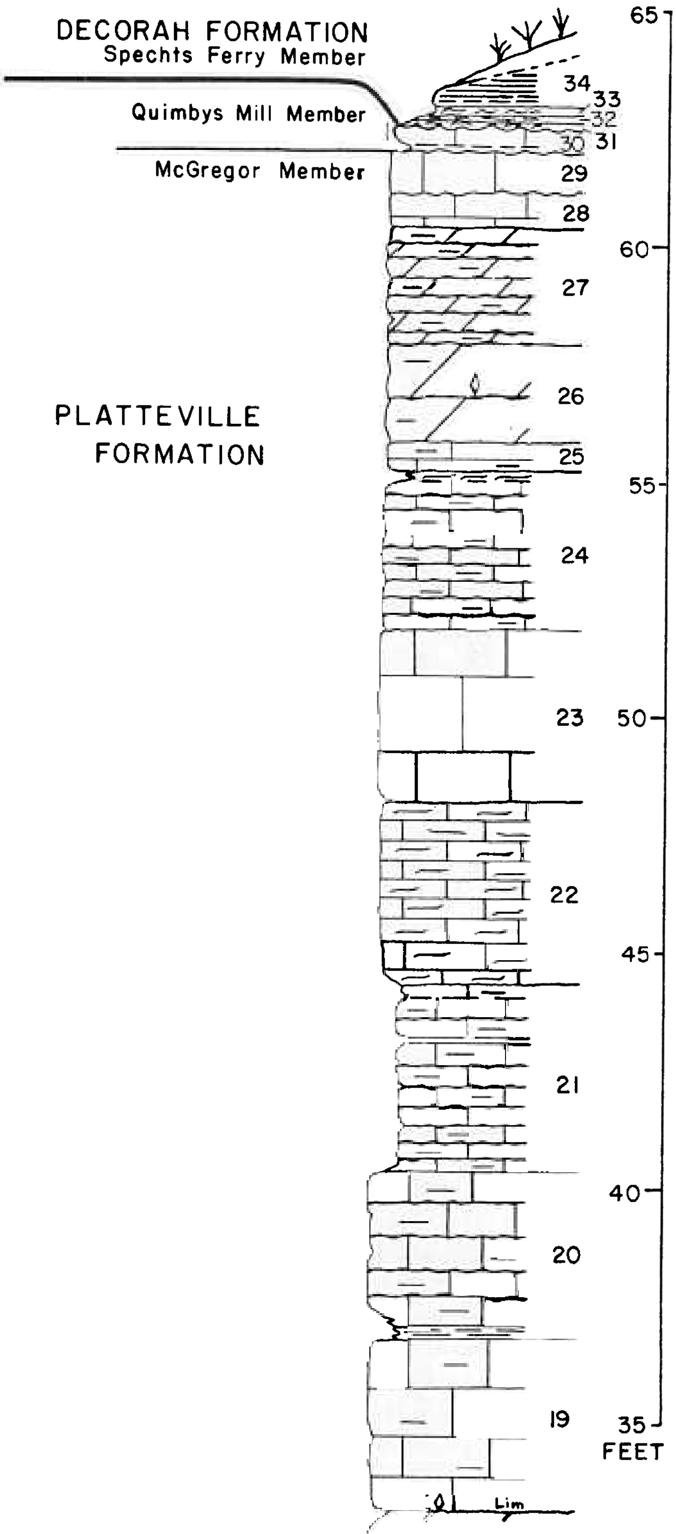


Figure 1. Location of exposure of Middle Ordovician Platteville Formation in roadcut at Hoadley Hill, Wisconsin.

3.5 ft (1 m) thick in the western part of the district and is absent in the east. The Harmony Hill Member is conformable with the overlying Hennepin Member, which consists of brownish and locally calcareous shale with scattered phosphatic nodules and clear rounded quartz sand grains. It is 5 ft (1.5 m) thick in the western part of the district and thins to disappearance in the east.

The Platteville Formation overlies and is conformable with the Glenwood Formation. It is one of the mineralized formations in the southwest Wisconsin zinc-lead mining district and was named by Bain (1905, p. 19) for exposures in the vicinity of Platteville, Wisconsin. It is known throughout the district, and within the Driftless Area, from exposures at the surface and in mines and from drill cuttings. In the district the Platteville Formation ranges in thickness from about 55 ft (17 m) in the west to near 75 ft (23 m) in the east, near Shullsburg.

The Platteville Formation consists of three members, which in ascending order are the Pecatonica, McGregor, and Quimbys Mill. The Pecatonica Dolomite Member was named by Hershey



Ordovician System
Champlainian Series
Sinnipee Group

34. Shale, bluish-green.

33. Bentonite, white; weathers orange brown.

32. Shale, yellowish-green above to bluish-green below.

31. Shale, brown and olive, soft.

30. Limestone, dark purple, fine-crystalline, dense, conchoidal fracture; very wavy upper surface; thin, dark-brown to black, fossiliferous platy shale parting at base.

29. Limestone, light-gray, very fine crystalline, very dense, conchoidal fracture like "glass rock" above, fairly massive, very fossiliferous; wavy upper surface.

28. Limestone as next above but less dense, medium-bedded above to thin-bedded below, fossiliferous; wavy upper surface.

27. Dolomite, light olive drab, fine crystalline, "sugary," argillaceous, very thin-bedded; nodular.

26. Dolomite as above but thick-bedded; calcite near middle.

25. Limestone, thin-bedded yet stands massively as one unit; light greenish gray brown; weathers brown, with a few argillaceous streaks; sparingly fossiliferous, but with fossils and fucoids on top surface.

24. Limestone, thin-bedded as above but the beds are distinct; nodular beds and shaly partings; upper 0.3 ft (10 cm) is argillaceous and very fossiliferous.

23. Limestone, light buffish gray, in medium to thick beds; in places gradational into above unit.

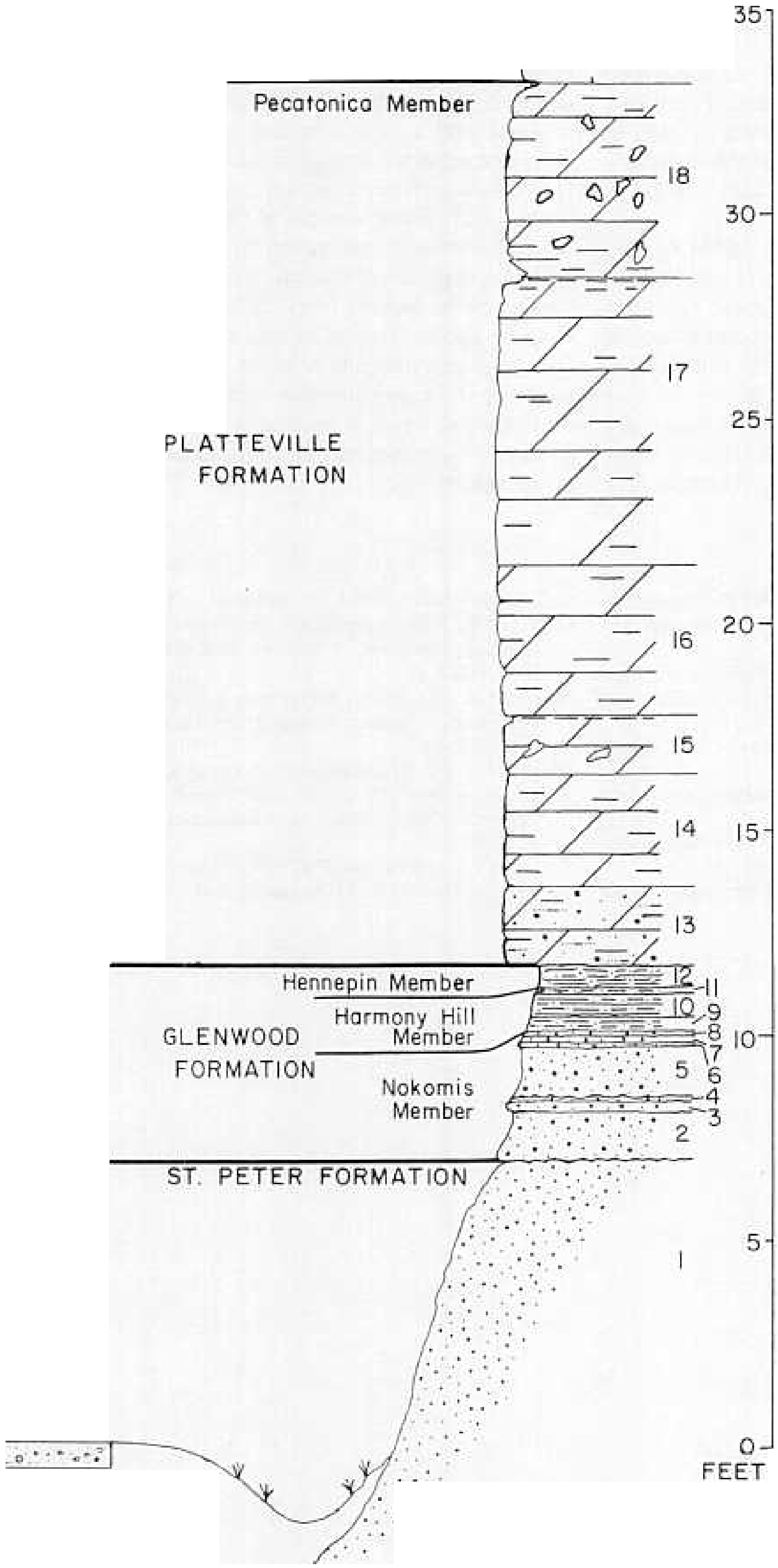
22. Limestone, light greenish to bluish gray, composed of thin beds but appears massive; abundant shaly material in wavy bands; fairly fossiliferous, argillaceous; a peculiar mottled light gray and darker gray 0.1-foot (3 cm) zone, 1 ft (30 cm) below top.

21. Limestone, light gray, very fine crystalline, very dense sublithographic, in extremely thin and nodular beds with thin calcarous shaly partings which become thinner below; the shale beds are light grayish blue, mottled, very fossiliferous; weathers slightly recessed.

20. Limestone, as above, but beds are not quite as thin; fossiliferous; poor gastropod zone 1.7 (50 cm) above base; shaly zone at base.

19. Limestone, dolomite, light-gray, fine crystalline, very slightly argillaceous, very fossiliferous, medium-bedded; indistinct argillaceous partings, not wavy; calcite and limonite, especially in basal 0.6 ft (18 cm).

Figure 2 (this and facing page). Section exposed in roadcut at Hoadley Hill, Wisconsin.



18. Dolomite, light grayish brown, very coarse crystalline and vuggy, upper 2 ft (60 cm) a mixture of lithology and a somewhat argillaceous fine crystalline "sugary" laminated dolomite; a 1-ft (30 cm) bed of very vuggy dolomite from 1.8 to 2.8 ft (55 to 85 cm) above base; shaly in lower part; stylolitic partings 1 ft (30 cm) above base.

17. Dolomite, medium gray, laminated, somewhat argillaceous, fine-crystalline "sugary," fossiliferous, especially in lower 0.9 ft (27 cm); medium- to thick-bedded; shaly at top; weathers brownish in lower 25 ft (75 cm).

16. Dolomite, medium gray, laminated, argillaceous; very fossiliferous partings.

15. Dolomite, light grayish brown, very coarse crystalline and vuggy; thin brownish gritty dolomitic and platy shaly parting at top.

14. Dolomite, medium gray, laminated, argillaceous, fine crystalline.

13. Dolomite, medium gray, laminated, argillaceous, silty and sandy with fine to coarse quartz grains similar to those of the St. Peter Sandstone, phosphate nodules abundant (especially in two zones, one at base, the other 1 ft (30 cm) above base .

12. Shale, sandy with rounded quartz grains, khaki to drab, soft; phosphate nodules.

11. Shale, sandy, olive to grayish brown; mottled yellowish brown, friable.

10. Shale, sandy, medium- to dark-gray, olive, blocky, very hard.

9. Shale, medium-gray, blocky, hard, sandy; streak of carbonaceous material at top.

8. Sandstone, red and white; rounded; frosted, coarse to medium-grained.

7. Sandstone, gray pinkish, very friable.

6. Sandstone, brown, iron-stained, hard.

5. Sandstone, yellow to gray, fine to coarse grained, very friable, reworked and burrowed, with irregular lower surface.

4. Sandstone, light-gray, very friable.

3. Sandstone, yellow to dark-brown, laminated, hard.

2. Sandstone, gray and yellow, fine to coarse grained, reworked and burrowed; hard irregular lower surface.

1. Sandstone as above, but medium- to fine-grained.

Base of exposure in roadside ditch

(1894, p. 175) from exposures in the Pecatonica River valley in southwestern Wisconsin near the Illinois border. It consists predominantly of medium-grained, granular, thick- to thin-bedded dolomite. The lowermost bed, the Chana Member (Templeton and Willman, 1963) contains phosphatic pellets as well as rounded quartz sand grains similar to those in the underlying Glenwood Formation. The Pecatonica is from 20 to 25 ft (6 to 8 m) thick in the district.

The McGregor Limestone Member was named by Kay (1935, p. 286) from an exposure in a ravine 1 mi (1.6 km) west of McGregor, Iowa. The McGregor Member consists of uneven bedded, thin- to medium-bedded, light gray to buff argillaceous dolomite and limestone and is from 25 to 30 ft (8 to 9 m) thick. The McGregor Member contains commercial deposits of zinc-lead ore. The Quimbys Mill Member was named by Agnew and Heyl (1946, p. 1585) from an exposure in a quarry at Quimbys Mill located 5 mi (8 km) west of Shullsburg, Wisconsin. The member ranges from less than 1 ft (0.3 m) in thickness to more than 18 ft (5.5 m). It consists of light brown, thin- to medium-bedded, crystalline sublithographic limestone and finely granular dolomite. This member is locally called the "glass rock" because it breaks with a conchoidal fracture. The Quimbys Mill also contains commercial deposits of zinc-lead ore.

The Platteville Formation is overlain discomformably by the Spechts Ferry Shale Member of the Decorah Formation. The Decorah Formation was named by Calvin (1906, p. 61) from exposures in the city of Decorah, Iowa. The Decorah Formation consists of the Spechts Ferry, Guttenberg, and Ion members. Only the Spechts Ferry is exposed at this outcrop. The Spechts Ferry consists principally of bluish green to brown shale with nodules and discontinuous thin beds of limestone. A thin metabentonite bed, which is believed to be an alteration product of volcanic dust, occurs near its base and can be correlated on a broad regional scale.

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Bedrock stratigraphic units in Wisconsin

Outcrop nomenclature

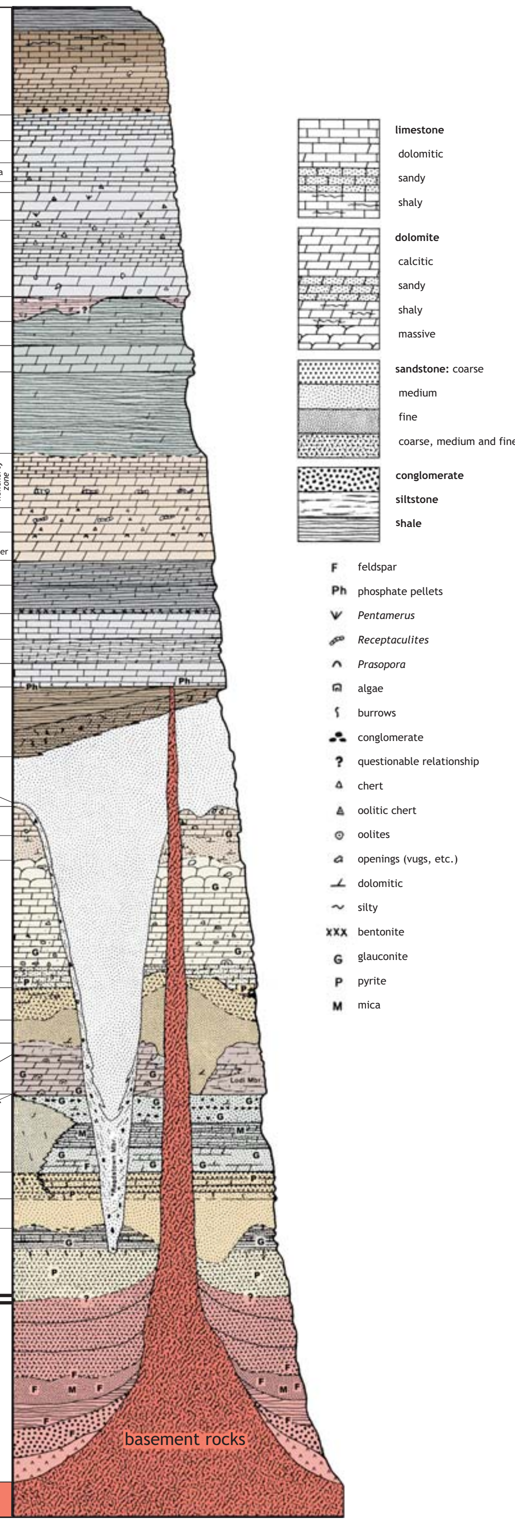
Geologic time / Time-stratigraphic

AGE*	ERA	PERIOD	EPOCH	SYSTEM	SERIES
354–370	Paleozoic	Devonian	Late	Devonian	Upper
370–417	Paleozoic	Devonian	Middle	Devonian	Middle
417–423	Paleozoic	Silurian	Late	Silurian	Cayugan
423–443	Paleozoic	Silurian	Early	Silurian	Niagaran
	Paleozoic	Silurian	Early	Silurian	Alexandrian
443	Paleozoic	?		?	
443–458	Paleozoic	Ordovician	Late	Ordovician	Cincinnatian
458–470	Paleozoic	Ordovician	Middle	Ordovician	Champlainian
470–490	Paleozoic	Ordovician	Early	Ordovician	Canadian
490–523	Paleozoic	Cambrian	Late	Cambrian	St. Croixian
1,050–1,110	Precambrian				
1,110–2,800	Igneous and metamorphic rock, undivided				

Rock-stratigraphic

GROUP	FORMATION	MEMBER
	Antrim	
	Milwaukee	
	Thiensville	
	Western / *Eastern* Waubakee	
	Hopkinton / Racine	
	Hopkinton / Manistique	Cordell (Waukesha)
	Hopkinton / Manistique	Schoolcraft
	Hendricks	
	Byron	
	Kankakee / Mayville	
	Edgewood / Mayville	
	Neda	
	Maquoketa	Brainerd
	Maquoketa	Fort Atkinson
	Maquoketa	Scales
Sinnipee	Galena	Dubuque (*noncherty zone*)
Sinnipee	Galena	Wise Lake: Stewartville, Sinsinawa (*noncherty zone*)
Sinnipee	Galena	Dunleith: *cherty zone*, Ion submember
Sinnipee	Decorah	Guttenberg
Sinnipee	Decorah	Sprechts Ferry
Sinnipee	Platteville	Quimbys Mill
Sinnipee	Platteville	McGregor
Sinnipee	Platteville	Pecatonica
Ancell	Glenwood	
Ancell	St. Peter	Tonti
Ancell	St. Peter	Readstown
Prairie du Chien	Shakopee	Willow River
Prairie du Chien	Shakopee	New Richmond
Prairie du Chien	Oneota	Hager City
Prairie du Chien	Oneota	Stockton Hill
Trempealeau	Jordan	Van Oser
Trempealeau	Jordan	Norwalk
Trempealeau	St. Lawrence	Lodi
Trempealeau	St. Lawrence	Black Earth
Tunnel City	Lone Rock / Mazomanie	Reno
Tunnel City	Lone Rock / Mazomanie	Tomah
Tunnel City	Lone Rock / Mazomanie	Birkmose
Elk Mound	Wonewoc	Ironton
Elk Mound	Wonewoc	Galesville
Elk Mound	Eau Claire	
Elk Mound	Mount Simon	
Keweenawan Supergroup – Bayfield	Chequamegon	
Keweenawan Supergroup – Bayfield	Devils Island	
Keweenawan Supergroup – Bayfield	Orienta	
Keweenawan Supergroup – Oronto	Freda Sandstone	
Keweenawan Supergroup – Oronto	Nonesuch	
Keweenawan Supergroup – Oronto	Copper Harbor Conglomerate	
Keweenawan Supergroup	Midcontinent Rift volcanics	

Dominant lithology



* Absolute age dates in million years are based on the Geological Society of America Geologic Time Scale, 1998.

Modified from Ostrom, M.E., 1968, Paleozoic Stratigraphic Nomenclature for Wisconsin: Wisconsin Geological and Natural History Survey Information Circular 8.

Wisconsin Geological and Natural History Survey Open-File Report 2006-06

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