Field Trip Guide Book
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M. E. Ostrom, State Geologist

QUATERNARY HISTORY OF THE DRIFTLESS AREA

ROAD LOG

QUATERNARY HISTORY OF THE KICKAPOO AND LOWER WISCONSIN RIVER VALLEYS, WISCONSIN
J. C. Knox

SPECIAL PAPERS

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GLACIATION OF THE DRIFTLESS AREA: AN EVALUATION OF THE EVIDENCE
D. M. Mickelson, J. C. Knox, and Lee Clayton

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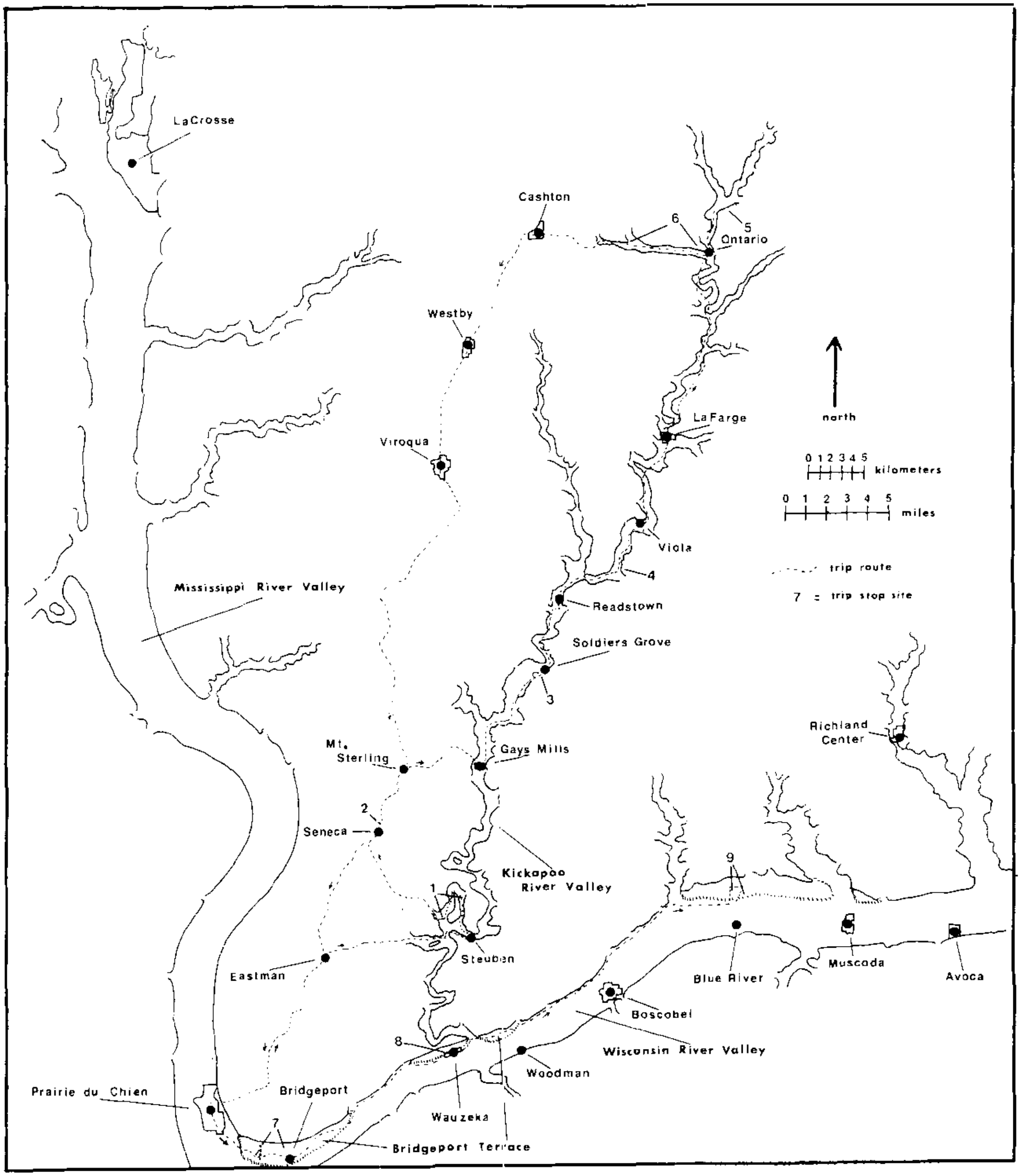

FIGURE 1.--Field-trip route map.

QUATERNARY HISTORY OF THE KICKAPOO AND LOWER WISCONSIN RIVER VALLEYS, WISCONSIN

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INTRODUCTION

The conference will emphasize stratigraphic evidence for various environmental conditions that occurred during the Quaternary in the south-central Driftless Area. Saturday, May 22, will be devoted to the Quaternary history of the Kickpoo valley. The early part of the day will focus on alluvial deposits contained in valley fills of the lower and middle Kickapoo valley. As the trip progresses to the headwaters of the Kickapoo, additional attention will be given to the Holocene alluvial chronology in the context of Holocene climate and vegetation changes. Pleistocene hillslope processes will also be examined at an upland site that includes a paleosol separating Woodfordian and pre-Woodfordian loess deposits.

The trip for Sunday May 23, consists of three stops in the lower Wisconsin River valley. The first stop presents evidence for glacial till on the Bridgeport terrace at the mouth of the Wisconsin River. The remaining stops examine the characteristics of pre-Wisconsinan deposits on the Bridgeport strath. Two pre-Wisconsinan outwash deposits are apparent on the strath and one suggests that the Wisconsin River was temporarily reversed relative to its present direction of flow. Except for the glacial till on the Bridgeport terrace at the mouth of the Wisconsin River, none of the Kickapoo or Wisconsin River valley deposits suggests that the Driftless Area was ever covered by continental glacial ice. The trip route and stops are shown on figure 1.

ROADLOG

MILEAGE

0.0 0.0

Start point (7:50 a.m., May 22, 1982) intersection of highways 18, 27, 35, 60, northeast side of Prairie du Chien. The city is built on a late-Wisconsinan terrace that averages about 12 m above present river level. The surface of the Prairie du Chien terrace probably was the floodplain of the aggrading Mississippi River until about 12,000 years ago when relatively sediment-free discharge from glacial lakes flowed through the Mississippi River and caused entrenchment (Clayton, this volume). The maximum thickness of outwash sediments in the valley is unknown, but wells in Prairie du Chien indicate approximately 50 m (150 ft) of fill.

Proceed east on highway 27 toward bluff of Mississippi gorge.

0.7 0.7

Base of bluff. Begin ascent of Driftless Area upland via steep tributary. The Ordovician rocks that crop out along the road are representative of the formations that dominate the southern one-third of the Driftless Area, especially south of the Wisconsin River. Ostrom (1970) described the outcrop from the base upward as including about 73 m (240 ft) of Prairie du Chien Formation (mostly dolomite) and about 8 m (26 ft) of St. Peter sandstone. Ostrom suggested that the exposure is excellent for examining the pre-St. Peter unconformity. Higher up the hill, around the bend in the highway, the Platteville Formation is exposed in a cutbank on the north side of the road. Ostrom (1970) found that the Platteville Formation was represented here by thick-bedded dolomites near the base and thinner-bedded argillaceous, nodular, calcitic limestones above.

2.1 2.8

Continue northeastward along the narrow upland stream divide. Examination of the quarry near the center of section 21 and 0.8 km north of the highway showed no evidence of glacial outwash gravels beneath the Wisconsinan loess as has been observed for a few upland sites near the western margin of the Driftless Area south of the Wisconsin River. The high-level outwash gravels to the south of here are thought to be of classical Nebraskan age (Willman and Frye, 1969).

The upland landscape in the area shows two prominent erosion surfaces. Trowbridge (1921) gave the name "Dodgeville" to the higher surface that is represented here by the points of highest elevation along highway 27 on the ridge crest. He gave the name "Lancaster" to the surface that is here about 30 m (100 ft) lower and dominates the secondary interstream divides. Later, Trowbridge (1954) recognized only the Lancaster as a relict erosion surface. Whatever the number, these Tertiary erosion surfaces that Trowbridge identified as peneplain remnants appear to represent the relief of the landscape that existed in the Driftless Area at the beginning of the Pleistocene. Evidence for deep incision of the landscape in Pleistocene time occurs across the Mississippi River in northeastern Iowa where Trowbridge (1966) showed that classical Nebraskan till is found only on the upland drainage divides but that classical Kansan till is found at much lower elevations including valley sides and certain stream terraces. Willman and Frye (1969) concluded from a study of high level glacial outwash in the Driftless Area of northwestern Illinois that the Mississippi River was not in its present position until after Nebraskan glaciation. They also concluded that the deep incision of the Driftless Area landscape had largely occurred by classical Kansan time. I support Willman and Frye's conclusions because glacial till on the Bridgeport terrace at the mouth of the Wisconsin River has a clay mineralogy that is nearly identical to the clay mineralogy of classical Kansan tills of northeastern Iowa (Knox, Attig, and Johnson, this volume). The dolomite strath under the Bridgeport till is from 125 to 150 m (400 to 500 ft) lower than the upland drainage divides. Significant additional incision occurred after development of the Bridgeport strath, however, because the adjacent bedrock valley floor is known to be at least 60 m (200 ft) below the strath surface.

9.1 11.9

Enter village of Eastman on highway 27. Continue northerly through village.

0.8 12.7

Junction of highways 27, 179, and D Turn right to east on highway 179 to Steuben. Descend upland into drainage of Otter Creek.

The west-to-east main valley segment of Otter Creek was described by Ludvigson (1976) as part of the Du Charme Creek "linear", a system of straight drainage segments trending N. 85 ° E. and represented by Du Charme, Otter, and Pine Creeks as well as a segment of the Kickapoo valley near Steuben and an un-named tributary east of Steuben.

1.3 14.0

Eastman rock quarry on left. Ludvigson (1976) described contacts between the St. Peter sandstone and the Shakopee and Oneota Members of the Prairie du Chien Formation for this site. Ludvigson argued that the irregular surface on the Prairie du Chien Formation is of karst origin because channel fills in the basal St. Peter sandstone terminate abruptly at the contact with underlying weak facies. He also, as others have before him, described an anticlinal structure that is apparent in the northwest corner of the quarry.

The Driftless Area often is perceived as a structureless region represented by dendritic drainage networks superimposed into nearly horizontal or very gently dipping bedrock. Actually, the drainage patterns and the ridge orientations are very strongly controlled by fracture orientations in the bedrock and by broad anticlinal and synclinal structures that often range from 30 to 50 km (15 to 30 mi) in length, 5 to 10 km (3 to 6 mi) in width, and from 30 to 60 m (100 to 200 ft) in amplitude (Heyl, and others, 1959).

1.5 15.5

Observe that bases of valley sides preserve remnants of colluvial and fluvial sediments that were much more extensive during late-Wisconsinan Woodfordian time. In adjacent Grant County I found that organic sediments near the base of colluvial and fluvial sediments in a similar setting were 20,270 ± 650 B.P. (ISGS-558). Periglacial climatic conditions occurred during at least part of the Woodfordian, and mass movement of materials from steep hillslopes was extreme then. Many of the eroded colluvial terraces show large angular dolomite boulders embedded within a deposit that consists mostly of reworked loess.

The erosion of the colluvial and fluvial sediments probably began shortly after about 12,000 B.P. when downcutting occurred in the Mississippi and Wisconsin Rivers and then extended into the tributaries by headward erosion (Knox, McDowell, and Johnson, 1981). The initial cause of incision was discharge of relatively sediment free water from glacial Lakes Agassiz and Duluth (Clayton, this volume).

0.4 15.9

Note large alluvial fan on south valley wall tributary. Most of the large alluvial fans now visible in the Driftless Area were constructed during Woodfordian time.

0.8 16.7

A rock pinnacle occurs on the lower portion of the north valley side. Although sandstone pinnacles can be formed by natural processes in a few thousands of years, it seems that much longer periods would be required when dolomitic rocks are involved as here.

1.0 17.7

Cross Pine Creek, tributary to Otter Creek. Continue east on highway 179. Note the rather extensive colluvial bench along the base of the north valley side.

1.1 18.8

Enter upland sag, southwest quarter of section 7 about 3 km (2 mi) west of Steuben (fig. 2). The upland sag appears to be the north section of a cut-off valley meander marking a former course of the Kickapoo River. Scars of former valley meanders are abundant in the lower Kickapoo valley (fig. 2)

The elevation on the floor of the upland sag is between 225 and 230 m (740 and 760 ft) above sea level. Since the elevation on the top of the eastward dipping outwash from the Bridgeport till is approximately 225 m (740 ft) above sea level at the mouth of the Kickapoo River near Wauzeka, the apparent cut-off probably occurred in response to sedimentation and valley aggradation in the lower Kickapoo as the base level at the mouth of the Kickapoo was raised by the Bridgeport outwash (Knox, Attig, and Johnson, this volume).

0.3 19.1

Crest of upland sag. Begin descent to valley floor of Kickapoo River.

0.8 19.9

Intersection of county highway E with state highway 179. Continue east on 179.

0.1 20.0

Cross Kickapoo River. The large mound on the floodplain adjacent to the left bank of the Kickapoo River is thought to be an Indian mound rather than a terrace remnant.

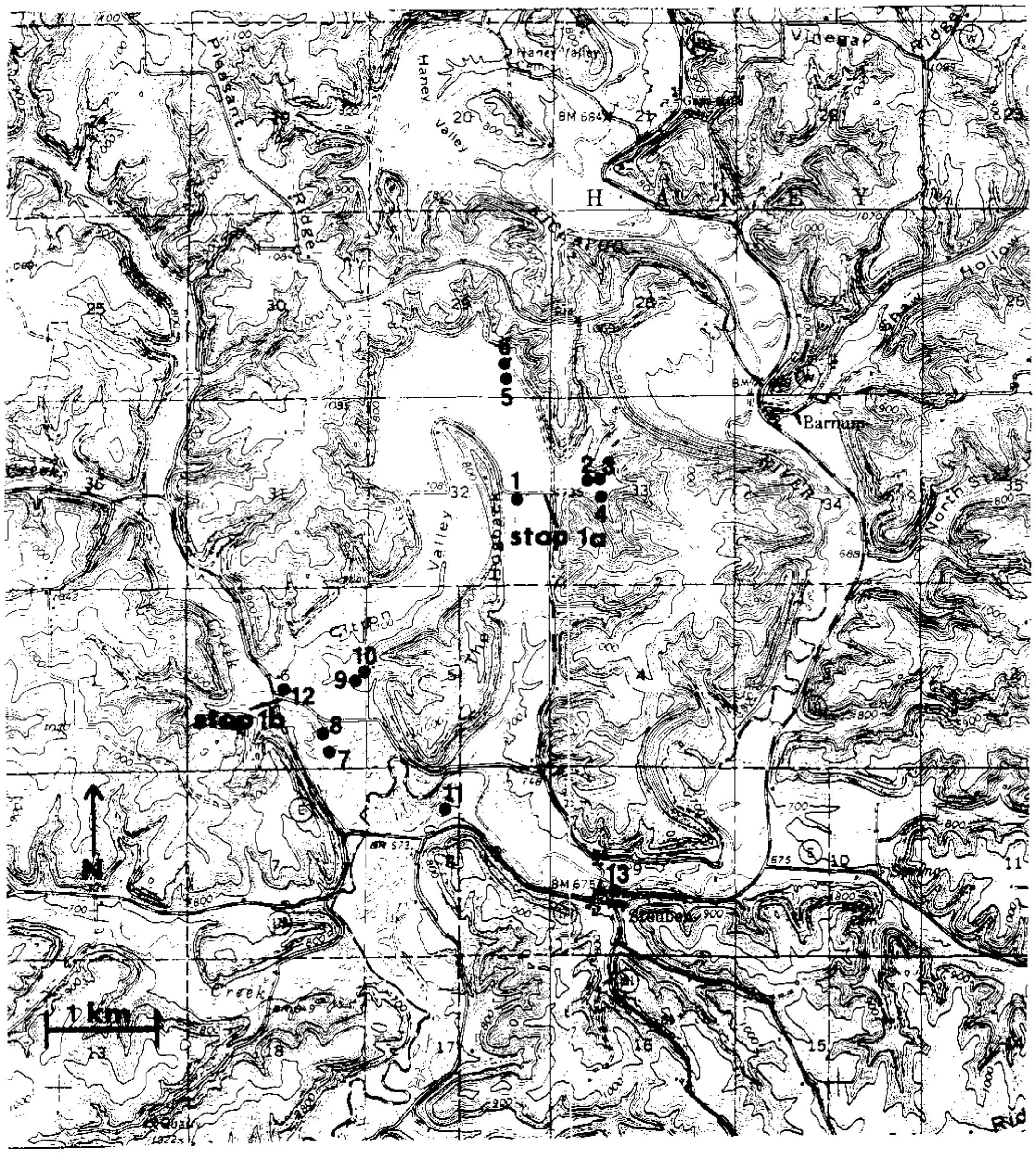

FIGURE 2.--Topographic map of the Citron Valley area, Crawford County, Wisconsin. The numbered dots represent drill hole locations.

0.3 20.3

Ascend high Woodfordian terrace in the gorge cut-off in front of Citron Valley (fig. 2). The terrace surface represents the maximum level of valley aggradation that prevailed between about 20,000 and 14,000 years ago. As may be recalled from earlier discussion, the major degradation of the aggraded valley floor probably began shortly after 12,000 years ago and may have continued until about 9500 years ago.

The average elevation of the high Woodfordian terrace in the Citron Valley area is about 210 m (690 ft) above sea level. Drill hole number 11 was placed in the floodplain below the high terrace in the gorge to determine the relationship between the fill in the gorge and the fill in the cut-off of Citron Valley (fig. 2).

The Pleistocene deposits of the Citron Valley area are shown schematically in figure 3. The drill holes that were used for construction of figure 3 can be located on the topographic map of figure 2.

The sedimentology associated with drill hole 11 indicated four major stratigraphic units in the alluvial fill (fig. 4) The top 3 m included a mixture of recent fill for a railroad grade and Holocene silty sand. The Holocene unit was underlain by approximately 7 m (23 ft) of Woodfordian age sand with chert gravel. The base of the Woodfordian occurs at an elevation of about 194 m (640 ft) above sea level. Since the top of the high Wisconsinan terrace is 210 m (690 ft) above sea level, the magnitude of Woodfordian aggradation in the gorge was about 16 m (52 ft).

The yellowish brown (2.5Y 5/4) Woodfordian cherty sands are resting on a paleosol that is developed in a leached and oxidized silt (fig. 4). The buried soil was black (7.5Y 2/2) at its surface and became dark gray (5Y 4/1) 30 to 60 cm (1 to 2 ft) below. The leached silts became dark yellowish brown (10YR 4/6) at a depth of about 11.3 m (37 ft) below the top of the drill hole and about 2.1 m (6.9 ft) below the top of the paleosol.

The age of the soil is unknown at this time. Its stratigraphic position indicates it is pre-Woodfordian. The depth of leaching and the increased concentration of clay with depth in the silt suggests that it might even be pre-Wisconsinan.

Sandy alluvium occurs under the oxidized and leached silt. The percentages of sand, silt, and clay in this pre-Woodfordian sandy unit are similar to percentages determined for the Woodfordian alluvial sands (fig. 4). However, when the sand fractions were analyzed at one-phi intervals, it was found that the pre-Woodfordian sands tend to be finer than the Woodfordian sands (see graphs of percentage of total sand retained on the 2-phi and 4-phi sieves, fig. 4).

Continue east on 179 into Steuben.

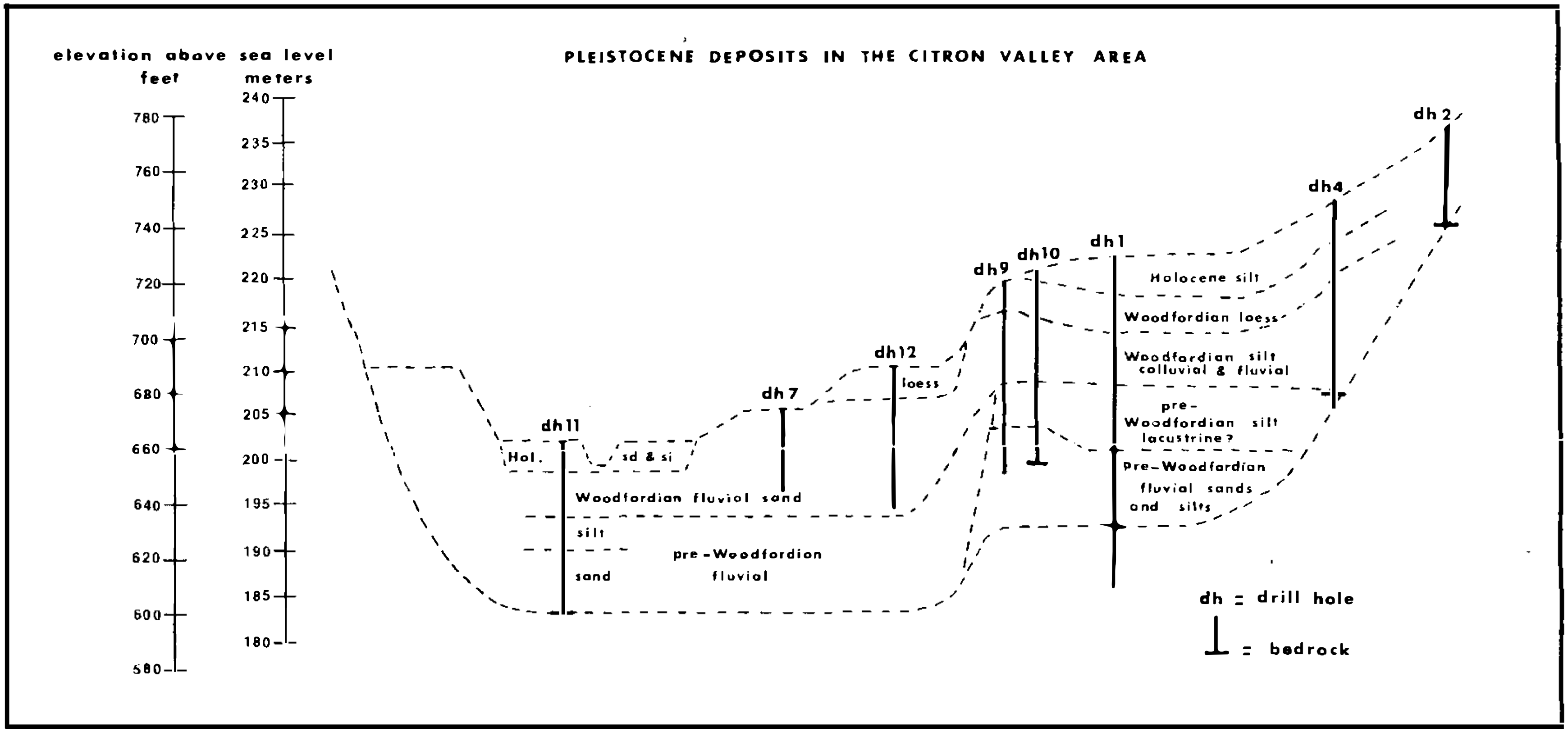

FIGURE 3.--Schematic diagram showing stratigraphic relationships of valley-fill deposits in the Citron Valley area.

7

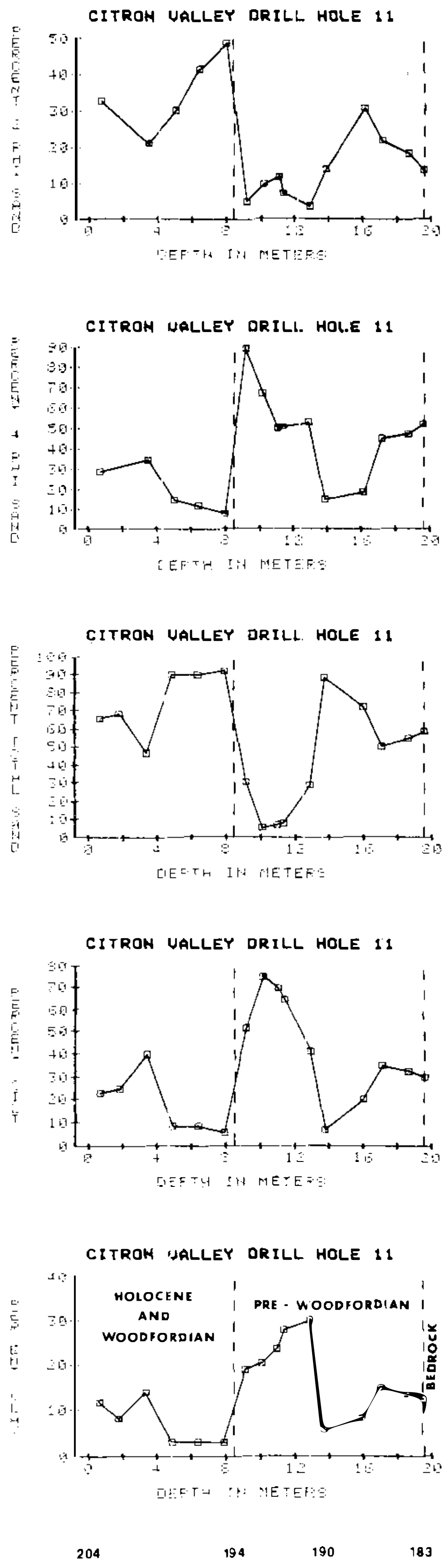

FIGURE 4.--Stratigraphy and sedimentology at drill hole 11. See figure 2 for location.

1.2 21.5

Intersection of highways 179 and 131. Highway 179 ends. Turn left to north on highway 131. Continue north one block and turn left to west in front of Steuben village park.

Drill hole 13 is located adjacent to the river bank at the northeast corner of the park. No buried soil or leached and oxdized silt was observed here as at drill hole 11. At the time of compilation of the road log, laboratory analyses had not been completed on samples from drill hole 13. Preliminary results suggest that the base of the Woodfordian occurs approximately 8 to 9 m (28 ft) below the surface where a gravel zone occurs. Silt derived from Woodfordian loess also appears to disappear below a depth of about 8.5 m (28 ft). Depth to bedrock is 15.1 m (50 ft).

The above-sea-level elevations on the bedrock valley floor are 183 m (602 ft) and 189 m (620 ft) respectively for drill holes 11 and 13. Comparison of these valley floor elevations with the elevation of the bedrock valley floor in Citron Valley at drill hole 1 indicates that the floor of the gorge is lower. The differences are 4 m (13 ft) between drill holes 1 and 13 and 10 m (33 ft) between drill holes 1 and 11. These differencs indicate that modest valley incision into bedrock occurred sometime after Citron Valley was abandoned as a course of the Kickapoo River (fig. 3).

Continue west past Steuben park. Turn right to north at the T intersection and cross Kickapoo River.

0.3 21.8

Ascend high Woodfordian terrace on north side of Kickapoo River.

0.5 22.3

Intersection with Citron west township road on left. Continue straight ahead (north) into the east section of the Citron Valley cutoff meander.

1.0 24.0

Intersection of Hughes Road and Citron Valley Road. Turn left onto Hughes Road.

0.1 24.1

<u>Stop 1A</u>. Valley fill stratigraphy represented in drill holes 1 through 6 (fig. 2).

The depth to bedrock is 30 m (98 ft) at drill hole 1, 22 m (70 ft) at drill hole 2, and 11 m (34 ft) at drill hole 4. The 30 m of valley fill is unusually deep for small valleys that are internal to the Driftless Area. The deep fill has resulted because Citron Valley has trapped most of the sediment derived from uplands throughout much of the late Pleistocene.

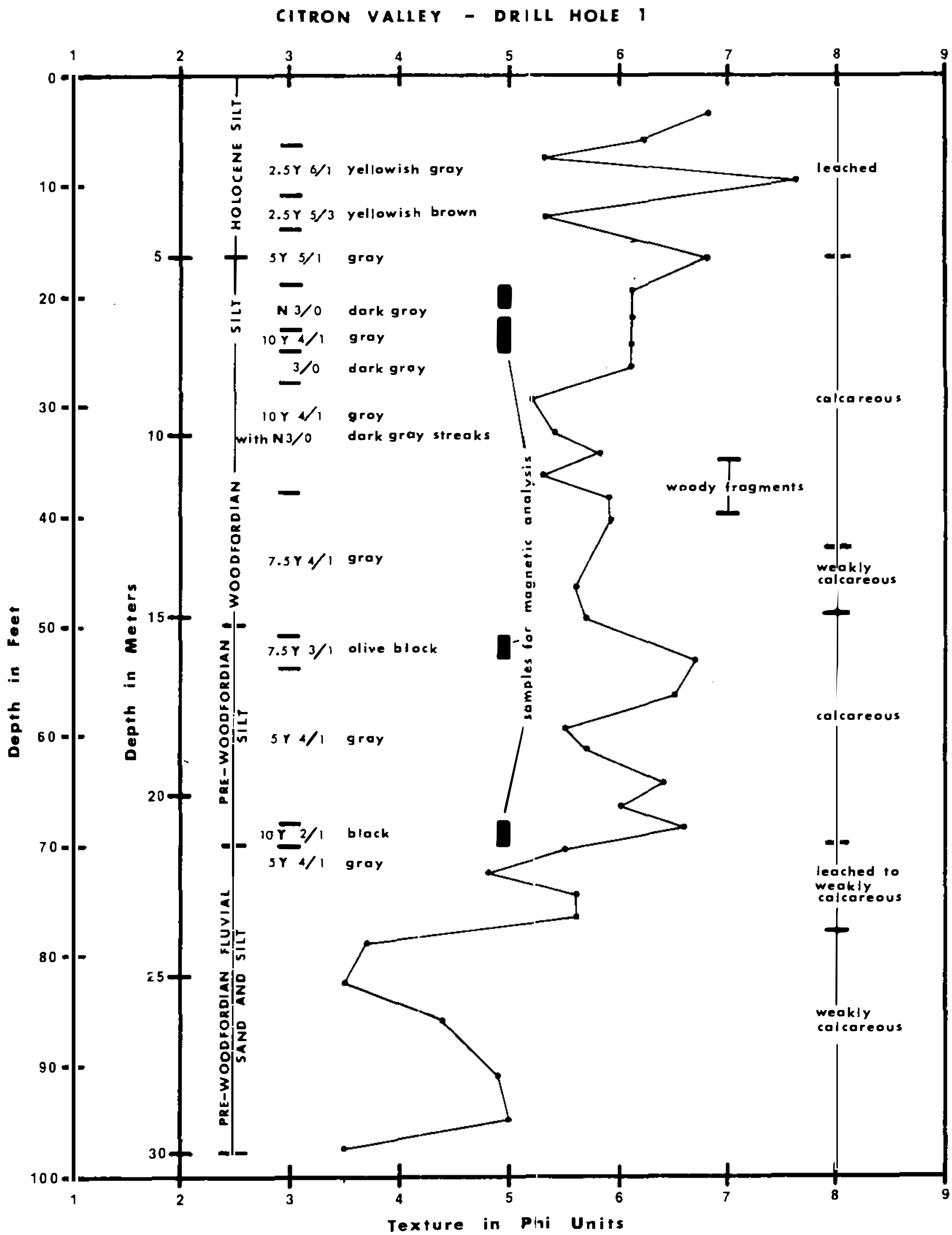

FIGURE 5.--Median texture and other physical properties of stratigraphic units at drill hole 1. See figure 2 for location.

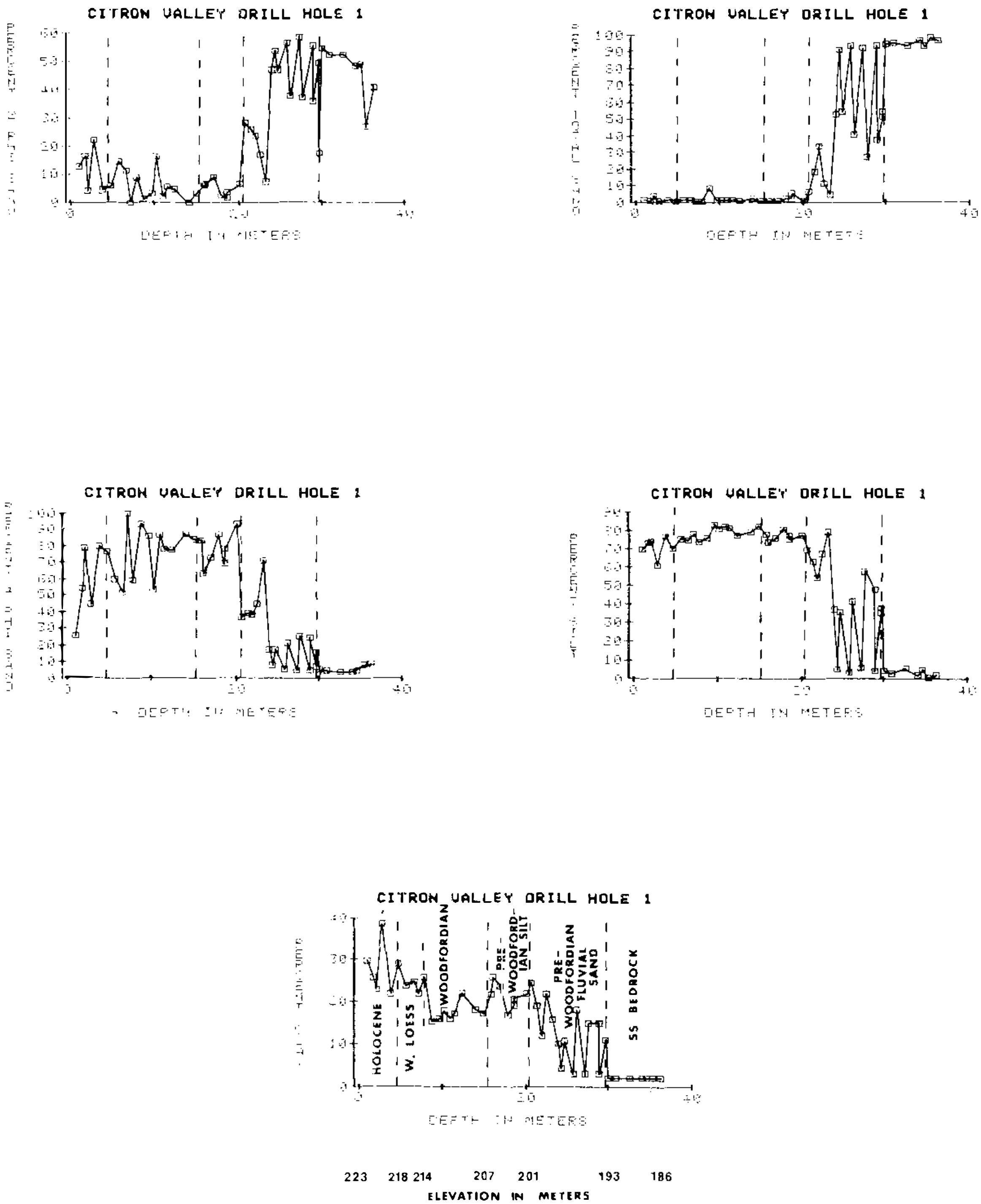

FIGURE 6.--Stratigraphy and sedimentology at drill hole 1. See figure 2 for location.

The stratigraphy in drill hole 1 is presented in figures 5 and 6, which show four major depositional units. The basal unit consists of interbedded silt and sand. The sand contains pebbles below a depth of 25 m (80 ft), which suggests that the basal unit probably was deposited by river processes. The top of the basal unit occurs at a depth of approximately 21 m (70 ft) and is leached to weakly calcareous in the upper 2.5 m. Silty sand and sandy silt in the basal unit is gray (5Y 4/1). The gray silty deposits are capped with a black (10Y 2/1) fine to medium silt that may represent a former soil horizon. The calcareous character of the black silt may have resulted from downward movement of calcareous fine sediments from the overlying unit. The age of the basal unit is unknown. It is possibly as old as classical Kansan. A Kansan age assignment assumes that the silts which are interbedded with the fluvial sand and gravel resulted from ponding when aggradation of the Wisconsin River valley by Bridgeport outwash caused the base level at the mouth of the Kickapoo River to rise. I have traced the up-valley distribution of meander scars on the bedrock valley walls that represent relict courses of the Kickapoo River (fig. 7). The up-valley distribution of the meander scars extends to an elevation of approximately 225 m (740 ft) above sea level. Since the maximum elevation of the Bridgeport outwash surface near Wauzeka at the mouth of the Kickapoo River is also approximately 225 m (740 ft) above sea level, it appears that the initial phase in the development of cut-off valley meanders is related to the Bridgeport outwash. The relationship of the meander scars with the Bridgeport till and outwash therefore supports an age assignment of classical Kansan (older than about 730,000 years) for the cut-offs. However, if Trowbridge (1954) was correct in his interpretation that the rock strath under the Bridgeport till represents the level of river incision in Kansan time, then the basal unit in Citron Valley must be younger since the bedrock floor of Citron Valley is approximately 20 m (70 ft) lower than the surface on the rock bench near Wauzeka. Tests for remnant magnetism were also conducted for several horizons in drill hole 1 (fig. 5). Since the last major reversal from the Matuyama (reversed) to the Bruhnes (normal-present) occurred about 730,000 years ago, finding of reversed remnant magnetism in Citron Valley sediments would support an age assignment to the Kansan. Unfortunately, it was not possible to extract a core from the basal fluvial unit, but cores were taken at several different horizons in overlying units. The polarity of the detrital remnant magnetism was analyzed with a spinner magnetometer. All samples were found to have normal polarity, including a sample from the fine to medium black silt at a depth of 21 m (68 to 70 ft). These tests therefore at least imply a post-Kansan age for all of the sediments overlying the basal unit. The presently available information therefore leads to a conclusion that the initial development of the cut-off valley meanders occurred when the Bridgeport outwash caused sedimentation in the lower Kickapoo valley. These sediments apparently accumulated in sufficient depth to overtop low sags on bedrock ridges on the inside bends of valley meanders. The incision of Citron Valley below the Bridgeport strath and the apparent post-Kansan age for the bulk of the valley fill in Citron Valley implies that at least some of the cut-off valley meanders may have been reoccupied by the Kickapoo River in post-Kansan time and that the

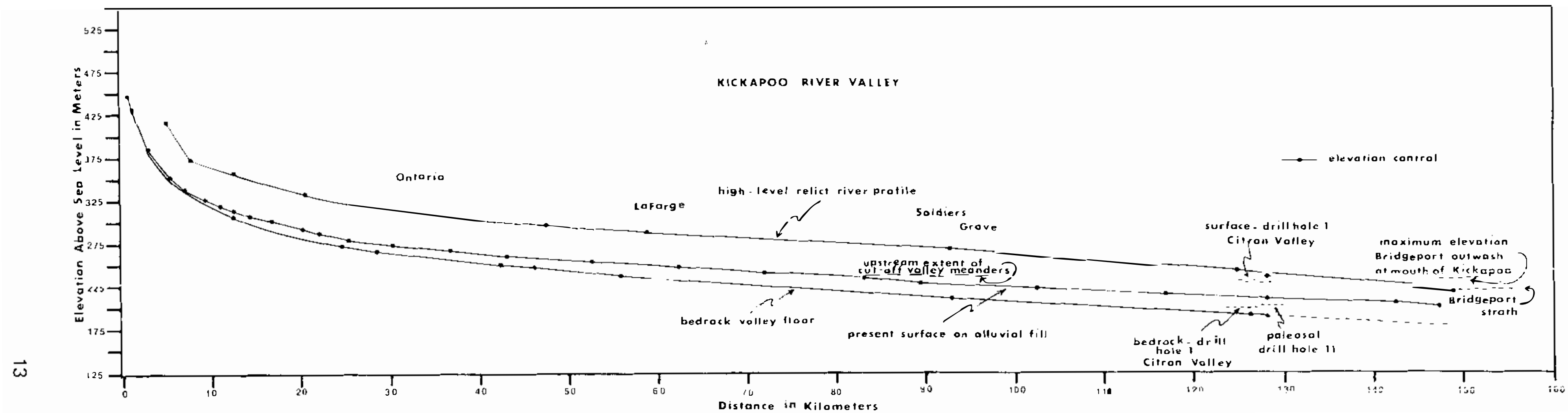

FIGURE 7.--Longitudinal profiles of geomorphic surfaces in the Kickapoo valley. Elevations were determined from 1:62,500 scale topographic maps with 20 ft (6 m) contour intervals. Depth to bedrock was determined by drilling. Note that the high-level relict river profile appears to be graded to the Bridgeport strath surface. Note also that the up-valley extent of scars representing relict valley meanders ends at about the same elevation as the Bridgeport outwash. The relationship suggests that the base level rise at the mouth of the Kickapoo during the time of Bridgeport outwash caused major aggradation in the lower Kickapoo valley and was the initial factor leading to the development of cut-off valley meanders.

lower Kickapoo valley was again subjected to major aggradation in post-Kansan time.

A second episode of aggradation of Citron Valley is represented in drill hole 1 by medium and fine gray (5Y 4/1) silts that are capped by olive black (7.5Y 3/1) fine to medium silt (fig. 5). The olive black sediments included horizons with alternating gray and black beds of 1 to 3 mm thickness. Organic fibers were common along the bedding planes. The general absence of sand and coarser textured sediments from this unit indicates that deposition occurred under very low energy conditions, possibly in a shallow lake. Tests of remnant magnetism imply either an Illinoian or Wisconsinan age for the sediments. They are at least pre-Woodfordian because Woodfordian sediments rest on top of them. If these fine and medium silts were deposited in a shallow lake as hypothesized above, then they most probably are of Illinoian age because ponding in the upper Mississippi River and its tributaries probably occurred when Illinoian glacial ice advanced across the Mississippi River into southeastern Iowa. Alden and Leverett (Alden, 1918, p. 172) speculated that elevations of diversion channels along the margin of the former Illinoian ice front in Iowa imply ponding up to an elevation of 200 m (720 ft) above sea level. Recent work by Updegraff (1981) has shown that present surface elevations in the diversion channels can be misleading because they may be filled with post-Illinoian deposits. Updegraff concluded that deposits in the Goose Lake diversion channel, which had previously been recognized as related to Illinoian ice, presented no evidence to substantiate the idea that Goose Lake valley in eastern Iowa was utilized by the Mississippi River during the Illinoian. It seems likely that some ponding occurred in response to blockage of the Mississippi River, as has been summarized by Willman and Frye (1970), but it is unclear at this time what the depth and duration of ponding may have been. It is also possible that the fine and medium silts were deposited by sheetwash and mass wasting processes which appear to have dominated for the overlying Woodfordian sediments. The higher percentage of clay in the lower silts compared to the overlying Woodfordian silts suggests that processes responsible for the two units were different, although the smaller clay percentages in the Woodfordian may simply reflect the greater importance of loess derived sediments there. If the lower silts are not of a lacustrine origin, then the buried paleosol at an elevation of 194 m (635 ft) above sea level at drill hole 11 may correspond in age with the weakly developed weathering zone at an elevation of about 207 (680 ft) above sea level in drill hole 1 (fig. 3). The latter interpretation seems unlikely, however, because the weathering zone that separates the pre-Woodfordian and Woodfordian silts occurs at nearly the same elevation throughout Citron Valley. Drill holes 9 and 10 are located in a position where alluvial fan deposition occurred during Woodfordian time and should have also occurred during the time of deposition of the underlying fine and medium silts if they were deposited in a non-lacustrine environment. Figure 3 shows that no elevated alluvial fan surface is present in the subsurface. Therefore, it is apparent that the hillslope processes which were characteristic of the Woodfordian were not dominant when the fine and medium silts were being deposited.

The base of the Woodfordian sediments occurs between about 207 and 209 m (680 and 685 ft) above sea level in Citron Valley. The base is defined by the change from strongly calcareous conditions above the contact to weakly calcareous or leached conditions below. At drill hole 9, where drainage is good (fig. 2), a well-developed soil occurred at the base of the Woodfordian. The top of the Woodfordian sediments vary from the modern surface at drill hole 9 to subsurface positions under Holocene sediments elsewhere. At drill holes 1 and 2 there are 5 and 6 m (17 and 20 ft) of leached Holocene sediments overlying the Woodfordian. The depth of leaching increases to 6 to 7 m (20 to 23 ft) at drill hole 4 in the side valley tributary (fig. 2). The slight trend toward increasing depth of leaching into the tributary indicates that the Holocene sediments probably experienced most of their leaching when they were located on hillslopes and have since been washed onto the floor of Citron Valley.

The Woodfordian sediments in drill hole 1 tend to be darker grays than the underlying fine and medium silts (fig. 5). The occasional occurrence of small woody fragments in the lower section of the Woodfordian and the common occurrences of organic rich dark gray (N 3/0) horizons in these silt dominated sediments imply accretion under very low energy conditions. I envision the environment of the valley floor to have been poorly drained shrub, sedge, and grass in the early Woodfordian about 20,000 years ago. As the climate became progressively more harsh when glacial ice lobes extended southward into central Iowa on the west and into southern Wisconsin and northern Illinois on the east, the shrubs probably disappeared from the valley floor and mass wasting of sediments from hillslopes probably increased. The coarsening of the sediments in drill hole 1 between depths of 8.5 and 11 m (28 and 37 ft) are interpreted as an index of a shift toward periglacial climatic conditions. Unfortunately pollen is too poorly preserved in the sediments to allow meaningful interpretations of past climate condtions. Radiocarbon dates associated with Des Moines lobe deposits in Iowa (Ruhe, 1969) and with ice marginal Woodfordian deposits on the eastern margin of the Driftless Area at Devils Lake (Maher, this volume) suggest that cold climate conditions extended to at least 14,000 years ago and possibly as late as 12,500 years ago. During this cold phase mass wasting and sheet wash appear as the dominant hillslope erosion processes. The mean frontal zone representing the northern boundary of moist air from the Gulf of Mexico probably was always south of the Driftless Area in all seasons of the year. Southward displacement of the frontal zone is suggested by the fine textures of sediments in Woodfordian alluvial fans. The streams that feed the alluvial fans cut through dolomite of the Prairie du Chien Formation which provides an abundance of pebbles, cobbles, and boulders under present-day conditions. The Woodfordian sediments of drill holes 4 and 6 (fig. 2) showed very little gravel and most of the Woodfordian section at each site was dominatd by reworked loess. The general absence of sediments coarser than sand implies that high intensity thunderstorms must have been extremely rare during the Woodfordian. Snow and low intensity frontal rainfall must have been the dominant precipitation processes.

The upper 2.5 m (8 ft) of the Woodfordian sediments in drill hole 1 are very homogeneous medium silts (fig. 5). The unit is traceable through several drill holes in Citron Valley (fig. 3). It appears that the medium silts represent an episode of intensified loess deposition toward the end of the Woodfordian. Although the silts probably are derived in part from reworked loess transported from hillslopes, I believe that much the accumulation may have been deposited *in situ*. Evidence to support *in situ* accumulation is suggested by the remarkably uniform thickness of the medium silts between drill holes, except for those in alluvial fans where the thicknesses are much greater due to hillslope contributions. For example, in drill hole 6 a major increase in loess contribution occurs at a depth of about 8.5 m (28 ft) below the surface. No gravels were observed above 8.5 m depth and the sand fraction was dominated by fine sand (fig. 8). Because the sediments at drill hole 6 were leached to a depth of 4.6 m (15 ft). The upper 3 to 4 m probably represent Holocene sedimentation. More discussion about Woodfordian sedimentation will follow at stop 1B.

The topmost major stratigraphic unit in drill hole 1 is Holocene in age. The lower boundary is set at about 5 m depth, which corresponds with the depth of leached sediments. It is unlikely that *in situ* leaching would be very extensive on the poorly drained floor of Citron Valley, a condition that existed in the Woodfordian and during most of the Holocene. Therefore, the leached sediments probably are derived from better drained sites on the surrounding hillslopes and have been transported onto the valley floor by running water. Assuming that the placement of stratigraphic boundaries are reasonably accurate, it is interesting to note that the thickness of Holocene sedimentation represents about one-third of the thickness of Woodfordian sedimentation in the area of drill holes 1 and 2 (figs. 2, 3, and 5). It is therefore apparent that a considerable quantity of sediment has been removed from the hillslopes, even under vegetation and climate conditions of the Holocene. The relatively thick loess on the uplands and the steepness of hillslopes in the Citron Valley watershed may account for the very high magnitudes of Holocene sedimentation. The proportions may not be representative of other parts of the Driftless Area.

Continue west on Hughes Road through the cutoff Citron Valley.

0.1 24.2

Drill hole 1 was located in field adjacent to south side of road.

Follow township road northwesterly along base of Hogback Hill. Note the large alluvial fans on the opposite valley wall. The major portion of these alluvial fans accumulated between 20,000 and 12,000 B.P.

Continue on the township road through the western limb of Citron Valley. Note the smooth concave profiles developed on colluvium and alluvium at the bases of hillslopes. These deposits illustrate the large quantity of sediments that were transported from hillslopes during periglacial climatic conditions of the Woodfordian.

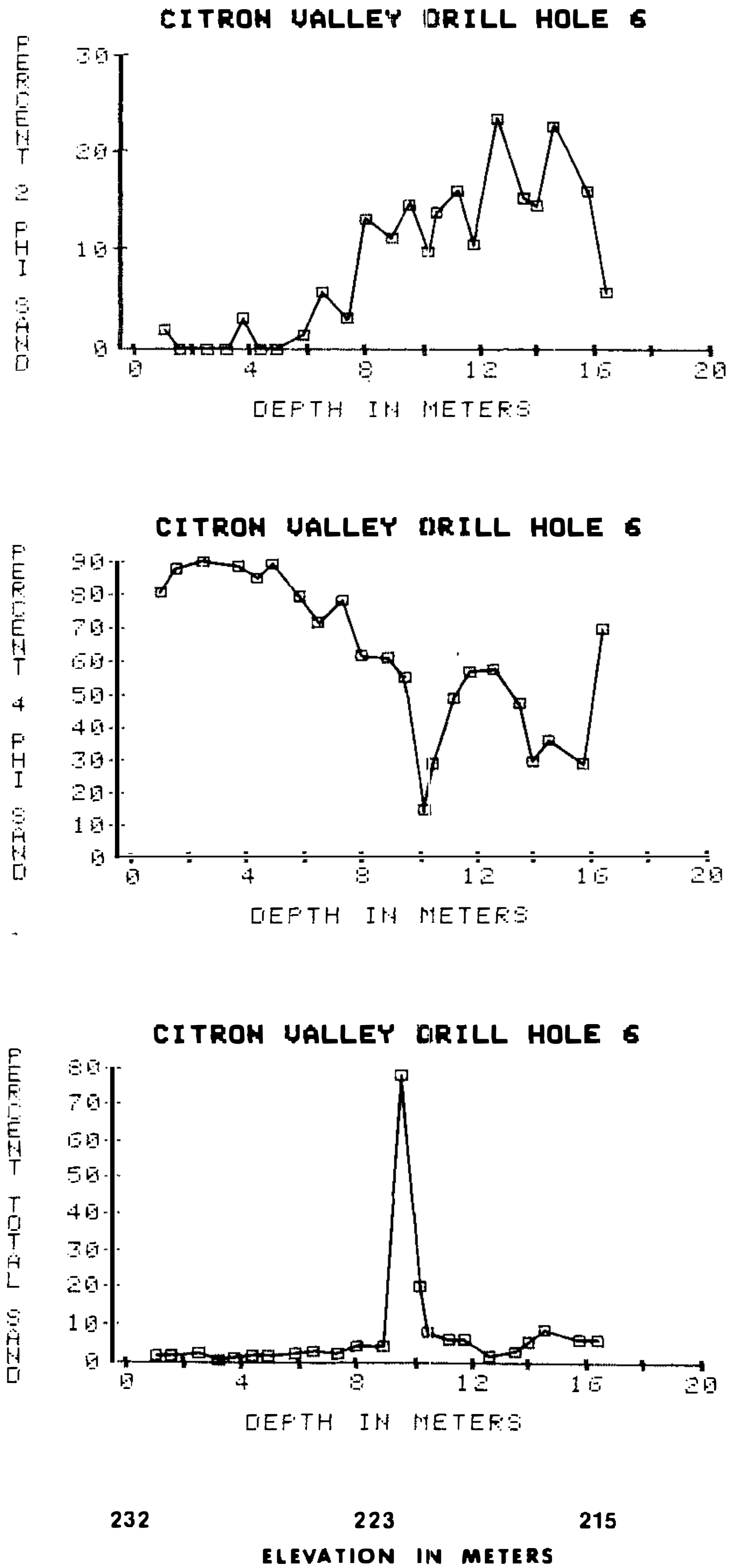

FIGURE 8.--Sedimentology at drill hole 6. See figure 2 for location.

2.3 26.5

Drill holes 9 and 10 are located here near the edge of the high surface in Citron Valley (fig. 2). The small steep tributary to the immediate east of the sites indicates that drill holes 9 and 10 are sensitive as indicators of past alluvial-fan deposition.

The stratigraphic sequence of sediment textures at drill hole 9 is presented in figure 9. The surface unit is interpretated as Woodfordian loess and is correlated with the upper Woodfordian unit that immediately underlies the Holocene sediments in drill hole 1 (fig. 3) There is little or no Holocene section represented at drill hole 9. The absence of a Holocene record probably results from trenching of the alluvial fan (near the present township road along the base of the hill) between 12,000 and 9500 years ago. The sedimentology of the Woodfordian sediments is generally similar to those at drill hole 1 in that dominance by reworked Woodfordian loess gives the unit a very high silt composition. The Woodfordian at this site has considerably more sand than the Woodfordian at drill hole 1. More sand should be expected here because of the close proximity to the steep tributary that cuts across sandstone bedrock. Note that the sand fraction increases to about 90 percent in the basal section of the Woodfordian. The very high sand percentage implies that in the early Woodfordian before much loess had blanketed the landscape, frost-dominated climate conditions probably were quite effective in weathering the sandstone rock. Relatively little loess would have been available to dilute the sand as it was transported onto the alluvial fan. The Woodfordian at drill hole 9 is leached to a depth of about 1.4 m (4.5 ft). Colors range from dull yellowish brown (10YR 5/4) to yellowish brown (2.5Y 5/3). The unit is strongly calcareous below the surface leached zone.

The sandy unit in the basal Woodfordian at drill hole 9 overlies a leached paleosol. The elevation of the paleosol is 209 m (685 ft) above sea level, an elevation that is nearly identical with the elevation for the base of the Woodfordian at drill hole 1 (fig. 3). The soil was dark brown (10YR 3/4 to 4/4) and leached. The similarity of elevations for the Woodfordian base suggests that a large "Woodfordian-type" alluvial fan is not present in the upper pre-Woodfordian sediments at drill hole 9.

Figure 9 shows that the pre-Woodfordian sediments contain on the average more sand and clay and less silt than the overlying Woodfordian sediments. Their average color is dull yellowish brown (10YR 5/4) and they were leached throughout the depth of drilling to 21 m (70 ft) below the ground surface. As indicated in the discussion for drill hole 1, the sediments are interpretated as shallow lacustrine. The greater variability in sand, silt, and clay percentages here, compared to drill hole 1, probably reflects the local influence of the steep tributary.

Further support for sediment accumulation in shallow water conditions is indicated by figure 10 which presents a higher resolution evaluation of vertical variations in sediment textures. Note that the vertical line separating Woodfordian and pre-Woodfordian sedi-

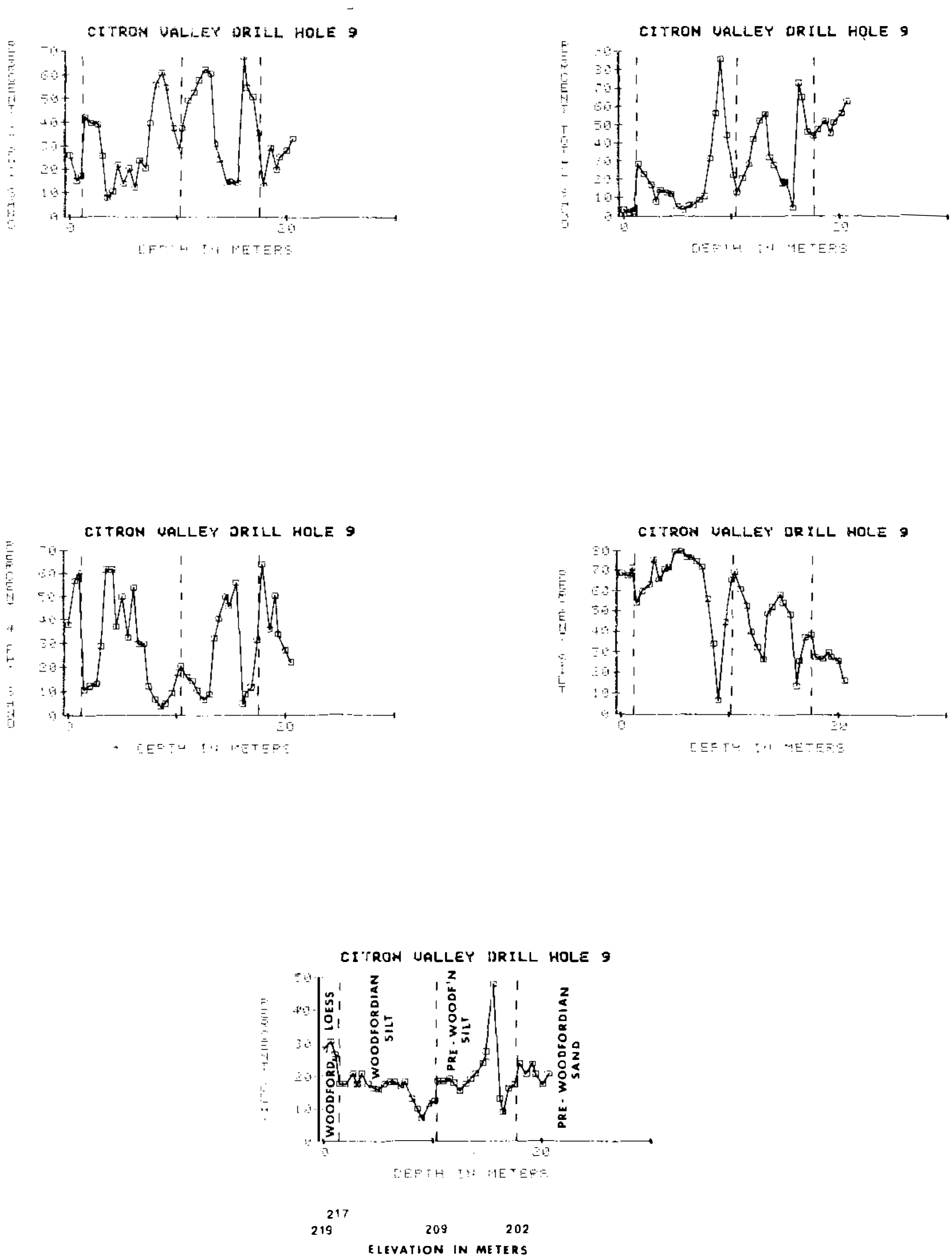

FIGURE 9.--Stratigraphy and sedimentology at drill hole 9. See figure 2 for location.

ments corresponds with an elevation of 209 m (685 ft) above sea level. The data on figure 10 therefore correlate with the data centered on about 10 m depth on figure 9. The pre-Woodfordian sediments at drill hole 10 are slightly to moderately calcareous below a surface leached horizon of about a meter thickness. The pre-Woodfordian sediments between about 12.5 and 16.5 m (40 and 54 ft) at drill hole 10 contain very little sand and high percentages of silt and clay. Sediment cores showed variations in bedding that ranged from finely-laminated to massive. The sediments appeared to have been deposited in ponded waters. Not all of the pre-Woodfordian sediments in drill hole 10 were as fine-grained as those represented in figure 10. Below a depth of 16.5 m (54 ft) the sediments changed to sandy silts and silty sands. Below a depth of about 18 m (60 ft) occasional chert pebbles were observed. The sand and pebble zone below 16.5 m may correspond with the basal unit at drill hole 1. For example, the beginning of pebble occurrences at a depth of 18 m (60 ft) corresponds with an above sea level elevation of 203 m (665 ft). The top of the basal unit at drill hole 1 occurs at 201 m (660 ft) elevation above sea level. Bedrock was encountered at a depth of 20.7 m (68 ft). The higher elevation of bedrock here compared to bedrock elevation at drill hole 1 probably relates to drill hole 10'S location on the inside of a valley meander, whereas drill hole 1 is located on a relatively straight reach of valley.

Continue south on township road to the T intersection.

0.2 26.7

T intersection. Turn right to west. The township road is located on a low terrace that is about 3.7 (12 ft) above the floodplain of modern streams. The 3.7 m terrace is about 4.6 m (15 ft) lower than the high Wisconsinan terrace that marks the maximum level of Woodfordian valley aggradation (fig. 3). The average elevations above sea level for the low and high terraces are 206 m (675 ft) and 210 m (690 ft). These terraces are inset against the fill of Citron Valley (fig. 3). Since the average elevation on the floor of Citron Valley is about 220 m (720 ft) above sea level, the high Wisconsinan terrace is therefore about 10 m (33 ft) lower.

0.4 27.1

Ascend high Wisconsinan terrace.

0.1 27.2

Drill hole 12 is located in the pasture on the north side of the road.

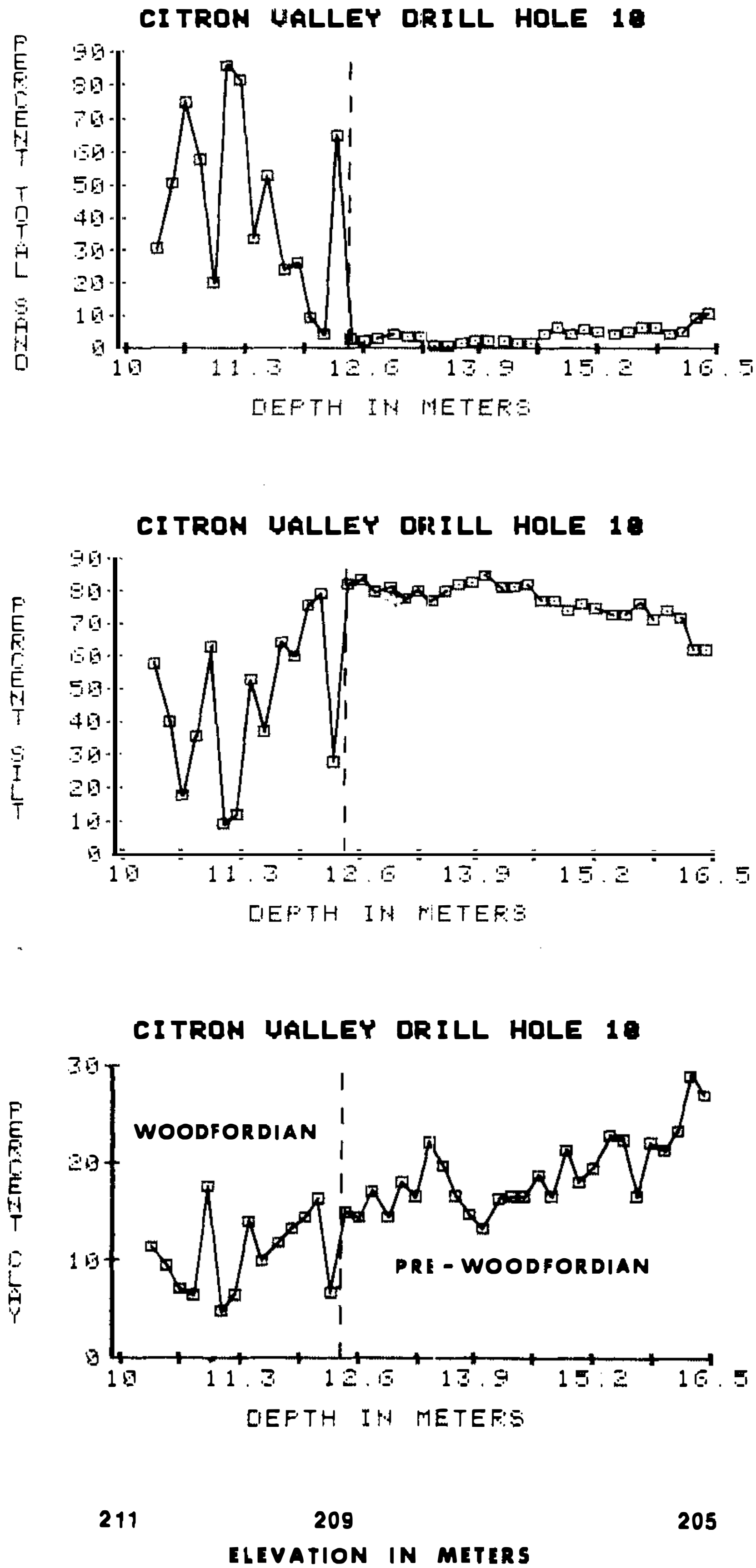

FIGURE 10.--Stratigraphy and sedimentology at drill hole 10. The Woodfordian sediments are interpreted as fluvial slopewash and colluvium. The pre-Woodfordian sediments appear to have been deposited in standing water. See figure 2 for location.

0.1 27.3

Intersection with county highway E.

Stop 1B.

Stop 1B is to examine the relationships of the two late-Wisconsinan terraces with each other and their relationship to the fills in Citron Valley.

A comparison of the stratigraphic sequences represented in drill holes 7 and 12 (figs. 11 and 12) show that the low terrace is erosional and cut into the high Wisconsinan terrace. Note, for example, the percentages of 2-phi and 4-phi sand fractions agree closely for a comparable datum. Since the surface elevation at drill hole 7 is 204 m (670 ft) above sea level, and the surface elevation at drill hole 12 is 210 m (690 ft) above sea level, the 9 m thick stratigraphic sequence that is represented in drill hole 7 correlates with sediments at depths between 5 and 14 meters below the surface in drill hole 12.

The top 2.4 m (8 ft) of sediment at drill hole 12 is interpreted as mostly *in situ* loess because it averages nearly 70 percent silt and contains very little sand. If this unit were of fluvial origin, or from hillslope mass wasting, it should have a much higher percentage of sand than what is represented by the data shown in figure 12. The average composition of 65 to 75 percent silt and 25 to 30 percent clay is nearly identical to the percentages of these sediment fractions in the topmost 2.5 m (8.2 ft) of the Woodfordian at drill hole 1 and in the topmost 1.3 m (4.3 ft) of the Woodfordian at drill hole 9 (figs. 6 and 9). Since the youngest Woodfordian deposits at drill holes 1 and 9 were also interpreted as representing a strong *in situ* loess component, the top 2.4 m of sediment at drill hole 12 on the high Wisconsinan terrace is correlated in time with these deposits (fig. 3).

Assuming the loess designation is correct, the loess must have been deposited before about 12,000 B.P. because that is when incision of the valley fills began (Clayton, this volume). A very silty unit in the high Wisconsinan terrace at a depth of 6 to 8 m (20 to 26 ft) suggests that loess fall may have been intensive then too. However, the lower silt in drill hole 12 contains from 10 to 20 percent sand, which indicates that sediment movement from the hillslopes was also occurring. The relatively low percentages of silt and clay in the lower sequence of sediments at drill hole 12 suggests that loess fall was less intense when those sediments were aggrading the valley floor.

0.0 27.3

Turn right (north) onto county highway E and continue northwesterly up the valley of Citron Creek. Note level of Woodfordian colluvial terrace on valley sides.

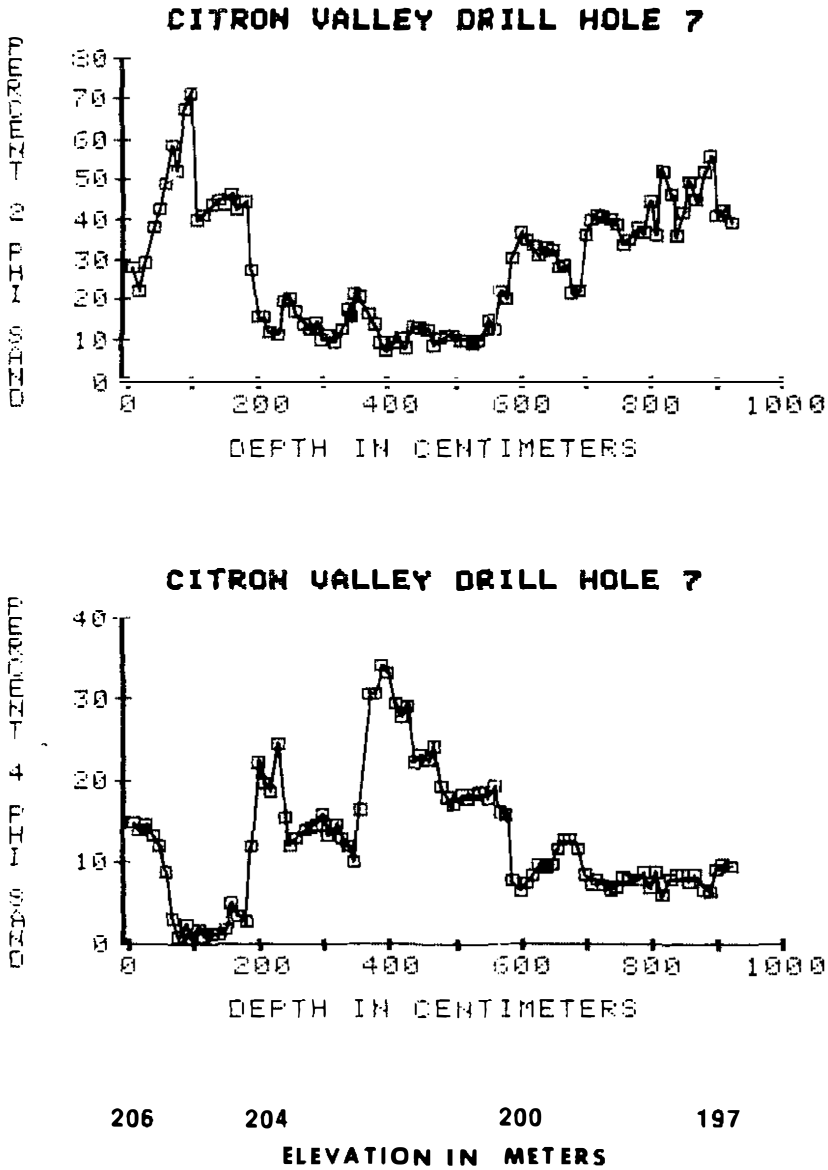

FIGURE 11.--Vertical variation in the quantity of medium and fine sand at drill hole 7. See figure 2 for location. The variation in fine and medium sand at drill hole 7 closely matches the variation in fine and medium sand at drill hole 12 (fig. 12) at equivalent elevations. The similarity indicates that the terrace on which drill hole 7 is located is erosional and cut into the deposits of the high terrace at drill hole 12.

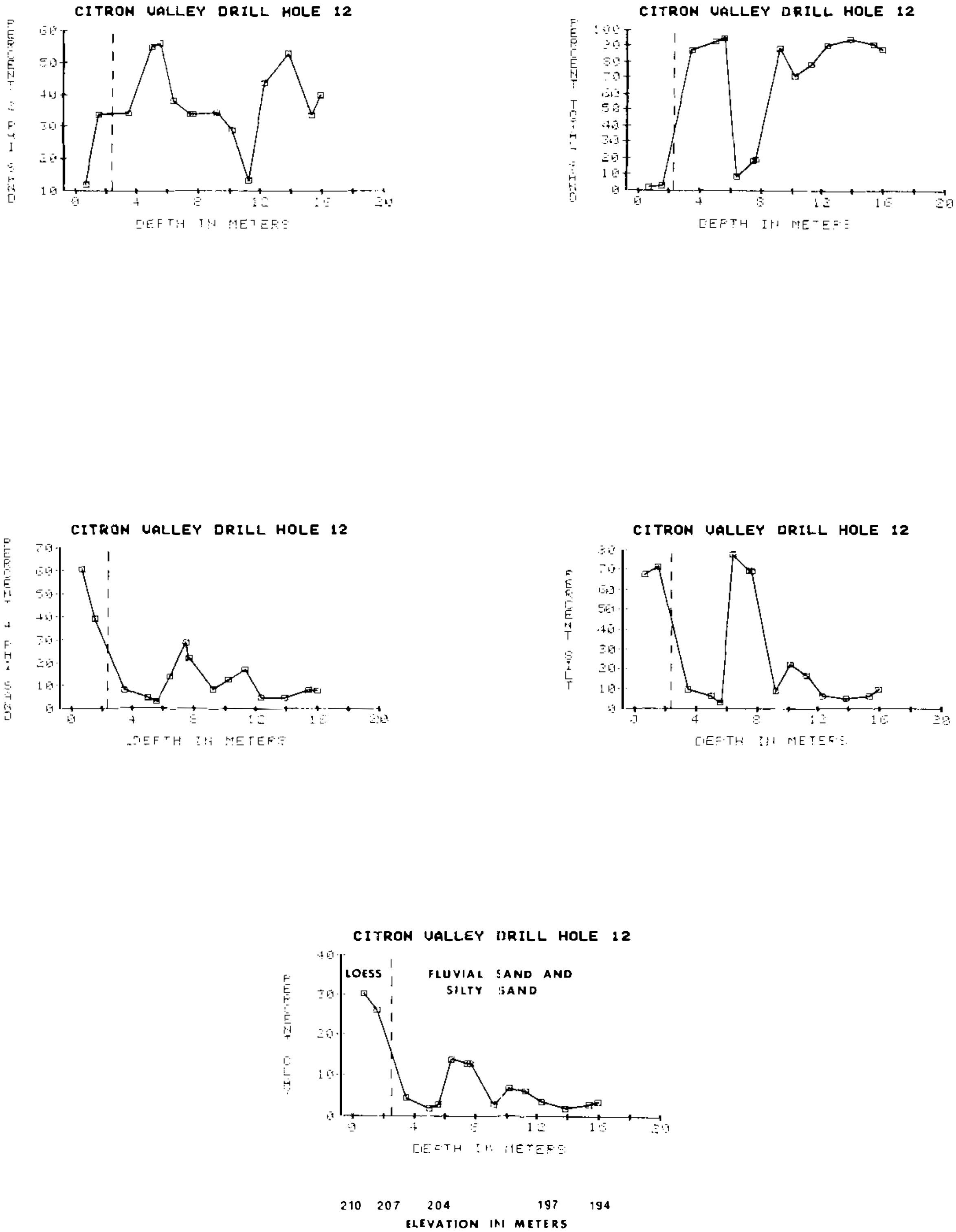

FIGURE 12.--Stratigraphy and sedimentology of the high Woodfordian terrace at drill hole 12. See figure 2 for location.

1.3 28.6

Highway makes a sharp curve to west. Begin ascending upland via Duffy Creek tributary. Note that the bedrock changes from sandstone in the lower hillslopes to dolomite in the upper hillslopes.

2.0 30.6

Outcrop of Pairie du Chien Formation.

0.6 31.2

The surrounding landscape provides a classic example of the two erosion surfaces described as peneplains by Trowbridge (1921) and others. The Lancaster erosion surface is illustrated on the lower upland interstream divides to the northeast of the highway.

0.3 31.5

This highest position on the main watershed divide represents the Dodgeville erosion surface. The Dodgeville and Lancaster surfaces are present on both sides of the highway along the next 0.8 km (0.5 mi).

2.1 33.6

Intersection of county highway E with state highway 27. Turn right (northeast) and continue through the village of Seneca.

0.4 34.0

Enter village of Seneca. Stay on Highway 27 through Seneca and to stop 2 about 0.6 km (0.4 mi) north of the village.

1.1 35.1

Stop 2. Seneca gravels (fig. 13).

The Seneca gravels are considered part of the Windrow Formation. Thwaites and Twenhofel (1921, p. 294) defined the Windrow Formation as consisting of "...quartz and chert pebbles in a matrix of quartz sand and brown iron oxide, iron oxide cemented sandstone, concretionary limonite, and at some localities blue and white sticky clay." The Windrow Formation is thought to be of Cretaceous age based on its distribution in Iowa and Minnesota.

The description provided by Thwaites and Twenhofel provides a reasonably accurate description of the Seneca gravels at stop 2 where they occur overlying dolomite of the Galena Formation. Andrews (1958) conducted a detailed examination of the Windrow Formation and subdivided it into an older Iron Hill Member consisting of iron oxide replacement deposits and a younger East Bluff Member consisting of clastic sediments of a fluvial origin. He recognized both members at the Seneca site. Andrews (1958, p. 616) concluded that the heavy-mineral suite of the Windrow Formation suggests a source of igneous

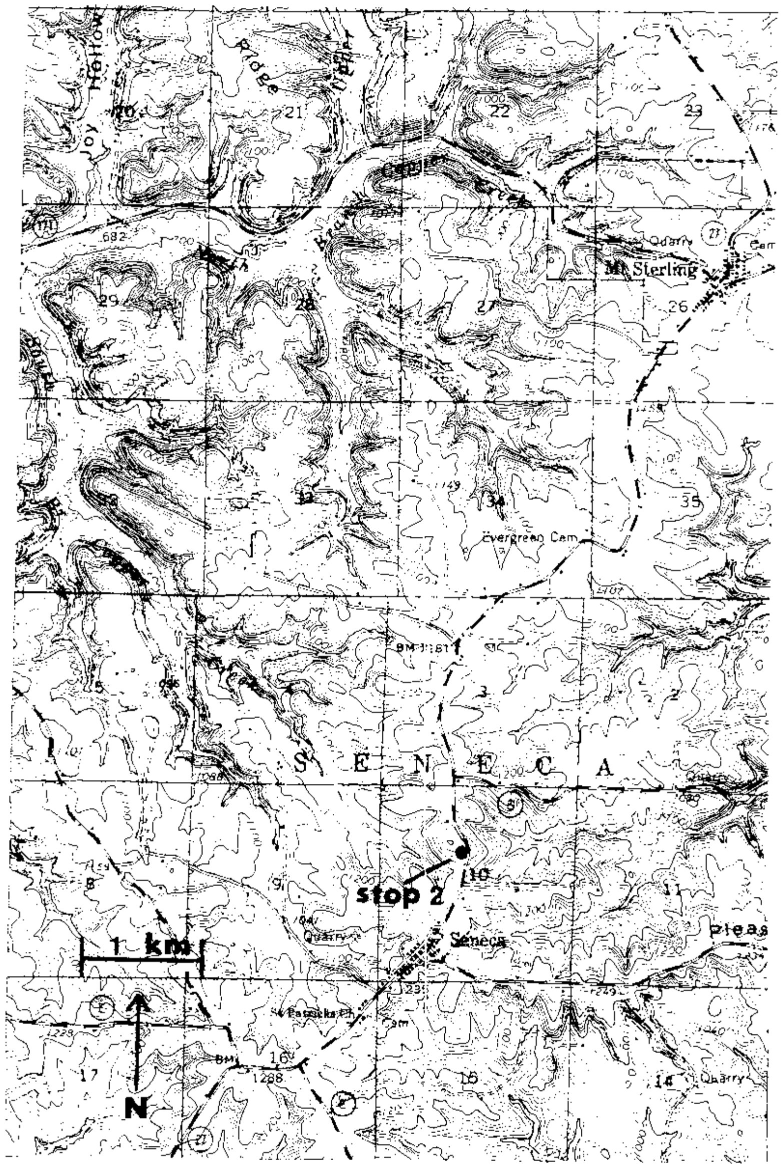

FIGURE 13.--Topographic map of the Seneca area, Crawford County, Wisconsin. The upland site at stop 2 exposes deposits of Windrow gravels.

and metamorphic rocks in addition to the various Paleozoic sedimentary rocks of the region.

The highly polished quartz pebbles in the Windrow Formation have been interpreted by many as stream deposits of channels that flowed on one of the upland erosion surfaces. Trowbridge (1921, p. 111-113) originally thought that the Windrow gravels were of a Tertiary age and accumulatd on the Dodgeville erosion surface. Later, he (1954, p. 802) suggested that only the Lancaster surface was a true erosion plane and he then accepted a Cretaceous age for the gravels. Andrews (1958, p. 619) concluded that the Dodgeville erosion surface at its type locality is much younger than the surface upon which the Windrow Formation rests.

The Windrow gravels are widely distributed throughout the Driftless Area. They are found under Pleistocene loess deposits and in caves and crevices within the Paleozoic bedrock formations. It would be easy to misinterpret the highly polished quartz pebbles as glacial erratics.

Continue northward on highway 27 to the village of Mt. Sterling. The thickness of reddish clay-rich soils that occur between the Woodfordian loess and bedrock varies considerably between sites in the Driftless Area as is apparent along this ridgeline. Frolking (this volume) found that the red clay soils are mineralogically uniform. Both within profiles and among different sites. He found that (1) smectites and mixed layer intergrades dominate both the fine-medium and coarse clay fractions, (2) kaolinite and quartz occur in moderate amounts, and (3) mica and vermiculite are present in small quantities. Frolking's analyses showed that the Woodfordian loess, except for greater amounts of vermiculite, exhibits a clay mineralogy similar to that of red clay. His textural, mineralogical, and thin section analyses led him to conclude that the high clay content of the red clays does not result solely from weathering of dolomite residuum or loess since the silt and sand fractions of these materials are largely quartz. Frolking suggested that the texture of the clays must result in part from an influx of clay and that much of the fine-medium clay could be illuvial.

Pleistocene erosion has removed much of the red clay residuum at most sites in the Driftless Area. The amount of red clay residuum that remains at a given site appears to be closely related to the amount of chert that is provided from the underlying bedrock. Red clay residuum tends to be very thin to absent at sites where chert is absent in the underlying bedrock. The critical geomorphic significance of the chert relates to its tendency to form a resistant lag deposit (stone line) on the surface and thereby slow the rate of fluvial erosion.

Note outcrop of St. Peter sandstone on ridge to left (west) side of highway 27.

2.2 38.7

Enter village of Mt. Sterling. Continue north on highway 27 to the intersection with state highway 171.

0.1 38.8

Intersection of highways 27 and 171. Turn right (northeast) onto highway 171 and proceed to the village of Gays Mills.

0.8 39.6

Small quarry on left (north) side of road is in St. Peter sandstone.

0.7 40.3

Descend upland and return to the Kickapoo valley floor via tributary of Caswell Hollow. Note passage through the dolomite of the Prairie du Chien Formation in the upper hillslopes. Below about 290 m (950 ft) above sea level, the valley walls are underlain by sandstone of the Trempealeau Group.

3.0 43.3

Entering Kickapoo valley floor. Note colluvial bench on valley sides.

0.4 43.7

Enter village of Gays Mills.

0.2 43.9

Cross Kickapoo River. The downtown section of Gays Mills is constructed in part on a low terrace. At this time it is unknown whether the terrace is correlative with the low terrace at drill hole 7 in Citron Valley or whether it is a younger Holocene terrace.

Continue east through Gays Mills to Junction with state highway 131 near base of hill on the east side of the village. Turn left (north) and continue on Highway 131.

1.7 45.6

Note sand exposure on right (east). The lower middle reaches of the Kickapoo valley, especially between Gays Mills and Soldiers Grove (fig. 14), are characterized by thick deposits of sand on the eastern side of the valley.

0.3 45.9

A drill hole was placed here adjacent to the west side of highway 131 (fig. 14). The elevation above sea level at the top of the drill hole was estimated to be about 232 m (760 ft) above sea level.

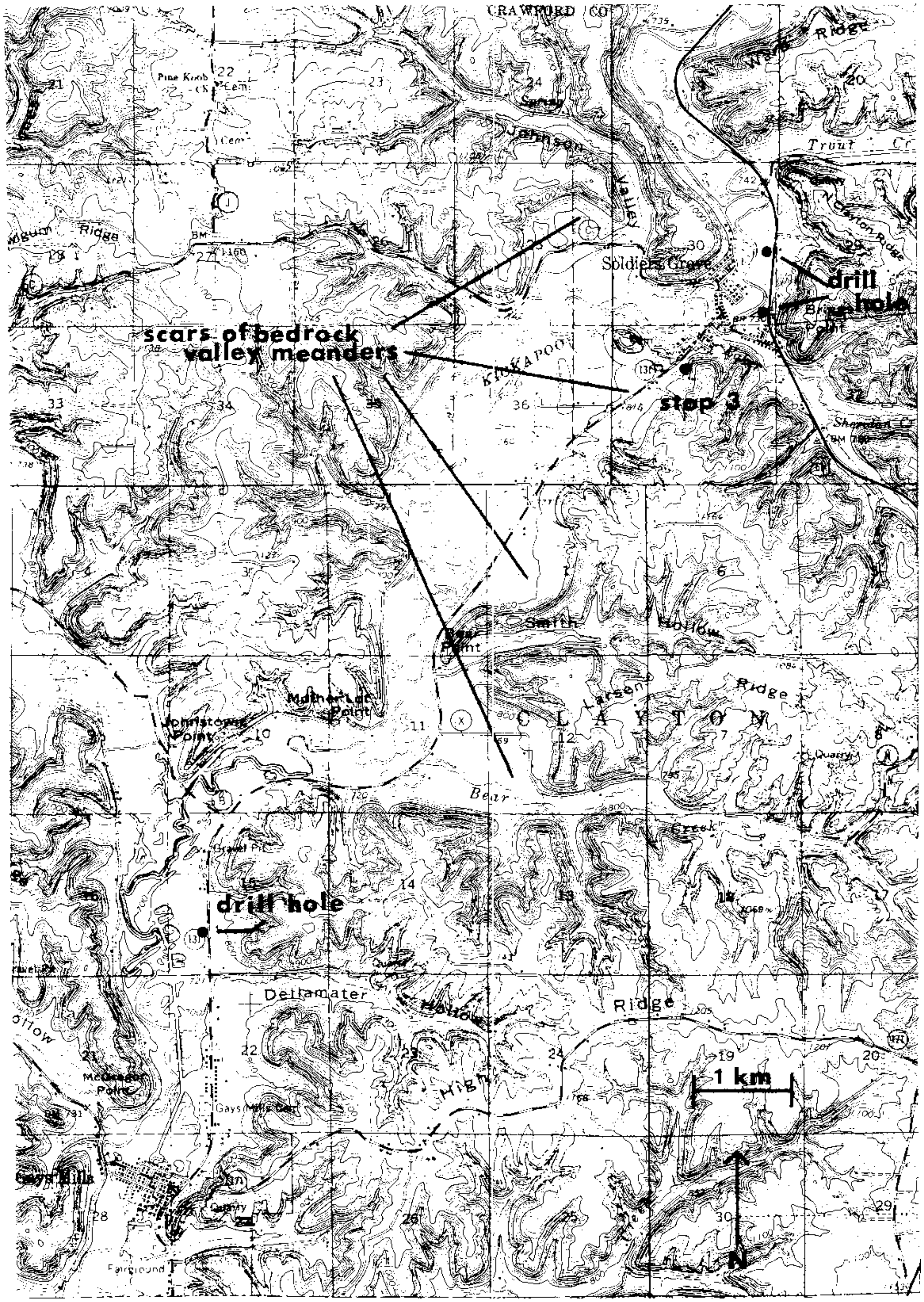

FIGURE 14.--Topographic map of the Gays Mills to Soldiers Grove reach of the Kickapoo valley, Crawford County Wisconsin. Note the abundance of scars on the valley sides representing relict courses of the Kickapoo River. The Gays Mills bench drill hole in a deep sand deposit is located in the lower left part of the map. Data for the sand are given in figure 15.

The elevation at the top of the drill hole is therefore about 6 m (20 ft) higher than the base level represented by the top of the Bridgeport outwash in the Wisconsin River valley at the mouth of the Kickapoo River (Knox, Attig, and Johnson, this volume). It is possible that the abundance of sand in this reach of the valley is related to the reworking (by wind action) of sand that may have accumulated at the head of a lake formed in the lower Kickapoo vally when the mouth of the river was blocked by outwash from the Bridgeport ice lobe. Another possibility relates to the contribution of the highly erodable Trempealeau Group of sandstone formations that are dominant in the valley walls of the lower middle Kickapoo valley.

Bates (1939, p. 868-870) categorized the section of the lower middle reach of the Kickapoo valley north of Gays Mills as particularly vulnerable to erosion. He divided the Kickapoo valley into four sections based on preservation of valley meanders. He stated (p. 868-870): "South of Gays Mills the base of the dolomite extends down to within less than 75 feet of the present stream level. It apparently has been sufficiently resistant to retard the destruction of meander-cores and inter-meander spurs by lateral and down-stream migration of the meander....The next section of the valley is...between Gays Mills and LaFarge....Here the valley has cut to a greater depth in the weak sandstones underlying the Prairie du Chien dolomite, and meander forms are not well preserved....The third part of the valley extends from La Farge to 3 miles north of Ontario....The outstanding feature of this section is the development of a rock terrace which forms the floor of an outer valley 400 feet deep and a mile in width, and below which the stream is incised 100 to 150 feet in large well developed meanders....Exposures along the inner valley of the stream show that the rock terraces are underlain by the Ironton sandstone member of the Franconia formation....This is a resistant stratum within the generally non-resistant...sandstones that underlie the Prairie du Chien dolomite....The valley above Ontario comprises the fourth section and is quite different from the others....Absence of the broad outer valley is explained by the fact that near Ontario the Kickapoo is joined by two tributaries, each of which is more than half the size of the mainstream....Apparently the smaller streams above this point have not been able to develop the wide meandering pattern which made possible the extensive removal of material above the resistant sandstone."

The median particle size averages about 2.4 phi (fine sand) for deposits that occur in the extensive sand bench north of Gays Mills (fig. 15). Graphical moments statistics for sorting, skewness, and kurtosis (Inman, 1952) indicate that the sands were not deposited by fluvial action (fig. 15). The sand is moderately well sorted, and the phi-distributions of the particle sizes are somewhat platykurtic and symmetrical about the median. Neither do the sands seem to be of colluvial origin because there is too little silt and clay associated with them. Total silt and clay in the samples associated with figure 15 were typically in the range of 4 to 6 percent by weight.

The particle size data presented in figure 15 suggest that wind action probably is responsible for deposition of most of the sand on the Gays Mills Bench. At least two episodes of sand deposition are

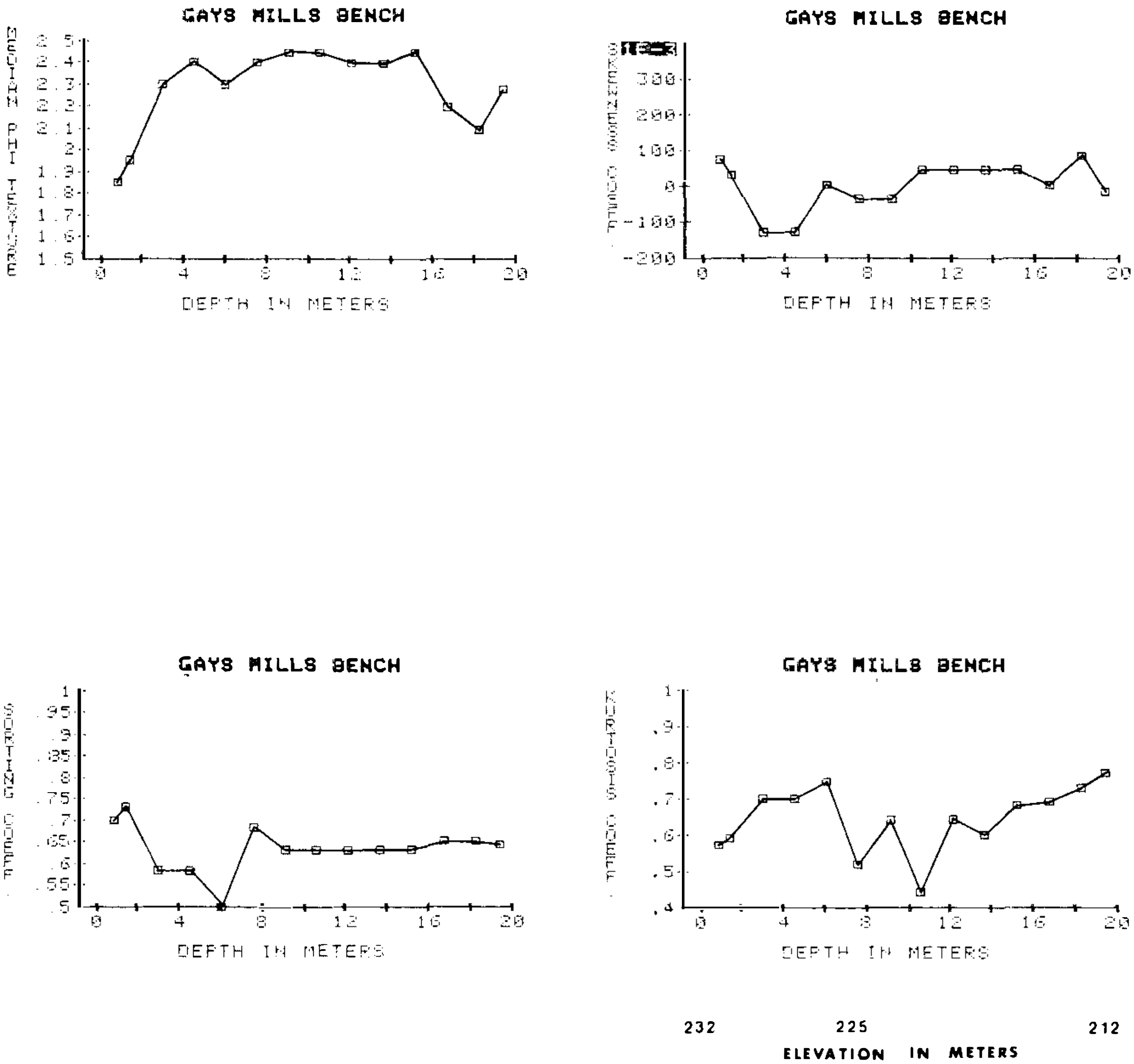

FIGURE 15.--Statistical properties of sediment textures of sand at the Gays Mills bench drill hole. The statistical parameters were determined by the graphic moments method. The statistical properties suggest that the sand deposit accumulated by aeolian processes. See figure 14 for location.

apparent, however. When drilling the site on the Gays Mills bench, a weak soil horizon was encountered at a depth of 7.6 to 8.2 m (25 to 27 ft). The soil was olive brown (2.5Y 4/3) and darker than the sands above and below which commonly were dull yellow brown (10YR 5/3) to yellowish brown (2.5Y 5/3). The sands above the soil were leached, as were the sands below the soil to a depth of about 12 m (40 ft). The sands were found to be very weakly calcareous below 12 m to the bottom of the drill hole at 19.5 m (64 ft). The sand became coarser below about 15 m (50 ft) depth, and collapse of the drill hole prevented drilling deeper than 19.5 m (64 ft). An electrical resistivity transect was conducted at the drilling site, and the results suggested a stratigraphic break at about 24 m (80 ft) below the surface. The elevation of the stratigraphic break is about 207 m (680 ft) above sea level. The elevation of 207 m is approximately 9 m (30 ft) higher than the elevation of the bedrock floor of the valley at this location (fig 7). Therefore, the stratigraphic break at 24 m (80 ft) below the surface may only represent a change in moisture content or a change in texture. It is also possible that it represents bedrock of a strath surface that is comparable in age to the floor of Citron Valley. It may be recalled that the bedrock floor of Citron Valley varied between 4 and 10 m (13 and 30 ft) higher than the bedrock floor of the adjacent gorge cutoff.

Continue north on highway 131 to Soldiers Grove.

2.0 47.9

Observe the prominent late-Wisconsinan terrace in Bear valley tributary to your right (east). The very large width of Bear valley tributary at its mouth results because the mouth area is a former valley meander of the Kickapoo River (fig. 14).

1.6 49.5

You are entering another valley meander scar. The area below he scar is also filled with an extensive sand deposit (fig. 14). Continue northeast on Highway 131 to stop 3 at the south edge of Soldiers Grove.

1.7 51.2

Stop 3. The large sand deposit on the east side of highway 131 at stop 3 was investigated by Akers (1965), a student of R.F. Black. I interpret the deposit as representing a sand dune of Woodfordian age, but Akers concluded that glaciation best explains the deposit. Akers (1965, p. 82) stated: "The deposit is near-deltaic in structure, yet without a source--a wasting ice block could provide that source." Akers suggested that sand grains in the deposit at stop 3 are less frosted and more angular than sand grains in the local bedrock or in the extensive sand deposits on the low bench along the eastern side of the valley. To support his glacial interpretation, Akers (1965, p. 82) also stated: "The deposit has fresh glauconite even where it lies above the highest glauconite-bearing bedrock--an ice block could easily have carried glauconitic sands to such a high position. The washing of the material, the less pitted appearance of the quartz, and the higher percentage of ilmenite and magnetite, all could result from the transport by and wastage of this material from ice." It is true that glacial processes could account for the phenomena Akers describes, but it is also true that the sand deposit at stop 3 could easily result from aeolian processes. As indicated above, I support the latter interpretation.

Continue on highway 131 through Soldiers Grove to junction between highways 131 and 61 on the southeast edge of the village.

1.0 52.2

Intersection of highways 131 and 61. Turn left (to north) and continue on highway 131. To combat the problems of annual flood damage, Soldiers Grove elected to move much of the business district away from the low ground near the river to a place of higher elevation. The new location is along highway 61 at the southeast edge of the present village. The move was supported in part by a federal grant. Most of the new buildings are heated by solar energy.

0.1 52.3

A drill hole was placed in the floodplain on the northeast edge of Soldiers Grove to determine the depth of Woodfordian fill and depth to bedrock (figs. 14 and 16). The sediments were represented by silt and sandy silt of Holocene age to a depth of about 2.75 m (9 ft). Woodfordian sediments are dominantly sand with minor silt and occasional gravel horizons. The Woodfordian sediments were leached and were typically dull yellowish brown (10YR 5/4) to yellowish brown (2.5Y 5/4) in color. The Woodfordian base is tentatively placed at a depth of 8.8 m (29 ft) where the sediments sharply increase in gravel content. The sedimentological measures presented in figure 16 show this break clearly in the percentage of the sand fraction that is retained on the 2-phi sieve. About 60 to 65 percent of the Woodfordian sand falls in the range from 1 to 2 phi, whereas the corresponding percentages for underlying pre-Woodfordian sand are between 40 and 50 percent (fig. 16). The pre-Woodfordian sands were somewhat lighter than the Woodfordian sands. Colors of 10YR 6/4 (dull yellow orange) were common. A slight calcareous reaction was also recorded for several of pre-Woodfordian horizons, especially those which were more gravel-rich. Weathered bedrock was encountered at a depth of 15 m (49 ft).

0.4 52.7

Note that the highway cuts through part of the colluvial bench extending out from the east valley wall. An open pit (summer 1981) in the lawn near the house at Everson's farm showed less than a meter of loess overlying relatively clean stratified sand. A drill hole was placed on the extension of the colluvial bench in Everson's field west of the highway (fig. 14). The drilling showed approximately 1 m (3 ft) of silt that graded into silty sand and sandy silt between 1 and 1.9 m (3 and 6.2 ft). Relatively clean sand became dominant below about 2 m depth. A gravel horizon occurred between 2.7 and 3.3 m (9.5 and 11 ft). Many of the sand horizons were of medium to

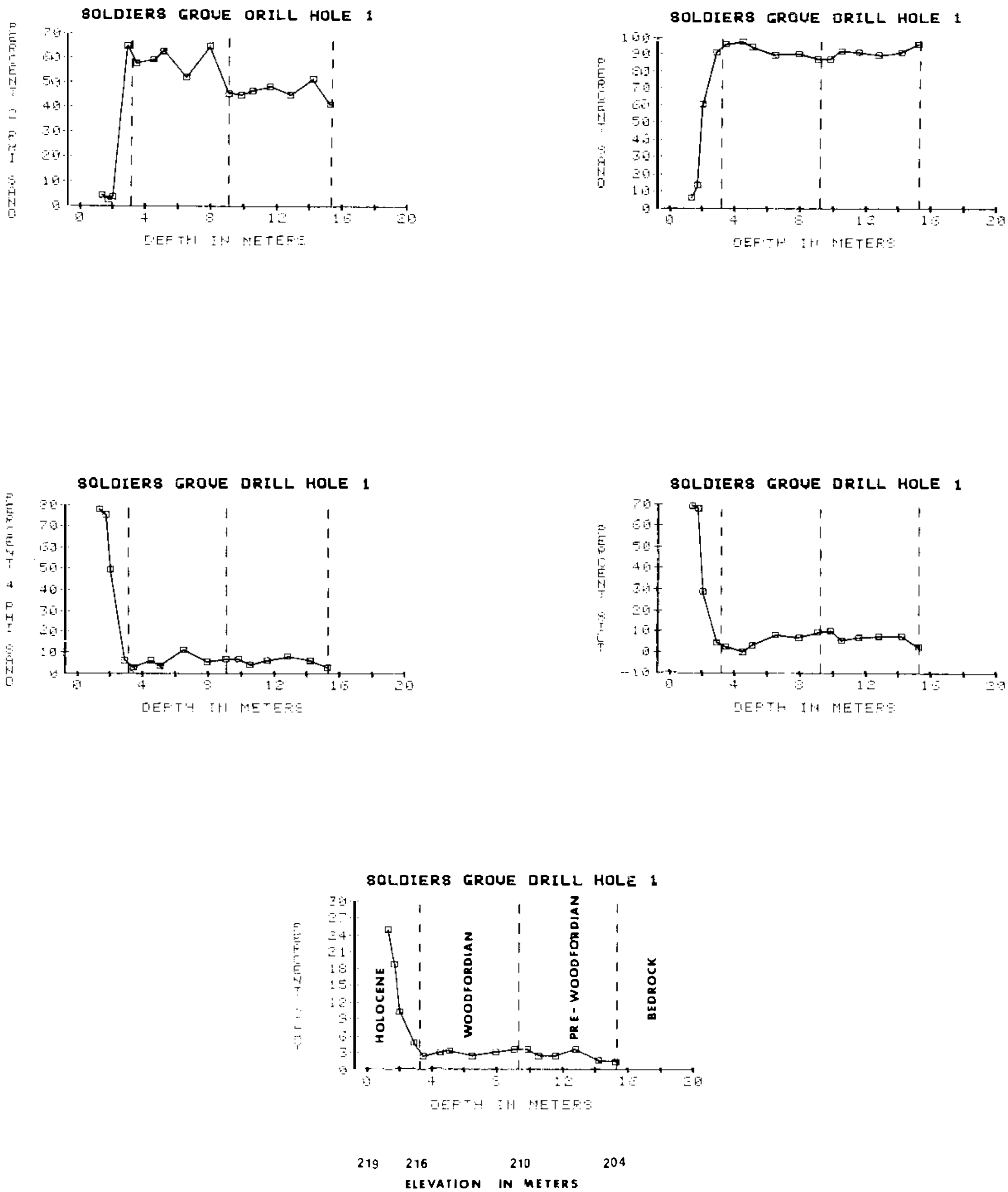

FIGURE 16.--Stratigraphy and sedimentology of drill hole 1 on the Kickapoo floodplain near Soldiers Grove. See figure 14 for location.

coarse texture, but ranges from fine to coarse were observed. The sands were commonly yellowish brown to dull yellowish brown. Sediments were found to be leached to a depth of 6.7 m (22 ft) below which they were slightly calcareous in a horizon of medium to coarse sands with common pebbles. The stratigraphy is unknown below the drill hole depth of 9 m (30 ft). The sediments above 9 m depth are thought to be entirely Woodfordian age because no gravelly zone was observed as at the drill hole on the floodplain near Soldiers Grove. Also, the surface elevation at the drill hole site in Everson's field was estimated to be from 4 to 6 m (15 to 20 ft) higher than the adjacent floodplain. Therefore, given the expected depth of Woodfordian fill, it is unlikely that a drill hole of only 10 m (30 ft) depth would have been sufficient to extend through the Woodfordian sediments.

Continue northerly on Highway 131 to Readstown.

3.5 56.2

Intersection of highways 14, 61, and 131 at Readstown. Turn right (east) and cross Kickapoo River.

0.3 56.5

Intersection. Turn left (north) and follow highway 131 through Readstown. The village is built on the Woodfordian colluvial bench.

Continue northeasterly from Readstown on highway 131.

2.4 58.9

County highway S from the west fork tributary joins on north side of highway. Continue easterly on highway 131.

2.5 61.4

Stop 4 (optional--depending on time). Stop 4 is the Kickapoo Center drill hole site (fig. 17). The drill hole is located on a colluvial bench along the west side of highway 131 near the northeast corner of the cemetery. The surface of the colluvial bench at the site is estimated to be about 220 m (725 ft) above sea level. The bench surface is about 6 m (20 ft) higher than the adjacent floodplain of the Kickapoo River. The bottom of the drill hole at 18.3 m (60 ft) below the surface appears to be in weathered bedrock.

The characteristics of sediment textures observed in the Kickapoo Center drill hole are presented in figure 18. The sedimentary sequence shows three prominent divisions. A rather sharp change in sediment textures occurs at a depth of about 8 m (26 ft). Above this depth the sediments average 80 to 90 percent sand, but below 8 m to a depth of about 14 m (45 ft) the sediments average more than 70 percent silt. The bottom unit is medium and coarse sand with common chert.

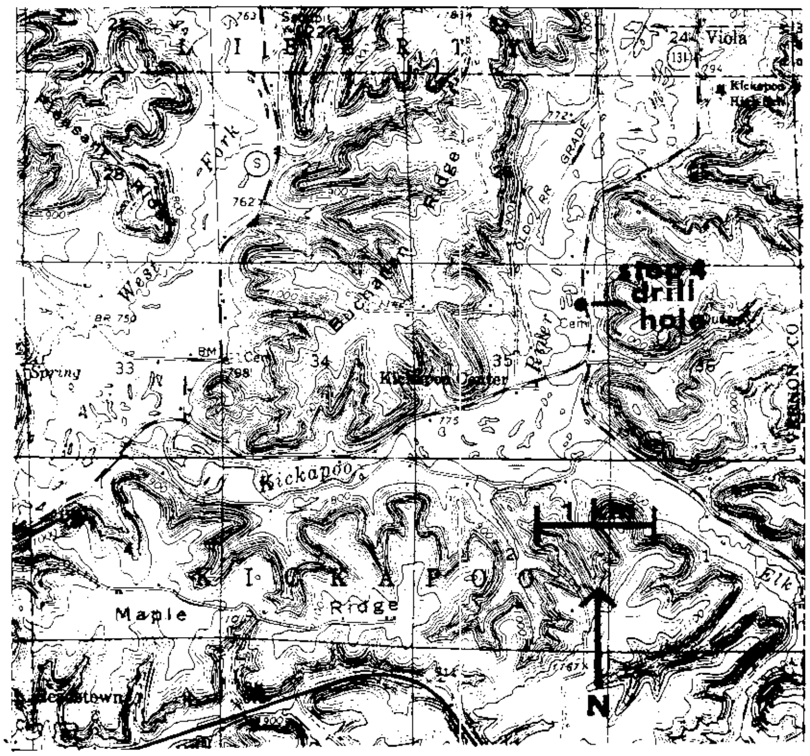

FIGURE 17.--Topographic map of the Kickapoo Center area, Vernon County, Wisconsin.

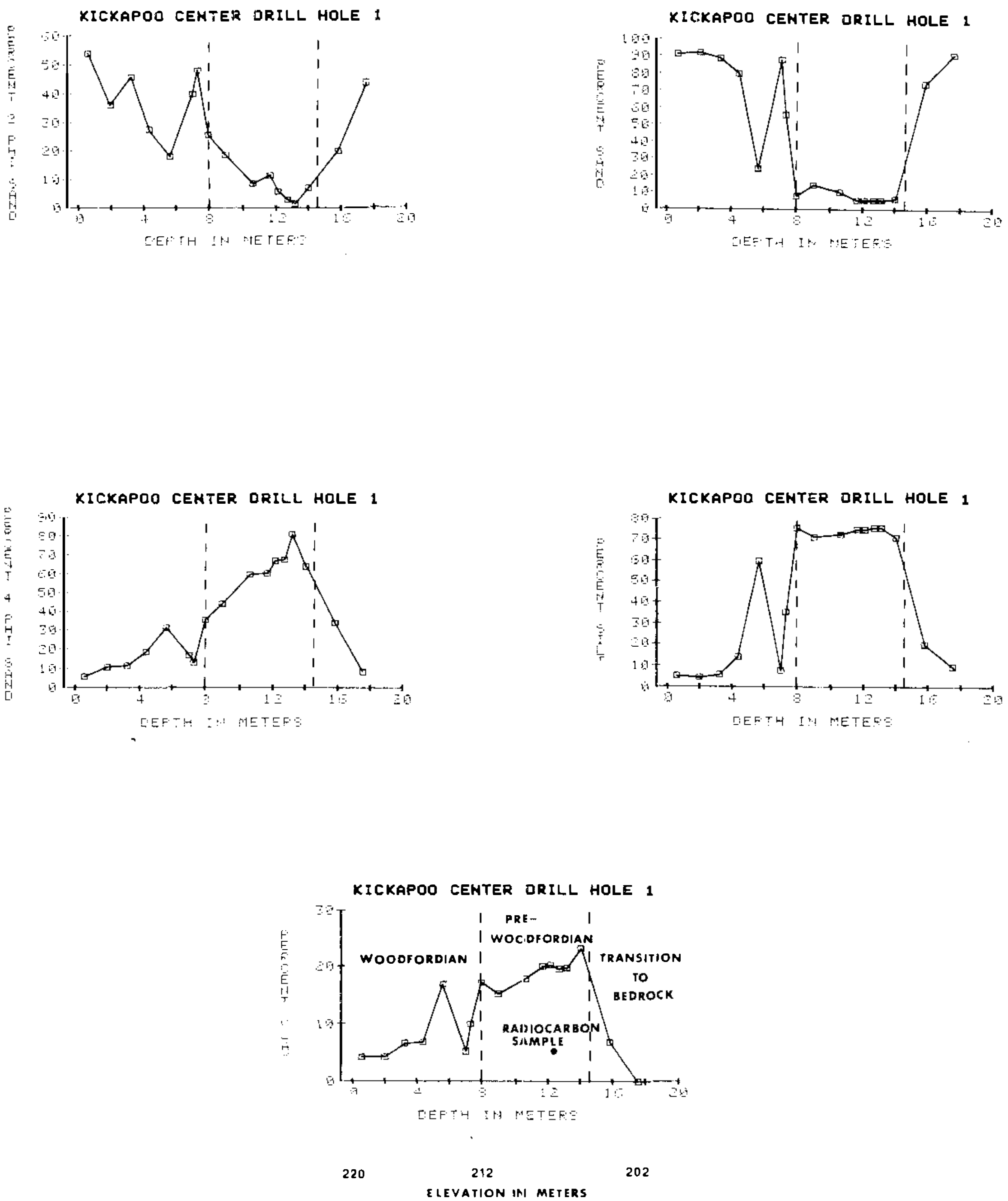

FIGURE 18.--Stratigraphy and sedimentology at drill hole 1, Kickapoo Center. The pre-Woodfordian sediments contain a peaty section at a depth of about 12 m (40 ft). The peat has been submitted for radiocarbon dating. The site location is given on figure 17.

The upper sandy unit is leached and its colors are generally in the range between dull yellowish brown (10YR 5/4) and yellowish brown (2.5Y 5/4). The sand fraction in the upper unit is dominated by medium textures (fig. 18). The silty middle unit is commonly represented by thinly interbedded sandy silts and silty sands that appear to have been derived from slope wash. Compared to the overlying sandy unit, colors in the middle unit are grayer. Except at the top of the unit where the color is brown (7.5YR 4/4), colors are brownish gray (10YR 5/1) to grayish brown (10YR 5/2). The middle unit is slightly calcareous to calcareous. It is least calcareous near its base between depths of 12.2 and 14.9 m (40 and 49 ft) where a silty peat horizon occurs. The silty peat is brownish black (7.5YR 3/2). A sample of the peat has been submitted for radiocarbon dating.

The basal unit at the Kickapoo cemetery drill site has stronger yellowish hues than overlying units. Its colors vary from yellowish brown (2.5Y 5/4). To bright yellowish brown (2.5Y 6/6). The unit is leached throughout. Sediment textures include abundant medium and coarse sand with chert pebbles. Silt and clay comprise less than about 25 percent of the total sediments (fig. 18).

The base of the Woodfordian is tentatively placed at a depth of about 12 m (40 ft) where a weakly developed soil and leached zone occurs. Note on figure 18 that percentages of clay are highest between 12 and 14 m depth, but decline steadily above 12 m depth. Although it is tempting to assign the base of the Woodfordian to the major stratigraphic break at about 8 m (26 ft) depth, the percentages of silt in the middle unit between 8 and 12 m (26 and 40 ft) are typical of those associated with Woodfordian loess in the region. Note also on figure 18 that the percentages of medium and fine sand, as represented by the 2 and 4 phi sieve fractions, show relatively progressive changes from the depth of about 12 to 14 m to the surface. The 12 m thickness of Woodfordian sediments at the Kickapoo Center drill site is greater than amounts recognized at drill sites discussed earlier. The larger quantity at the Kickapoo Center site is attributed to its close proximity to the base of the hillslop. The sediments at the Kickapoo Center site appear to be mostly derived from local slope wash rather than from having been transported to the site by river processes. Bates (1939, p. 872) also concluded that aggradation of the Kickapoo Valley was accomplished primarily by wash from the valley sides.

Continue northerly on highway 131.

1.9 63.3

Enter village of Viola. Continue on highway 131 to village park for lunch stop.

0.6 63.9

Cross Kickapoo River. Lunch Stop in park to right.

Continue northerly on highway 131.

0.1 64.0

Intersection of highways 56 and 131. Turn left (northeast) and continue northerly on highway 131.

3.8 67.8

Tunnelville roadcut. Observe on figure 19 that the bedrock meander core that extends northwestward from the Tunnelville roadcut was not cut off as occurred downstream at other sites. The elevation of the present floodplain surface is about 238 m (780 ft) above sea level. Since the maximum level of the Bridgeport outwash at the mouth of the Kickapoo River near Wauzeka is about 225 m (740 ft) above sea level, aggradation related to that event would have been insufficient to overtop and cut off the bedrock valley meander at Tunnelville. The Tunnelville meander core represents a landform that was once common throughout the lower reaches of the Kickapoo River.

Continue northerly on highway 131 to the village of LaFarge. New roadcuts that were open during the summer of 1981 along the base of the southeast valley wall between Tunnelville and LaFarge showed common thin beds of silt interbedded with thin beds of sand. The sediments appeared to be derived from slopewash.

2.1 69.9

Enter village of LaFarge.

0.2 70.1

Ascend low terrace. The terrace is similar in elevation to early Holocene (10,000 to 7000 B.P.) terraces in the upper Kickapoo drainage, but it may also be correlative to the low Citron Valley terrace, which would place it between 9500 and 12,000 B.P.

0.1 70.2

Intersection of highways 82 and 131. Turn left (west) and continue on highway 131.

0.3 70.5

Intersection of highways 82 and 131. Turn right (north) and continue on highway 131.

0.8 71.3

The rock benches between elevations of about 275 to 285 m (900 to 935 ft) elevation above sea level contain river gravel deposits. An opportunity to see these gravels will be presented at stop 5. The location of highway 131 north of LaFarge has been moved since the topographic map in figure 19 was printed. The new highway passes along the intermediate elevation rock benches and hillslopes on the east side of the Kickapoo River.

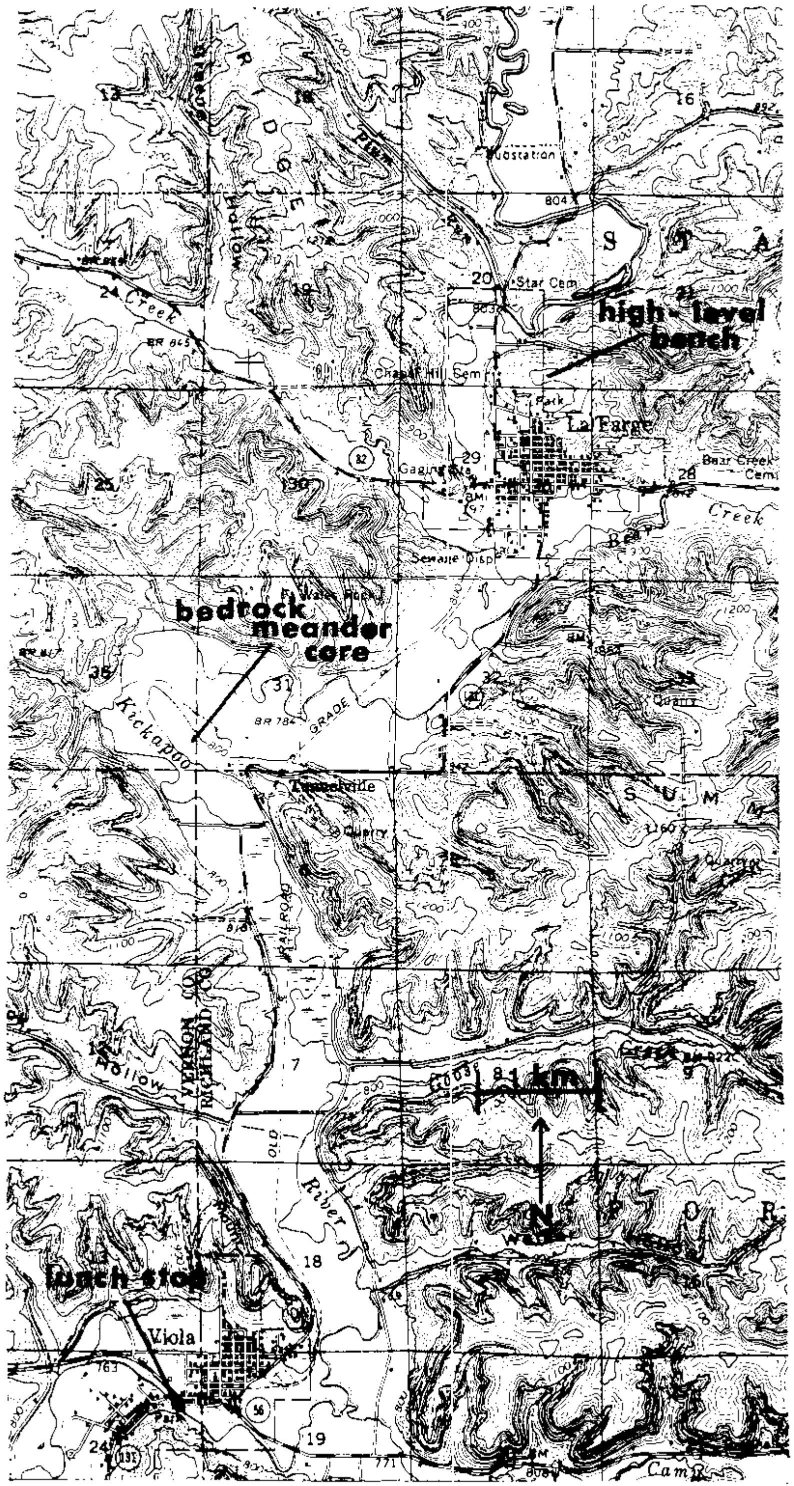

FIGURE 19.--Topographic map of the middle Kickapoo valley, Vernon County, Wisconsin. Scars of relict valley meanders generally are not apparent north of Viola. The bedrock meander core near Tunnelville probably represents a landform condition common to the lower Kickapoo valley before aggradation over-topped the low sags and subsequently resulted in cut-offs when trenching was renewed. The initial phase of aggradation is correlated with base level rise at the mouth of the Kickapoo when the Wisconsin River valley was aggraded with Bridgeport outwash sediments during classical Kansan time.

0.6 71.9

Note the partially completed dam on the Kickapoo River on your left (west). The dam and related facilities were nearly completed in the early 1970s when the project was challenged on economic and environmental grounds. One of the environmental issues related to rare plants growing on sandstone bluffs along the river. These plants have been viewed as evolving from unique conditions of a Driftless Area refugium while the rest of the region experienced glaciation. Many of the plants would have been destroyed by the ponded waters above the dam. The most serious environmental issue related to the expected quality of water in the future lake. Because of the intense agricultural land use in the watershed stream runoff was expected to have transported large quantities of sediment and nutrients into the lake.

Continue northerly on highway 131. The pine plantations on the hillslopes were planted by the U.S. Army Corps of Engineers in the 1970s as part of the LaFarge dam project.

4.0 75.9

Highway makes an abrupt left turn (to west) to cross Kickapoo River. Note graffiti on railing of south side of bridge.

0.3 76.2

Village of Rockton. Continue north on highway 131 to village of Ontario. Note the prominent intermediate elevation rock bench that averages 60 to 90 m (200 to 300 ft) lower than the surrounding upland ridge crests. The rock bench contains relict stream gravels on its surface at many sites.

5.2 81.4

Note to your right (southeast) the prominent low terrace containing an abandoned barn on its surface. The terrace is interpreted as early Holocene on the basis of similarities with a radiocarbon dated terrace a few kilometers upstream.

1.5 82.9

Enter village of Ontario (fig. 20).

0.5 83.4

Intersection of highways 33 and 131. Follow 131 northerly through Ontario.

0.6 84.0

Highway 131 crossing Kickapoo River near north edge of Ontario. Continue northerly on 131.

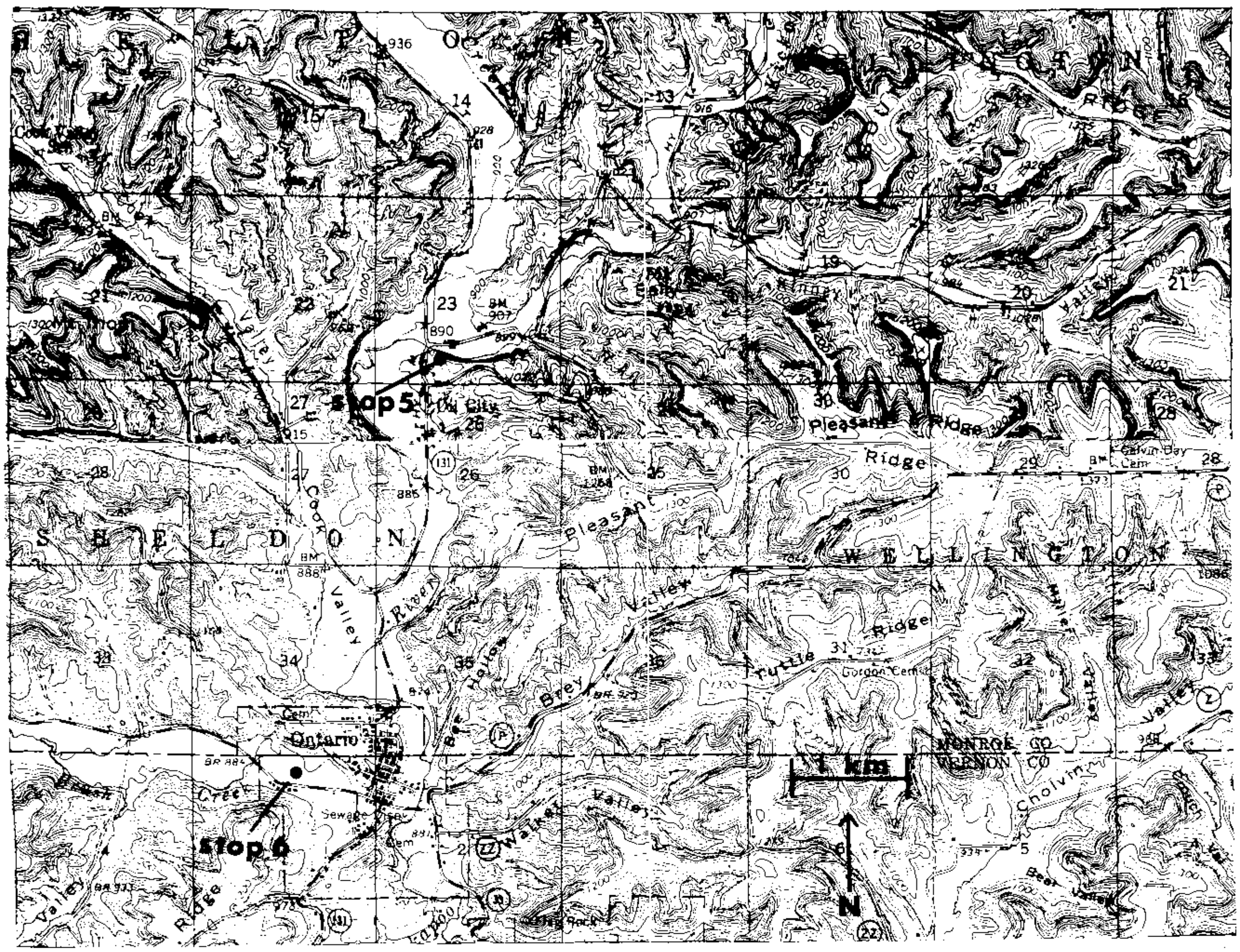

FIGURE 20.--Topographic map of the upper Kickapoo valley in the Ontario and Oil City region, Vernon and Monroe Counties, Wisconsin. Note the prominent high-level benches. Some of the high-level benches contain relict river gravels such as at the Oil City roadcut (fig. 21).

1.9 85.9

Stop 5. Oil City roadcut (fig. 20).

This section is extremely dangerous because the critical stratigraphy is located at the top of the cut high above the road. Be careful not to lose your balance while viewing the section. Before inspecting the exposure on the east side of the highway, please climb up the backslope of the surface on the west side of the highway. The vantage point provides a convenient position for discussing and photographing the entire sequence of Pleistocene deposits at this site.

Stratigraphic interpretations that have been applied to the exposures along the east side of the highway are shown in figures 21 and 22. An overview which shows the relative positions of the sections in figures 21 and 22 is given in figure 23. Note that the section at position A includes two loesses overlying relict stream gravels. The loess units are separated by a paleosol. The section at B represents a gully fill. Note that the gravels have been truncated higher upslope and have been transported to form a stone line over the gully fill. Note also that the stone line occurs above the paleosol.

The age of the relict river gravels is unknown. They may be of early Pleistocene age because the rock bench on which they rest appears to grade into the Bridgeport strath in the Wisconsin River valley (fig. 7) The profile elevations were estimated from contours on 1:62,500 scale topographic maps with 20ft (6 m) contour intervals. North of LaFarge the elevations of the bench were easily determined because the surface is extensive. South of LaFarge the task was more difficult because the rock benches were either absent or less prominent. The profile south of Soldiers Grove represents elevations on the floors of high-level meander scars such as the one immediately west of Soldiers Grove near the center of section 25 (fig. 14). If Trowbridge (1954) was correct in his interpretation that the surface of the Bridgeport strath was the level of valley incision at the time northeast Iowa was invaded by classical Kansan ice, and if the profile of figure 7 is correct, then the gravels at the Oil City section may be of early Pleistocene age.

Textures of sediments overlying the gravels at section A in the Oil City roadcut are shown in figure 24. The sediments above line A on figure 24 represent mostly Woodfordian loess. A thin horizon of Altonian loess might be represented just above the buried soil, but the increased density at that horizon might also be due to illuviation of clays (fig. 24). An older loess is represented under the line A on figure 24. It has a well-developed soil in its surface horizon. The upper several centimeters, in the soil horizon, has many vertical cracks that are filled with fine sand. It is the sand in the cracks that accounts for the anomalous bulge on the sand curve in figure 24. The cracks may represent desiccation at some time in the past, but it is more likely that they represent cracking of the ground under conditions of extreme cold. The geomorphic effects of periglacial climates during Woodfordian time have been discussed

FIGURE 21.--Stratigraphic sequence at the Oil City roadcut, stop 5. See Figure 20 for location. The river gravels appear to rest on a surface that represents a relict river profile that is graded to the Bridgeport strath (fig. 7).

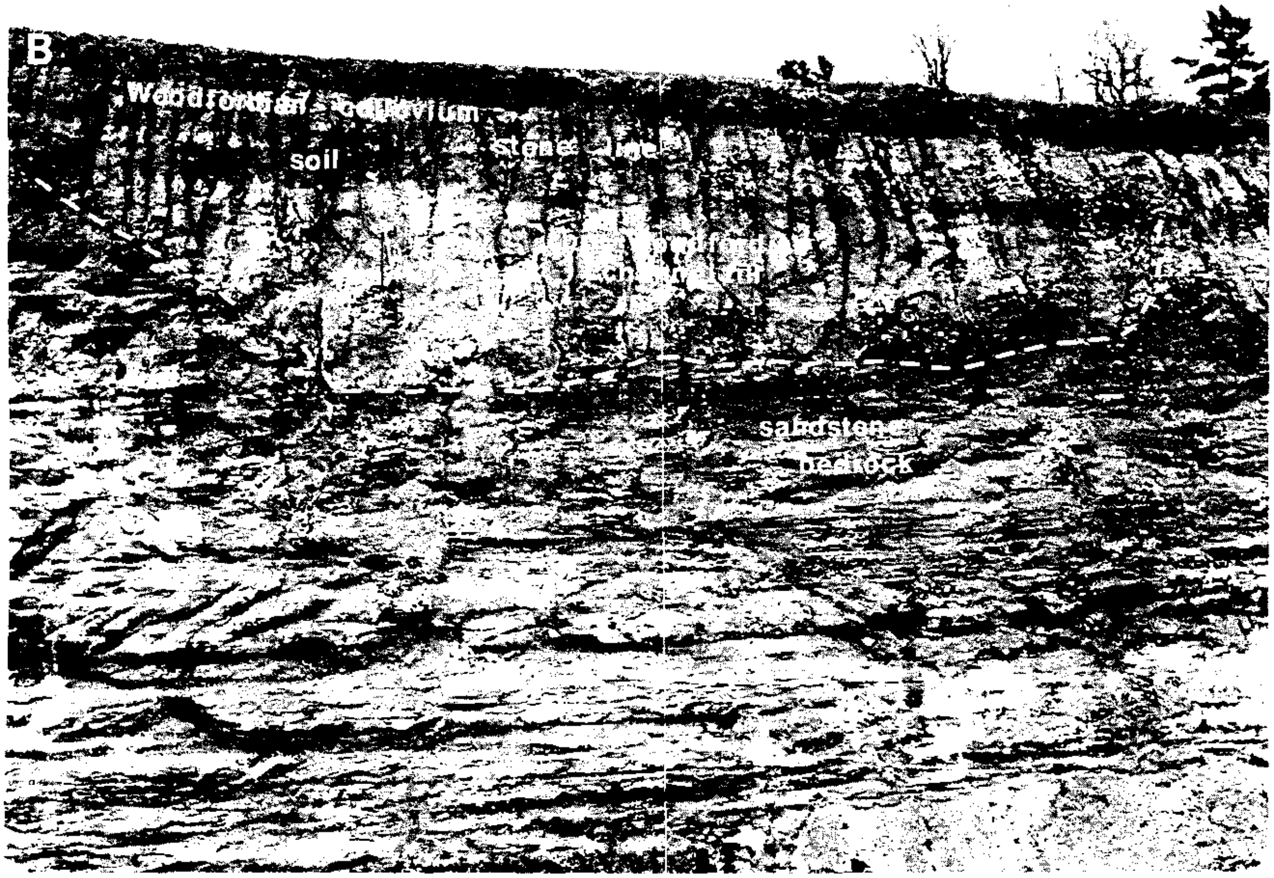

FIGURE 22.--Relict filled channel at the Oil City roadcut. See figure 20 for location. Note that the channel fill is capped by a paleosol. The paleosol is in turn overlain by a stone line and Woodfordian colluvium that represent the destabilization of hillslopes during cool moist climates of the late Wisconsinan.

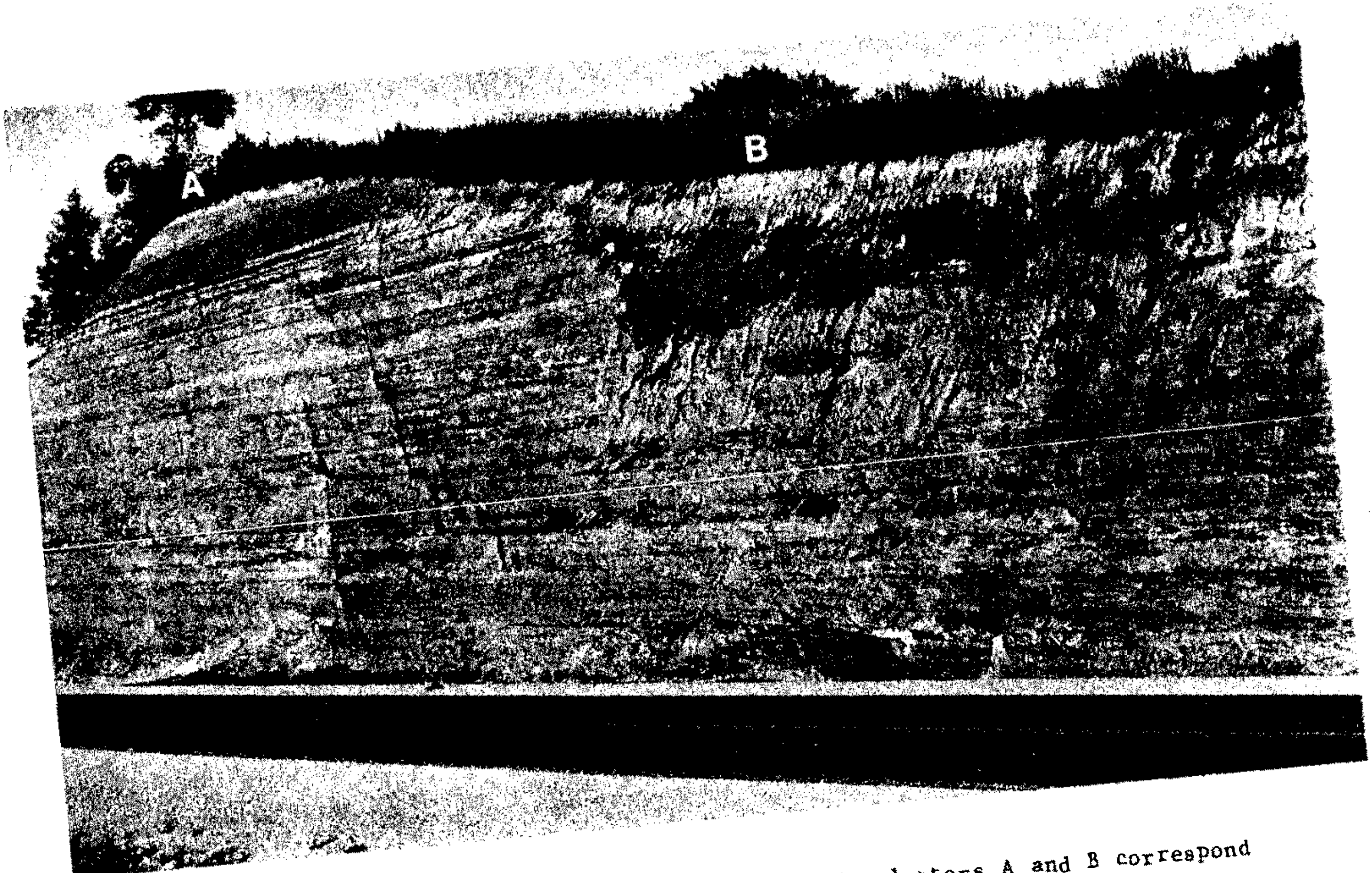

FIGURE 23.--Overview of the Oil City roadcut. The letters A and B correspond with the stratigraphic sections shown in figures 21 and 22.

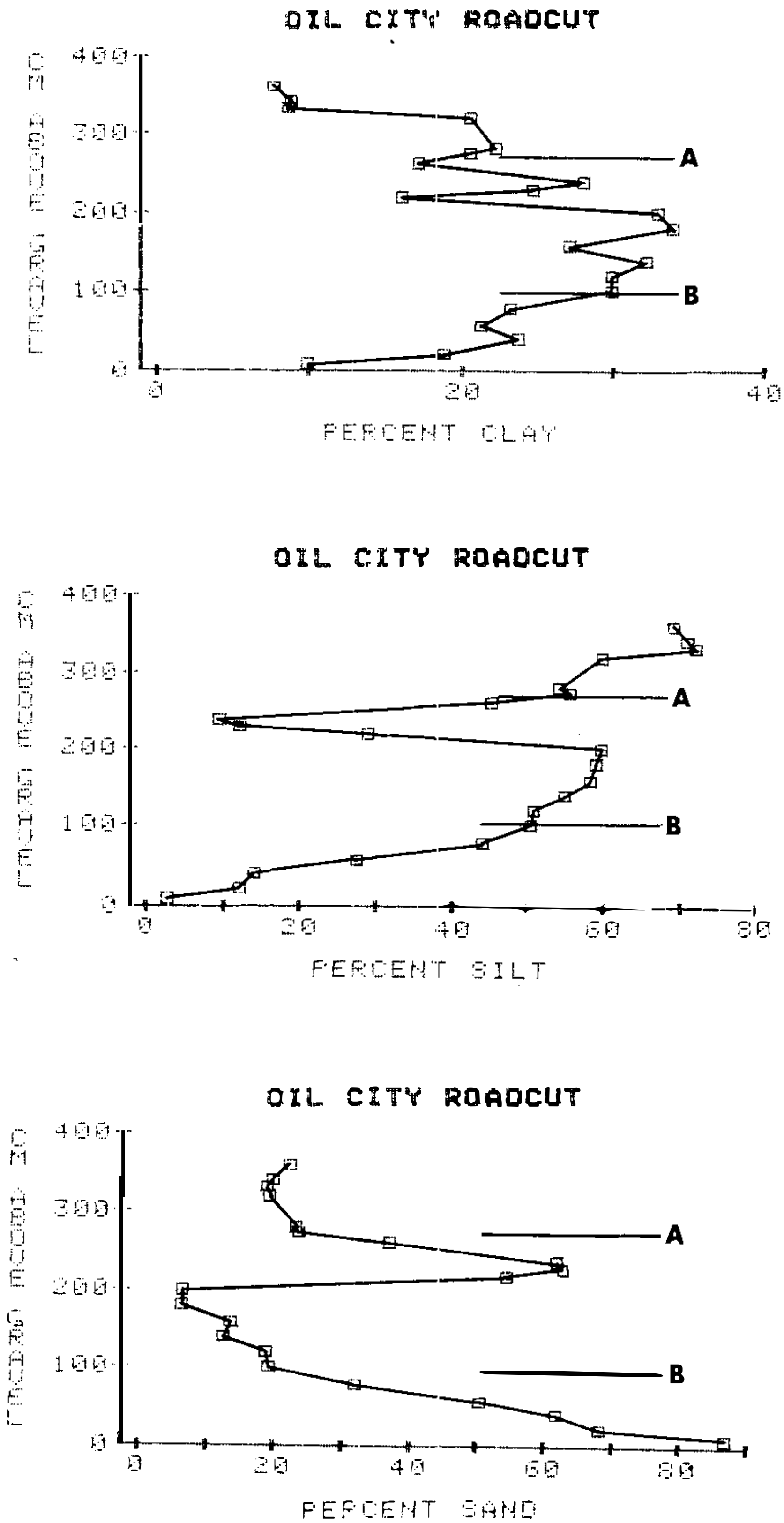

FIGURE 24.--Textural properties of sediments at the Oil City roadcut. For location see figure 20. The sediments above line A are Woodfordian loess. A paleosol occurs immediately below line A. Vertical cracks in the paleosol are filled with fine sand and account for the sand bulge at that horizon. The sediments between lines A and B represent a pre-Woodfordian loess. Sediments below line B represent point bar sands of a river that formerly flowed at the elevation of the upland bench.

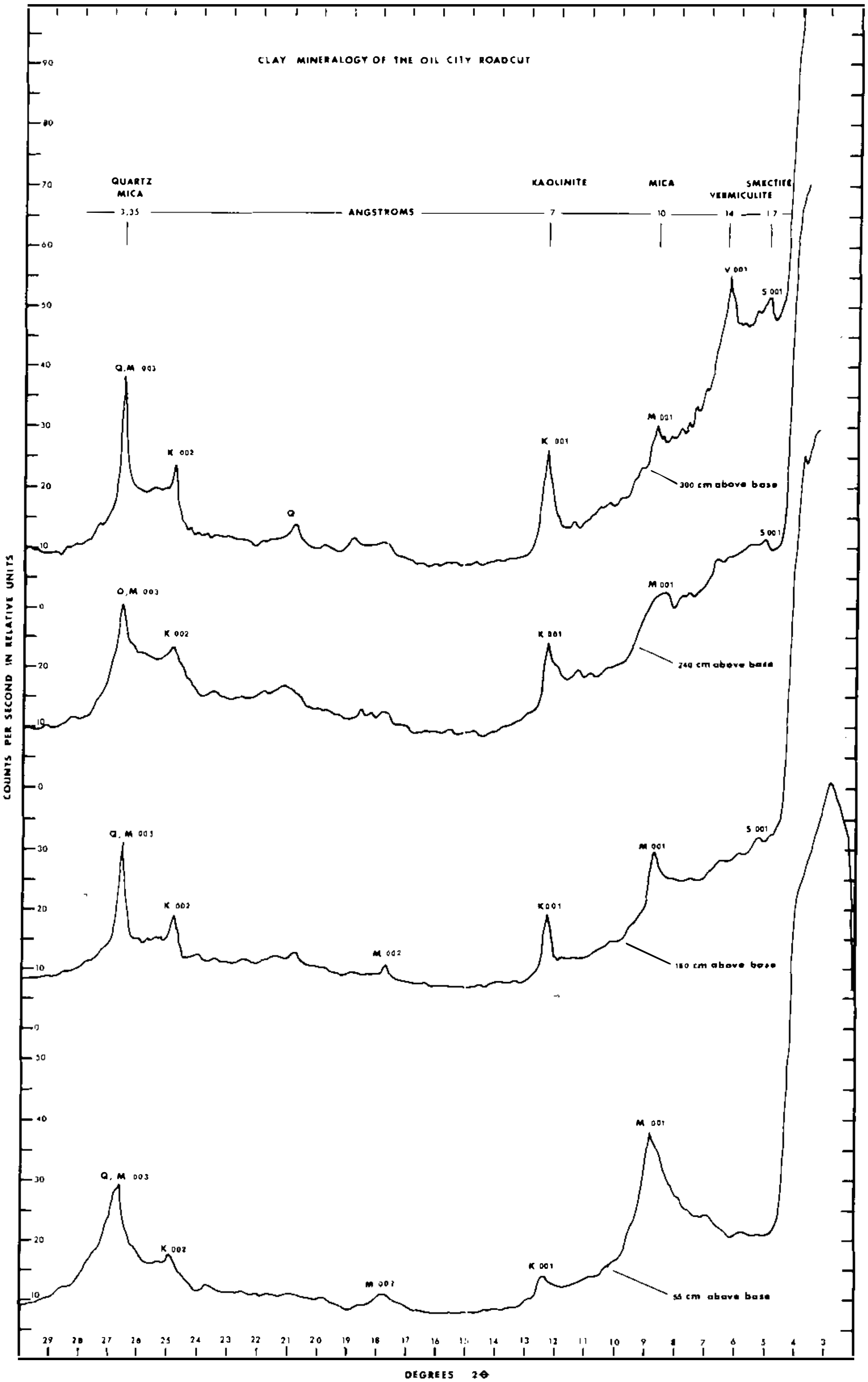

FIGURE 25.--Clay mineralogy of sediments from the major stratigraphic units at the Oil City roadcut (figs. 20 and 21). The positions of the samples are measured in centimeters above the surface of the river gravels. The 55-cm sample represents the Point Bar sands; the 160-cm sample represents the lower portion of the pre-Woodfordian loess; the 240-cm sample represents the paleosol horizon; and the 300-cm sample represents the basal component of the Woodfordian loess.

earlier in this log. Severe conditions leading to hillslope erosion during the early Woodfordian is indicated at the Oil City section by the beveling of the fluvial gravels and their subsequent downslope movement to bury the paleosol overlying the channel fill (figs. 22 and 23). The age of the loess with the well-developed soil is also unknown. It may be of early Wisconsinan age, but the paleosol at its surface seems too well-developed for that assignment. Possibly, it is of Illinoian age.

Clay mineralogy was also determined by x-ray analysis on several samples from section A at the Oil City roadcut. Four of the x-ray defraction curves representing the principal units are presented in figure 25. Clay samples were withdrawn at 5 cm depth in a 1000 ml cylinder after shaking and four hours suspension. Sodium hexametaphosphate was used as a dispersing agent. The extracted sample was dripped onto a heated slide to minimize the effects of differential settling of clay particles. The samples were saturated with ethylene glycol by placing them over a glycol solution in a closed container. The container and samples were heated to 60°C overnight and then x-rayed. The sample slides were subsequently heated to 550°C for two hours and then x-rayed again to aid in the identification of certain minerals.

The x-ray analyses indicate differences in degree of weathering and clay mineralogy between the principal stratigraphic units. The Woodfordian loess has strong vermiculite and kaolinite peaks, but the mica and smectite peaks are relatively subdued. The vermiculite identification was based on its position at 14 angstroms for the glycolated analysis and its collapse to 10 angstoms after heating to 550°C for two hours. The defraction curve representing 240 cm above the fluvial gravels is from the upper part of the paleosol (fig. 24). Note that the pattern shows a broad rise rather than well defined peaks in the range from 10 to 17 angstroms. This may be due to randomly interstratified mica-smectite, suggesting pedogenic alteration of mica to smectite. This broad rise collapses to 10 angstroms after heating to 550°C for two hours. At greater depth, the defraction curve representing sediments from 160 cm above the gravel surface shows that the 10 angstrom mica peak is sharper, indicating that the mica has not been as severely altered to smectite as above. The bottom defraction curve represents the clay fraction of a silty sand unit that may be point-bar sediments associated with the underlying river gravels. A very strong mica peak is characteristic of these sediments and the strength of the peak suggests little mica alteration by weathering.

Silt in the Woodfordian loess is slightly coarser than silt in the loess below the surface of the paleosol. The percentage of Woodfordian silt that fell between 0.06 and 0.03 mm diameter averaged 24.2 percent with a standard deviation of 4.0 percent. The percentage of underlying pre-Woodfordian silt that fell between 0.06 and 0.03 mm averaged 20.8 percent with a standard deviation of 2.9 percent. The finer texture of the pre-Woodfordian loess might imply that it was transported from a location farther away than the source of the overlying Woodfordian loess. Alternatively, the differences in texture might be due to alteration by weathering. The textures of silt in both loesses do not compare closely with the texture of silt in the underlying point bar where the percentage of the silt that was between 0.06 and 0.03 mm diameter averaged 54.7 percent with a standard deviation of 16.8 percent.

No analyses have yet been done to determine if the silt in the channel fill (fig. 22) is derived from the pre-Woodfordian loess at the top of the slope.

The deposits at the Oil City roadcut are at great odds with the suggestions by Black (1970) that the Driftless Area has been glaciated. The presence of a well-developed paleosol separating Woodfordian and pre-Woodfordian loess deposits especially challenges Black's conclusion that the region may have been glaciated as recently as about 30,000 B.P. The river gravels on the narrow Oil City interfluve probably have been there since at least classical Kansan time.

Turn vehicles around and return to village of Ontario via highway 131.

1.9 87.8

Kickapoo River bridge near north edge of village of Ontario. Continue through Ontario to the intersection with highway 33 on the south side of the village (fig. 20).

0.6 88.4

Intersection with highway 33. Turn right (west) onto highway 33. Continue westward into Brush Creek tributary.

0.2 88.6

Cross Brush Creek tributary.

0.3 88.9

Stop 6. Powell site.

The upper Kickapoo drainage system has been the focus of intensive Holocene paleohydrologic studies during the past few years (Knox, McDowell, and Johnson, 1981; McDowell, this volume). Brush Creek tributary represents an area of the drainage system where studies were especially concentrated.

The fluvial incision on the Mississippi and Wisconsin Rivers caused by glacial meltwaters apparently worked its way headward to the upper reaches of the Kickapoo system and is the first major event in Holocene paleohydrology. The interpretation is based on the observation that Woodfordian colluvial benches stand much higher than Holocene alluvial terraces that date as early as 9000 B.P. Since that time, climate appears to be the dominant factor controlling fluvial episodes.

Climate conditions at the beginning of the Holocene about 10,000 B.P. were relatively cool and moist but were rapidly becoming warmer and drier (Bartlein and Webb, this volume). The relatively fine textures of early Holocene sediments indicate that large floods were relatively rare during this period, probably because moist air masses were rarely able to penetrate the region prior to 7000 to 6000 B.P. (Knox, 1982). Early Holocene alluvium usually is represented by less than 1 m thickness of channel lag gravels that are overlain by about 2 to 3 m (6.5 to 10 ft) of sandy silt sediments. The sandy silts usually have a yellowish brown (10YR 5/4) appearance because they have been oxidized. Maximum warmth and dryness occurred about 7200 B.P., when prairie expanded into woodlands of the area. Between about 7500 and 6000 B.P. convectional thunderstorms probably were the dominant cause of floods, because erosion and deposition of alluvial sediments were concentrated in small watersheds and on alluvial fans. Since about 6000 B.P., climate has generally followed a cooling trend, and sedimentary sequences indicate episodes when floods were much larger than those of the earlier Holocene. These floods contributed to intensive lateral channel migration and slight channel incision, especially during 6000 to 4400, 3100 to 1800, and 1200 to 800 B.P. Although channel conditions at other times were apparently more stable, some lateral migration occurred continuously through the Holocene. The tendency for episodic variations in the intensity of fluvial activity on the valley floor has resulted in a sequence of Holocene surfaces each having relatively unique sedimentological and pedological properties (McDowell, this volume). The fine textured early Holocene deposits represent the highest alluvial surface. They tend to have well-developed soil profiles with B-horizon development. Holocene deposits that are younger than about 6000 B.P. tend to include very sandy point bars with many cobbles and small boulders. Soils on deposits younger than 6000 B.P. tend to have A/C profiles. The poorly developed profiles on the late Holocene alluvium reflects the poorer drainage of these lower elevation sites and the shorter period of time that they have been exposed to weathering processes.

The relatively gradual changes in Holocene vegetation implied in pollen diagrams of the region (Maher, this volume) correspond poorly with the relatively abrupt changes apparent in the alluvial record. The abruptness of change in fluvial activity appears to be better related to long-term variations in recurrence intervals of large floods (fig. 26). The magnitudes of high frequency floods of the Holocene were estimated from the bankfull dimensions of Holocene paleochannels. Floods having a recurrence interval of about 1.58 years (annual duration series records) correlate closely with the bankfull stage in alluvial channels. Magnitudes of contemporary floods of 1.58-year recurrence frequency were used as a baseline to assess relative magnitudes of 1.58-year floods during the past Holocene time. Magnitudes of floods in Holocene channels were calculated from the application of empirical equations that describe the functional relationship between contemporary bankfull channel capacities and 1.58-year flood magnitudes. Holocene paleochannel dimensions were reconstructed by cross-profile drilling at intervals of a few meters. Channels were dated either directly by radiocarbon

FIGURE 26.--The sizes of 1.58-year (high frequency) floods determined for Holocene paleo river channels at sites in the upper Kickapoo watershed indicate departures ranging from 40 % smaller to slightly larger than contemporary floods of the same recurrence frequency. Small floods were especially characteristic between about 8000 and 6000 B.P. and 4000 and 3000 B.P. Large floods were especially characteristic between 6000 and 5000 B.P., 3000 and 2000 B.P., and for a brief period after 1200 B.P.

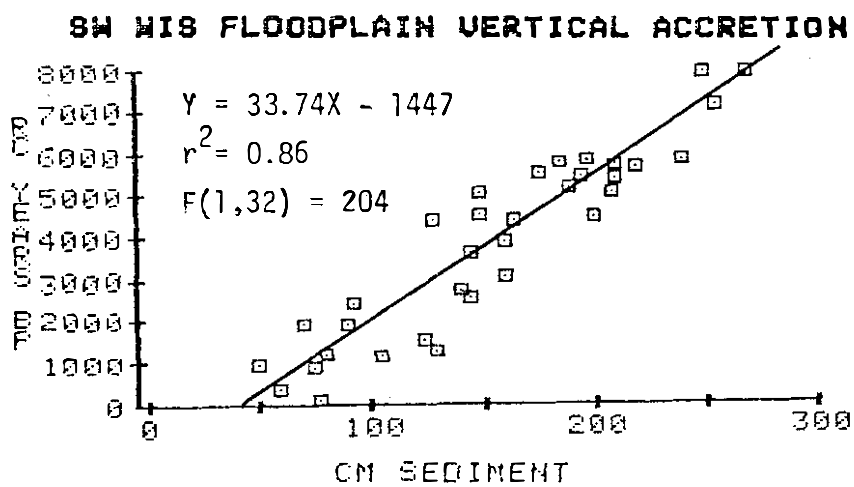

FIGURE 27.--Relationship between radiocarbon age and overlying pre-agricultural sediment thickness for sites in the Driftless Area of southwestern Wisconsin. The relationship indicates that the long-term vertical accretion rate is about 1 cm per 34 years, but clustering of the data points indicate that accumulations probably occurred episodically.

analysis or indirectly by estimating the relative age from a graph representing thickness of sediments in the channel margins versus radiocarbon age (fig. 27). Tentative results shown in figure 26 indicate that Holocene magnitudes of 1.58-year-frequency floods have varied from nearly 40 percent smaller to perhaps 55 percent larger than magnitudes of contemporary floods. Large floods were characteristic between about 6000 to 5000 B.P., 3000 to 2000 B.P., and during a brief interval after 1200 B.P. Small floods apparently were dominant between about 8000 and 6500 B.P., 4000 and 3000 B.P., and about 2000 and 1500 B.P. Tentative calculations suggest that the Holocene magnitudes of high frequency floods probably ranged from about 10 percent smaller to perhaps more than 100 percent larger than magnitudes of 1.58-year floods prior to settlement in the region. It is concluded that climatic influence on magnitudes and frequencies of floods was the most important cause of the episodic intensity of fluvial activity. Although the radiocarbon age versus sediment thickness curve of figure 27 suggests progressive aggradation, closer inspection shows that the dates are clustered and that the breaks between clusters tend to correspond with the breaks between fluvial episodes identified above. Figure 27 might be better characterized as a step-function sequence.

Continue west on highway 33.

4.2 93.1

Alternate stop 6 (Helmuth site, fig. 28).

The Helmuth site shows the contact relationship between early and late Holocene alluvium. The sedimentology of this site is described in Knox, McDowell, and Johnson (1981).

Continue west on highway 33 to village of Cashton.

6.4 99.5

Intersection of highways 27 and 33 in village of Cashton. Turn left (south) onto highway 27 and return to Prairie du Chien.

End of road log for day 1.

DAY TWO

Trip participants are requested to read the paper "Pre-Wisconsinan Deposits in the Bridgeport Terrace of the Lower Wisconsin River Valley" by Knox, Attig, and Johnson, in this volume. Much of the evidence for reconstructing the Pleistocene history of the lower Wisconsin River valley is presented in this paper. Hereafter, for brevity, reference to information in the Knox, Attig, and Johnson paper will be given as "Bridgeport paper".

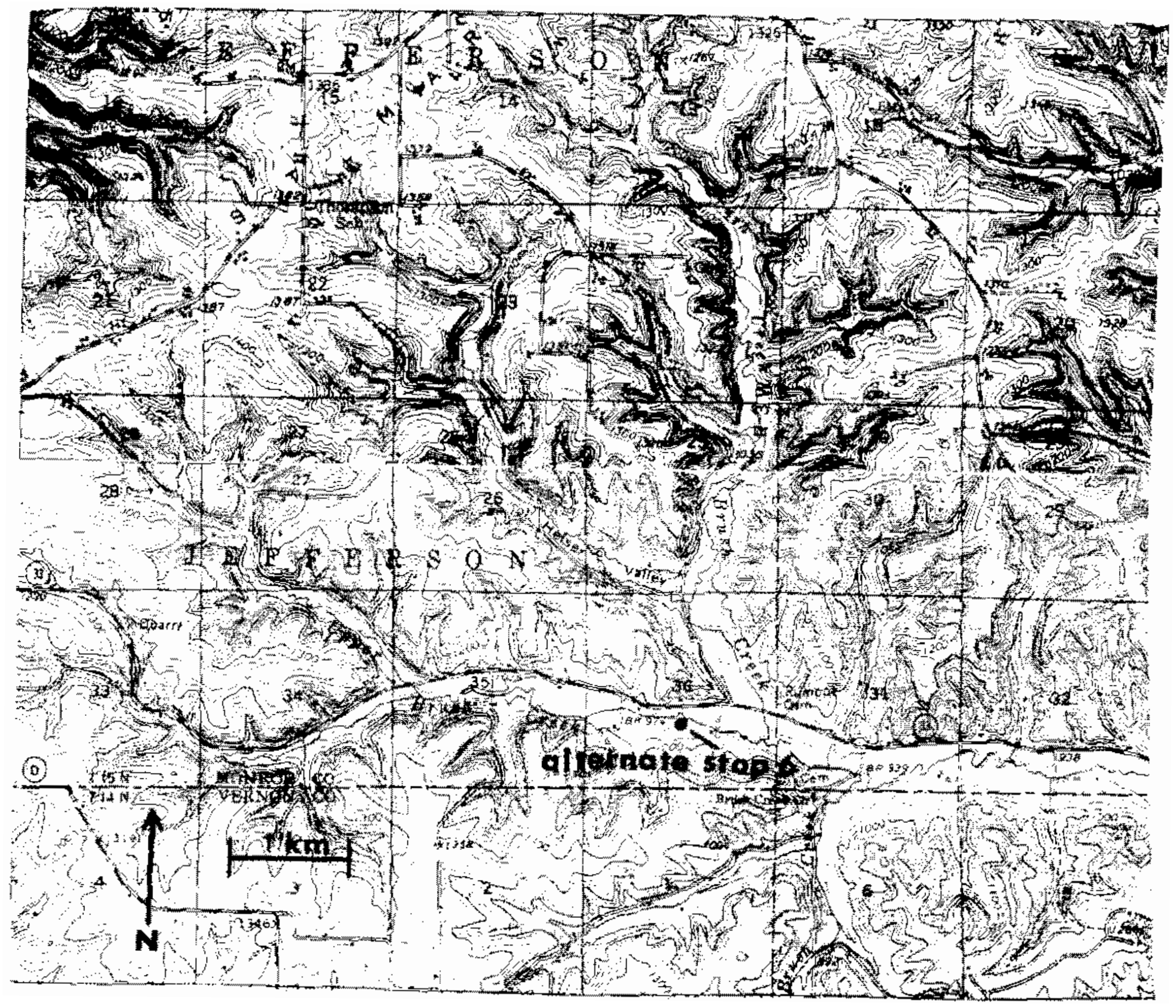

FIGURE 28.--Topographic map of Middle Brush Creek drainage basin, Vernon and Monroe Counties, Wisconsin.

MILEAGE

0.0 0.0

Start point (8:30 a.m., May 23, 1982) intersection of highways 18, 27, 35, & 60, northeast side of Prairie du Chien.

Proceed south on highway 18, 35, and 60.

1.6 1.6

Pass Hidden Valley Lodge (Friends of Pleistocene headquarters for conference) on right (west). Continue southeasterly on the high Wisconsinan terrace. The maximum elevation on the Wisconsinan terrace is about 195 m (660 ft) above sea level. It may be recalled from yesterday's road log that at least 50 m (150 ft) of alluvial fill underlies the high Wisconsinan surface in this area.

1.8 3.4

The highway begins to ascend the Bridgeport terrace. Note the outcrops of Prairie du Chien dolomite under the Pleistocene deposits on the Bridgeport terrace. The bedrock surface underlying the Pleistocene deposits on the Bridgeport terrace is recognized as a strath. Trowbridge (1954) suggested that the Bridgeport strath "...seems to represent the depth to which the Mississippi and Wisconsin Rivers had cut their valleys by the time the Kansan glacier reached this position." A driller's log for a well on the high Wisconsinan terrace and adjacent to the Bridgeport terrace indicates 27 m (90 ft) of alluvial fill. The close proximity of the well to the bedrock of the Bridgeport strath indicates that the Mississippi valley descends abruptly at the contact with the Bridgeport strath.

0.2 3.6

Junction of old highway 18 on right (south) with highway 18, 35, and 60. Turn right to southeast on old highway 18.

0.9 4.5

Stop 7. White Site.

The location of the White site is indicated by the letter X near the eastern edge of section 9, Bridgeport Township, shown on figure 2 of the Bridgeport paper. The drill hole at the White site is about 3 km (1.5 mi) upstream from the mouth of the Wisconsin River. Results from drilling (see appendix I, Bridgeport paper) indicate that a sandy till of at least 6.4 m (21 ft) thickness occurs at the White Site. The till is very sandy as might be anticipated at a site where ice apparently overrode outwash sediments in the Mississippi valley. Sand:silt:Clay averages are 50:30:20 for the upper part of the till sequence and 51:24:25 for the lower part of the till sequence. Quartz grains in the sand fraction are predominantly angular. Their angularity stands in sharp contrast to outwash sediments observed elsewhere in the lower Wisconsin River valley (appendix I, Bridgeport paper).

The hypothesis that Bridgeport till at the White Site was deposited by an ice lobe that advanced from the west in northeast Iowa was tested by comparing clay mineralogy of Iowa samples with Bridgeport samples. To maintain comparability of laboratory procedures, five samples from the Bridgeport till were sent to the Iowa Geological Survey for processing. Semi-quantitative calculations showed that the clay mineralogy of the five samples averaged 67% expandables, 16% illite, and 17% kaolinite plus chlorite (table 1, Bridgeport paper). These results agree closely with average values of clay minerals in the Iowa Wolf Creek Formation where the mineralogy is 62% expandables, 17% illite, and 21 kaolinite plus chlorite (Hallberg, 1980, p. 15). My own analyses of clay mineralogy indicated significantly higher percentages of mica (illite) than suggested by the Iowa analyses. Differences appear to be related to laboratory procedures. I used a very rapid drying procedure in slide preparation by comparison to the Iowa procedure (Hallberg, 1978).

The Wolf Creek Formation of Iowa corresponds with classical Kansan time. It is interesting that the clay mineralogy of the Wolf Creek Formation is broadly similar to the clay mineralogy of the Hersey till in Wisconsin on the northwestern margin of the Driftless Area (Baker and Simpson, 1981). A listing of average data is given in table 1 of the Bridgeport paper. Baker and others (1982) present results of paleomagnetic analyses that indicate the Hersey till was deposited during the Matuyama Reversed Epoch prior to 0.7 million years ago and is therefore classical Kansan in age.

Continue east on old highway 18 to intersection with highway 18, 35 and 60.

0.3 4.8

Intersection with highway 18, 35, 60. Turn right (east). The intersection occurs on a low north/south trending drainage divide on the Bridgeport terrace. This divide is tentatively identified as the terminal moraine of the Bridgeport ice lobe that advanced from the west. Alden (1918) and MacClintock (1922) reported frequent observations of striated pebbles in sediments underlying this low divide. Alden (1918, p. 172) also observed a quartzite boulder on the surface near this intersection that was 60 cm (2 ft) diameter on the intermediate axis. He concluded the quartzite boulder was deposited by an ice lobe that advance across the Mississippi River from Iowa.

0.5 5.3

Intersection with Bouska Road on left (north). Stop 7B (alternate, depending on time and weather conditions).

The gully in the northeast corner of section 10, Bridgeport Township (see fig. 2, Bridgeport paper) cuts through thick sandy outwash sediments.

Continue east on highway 18, 35 and 60.

0.6 5.9

Intersection of highway 60 with highway 18 and 35. Turn left (east) onto highway 60. Proceed easterly over the highly dissected Bridgeport terrace.

0.6 6.5

The very well preserved flat terrace at about 200 m (660 ft) above sea level in the unnamed tributary of section 11 northeast of Bridgeport is the high Wisconsinan terrace (fig. 2, Bridgeport paper). Valleys appear to have been graded to this level until about 12,000 years B.P. (Clayton, this volume).

Continue easterly on highway 60 to stop 8 at Wauzeka.

Much of the Bridgeport terrace between Bridgeport and Wauzeka has been removed by erosion (fig. 1, Bridgeport paper). Where it is preserved the Pleistocene sediments overlying the strath on the Prairie du Chien Formation are primarily represented by Woodfordian loess and Bridgeport outwash sediments. Colluvium of unknown age also occurs at many sites. The gradient on the surface of the outwash slopes relatively steeply eastward and indicates that the Wisconsin River was reversed, relative to its present position, when ice blocked the valley at the mouth (fig. 3, Bridgeport paper). Former eastward flowage of the Wisconsin River is also suggested by the tendency for the median particle size of sand in the Bridgeport outwash to become finer in an eastward direction (fig. 5, Bridgeport paper).

3.8 10.3

Note effigy Indian mounds in wayside park on left (northwest side of highway 60).

0.3 10.6

Cross Gran Grae Creek. The high Wisconsinan (Woodfordian) terrace is very evident in most tributaries as is the situation here. A lower terrace that is about 4.6 m (15 ft) above the floodplain of the Wisconsin River is also preserved at selected localities along this reach of the Wisconsin River valley and its tributaries. The low terrace appears to be of erosional origin and equivalent to the low terrace described for drill hole 7 at stop 1 of this trip.

2.4 13.0

The basal hillslopes are now underlain by Cambrian sandstone and siltstone stratigraphy.

1.5 14.5

Cross Little Kickapoo River.

0.2 14.7

Note sign on left (north) advertising the Kickapoo Caverns. The cave occurs high on the valley wall about 90 m (300 ft) above the present level of the Wisconsin River floodplain. The cave apparently developed before entrenchment of the Wisconsin River.

2.1 16.8

Intersection with county highway N on left (north). Continue east on highway 60. Highway 60 is constructed on the low late Woodfordian terrace.

0.3 17.1

Turn left (north) on side street in Wauzeka. Proceed two blocks ~~T intersection and turn right (east).~~

0.1 17.2

Stop 8. Wauzeka quarry.

The Wauzeka section shows Bridgeport outwash gravels overlying bedrock. The elevation at the site is about 215 m (700 ft) above sea level. An exposure of outwash gravel in a tributary northwest of the village indicates that the gravels extend upward to 225 m (740 ft) above sea level (Alden, 1918). The 225-m level is therefore recognized as the approximate level of aggradation of the Wisconsin River valley when it was transporting Bridgeport outwash. Since the Bridgeport till appears to be of classical Kansan age, the elevation range between 215 and 225 m (700 and 740 ft) above sea level apparently represents the local base level of the adjacent Kickapoo River for at least a brief period of the early Pleistocene. The reader may recall from prior discussion associated with stops 1 and 5 that scars of cut-off and relict valley meanders in the Kickapoo valley extend upstream to about the same elevation above sea level. The reader may also recall that the longitudinal profile of the surface containing the fluvial gravels at the Oil City roadcut (stop 5) and bottom elevations of relict high-level valley meander scars grade to the surface on the strath at Wauzeka (fig. 7). These relationships suggest that the initial phase in the development of cut-off valley meanders began in classical Kansan time and that major valley incision has occurred since then.

Return to highway 60 and proceed eastward. Note the two levels of the Woodfordian terraces in Wauzeka.

1.8 19.0

Cross Kickapoo River.

0.1 19.1

Intersection of highways 131 and 60. Continue eastward on highway 60. The high late Woodfordian terrace occurs just ahead along the base of the hill on the north side of the road. A short distance beyond. The highway returns to the higher elevation of the Bridgeport terrace (fig. 29).

2.4 21.5

The Roth Site drill hole (Bridgeport paper) is in the pasture south of the sharp bend in the highway (NW1/4NE1/4 sec. 11, fig. 29, and Bridgeport paper, appendix I). Depth to bedrock at the site is 10 m (32 ft), and all of the sediments above the weathered bedrock are leached. The Bridgeport outwash is represented by clayey gravelly sand. The Bridgeport outwash is overlain by 1.7 m (5.5 ft) of clayey silt. Textural analyses of four samples between about 3 and 9 m (10 and 28 ft) depth indicated average percentages of sand, silt, and clay of 74:9:17.

0.2 21.7

Begin descent from Bridgeport terrace. Observe that both the high and low Wisconsinan terraces are preserved along the base of the Bridgeport terrace on the descent into Boydtown tributary (fig. 29).

0.3 22.0

Bridge crossing Boydtown Hollow tributary.

1.0 23.0

The highway passes along the edge of the Bridgeport terrace. Note the presence of the strath and bedrock in the roadcut on the north side of highway 60.

Continue northeasterly on highway 60. The next several miles provide scenic views of the braided lower Wisconsin River. The bedrock in the lower segments of hillslopes is Cambrian sandstone and siltstone, including the Jordan, St. Lawrence, and Lone Rock Formations.

4.7 27.7

Intersection of highways 60 and 61 at Wisconsin River bridge. Continue straight ahead (northeasterly) on highway 60. Note the colluvium overlying the Cambrian bedrock in the roadcut along the north side of the highway.

1.2 28.9

Intersection of highways 60 and 61. Turn right (east northeast) and continue on highway 60. Figures 1, 3, and 4 of the Bridgeport paper should be examined in preparation for stop 9.

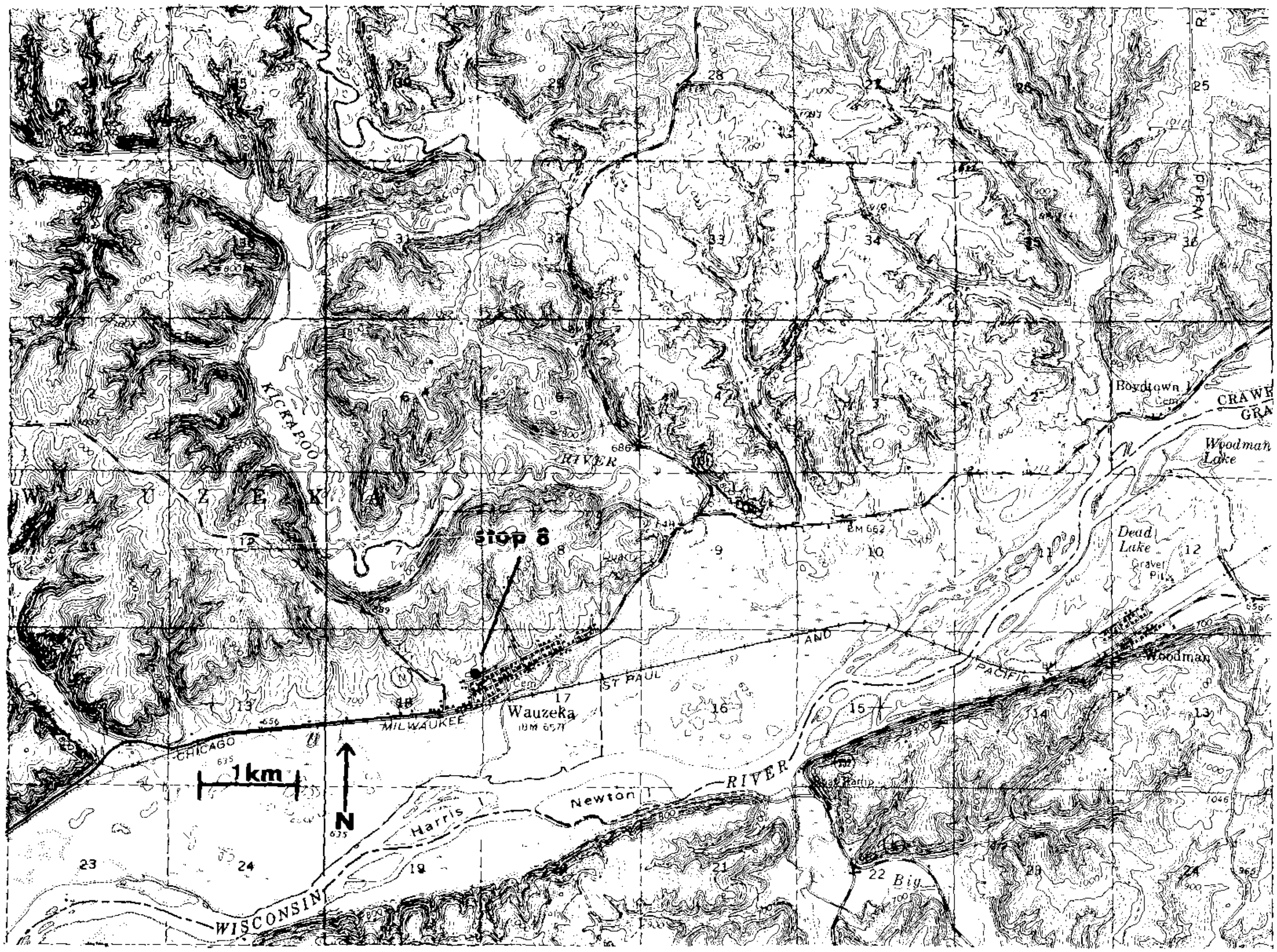

FIGURE 29.--Topographic map of the Wauzeka area. Crawford and Grant Counties, Wisconsin.

0.2 29.1

The cemetery to the left (north) of the highway is on the high Wisconsinan terrace.

4.1 33.2

Crossing Knapp Creek. The Wisconsinan terraces are very sandy as is apparent from the sand pit to the south of highway 60 immediately east of Knapp Creek.

0.2 33.4

Low Wisconsinan terrace.

0.2 33.6

Ascend high Wisconsinan terrace. Continue east on highway 60 to Port Andrew (fig. 4, Bridgeport paper).

3.1 36.7

Intersection of county highway X with highway 60 in Port Andrew. Turn left (north) onto county highway X. Proceed north on X to the Greenwood Cemetery drill hole (fig. 4, Bridgeport paper).

0.9 37.6

Stop 9A, Greenwood Site.

A description of the stratigraphic sequence for the drill hole at the Greenwood Site is given in appendix I of the Bridgeport paper. A generalization of the stratigraphy at the Greenwood Site is given in figure 3 of the Bridgeport paper.

Sediments at the Greenwood Site can be grouped into four categories. The top 5.6 m (18 ft) are represented by Woodfordian loess. A paleosol occurs near the base of the loess. The paleosol is not leached, a condition that probably has resulted from downward movement and inclusion of the overlying highly calcareous Woodfordian loess.

The second major unit is sand with chert granules and minor crystalline erratics (appendix I, Bridgeport paper). The sand extends to a depth of about 7.3 m (24 ft) and is leached throughout. The occurrence of the chert granules indicates the sand is not wind deposited. When the sand textures and lithology at this site were compared with characteristics of the sand fractions at the Elder and Thiede Sites 10 to 13 km (6 to 8 mi) to the east, it suggested that the sand unit at the Greenwood Site is fluvial outwash sediments from an eastern and northern Wisconsin source.

The third unit extends from the base of the relatively clean sand at 7.3 m (24 ft) to the contact with the Bridgeport outwash at a depth of 10.4 m (34 ft). The third unit is composed of silty sand

and sandy silt. Loess appears to have been a major contributing source for the silt and fine sand fractions. The unit is also leached.

The basal unit of Bridgeport outwash extends from 10.4 m (34 ft) to weathered bedrock at a depth of 15.2 m (50 ft). The Bridgeport outwash is calcareous at this site, which might suggest that it was soon buried by loess following its deposition. The sand fractions of the western-derived Bridgeport outwash and the overlying eastern-derived outwash have both similarities and differences. There is a tendency for the modal fraction to cluster around 1.50 to 1.75 phi (medium sand) in both, although the Bridgeport sand has a finer mode at the Thiede Site (figs. 4 and 5, Bridgeport paper). The most distinctive difference occurs in the coarse sand fraction where the Bridgeport outwash is dominated by quartz and the eastern-derived outwash has little quartz and is dominated by chert fragments.

It is concluded that the stratigraphy determined from the three drill holes and from surface exposures along the north side of the Wisconsin River near Blue River and Muscoda suggests two pre-Wisconsinan outwash deposits. MacClintock (1922) also recognized two pre-Wisconsinan deposits in the lower Wisconsin valley, but his relative age assignments appear to have been incorrect. MacClintock noted that the Bridgeport terrace deposits near Bridgeport contained more carbonate clasts than Bridgeport terrace deposits near Blue River and Muscoda. He therefore concluded that the Bridgeport terrace near Blue River and Muscoda was older than the Bridgeport terrace near the mouth of the Wisconsin River at stop 7. MacClintock worked only with surface exposures and was unaware of the calcareous Bridgeport outwash gravels such as described for the basal unit at the Greenwood Site. The difference in carbonate content of surface sediments probably is explained by source regions of the sediments. The western source area for the basal eastward dipping Bridgeport outwash includes southeastern Minnesota and northastern Iowa where carbonate bedrock occurs commonly at the surface. The source region for the pre-Wisconsinan outwash sediments that overlie the eastward-dipping Bridgeport sediments apparently is central and northern Wisconsin where carbonate bedrock is less commonly exposed at the surface.

Continue north on county highway X to the intersection with county highway M.

0.1 37.7

Intersection of county highways M and X. Turn right (east) on highway M. Note that from the crest of the hill it is apparent that the surface of the Bridgeport terrace to the east is both lower and less dissected, an appearance that is confirmed by the topographic map of figure 4 (Bridgeport paper). The difference in elevation and dissection might be because two different age Bridgeport deposits are juxtaposed here, but is is more likely that the difference is related to thicker loess deposits on the higher and more dissected area. Observe from figure 1 of the Bridgeport paper that the Greenwood Site is located immediately opposite a long southwest-northeast trending reach of the Wisconsin River. The Greenwood Site is therefore ideally located to receive large contributions of loess deposition compared to the sites further east. Loess thickness was only about 1 m (3 ft) thick at the Elder Site (Bridgeport paper, appendix I).

0.7 38.4

Intersection of county highway M with highway 60. Turn right (south) onto highway 60 and return toward Port Andrew.

0.8 39.2

Stop 9B (optional), Glenwood Drive section.

Turn left onto gravel road of Glenwood Drive and proceed for 0.6 mi. The Glenwood Drive section shows the Bridgeport outwash overlying Cambrian sandstone. The site is a gully exposure on the north side of the road. The quality of the site is subject to change. During the summer of 1981 the section provided evidence of weak eastward dipping bedding. The section usually is good for gaining an impression of the composition and textural variations of the western-derived Bridgeport outwash.

End of trip.

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MAPS OF THE POLLEN DATA

Two major conditions must be met if the calibration procedures are to work: (1) the pollen changes that are calibrated in the fossil data should be variations that can be attributed to climatic causes and (2) modern pollen data must exist that match these variations and can therefore be used to calculate the necessary calibration functions. Mapping the fossil pollen data provides an excellent way to illustrate the patterns in these data and to aid the selection of those patterns and those pollen types that seem to be responding to climatic changes. Isopoll maps with isopleths of pollen percentages show the geographical patterns in the modern (Davis and Webb, 1975; Webb and McAndrews, 1976) as well as the fossil data (Bernabo and Webb, 1977; Webb, 1981). These maps illustrate the patterns in pollen percentages, and hence in the vegetation, at selected dates, such as today, 3000 and 6000 B.P. Difference maps between two time periods show the changes in these patterns. In eastern North America, when similar changes for one or several pollen types occur from New England to Minnesota, then climate can be postulated as a likely candidate for the cause of these changes (Webb, 1980). Finding "revertence" in the values of certain pollen types (that is, the return to high values after an intermediate period of low values) is a second feature in pollen maps that can help in identifying past variations that probably reflect climatic changes. Von Post (1946) first proposed that revertence in a pollen type is a behavior indicative of climatic changes, and isochrone maps, which show the location of a particular isopoll at different times, provide an excellent format for illustrating revertence.

When the pollen data from the Midwestern sites are mapped for dates between 10,000 B.P. and the present (fig. 1), isochrone maps for selected isopolls show broadscale movements and revertence in the distribution patterns for the major pollen types (Bernabo and Webb, 1977; Webb and others, in press). Both the revertence and broad-scale patterns suggest that climatic changes may be the main, ultimate cause for the mapped distribution changes. The 5 % contour of spruce (Picea) pollen shows that spruce populations moved northward into Ontario from 10,000 to 8000 yr ago (fig. 1a). Later, after 4000 B.P., the abundance of spruce trees increased southward into northern Minnesota and Michigan. The isochrones for oak (Quercus) pollen show a similar northward movement in the early Holocene and southward movement after 6000 B.P. (fig. 1b,d). The primary movement of the isochrones for 20 % prairie-forb pollen (Ambrosia + Artemisia + Compositae + Chenopodiaceae/Amaranthaceae) is east-west (fig. 1c), and these isochrones show that the prairie/forest border had moved eastward into southwestern Wisconsin by 8000 B.P. and then retreated westward after 6000 B.P. (Wright, 1971).

These mapped changes in the pollen data suggest that temperature increased from 10,000 to 6000 B.P. and then decreased after that time. The map of prairie-forb pollen suggests that precipitation decreased from 10,000 to 6000 B.P. with most of the decrease in Minnesota and the Dakotas. After 7000 to 6000 B.P., precipitation again increased. In support of this interpretation, cores from many Iowa and Minnesota lakes show that water levels were lower at 7000 B.P. than they are today (Watts and Winter, 1966).

THE RELATIONSHIP BETWEEN MODERN CLIMATE AND POLLEN DATA

By statistically analyzing the modern climate and pollen data, we gained regression equations that represent the pollen/climate relationships within these data. Maps of the modern data illustrate these relationships and provide useful information for choosing the calibration data set. In the Midwest today, the mapped patterns of mean annual temperature and precipitation show good correspondence with the map-

HOLOCENE CLIMATIC CHANGES ESTIMATED FROM POLLEN DATA FROM THE NORTHERN MIDWEST

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INTRODUCTION

In the Midwest, the regional patterns of Holocene vegetational and climatic change provide a context for understanding the fluvial events recorded by soil and alluvial stratigraphic records in the Driftless Area. A spatial network of over 40 radiocarbon-dated pollen diagrams permits mapping of the geographic patterns of vegetational and environmental changes (Webb and others, in press), and calibration-function techniques can be employed to construct paleoclimatic maps from the fossil pollen data (Webb and Bryson, 1972; Webb, 1980). We have used these techniques and mapped the past patterns in annual precipitation and temperature for 9000, 6000, and 3000 B.P. We have also plotted the changes in these climatic variables along two transects that cross the Driftless Area. The first transect lies north-south and shows the 500-year changes in annual temperature from southwestern Ontario to central Illinois. The second transect extends east-west and shows the 500-year changes in annual precipitation from eastern South Dakota to Indiana.

METHODS AND DATA

Recent research in the climatic calibration of pollen data has focused on using regression techniques to gain quantitative climatic estimates from fossil pollen data (Webb and Clark, 1977; Howe and Webb, 1977; Kay, 1979; Heusser and Streeter, 1980; Andrews and others, 1980). An equation of the form

$$C = PB \qquad (1)$$

is used , where C is some measure of climate, P is the pollen data, and B is a vector of regression coefficients. The pollen/climate relationships represented by B are gained from analysis of geographic networks of modern pollen and climatic data. If a number of statistical and ecological assumptions are met, then the resultant calibration functions will yield reliable climatic estimates for fossil data (Webb and Clark, 1977; Howe and Webb, 1977; Howe and Webb, in preparation; Bartlein and Webb, in press).

In our study, we used a data set of modern pollen and climatic data from the region within 40° to 50°N. and 85° to 105°W. (Webb and McAndrews, 1976). The precipitation equation was derived from data within this whole region, but the temperature equation was derived from just the data between 85° and 95°W. (Bartlein and Webb, in press). The reconstructed climatic values were calculated for 40 radiocarbon-dated pollen diagrams in the Midwest (Webb and others, in press). Webb and others (in preparation) will provide a detailed description of this analysis.

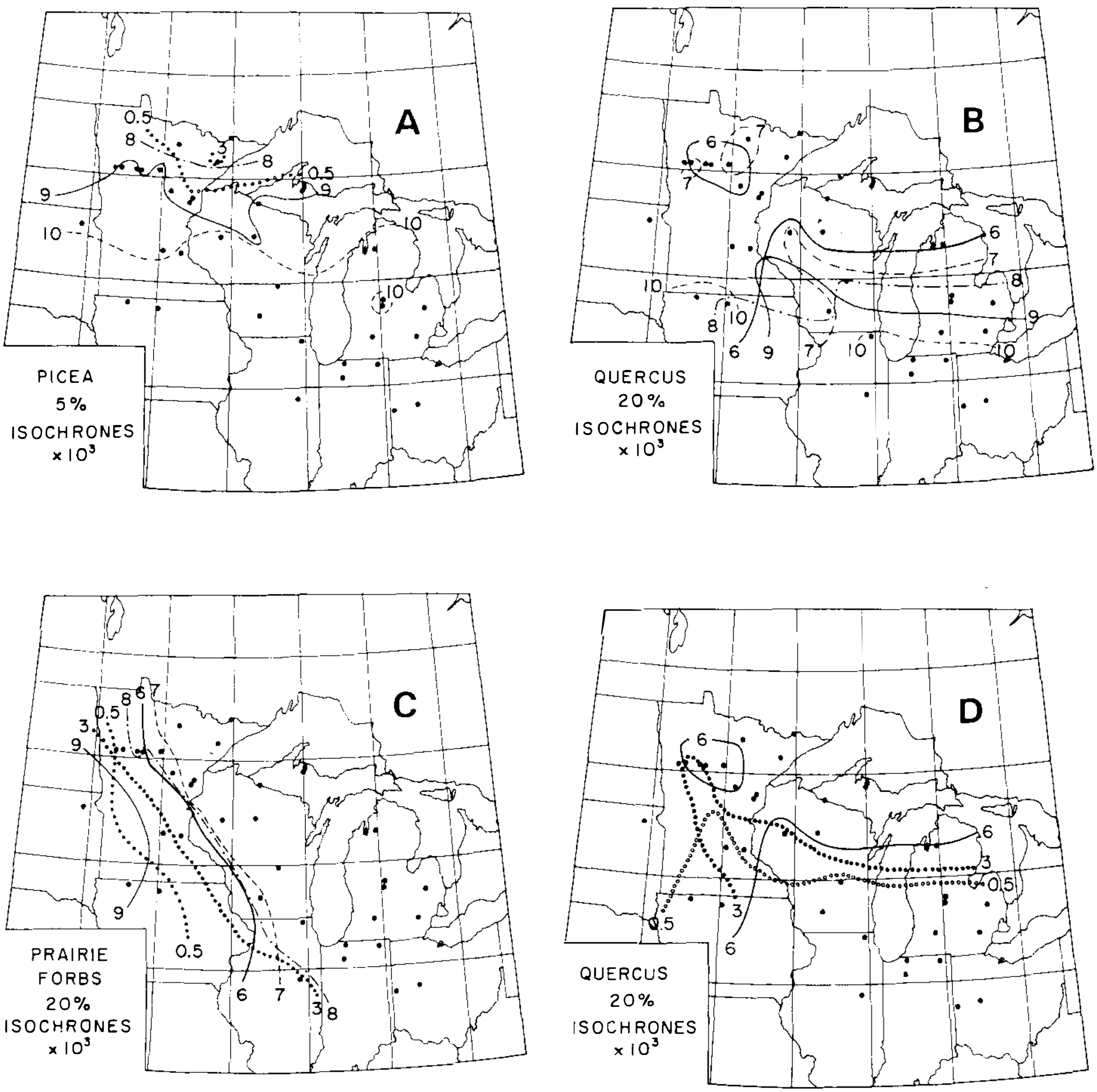


FIGURE 1.--Isochrone maps for 10,000 to 500 B.P. from radiocarbon-dated pollen diagrams in the Midwest: (a) map of *Picea* pollen in which 5 % or more spruce pollen is to the north of the isochrones, (b) and (d) maps of *Quercus* pollen in which oak pollen is to the south of the isochrones, and (c) map of prairie forb (*Artemisia* + *Ambrosia* + Compositae + Chenopodiaceae/Amaranthaceae) pollen in which 20 % or more of the pollen is to the west of the isochrones (from Webb and others, in press).

ped patterns of certain pollen types in the set of modern pollen data. The north-south gradient of temperature east of 95°W. is inversely related to the distribution of spruce pollen (fig. 2a and 2c) and is directly related to the distribution of oak pollen (fig. 2d). The general east-west gradient of precipitation between 40° and 50°N. (fig. 2b) is directly related to the distribution of oak pollen (fig. 2d) and inversely related to the distribution of prairie-forb pollen (fig. 2e). The mapped patterns of temperature and precipitation show that these climatic variables are essentially orthogonal over the Midwest, and we are therefore able to reconstruct two essentially independent components of the Midwestern climate.

The strong relationships between the climate variables and the selected pollen types are also evident on scatter diagrams for these variables (fig. 3). The scatter diagrams reveal pronounced nonlinearities in the relationships. These bivariate relationships may be linearized by transforming one or both of the variables. Transforming the percentage of oak pollen by raising it to the 0.25 power linearizes its relationship with mean annual temperature (fig. 3e), and raising the percentage of prairie-forb pollen to the 0.50 power linearizes its relationship with mean annual precipitation (fig. 3f).

Our aim when calculating a regression model is to minimize any violations to the statistical assumptions that ordinary least squares estimation (OLS) requires (Bartlein and Webb, in press). In particular, we have found that model-specification errors are the main source of assumption violations when calculating a calibration function. Specification error arises from a number of sources including (a) selection of too large or too small a geographic region for the modern data used in the analysis, (b) neglect of non-linear relationships between the dependent and predictor variables, and (c) omission of potentially useful predictors from the model. We have found that when specification error is minimized, other sources of assumption violations are minimized as well. The particular sequence of steps we use in model building includes

(1) selection of the pollen types with mean values greater than 1 % and recalculation of the pollen precentages using a sum of just these types;

(2) examination of bivariate scatter diagrams a) to help choose transformations for linearizing the pollen/climate relationship and b) to help screen for outliers and bad data values;

(3) estimation of a regression model with all candidate predictors included, and the calculation of various residual diagnostic statistics to help screen for further unrepresentative or overly influential data points and to assess the homogeneity of the region from which data is drawn;

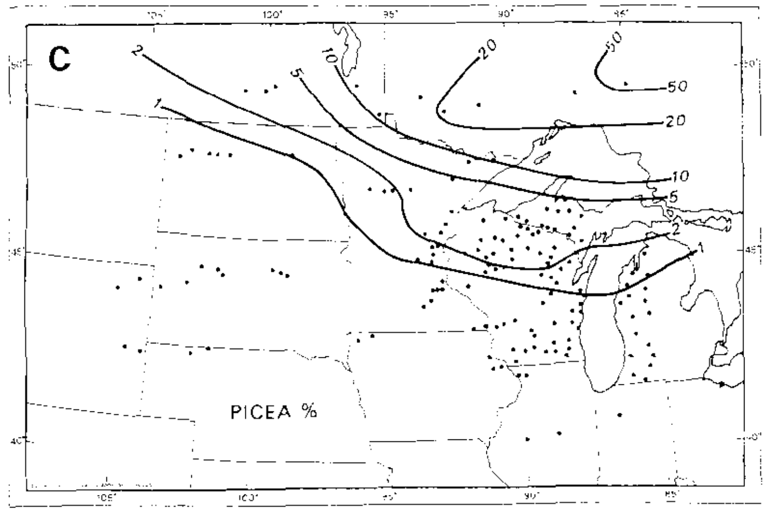
(4) estimation of a regression model with a subset of the candidate predictors chosen by a criterion that minimizes bias within the coefficients; and

(5) a formal analysis of the residual diagnostics for the chosen equation and modification of the equation, if necessary.

Howe and Webb (in prep.) and Bartlein and Webb (in press) describe the benefits that are derived in applying this sequence of steps.

The equations used for reconstructing annual temperature and precipitation appear in table 1. The summary statistics listed in this table are defined in Bartlein and Webb (in press).

Before interpreting the maps from 9000, 6000, and 3000 B.P., we first checked the performance of the calibration equations in estimating modern climatic values for the "core-top" samples at each site with a fossil data. We were concerned lest any biases appear among the climatic estimates. Bias might arise as systematic under- or over-estimation of the climatic values. To check for this type of bias, we regressed the climate estimates for the core-top data against the modern observed climate data. For both the temperature and precipitation estimates, these regressions yield intercepts insignificantly different from 0 and slopes insignificantly different from 1 (table 2). These results therefore revealed no evidence of bias among the modern samples from the fossil sites.



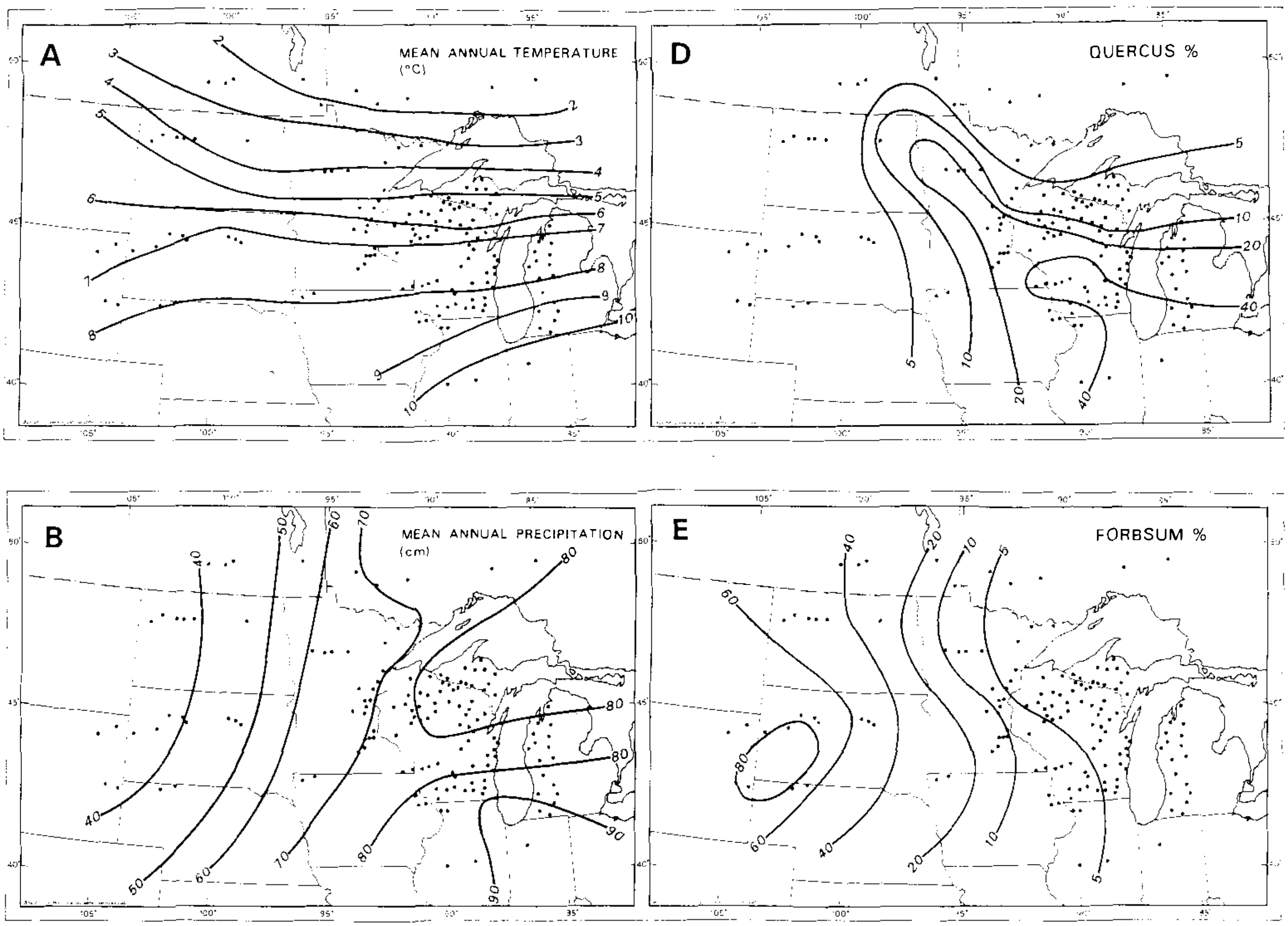
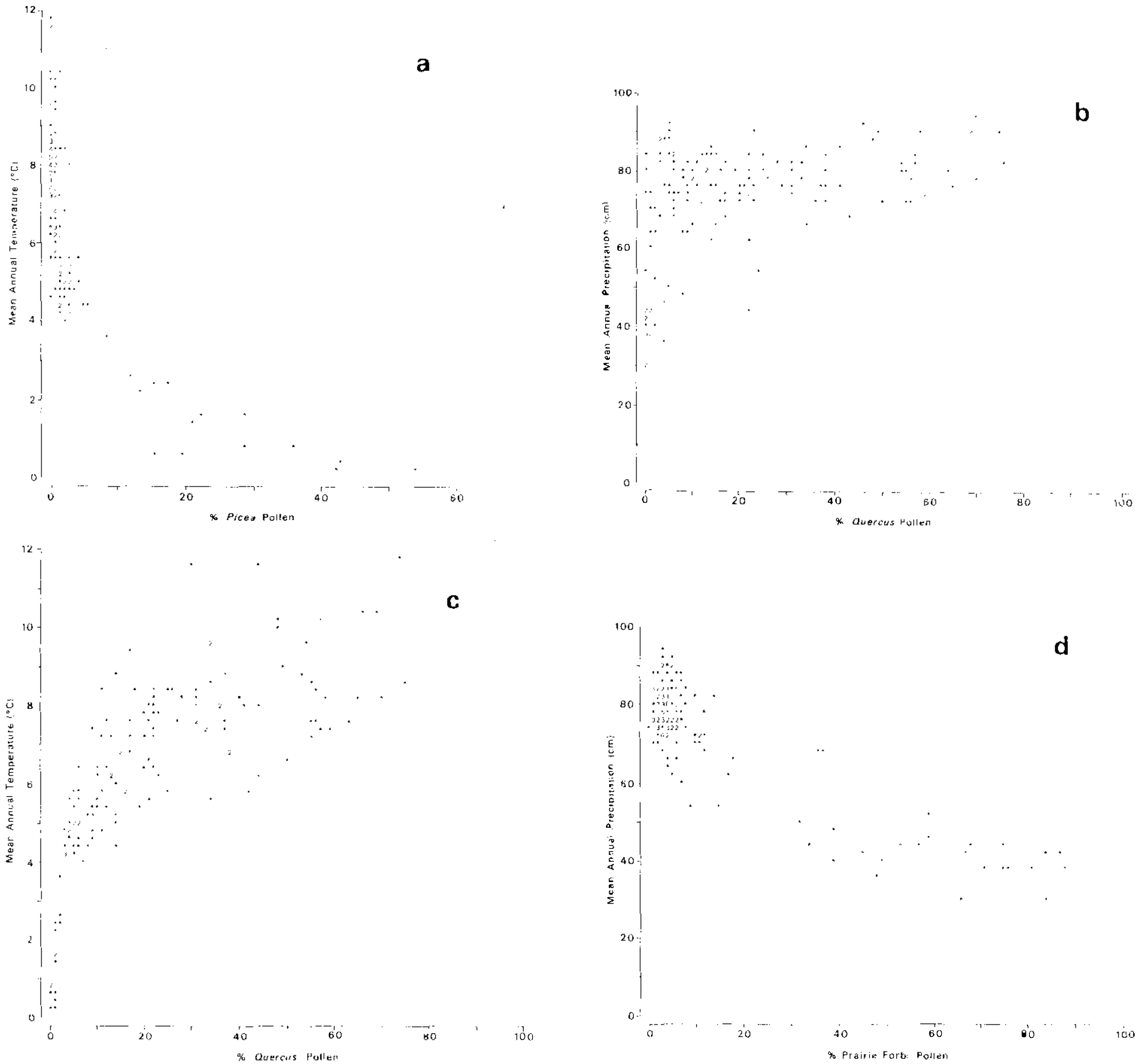


FIGURE 2.--Modern patterns of (a) mean annual temperature (°C), (b) mean annual precipitation (cm), (c) *Picea* pollen (d) *Quercus* pollen, and (e) prairie forb pollen.


a
Mean Annual Temperature (°C)
% Picea Pollen
b
Mean Annual Precipitation (cm)
% Quercus Pollen
c
Mean Annual Temperature (°C)
% Quercus Pollen
d
Mean Annual Precipitation (cm)
% Prairie Forb Pollen

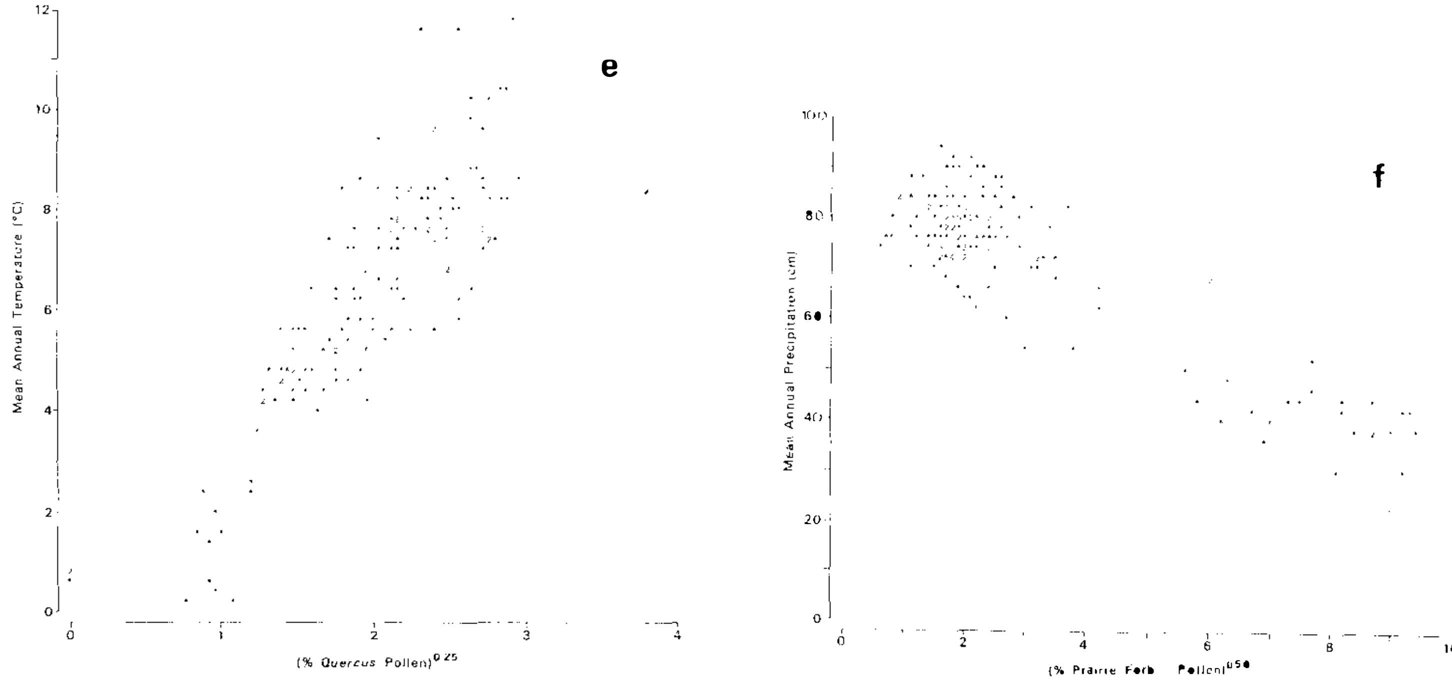


FIGURE 3.--Scatter diagrams for (a) mean annual temperature vs. the percentages of *Picea* pollen, (b) mean annual precipitation vs. the percentages of *Quercus* pollen, and (c) mean annual temperature vs. the percentages of *Quercus* pollen, (d) mean annual precipitation vs. the percentages of prairie forb pollen, (e) mean annual temperature vs. the percentages of *Quercus* pollen raised to the 0.25 power, and (f) mean annual precipitation vs. the percentages of prairie forb pollen raised to the 0.50 power. *Ambrosia* was deleted from prairie forb pollen for these scatter plots and for climatic calibration.

TABLE 1.--Calibration functions for mean annual temperature and precipation and summary statistics for the regression equations

— Reconstruction Equations —

(a) Mean Annual Temperature (°C)

$$\text{TMEANYR} = \underset{(.81)}{5.55} - \underset{(.17)}{.46}*\text{PICEA}.^{.25} - \underset{(.06)}{.17}*\text{PINUS}^{.50} - \underset{(.07)}{.29}*\text{BETULA}^{.50}$$

$$+ \underset{(.05)}{.20}*\text{FRAXINUS} + \underset{(.27)}{1.43}*\text{QUERCUS}^{.25} - \underset{(.02)}{.05}*\text{ULMUS}$$

$$+ \underset{(.06)}{.14}*\text{ACER} + \underset{(.09)}{.24}*\text{TSUGA}^{.50} - \underset{(.13)}{.29}*\text{ALNUS}^{.50}$$

(b) Mean Annual Precipitation (cm)

$$\text{PRECPYR} = \underset{(4.86)}{92.75} + \underset{(1.49)}{3.91}*\text{ABIES}^{.50} + \underset{(.81)}{2.08}*\text{Juniper}^{.50} - \underset{(.41)}{1.67}*\text{PINUS}^{.50}$$

$$= \underset{(1.35)}{2.48}*\text{FRAXINUS}^{.25} + \underset{(1.42)}{6.21}*\text{QUERCUS}^{.25} - \underset{(1.54)}{4.17}*\text{ULMUS}^{.25}$$

$$+ \underset{(.93)}{3.68}*\text{TSUGA}^{.25} - \underset{(1.28)}{5.12}*\text{ALNUS}^{.25} - \underset{(.43)}{5.87}*\text{FORBSUM}^{.50}$$

(Values in parentheses are the standard errors of the regression coefficients)

— Summary Statistics —

	(a) Temperature	(b) Precipitation
n (Number of observations)	108	135
p (Number of predictors)	9	9
R^2 (Squared multiple correlation coeff.)	84.6	88.2
adj. R^2 (R^2 adjusted for p)	83.2	87.3
C_p (Mallow's statistic)	8.39	10.09
Std. Error	.687	5.39
Condition Number	69	109
Statistics for test of		
inhomogeneity of residual variance	1.3478*	.7743*
non-normality of residuals	.997*	.994*
Spatial autocorrelation of residuals	.16**	.09**

* indicates that the statistic is not significant

** indicates that the test statistic is significant

Equation	Climate Variable	Dependent Variable 1/	Independent Variable	b_o intercept	t - test $b_0 = 0$	b_1 slope	t - test $b_1 = 1.0$	r^2	n
1	TMEANYR	Modern	0 yr B.P.	-0.0938	-0.205	1.0294	0.420	88.6	30
2	PRECPYR	Modern	0 yr B.P.	-8.6410	-1.170	1.1303	1.298	77.9	38
3	TMEANYR	Modern	500 yr B.P.	0.0857	0.170	1.0254	0.326	86.1	30
4	PRECPYR	Modern	500 yr B.P.	-9.9185	-1.092	1.1055	0.884	70.5	38

(all t - values are not significant)

TABLE 2.--<u>Cross validation regression summaries for 0 and 500 B.P.</u>

Another potentially more harmful source of bias in the reconstructions arises from Euro-American disturbance of the Midwestern landscape and its effect on the pollen rain (Van Zant and others, 1978). Despite the deletion of *Ambrosia* pollen from the analysis, herb pollen types including the other prairie forbs are higher in the modern data than the fossil data as a result of human activity. Reconstruction equations, such as the one for precipitation that depend heavily on the herb types, may thus produce biased reconstructions that are over-estimates of precipitation during the Holocene. One test for such bias is to compare the modern climate values with those estimated from pollen data representative of the time immediately before settlement. We used the pollen data interpolated to represent 500 B.P. and regressed their climate estimates against the modern observed climate data. Again, the intercepts are not different from 0 and the slopes not different from 1 (table 2). Although some climate variations have occurred during the past 500 years in the Midwest (Swain, 1978; Bernabo, 1981), these variations are modest when compared with the climatic variations during the rest of the Holocene, and are unlikely to mask any serious bias in the reconstructions caused by some pollen types being over-represented in the modern set.

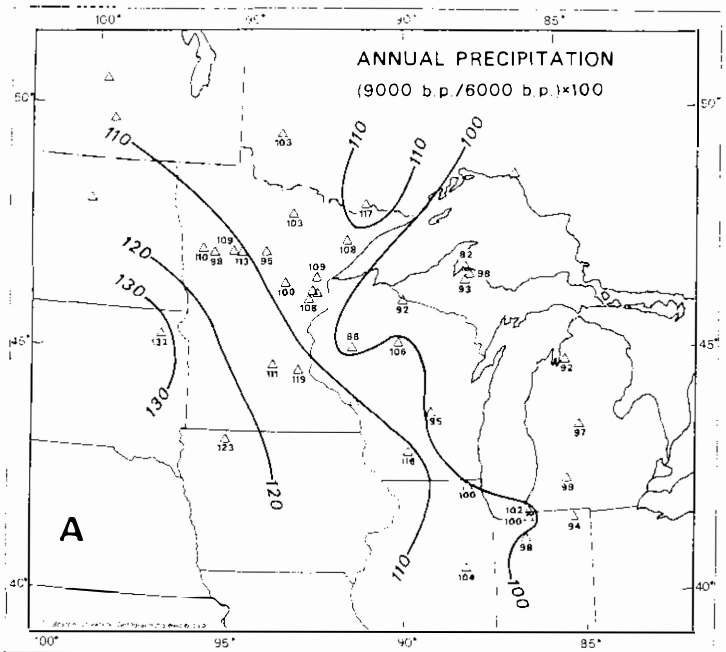
HOLOCENE CLIMATIC VARIATIONS

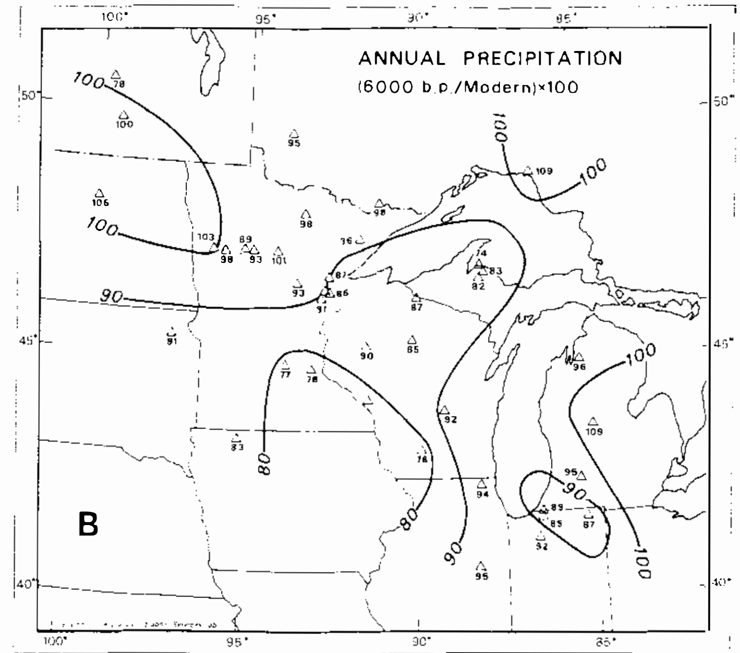
From 9000 to 6000 B.P., annual precipitation declined nearly everywhere across the network of sites in the Midwest (fig. 4a). Precipitation was over 20 % higher at 9000 B.P. than at 6000 B.P. in parts of western Minnesota and Iowa, and was at least 10 % higher in western Wisconsin. At 6000 B.P., precipitation was less than modern values nearly everywhere in the Midwest, and was less than 80 % of its modern values over parts of western Wisconsin, southern Minnesota, and Iowa (fig. 4b). These precipitation variations match the movements of the prairie/forest border between 9000 and 6000 B.P. (fig. 1). The precipitation variations are probably associated with a region-wide increase in the duration of mild, dry Pacific airmasses that resulted from an increase in the frequency of zonal-type circulation patterns (Wendland and Bryson, 1981). Our unpublished reconstructions of airmass durations support this interpretation.

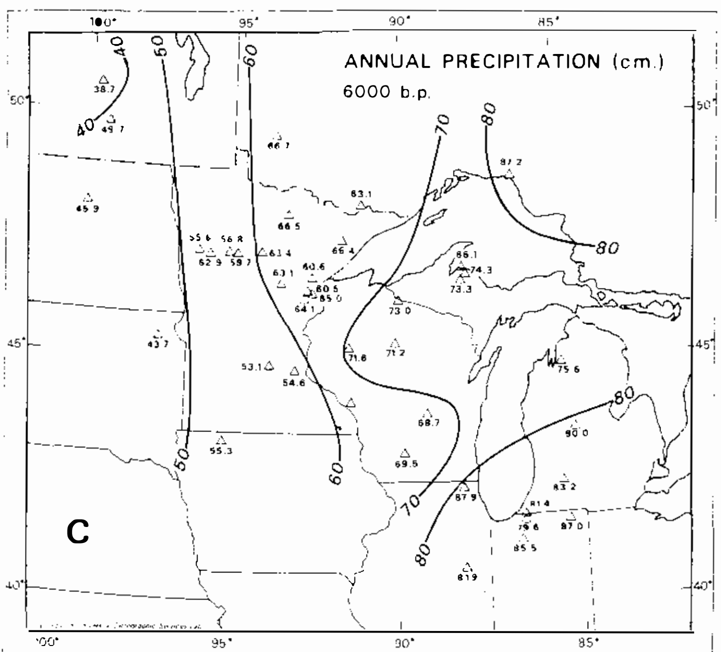
From 6000 to 3000 B.P., precipitation increased around 10 % nearly everywhere in the Midwest except from Iowa to northern Indiana (fig. 4e). The decrease in precipitation along the southern edge of the region is matched by a similar adjustment in the distribution of prairie-forb pollen (fig. 1) and is accompanied by a decrease in temperature across the northern half of the Midwest. These changes are probably related to an increase in the duration of Arctic airmasses in the north and a southward shift in the axis of the strongest westerlies. From 3000 B.P. to the present, precipitation increased across the southern edge of the region as the duration of the moist Atlantic (maritime Tropical) airmasses increased there (fig. 4f).

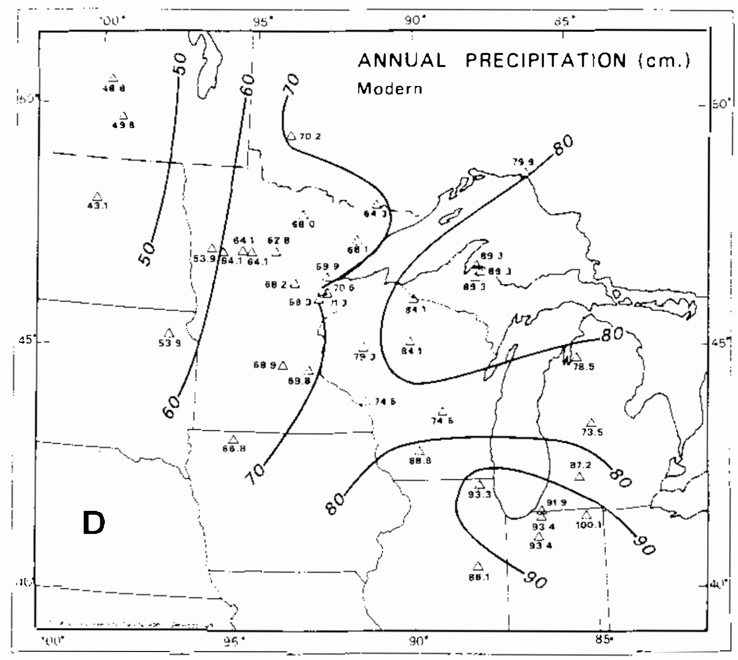
Temperatures in the Midwest were up to 2.5°C cooler at 9000 B.P. than at present, and increased between 1°C and 2°C nearly everywhere across the Midwest between 9000 B.P. and 6000 B.P. This temperature change is matched by the retreat of the boreal forest into Canada (fig. 1). In parts of northern Minnesota, temperatures were over 2.0°C warmer at 6000 B.P. than at present as Pacific air replaced the colder Artic air (fig. 5). Between 6000 B.P. and

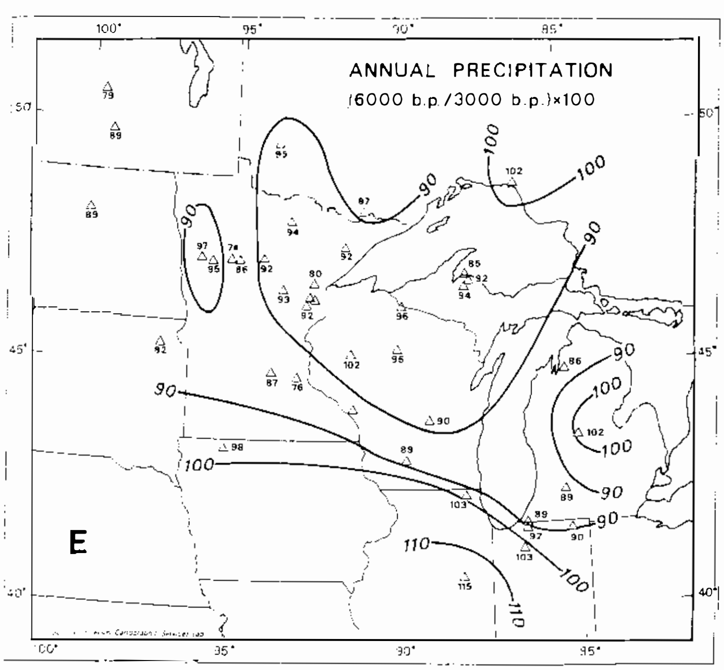
FIGURE 4.--Maps of mean annual precipitation for (a) the ratio between 9000 and 6000 B.P., (b) the ratio between 6000 B.P. and today, (c) 6000 B.P., (d) today, (e) the ratio between 6000 and 3000 B.P., and (f) the ratio between 3000 B.P. and today.

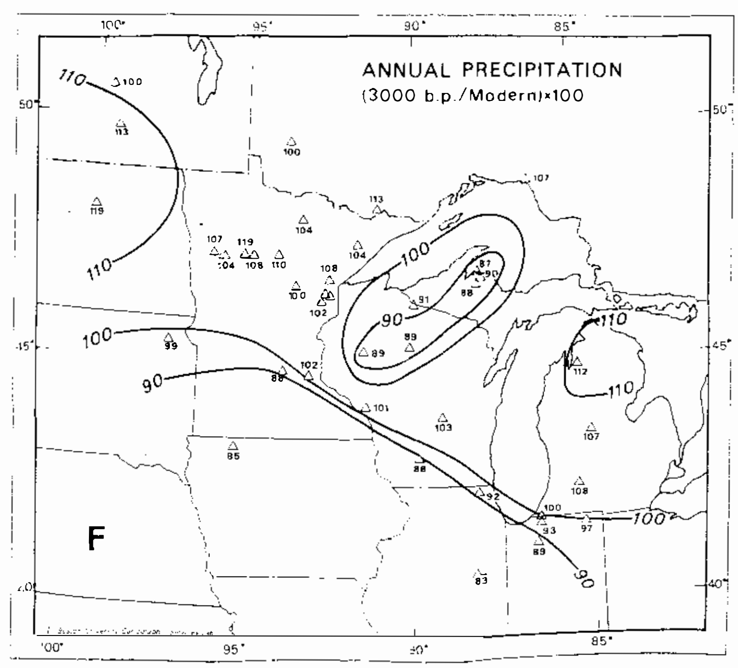

ANNUAL PRECIPITATION
(9000 b.p./6000 b.p.)×100
A


ANNUAL PRECIPITATION
(6000 b.p./Modern)×100
B


ANNUAL PRECIPITATION (cm.)
6000 b.p.
C


ANNUAL PRECIPITATION (cm.)
Modern
D


ANNUAL PRECIPITATION
(6000 b.p./3000 b.p.)×100
E


ANNUAL PRECIPITATION
(3000 b.p./Modern)×100
F

the present, temperatures decreased across northern Minnesota into northern Michigan, and increased to the south of that region.

Further insight into the character of the Holocene climatic variations can be gained by examining the climatic variations along two transects: an east-west transect of reconstructed precipitation values from Pickerel Lake in South Dakota to Pretty Lake in Indiana (fig. 6), and a north-south temperature transect from FRB Lake in Ontario to Chatsworth Bog in Illinois (fig. 7). The time/space cross section of precipitation clearly reveals the tongue of dry air and lower precipitation that extended eastward across the Midwest during the mid-Holocene (fig. 6). The maximum dryness was attained between 6500 and 6000 B.P. The time/space cross section for temperature (fig. 7) shows an apparently time transgressive temperature maximum or "Hypsithermal", with the maximum warmth occurring around 6000 B.P. in the northern Midwest, and around 4000 B.P. in the southern Midwest. The general warming from the early to the mid-Holocene was probably related to the replacement of the retreating Arctic air by the warmer (and drier) Pacific airmasses prior to 6000 B.P. The later temperature maximum in the southern Midwest reflects the replacement of the Pacific airmasses by the warmer Atlantic airmasses between 6000 and 3000 B.P.

One significant feature of the maps of reconstructed climates is that patterns at both a regional and a subregional scale appear on the change maps (fig. 4, 5). We may thus be able to derive "synoptic" scale maps of past climatic patterns with expansion of the reconstructions into the eastern United States and eastern Canada. It will then be possible to make inferences about the specific nature of the variations in atmospheric circulation that control regional variations in climate.

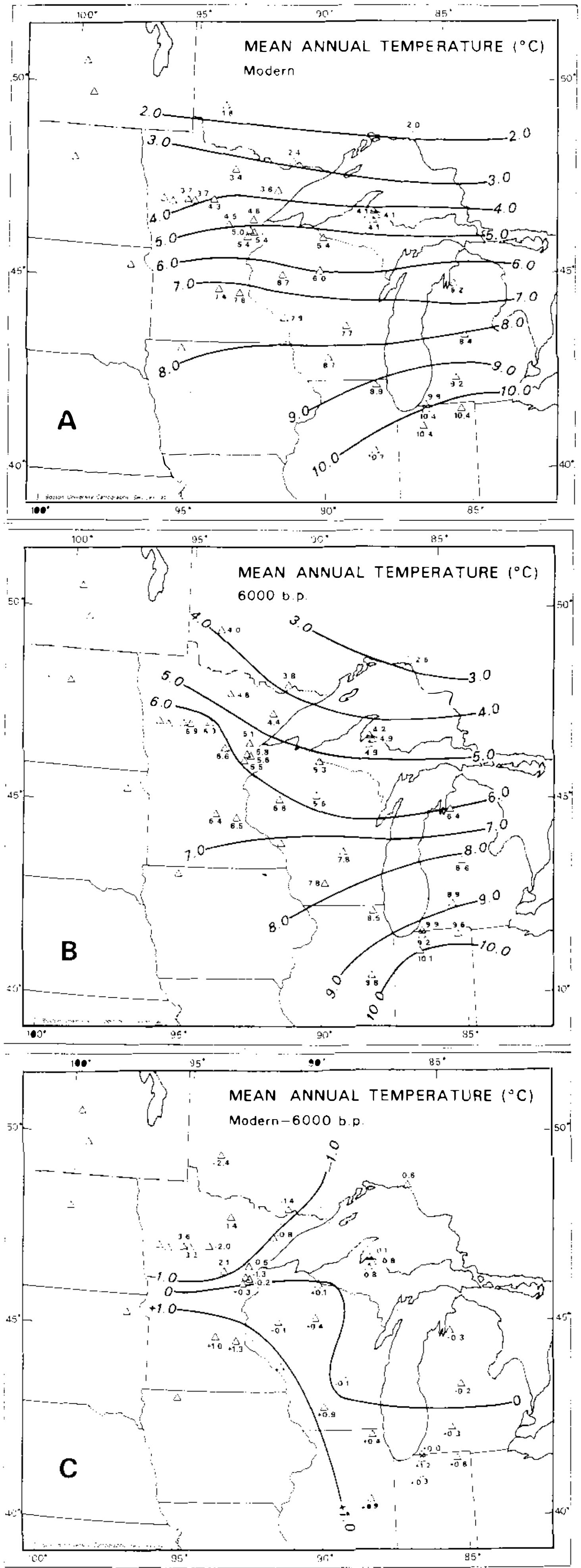


FIGURE 5.--Maps of mean annual temperature for (a) today, (b) 6000 B.P., and (c) the difference between 6000 B.P. and today. Negative values indicate lower temperatures today than at 6000 B.P.

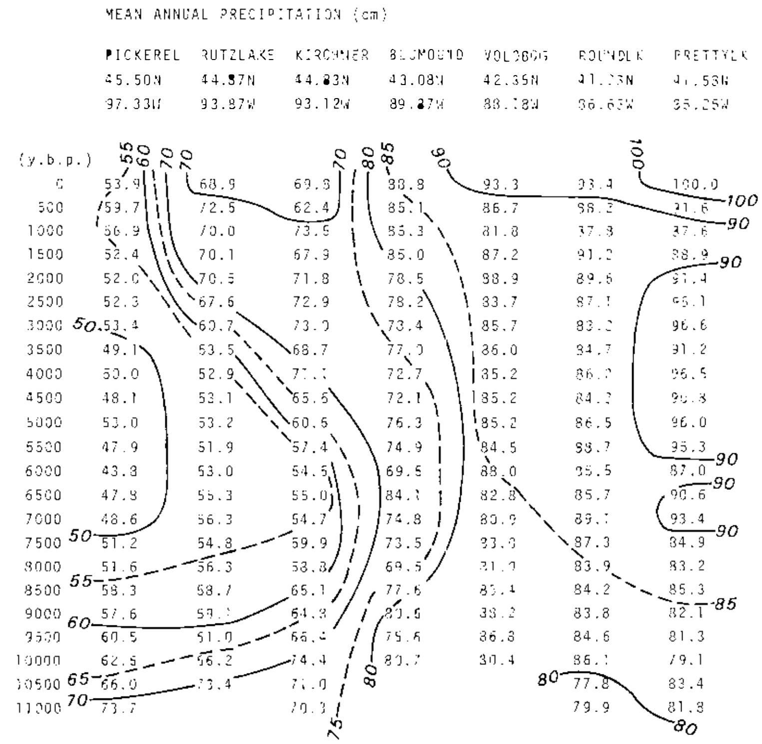


FIGURE 6.--West to east time/space cross section in 500-yr intervals for mean annual precipitation.

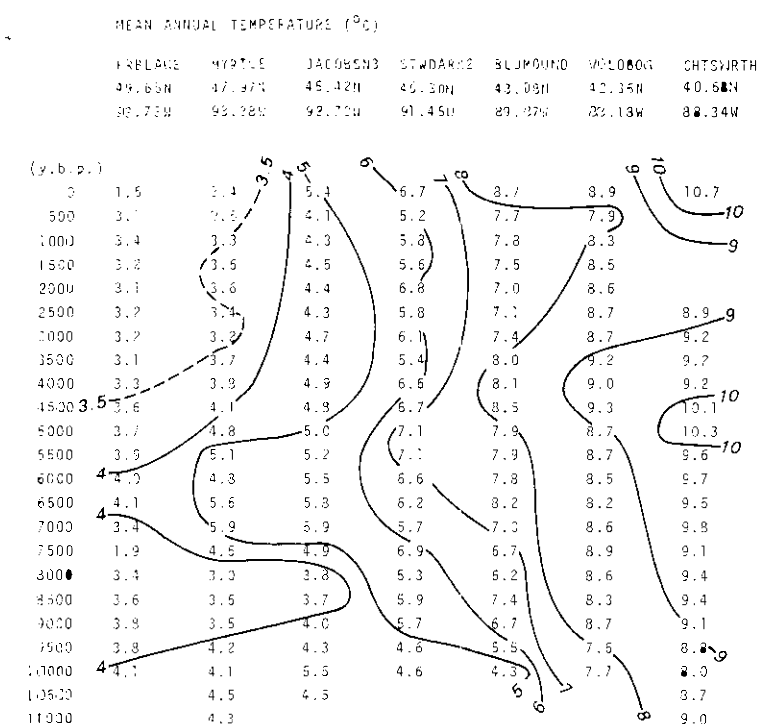


FIGURE 7.--North to south time/space cross section in 500-yr intervals of mean annual temperature.

SUMMARY OF RESULTS

Environmental change in the Midwest during the Holocene is recorded in fossil pollen data that describe the vegetational history of the region. The principal features of the vegetational history include the northward movement of the boreal forest from 10,000 B.P. to 8000 B.P., and the eastward movement of the prairie/forest border into Wisconsin by 8000 B.P. and its subsequent westward retreat after 6000 B.P. (fig. 1).

These vegetation changes as recorded in the fossil pollen can be transformed into quantitative estimates of past climates by use of regression equations calibrated from modern climate and pollen data (table 1). The reconstructed climatic variations are temporally and spatially coherent, and match the qualitative reconstructions based on pollen isochrone maps. From the early Holocene to 6000 B.P., precipitation decreased between 10 % and 20 % over much of the Midwest (fig. 4). At 6000 B.P. precipitation was less than 80 % of its modern values in parts of western Wisconsin, southern Minnesota, and Iowa. After 6000 B.P. precipitation generally increased across the Midwest. Temperature increased between 1°C and 2°C nearly everywhere across the Midwest from the early Holocene to 6000 B.P., when temperatures in parts of Minnesota were as much as 2°C above present values (fig. 5). After 6000 B.P. temperatures declined in the northern Midwest, but continued to slowly increase in the southern Midwest.

CLIMATIC VARIATIONS IN THE DRIFTLESS AREA

From the regional patterns of climatic variations, we can gain some idea of the Holocene climatic variations in the Driftless Area. Previous work (Davis, 1977) showed that climatic interpretation of pollen data from the Driftless Area is difficult because of the steep relief and consequent creation of microhabitats at certain sites. The Midwestern transects and maps show the climatic gradients across the Driftless Area and allow one to infer the climatic changes within it.

Our regional reconstructions show that temperatures generally increased there throughout the Holocene, with the maximum occurring between 6000 B.P. and 3000 B.P. (fig. 7). The maximum in the northern Driftless Area may have been earlier than the maximum in the south. Precipitation decreased during the early Holocene, reaching values 20 % lower than modern values at 6000 B.P. Since 6000 B.P. precipitation has increased steadily to its modern values (fig. 4,6). Geomorphic activity in the Driftless Area is broadly correlated with these climatic variations (Knox, 1972; Knox and others, 1981). In particular, the interval of maximum dryness immediately prior to 6000 B.P. corresponds to an interval of little or no alluvial deposition in the valley of Brush Creek, Wisconsin (McDowell, in preparation). As precipitation increased after 6000 B.P., significant changes in the nature of geomorphic activity occurred, resulting in widespread alluviation in Driftless Area valleys.

ACKNOWLEDGMENTS

National Science Foundation grants (ATM79-16234 and ATM81-11870) to COHMAP (Climates of the Holocene: Mapping based on Pollen Data) and a Department of Energy contract (DE-AC02-79EV10079) from the Carbon Dioxide Program supported this research at Brown University. We thank E. McClennen and M. Anderson for drafting the figures and maps and R. Arigo, E. Fleri, D. Gaudreau, R. McKendall, R. M. Mellor, S. Suter, and C. Thrall for technical assistance. J. H. McAndrews and R. Kapp kindly allowed us to use some of their unpublished pollen data.

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INFLUENCE OF AGASSIZ AND SUPERIOR DRAINAGE ON THE MISSISSIPPI RIVER

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From about 12,000 to 9500 B.P., drainage from Lake Agassiz and Lake Superior had a major influence on the Mississippi River and its tributaries (fig. 1), including, presumably, those of the fieldtrip area. Although few radiocarbon dates are available, Clayton and Moran (1982) and Clayton (Agassiz Symposium at the Geological Association of Canada meeting in Winnipeg, May 17 to 19, 1982; and Wisconsin Geological and Natural History Survey Report of Investigations, in press) have attempted to correlate events in the Agassiz and Superior areas; their chronology is presented in outline form here (fig. 2), but without the evidence given in those publications.

Whenever the Red River Lobe (in the Agassiz basin) or the Superior Lobe of the ice sheet terminated south of the continental divide (fig. 1) outwash was transported down to the Minnesota and St. Croix Rivers, causing them to aggrade. Whenever the lobes were north of the divide, water was ponded in the Agassiz and Superior basins and lake water spilled southward, causing the Minnesota and St. Croix Rivers to degrade because they no longer carried bedload from the glacier. The presumed effect of these events on the Mississippi River is shown in the center column of figure 2. The Mississippi River aggraded until the terminus of the Red River and Superior Lobes retreated north of the continental divide about 12,200 B.P. Then rapid downcutting occurred during the Two Creeks interval. Both the Red River and Superior Lobes again advanced across the divide about 11,500 B.P., causing renewed aggradation. This was followed by the main period of downcutting, until about 10,800 B.P. Another period

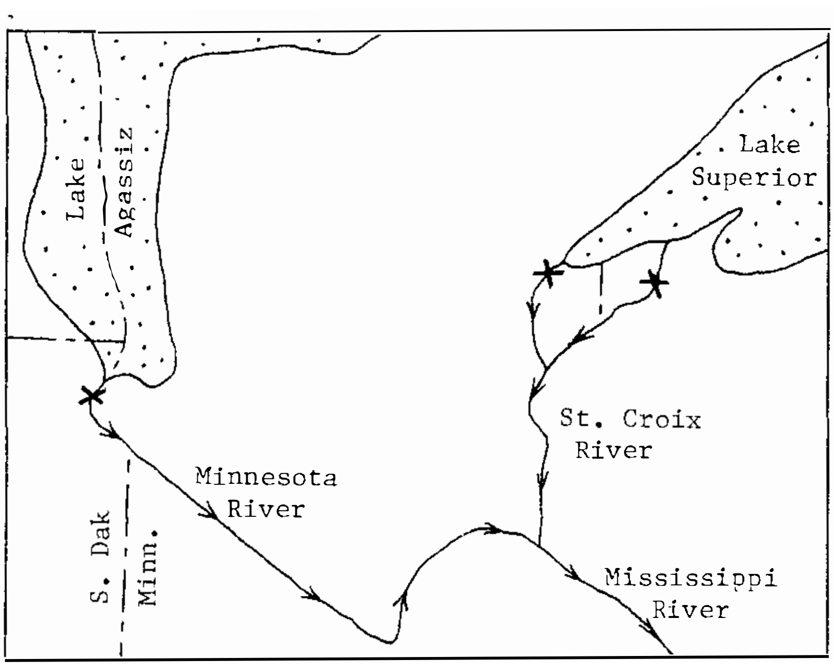

FIGURE 1.--Location of the southern spillways of Lake Agassiz and Lake Superior. X indicates the present-day continental divide.

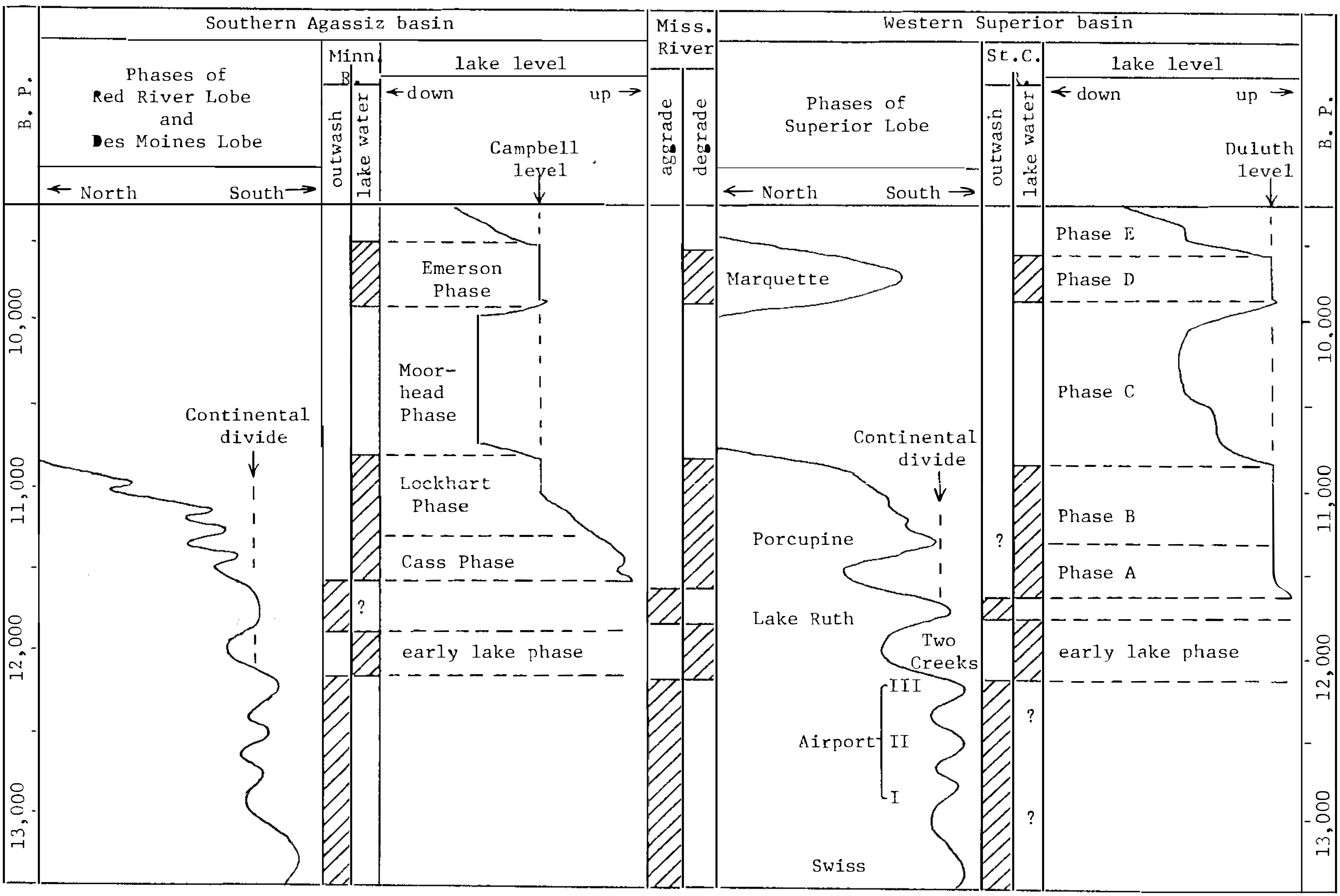

FIGURE 2.--Chronology of events in the Agassiz and Superior basins. Each named glacial phase represents a period of ice-margin advance and retreat.

of downcutting occurred between 9900 and about 9500 B.P. These events are described in more detail in the following paragraphs.

The Swiss Phase of the Superior Lobe (fig. 2) produced an outwash terrace down the St. Croix valley. This terrace flattens at the level of Lake Grantsburg (fig. 3), suggesting that it ended there at a delta. Lake Grantsburg was dammed in the St. Croix valley by the Grantsburg Sublobe of the Des Moines Lobe (the predecessor of the Red River Lobe). If this interpretation is correct, the Swiss Phase is equivalent to the Split Rock Phase of the Superior Lobe in Minnesota, which also produced outwash ending at a delta in Lake Grantsburg.

This was followed by the Airport Phases of the Superior Lobe (fig. 2), which produced outwash terraces that seem to correlate with the main high-level terrace down the St. Croix and Mississippi Rivers (fig. 3). Along the middle and lower reaches of the St. Croix River, this terrace consists of well-sorted sand overlain by eolian dunes, probably because the sediment was derived from the late Precambrian sandstone of the Superior basin. In contrast, later terraces consist of more poorly sorted, coarser sediment lacking dunes, probably because the sediment was derived from a variety of materials including till in the cutbanks of the spillway. Clayton (in press) has tentatively correlated the Airport Phase with the Nickerson Phase of the Superior Lobe in Minnesota and the Winegar Phases of the Chippewa Sublobe of Wisconsin and Michigan.

During the following period (the Two Creeks interval), both the Red River (Des Moines) Lobe and the Superior Lobe wasted across the continental divide, initiating early phases of Lake Agassiz and Lake Superior, causing the Minnesota and St. Croix Rivers to downcut. The upstream end of the Airport outwash surface was entrenched more than 30 m.

Mark Flock (Agronomy Department, University of Illinois, Urbana; in preparation, April 1982) has identified slack-water clay on the main high Mississippi terrace (Airport?) from the Chippewa River to St. Louis; layers of coarse, red, kaolinitic clay were probably derived from the Superior basin, and layers of fine, gray, montmorillonitic clay were probably derived from the Agassiz basin. He suggests that fluvial deposition on this terrace ceased and slack-water deposition began around 12,500 B.P., about the time loess deposition ceased in the region.

A readvance of the glacier across the continental divide briefly produced a new phase of aggradation in the St. Croix valley, during the Lake Ruth Phase (fig 3). Clayton (in press) has tentatively correlated the Lake Ruth Phase with the Marenisco Phase of the Chippewa Sublobe.

Both lobes again wasted back across the continental divide, shutting off the supply of outwash to the Mississippi River about 11,500 B.P. During the Cass and Lockhart Phases of Lake Agassiz and phases A and B of Lake Superior (fig. 2), lake water lacking bedload spilled southward down the Minnesota and St. Croix Rivers, causing them to rapidly downcut. Several proglacial lakes in Alberta, Saskatchewan, Manitoba, North Dakota, Minnesota, and Ontario spilled eastward and southward, sometimes catastrophically, causing flows of perhaps as much as 10^6 m^3/s down the Mississippi (see proceedings of the Agassiz Symposium at the Geological Association of Canada Meeting, May 1982). Clayton (in press) has shown that during much of the time that Lake Superior stood at the Duluth level, the upper St. Croix River was 1 km wide and 12 m deep; and much of the time that Lake Agassiz stood at or above the Campbell level, the Minnesota River was much bigger than the St. Croix.

At the end of the Lockhart Phase (fig. 2) Lake Agassiz dropped below the Campbell level, shutting off the flow

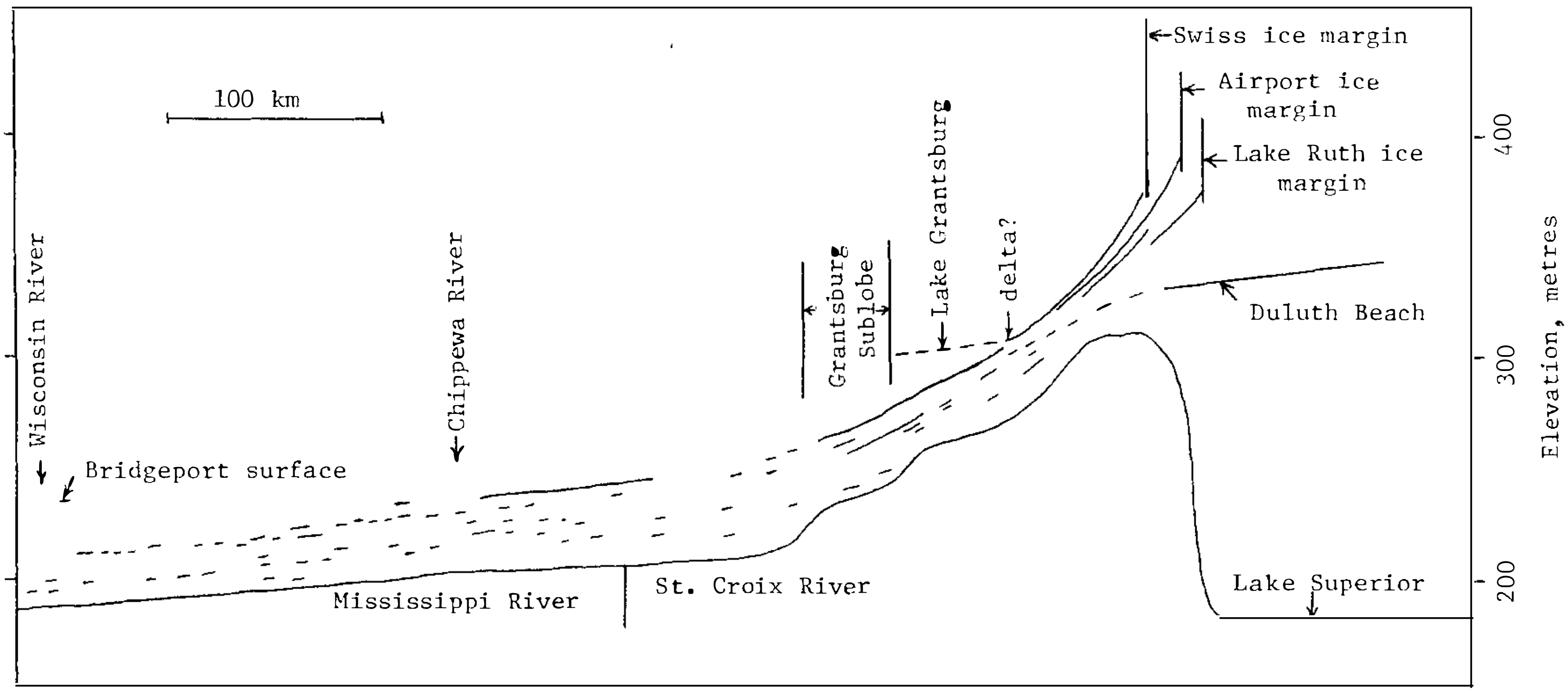

FIGURE 3.--Terraces along the St. Croix and Mississippi Rivers.

to the Minnesota spillway. This was caused by the opening of the eastern Agassiz outlets to Lake Superior as the Superior Lobe wasted from the Nipigon basin in Ontario. Clayton (in press) has argued that this occurred when Lake Superior stood below the Duluth level (fig. 2), because the Duluth level was above the Campbell level, perhaps when Lake Superior stood at the Algonquin level, about 11,800 B.P.

Both Lake Agassiz (Emerson Phase) and Lake Superior (phase D) again spilled southward to the Mississippi River during the Marquette Phase of the Superior Lobe, when it again blocked the Agassiz outlets through the Nipigon basin. The beginning of this event is well dated at 9900 B.P. The upper Mississippi River continued to downcut during this episode, probably to below its present level.

The Emerson Phase of Lake Agassiz ended when it again drained eastward through the Great Lakes about 9500 B.P. Like the flood at the start of the Moorhead Phase, it could not have come down the St. Croix River because the Duluth level was above the Campbell level (fig. 2)

I thank Mike Mudrey, Dave Mickelson, and Jim Knox for reviewing this note.

REFERENCE CITED

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THE GENESIS AND DISTRIBUTION OF UPLAND RED CLAYS IN WISCONSIN'S DRIFTLESS AREA [1]

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INTRODUCTION

Reddish clay-rich soils, which commonly occur in conjunction with weathered limestone and dolomite surfaces, are termed *terra rossa*. Because of close association, the red clays are sometimes thought to be weathered residual insoluble materials that accumulate as carbonate dissolution progresses, although fluvial-colluvial and eolian provenance are also indicated. In karst areas of southern Indiana much of the *terra rossa* is pedisediment which is derived from the backwasting of stratigraphically higher clastic rocks (Olson and others, 1980). Eolian contributions have been recognized by similarities in clay mineralogy (Ballagh and Runge, 1970) and by the divergence of oxygen isotopic ratios of associated fine quartz (Syers and others, 1969).

Buried reddish-brown clays discontinuously mantle the dolomite uplands of southwestern Wisconsin. The strong brown to yellowish red colors, clayey texture, strong angular blocky structure, and distinct lack of horizonation impart to these clays an appearance which contrasts with the overlying Peoria-loess derived soils. The red clays are very similar in appearance to those of the limestone uplands of central and southern Missouri and to the central European brown loams (*terra fusca*) described by Kubiena (1953).

Chert pebbles and cobbles are often present within the clay matrix and commonly are concentrated as stonelines at the boundary with the overlying loess. The boundary with the loess is generally quite distinct. The dolomite characteristically weathers to rhombs of sand size before complete dissolution. Localized mixing of clay and sandy dolomite as well as penetration of clays into dolomite pores and cracks is common. A thin zone of brown clay several millimeters to centimeters thick often lies directly over the dolomite boundary. Distinguishing true residuum from highly weathered loess is not possible in many places (Black, 1970). The similar clay mineral assemblages of loess and red clay and the high within-group variance of heavy mineral suites of dolomites and red clays have frustrated provenance studies (R.H. Akers, 1961, Clay minerals of glacial deposits of west-central Wisconsin: M.S. thesis, University of Wisconsin, Madison). This paper provides a genetic model that helps to explain the mineralogy, texture and profile characteristics as well as to assess the relative contributions of residual and eolian materials to red clay profiles.

FINE CLAY ILLUVIATION

The upland red clays are texturally distinct from both the dolomite insoluble material and the Peorian loess (table 1). The high clay content of the red clays cannot result solely from weathering of dolomite residuum or loess since the silt and sand fractions of these materials are largely quartz. Rather the texture of the clays must result from an influx of clay. The very high fine-medium clay content of the red clays suggests that much of the red clay could be illuvial. Thin sections of samples taken from the red clay-dolomitic sand contact reveal isolated dolomite rhombs within the basal red clay and oriented clays coating grains in the weathered dolomite zone. These latter clays cannot result from in situ dolomite dissolution since the unweathered dolomite contains only an average of 2.1 % clay. Slight increases in organic matter content and acid ammonium oxalate extractable iron near the base of several red clay profiles suggest that clay-organo-metallic complexes have been translocated to the clay dolomite contact (table 2).

The red clays are mineralogically uniform, both within profiles and among different sites (table 1). Smectites and mixed layer intergrades dominate both the fine-medium and coarse clay fractions. Kaolinite and quartz occur in moderate amounts and mica and vermiculite are present in small quantities. The loess, except for greater amounts of vermiculite, exhibits a clay mineralogy similar to that of the red clay. The mineralogy of clays present within the yellowish brown dolomitic sand resembles that of the overlying red clay (fig. 1). The calcareous environment of the dolomitic sand is not favorable for the almost complete alteration of mica to 2:1 expandables and kaolinite.

Textural, mineralogical and thin section evidence indicate a mechanism for red clay accumulation similar to that for Beta-B horizons (Bartelli and Odell, 1960). Fine textured, smectite-rich clays accumulate in the zone of dolomite dissolution. When proximal to dolomite, acid illuvial clays exchange hydrogen ions for Ca^{2+} and Mg^{2+}. The

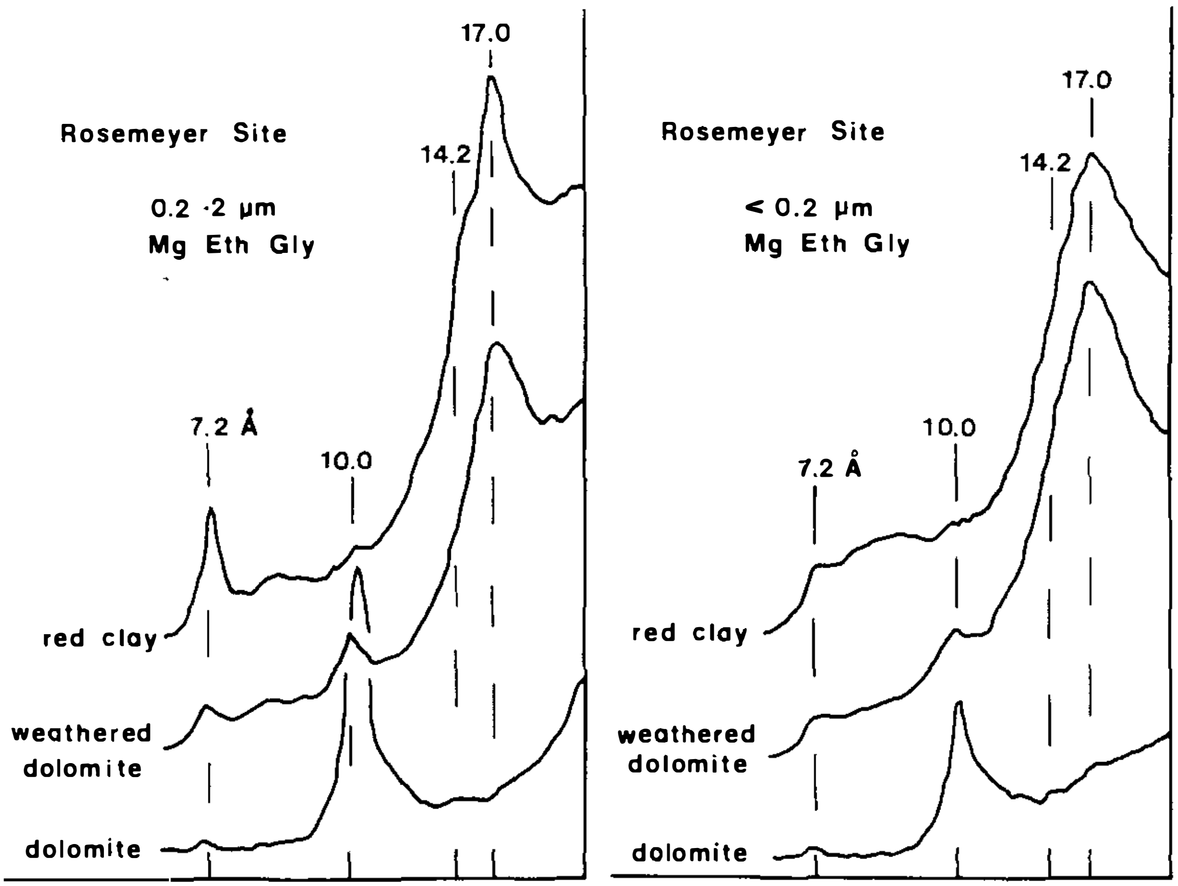

FIGURE 1--X-ray diffractograms of Mg-ethylene glycol solvated clays: (bottom) from dolomite (micaceous, 10 Å), (top) from lower red clay (mainly smectite and kaolinite, 17 Å and 7.2 Å), and (middle) oriented clay films on weathered dolomite rhombs (mainly smectite with some mica and kaolinite).

TABLE 1.--<u>Size distribution and mineralogy (scale 1 to 5) of loess, red clay and dolomite residuum</u>

	Texture				Mineralogy (<2 µm)				
Material	<0.2 µm	<2 µm	2-50 µm	>50 µm	S	V	M	K	Q
	%								
Loess	16 ±4.9(4)*	28± 4.4	70 ±2.7	1.4±1.8	4.7±0.5(6)	2.7±0.8	1.5±0.5	1.7±0.5	1.7±0.8
Red clay	70±10.4(15)	82± 9.1	14.7±8.1	5.2±4.5	4.5±0.5(24)	1.9±0.6	1.5±0.6	2.4±0.9	2.0±0.7
Dolomite	34 ±9.6(7)*	47±11.0	43 ±8.7	10.1±9.7	1.1±0.3(8)	1.1±0.3	3.6±1.2	1.0±0	4.2±0.7

*Numbers in parentheses are numbers of samples in group. Mineralogy represents an averaging of <0.2 µm & 0.2-2 µm XRD data: S=smectite, V=vermiculite, M=mica, K=kaolinite, Q=quartz (Feldspar & chlorite <5%).

TABLE 2.--<u>Ph, organic carbon, and oxalate extractable iron of red clay samples in upland profiles, near dolomite contact, and in subsurface solution tube clays</u>

Material	pH	Carbon	AAO Fe
		%	
Red clay(9)	6.0±1.0	0.27±0.11	0.26±0.11
Basal red clay(3)*	7.0±0.4	0.36±0.05	0.53±0.20
Solution tube clay(4)	7.4±0.1	0.14±0.03	0.22±0.04

*Within 20 cm of dolomite contact.

pH of upland red clays prior to the influx of the dolomitic loess can best be estimated from those profiles having a shallow loess cover, in which enrichment in Ca and Mg from loess would be minimized. Four red clay samples from two profiles having less than 40 cm of loess averaged pH 4.8. The neutrality of the basal red clays at the clay-dolomite macro-contact and the clays in solution tubes (table 2) results from ion exchange in the zone of dolomite dissolution.

Clay bodies also occur many meters below the bedrock surface and thus clays must have been carried by percolating waters through fine cracks and pores in the dolomite. The clay bodies, upon drying and wetting, filter migrating clays which, through their exchange acidity, promote further dissolution of dolomite. Even at the ambient pH of the (Ca,Mg) $(HCO_3)_2$ solution, Ca- and Mg-smectite has surface acidity (Mortland and Raman, 1968) which enhances the dissolution of dolomite. Thus dolomite dissolution is favored at the periphery of the clay pods and tubes, which therefore enlarge over time. The preservation of the linearity of chert bands extending from dolomite through red clay pods (fig. 2) requires that (i) the clay replaces dolomite on a volume for volume basis and (ii) the cohesion of the clay mass, aided by swelling pressure, furnishes support for the chert.

QUARTZ AS AN INDICATOR OF RED CLAY PROVENANCE

Quartz in dolomite is mainly authigenic microcrystalline chert and secondary low temperature vein and replacement quartz. Chert consists of aggregates of fine euhedral and subhedral quartz whereas cavities may be lined with euhedral and polyhedral microcrystalline and macrocrystalline quartz (Sayin and Jackson, 1975) with high oxygen isotopic ratios. Eolian quartz, derived from a mixed provenance, contains a significant percentage of igneous detrital quartz generally in the form of angular to rounded anhedral grains, with low oxygen isotopic ratios. Thus the characteristics of the quartz associated with the red clays should reveal whether the silt and sand fractions are largely residual from the dolomite or whether eolian and other transported additions have been significant.

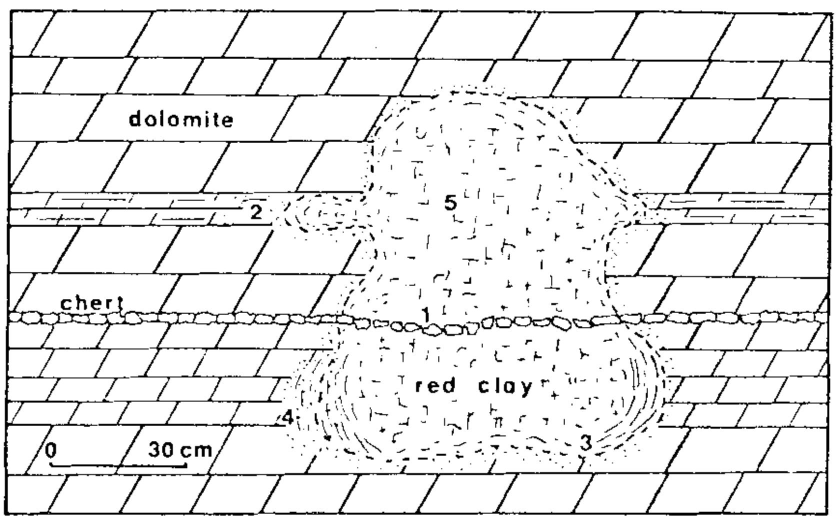

FIGURE 2--Generalized diagram of a clay pod, illustrating key features: (1) chert line extending across clay pod, (2) preferential carbonate dissolution in thinly bedded permeable strata, (3) laminar structure of clays proximal to dolomite, (4) broad dissolution zone of sandy weathered dolomite with clay films, (5) typical fine-medium angular blocky structure of red clays.

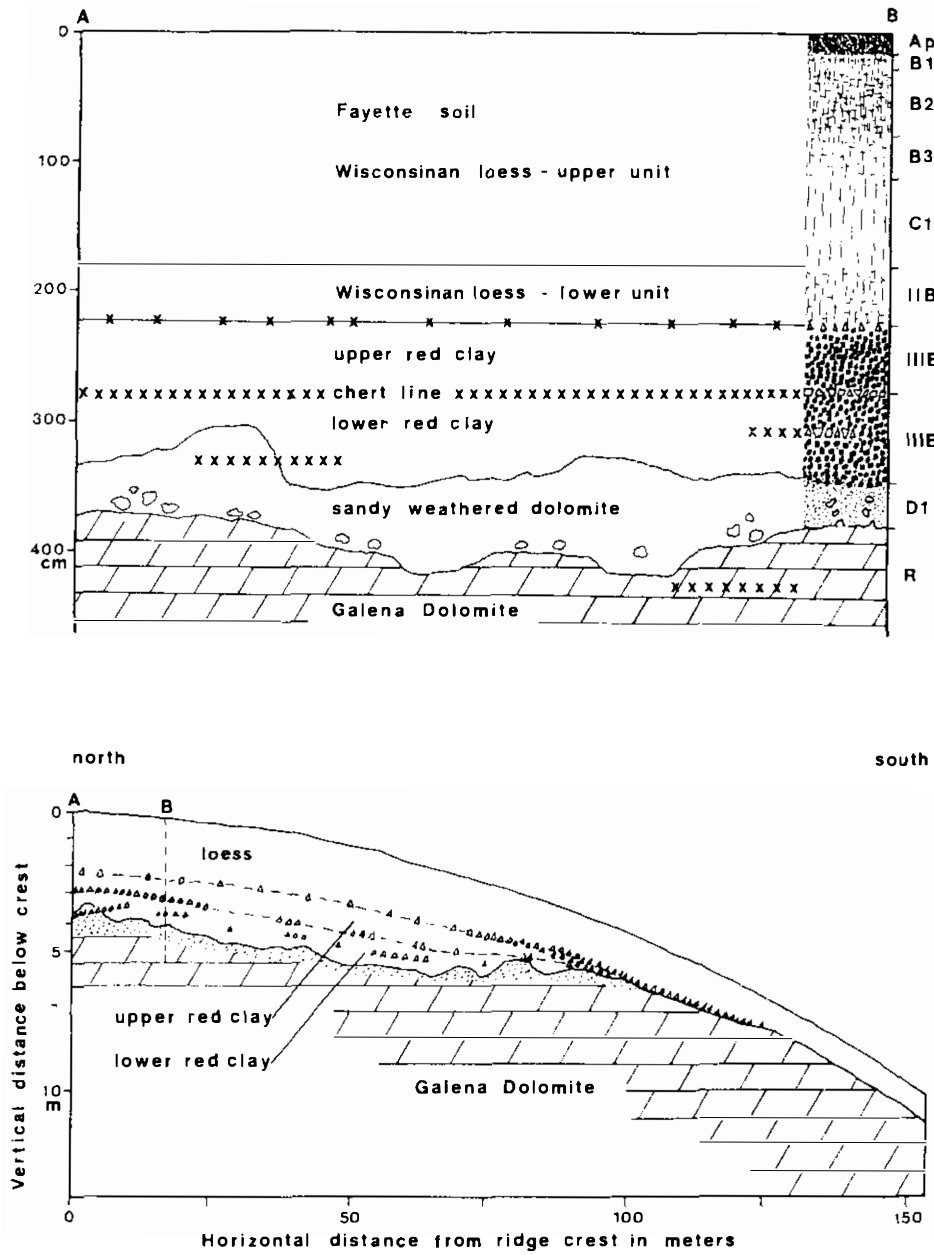

FIGURE 3.--Distribution of red clay and chert at the Rosemeyer Farm Site (modified from Knox and Maher, 1974, p. 40).

At the Rosemeyer Farm quarry, an _in situ_ chert band extends through the red clay separating an upper and lower red clay unit (fig. 3). Detailed analyses were made of the 1-10 μm quartz fraction from four samples: loess (R8); upper red clay (R15); lower red clay (R19); and dolomite (Rdol). The ratio of XRD peak heights of (100) quartz (d = 0.427 nm) and (101) quartz (d = 0.335 nm) for grains allowed to settle through an aqueous medium measures the relative concentration of euhedral grains in the sample (Eslinger and others, 1973). The I_{100}/I_{101} ratio for loessial quartz indicated randomly oriented grains, that is, a lack of euhedral crystals (table 3). The high value of the lower red clay compared to the dolomite may result from differences in detrital and authigenic quartz content or from the breakdown of chert aggregates into 1-10 μm euhedral grains in the red clay. The intermediate value for the upper red clay may indicate a mixing of dolomitic and loessial quartz. Mean I_{100}/I_{101} ratios for loess, red clay and dolomite 5-20 μm fractions revealed the same trend.

The oxygen isotope ratios ($\delta^{18}O$) of 1-10 μm quartz from dolomite and the lower red clay indicate a predominance of low temperature authigenic quartz (chert) in these samples (table 3). The mixed provenance of the loessial quartz yields an intermediate $\delta^{18}O$ value typical of midwestern loessial soils (Syers and others, 1969). The upper red clay has had significant additions of loess or of long-term aerosol dustfall.

Standardized trace element concentrations of 1-10 μm quartz show the same pattern as the I_{100}/I_{101} and oxygen isotope data (table 3). Trace element concentrations in the loessial quartz reflect a mixed provenance (that is, igneous, metamorphic, and low temperature quartz). The low trace element concentrations of the dolomitic quartz relative to the loess indicate a predominance of relatively pure low temperature authigenic quartz (chert). The low trace element concentrations of the lower red clay compares closely to the dolomite, with some weathering dissolution of amorphous cements which typically have higher trace element concentrations than the euhedral quartz crystals.

The I_{100}/I_{101}, oxygen isotope, and trace element data for 1-10 μm quartz are corroborative, suggesting the the 1-10 μm fraction of the lower red clay is residual from the dolomite. Intermediate values for the upper red clay agree with the texural data, all of which indicate admixture of loess.

The very fine and fine sand fractions revealed microscopic differences in the grain morphology of loess and dolomite insolubles (table 4). The loess contains many fractured, smooth-sided grains whereas euhedral grains and microcrystalline aggregates (chert) are prevalent in the dolomites. The quartz morphology in the red clays is very similar to that in the dolomites, indicating that the fine sand fraction is residual from the dolomite. The coarse quartz of the upper red clay from the Rosemeyer site had the highest percentage of angular-fracture grains (11 %) of the six red clay samples that were analyzed.

PROFILE CHARACTERISTICS

The very high clay content of the red clay suggests that silt, sand and coarser fractions present in the dolomite are diluted by illuvial clays as dolomite dissolution progresses. The provenance of the clay fraction is quite difficult to assess. Certainly

Sample	I_{100}/I_{101}	$\delta^{18}O$ (‰)	Trace elements* (relative mass)
Loess	0.23	18.6	2.15±0.40
Upper red clay	0.54	19.7	1.61±0.18
Lower red clay	1.02	26.9	0.78±0.10
Dolomite	0.70	26.4	1.0

*Trace elements are Sc, Co, La, Sn, Yb, Lu, Hf, Th, U.

TABLE 3.--Quartz grain shape, oxygen isotope ratio and trace element content for 1-10 μm fraction from four strata from Rosemeyer Site

Material	CBD Fe	50-250 μm Euhedral	Fractured
	%		
Loess	1.2±0.4(7)*	5.7±3.2(3)	34 ±20
Red clay	4.9±2.4(21)	16 ±4.0(6)	5.2 ±3.5
Dolomite†	8.6±4.6(8)	18 ±11 (3)	4.7 ±3.5

†Carbonate-free basis.

*Numbers in parentheses are numbers of sample in group.

TABLE 4.--CBD Fe and sand grain shapes

most of the clays present within the dolomite are incorporated into the red clay. Shale zones within the dolomite, eolian material and pedisediment could also be sources of clay. Over time, as dissolution and downwasting of the dolomite continue, clays would be recycled and mixed by the illuviation-replacement process. The results of this process are clay-rich profiles containing clays of mixed provenance. The abundance of cutans, particularly in the lower portions of red clay profiles, indicates that clays illuviate within the profile as well as within the zone of dolomite dissolution. As clays are removed from the upper solum by downward translocation and surface erosion, the coarse fractions remain at the surface as lag deposits.

The dolomite is a major source of iron for the red clay. The iron present as part of the dolomite, calculated on a carbonate-free basis (table 4), is higher than the iron content of the red clay. Finely disseminated aluminous iron oxides, derived from the oxidation of pyrite (FeS_2) and dolomite-ankerite ($(Mg,Fe)Ca(CO_3)_2$) in the presence of clay, form stable coatings on clay surfaces.

Red clay profiles lack readily identifiable horizons in terms of color, structure and clay mineralogy. No increase in kaolinite or Al-chlorite was found near the upper boundaries of profiles. The mineralogical uniformity of the profiles may result from a combination of low weathering rates due to the compact, nonporous nature of red clay peds and profile mixing over time. No remnants of A1 horizons in the red clay have been detected in the Driftless Area (Hole, 1976). This may be due to widespread stripping during the Wisconsinan or to post-burial degradation of salient A horizon features. In Europe, *terra rossa* and *terra fusca* soils typically have thin, weak mull-like moder horizons (Kubiena, 1953). The visible properties of the A horizon could soon be lost in a brightly hued soil with minimal organic incorporation that is subject to physical pedoturbation.

RED CLAY DISTRIBUTION

The patchy distribution of the upland red clays is nonrandom. Several factors, important at local to regional scales, have been put forth in explanation of their distribution. These include ridgetop breadth, slope angle, proximity to the Woodfordian ice front, proximity to loess source areas, bedrock surface morphology and availability of coarse lag deposits. For example, Black (1970) concluded that, over-all, thicker deposits are present on broad uplands as opposed to narrow ridgecrests and that red clay is generally absent on slopes greater than five percent. While the shortage of suitable exposures prevents a close examination of these various factors, available data show that bedrock surface morphlogy and the availability of chert, which are both to some extent stratigraphically controlled, outweigh the other factors.

Red clays are characteristically absent on ridgetops underlain by the upper Galena Dolomite (non-cherty unit) including several large upland tracts (Military Ridge from Mount Horeb to Montfort and the secondary divides extending south to Mineral Point and Livingston, (figure 4). Whether erosion has kept pace with soil formation through time or whether accelerated erosion during periglacial episodes in the late Pleistocene stripped these surfaces is not known. To the southwest of this area the lower cherty unit of the Galena Formation surfaces and red clays are common, generally in conjunction with jumbled lag chert. Red clays are also found in the southeastern Driftless Area (Green County) where the lower Galena Dolomite forms the uplands. Here the presence of cherty red clays on fairly narrow ridgetops close to the Woodfordian ice maximum indicates that chert armoring can prevent red clay removal even in erosion intensive areas.

Red clays appear to be most abundant in southern Monroe and Juneau Counties along the northern edge of the Driftless Area. The lower Prairie du

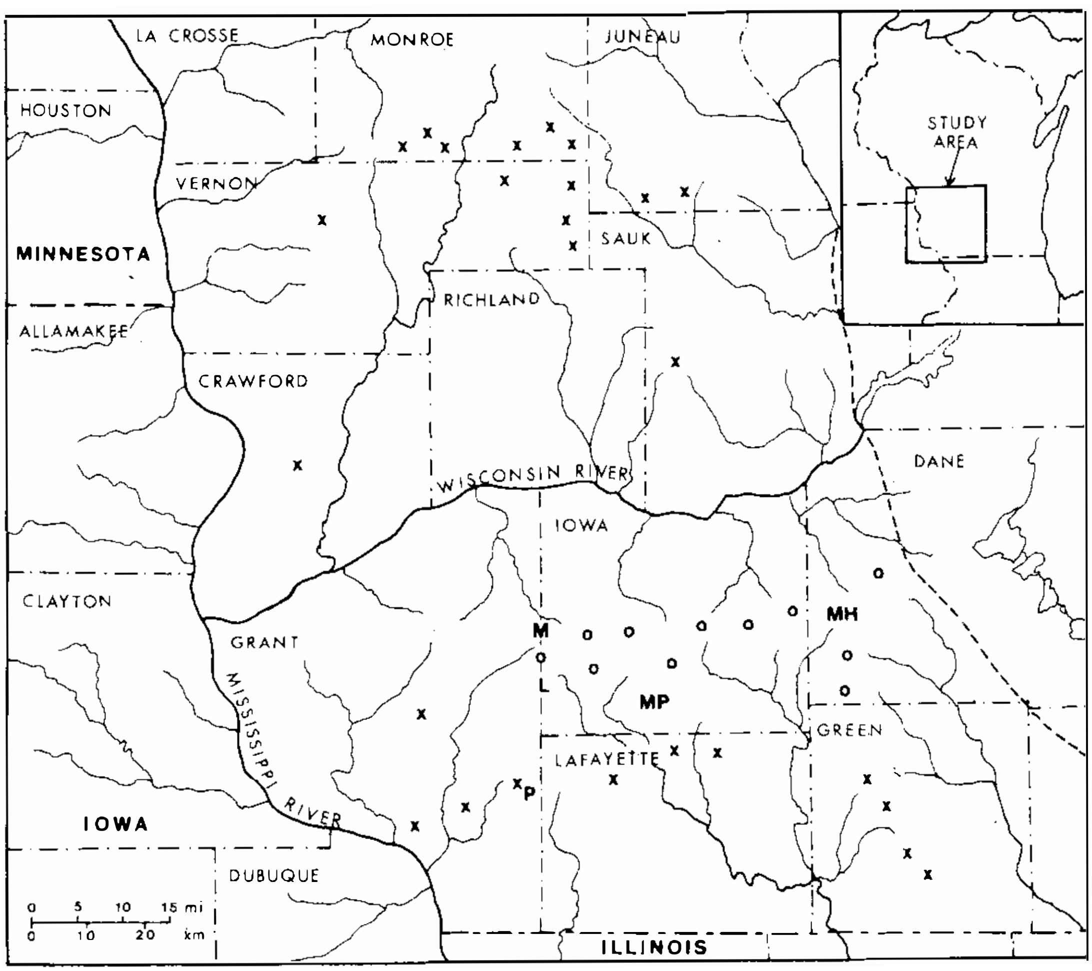

FIGURE 4.--Relative abundance of red clays at sites investigated by Frolking (1978); 'x' denotes an abundance of clays, 'o' denotes a scarcity of clays.

Chien Formation (Stockton Hill Member), which forms the uplands in this area, contains the most extensive chert and vein quartz deposits in the Ordovician sequence. The fact that this area is quite far from loessial source areas suggests that the quantity of eolian materials deposited over time bas not influenced the quantity of red clay remaining on the uplands.

Because of the lack of extensive exposures, the degree to which masses of upland clay lie protected in karst depressions is difficult to assess. A feedback mechanism exists in which the presence of a soil cover favors uneven bedrock dissolution (Sweeting, 1973), which in turn protects soil accumulations. In contrast, bare carbonate rock tends to weather evenly and thereby does not favor soil accumulation. Bedrock that underlies ridgecrests is typically more decomposed than shoulder or side slope bedrock. Therefore bedrock at shoulder positions may buttress red clay accumulations in broad ridgetop depressions. Bedrock bedding, jointing and porosity characteristics influence bedrock dissolution patterns and thus partially control the distribution of upland clays.

In southwestern Wisconsin the late Pleistocene was dominated by erosion (Knox and Johnson, 1974). The interfingerings and involutions of red clay and loess on moderate slopes (less than 10 %) and the interlayering of clays and chert masses on sandstone hillslopes attest to the predominance of mass wasting under periglacial conditions. The field identification of transported clays as opposed to *in situ* illuvial replacement clay proves difficult due to the lack of horizonation in red clay profiles and to the rapid redevelopment of structural features (Kubiena, 1953). In some areas clays are common on hillslopes of twenty to twenty-five percent. Much of this material has probably been transported. On hillslopes, the contact with the overlying loess is often quite diffuse indicating mixing during the early stages of loess deposition. The upper red clay at the Rosemeyer Farm Site probably is a mass wasted mixture of red clay and loess that became stabilized as the Peorian loessfall progressed.

SUMMARY

Smectitic reddish brown clays discontinuously mantle the dolomite uplands of southwestern Wisconsin. Their high clay content can be explained by a Beta-B type mechanism in which clays illuviate in the zone of dolomite dissolution. The mineralogical similarity of clays in this zone to those in the red clays above and the presence of clay films on dolomite grains support this mechanism. The occurrence of clay pods and undisturbed chertlines in red clay indicate that illuvial clay can replace dolomite on a volume for volume basis. Grain shape, oxygen isotope and trace element data show that the silt and sand fractions are residual from dolomite. Most red clay masses contain chert lag gravels. The distribution of red clays correlates strongly with the distribution of chert-rich dolomite units although other factors such as bedrock surface morphology are important. Interfingerings and involutions of red clay and loess suggest that mass wasting was the dominant form of hillslope erosion during the late Wisconsinan.

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SOILS OF CRAWFORD, VERNON, AND WESTERN RICHLAND COUNTIES, WISCONSIN

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The generalized map (fig. 1) of this area, based on a U.S. Geological Survey (1968) contour map of Wisconsin (1:500,000), shows major soil landscapes (A1, A5, etc.) as delineated on the state soil map (1:710,000) published in 1968 (Hole, 1976). Regional relief is more than 200 m (700 ft) and local relief diminishes up-tributary from about 170 m at bluffs of the Mississippi and Wisconsin Rivers.

The initial materials of the soils include several bedrock formations (Hole, 1976), cherty clay residuum (Frolking, 1978) from dolomites, loess, colluvium, outwash benches (terraces of valley trains), alluvium and peat. This part of the "Driftless Area" is occupied by a maturely dissected backslope of a cuesta having a regional slope southward of about 2 m (7 ft)/mi (1.2m/km) (Martin, 1965). The cuesta has, over Cambrian sandstone and siltstone strata, a cap of Ordovician Prairie du Chien dolomite on which are perched some outliers of St. Peter sandstone and Sinnipee dolomite (also Ordovician in age). The loess material is coarse silt loam in an 8-m cover near the Mississippi gorge and medium silt loam to the east, thinning to 1 m at eastern county lines (fig. 1). The loess overlies (1) cherty red clay residuum, which near Cashton is more than 2 m thick, locally; (2) sandstone; (3) old colluvium in high positions on major valley slopes; and (4) outwash terraces, particularly the most elevated ones. A few of these terraces consist of calcareous reddish brown silty clay materials, rather than the usual stratified sand and gravel. Robinson (1950) estimated ages of some Wisconsin soils developed in calcareous initial materials on the basis of depth of leaching, assuming a constant rate of leaching comparable to that known to remove agricultural limestone powder from cultivated soils currently. His estimates (not published elsewhere) are reported in figure 2 and table 1.

Some interesting soil series are briefly characterized as follows. Seaton and Fayette silt loams are forest soils (Typic Hapludalfs, fine-silty, mixed, mesic, according to the Soil Survey Staff, 1975) that constitute a west-east sequence developed in coarse and fine loess, respectively. These soils and the shallower (to bedrock) Palsgrove, Ashdale and Dubuque soils are found on slopes as steep as 45 % to 60 % with well developed argillic (clayey) B horizons, which indicates that the slopes have been stable for thousands of years. Fayette silt loam has been mapped in three landscape positions, namely upland, valley slope and bench. Tama silt loam (Typic Argiudoll, fine-silty, mixed, mesic) is the prairie equivalent of Fayette. Valton silt loam is transitional between a forest soil and a prairie soil and formed in 0.3 to 0.9 m (15 to 36 in.) of loess overlying more than two meters of cherty red clay residuum on dolo-

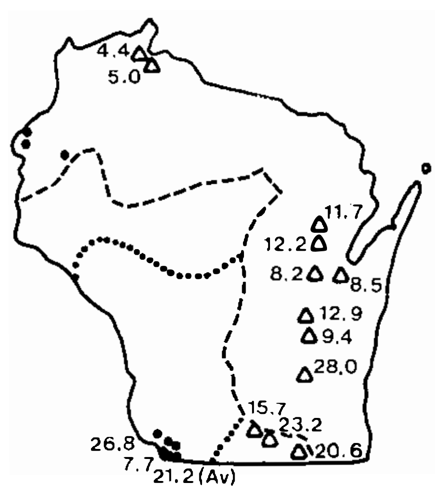
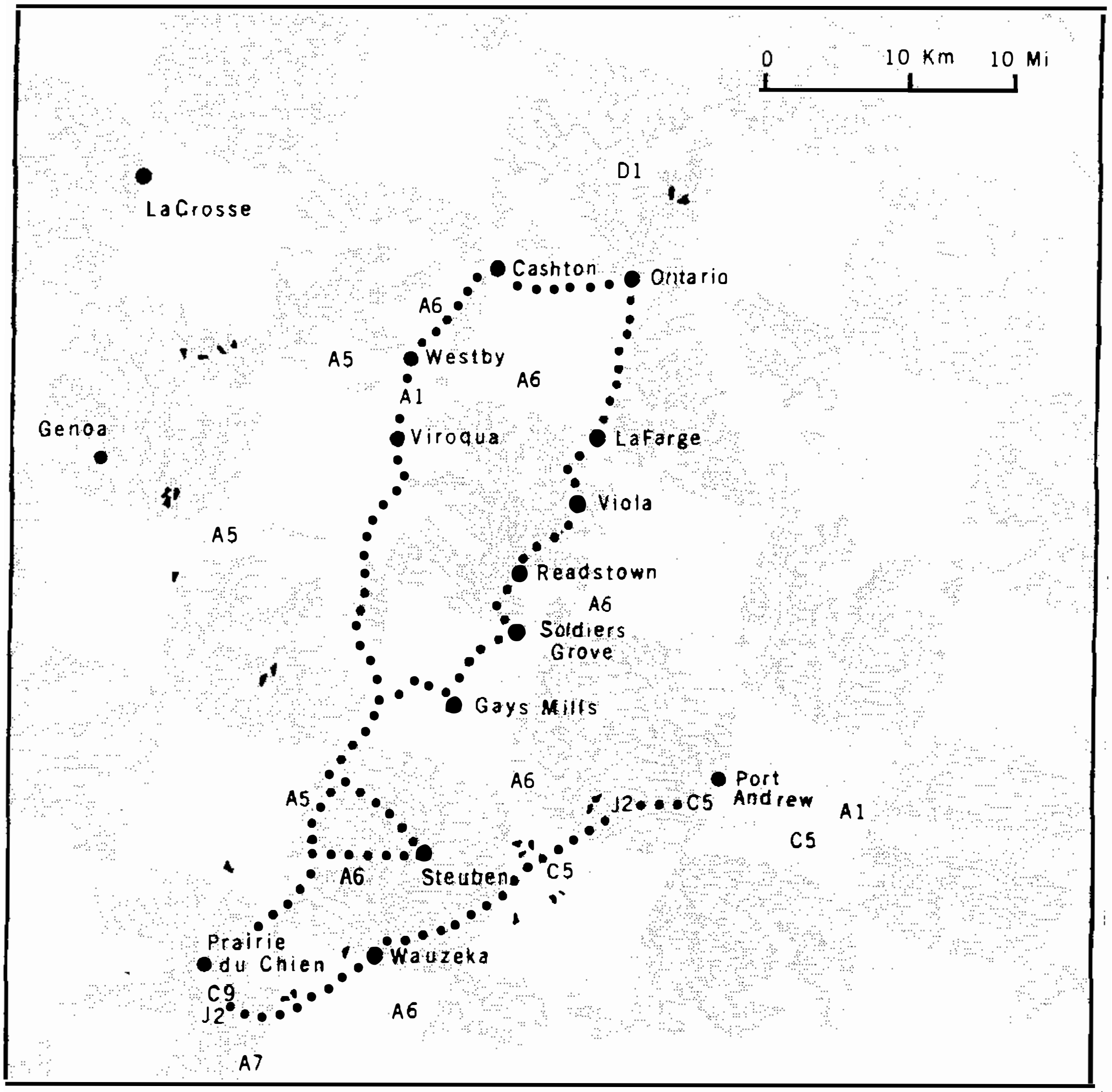


FIGURE 2.--Estimated relative ages in thousands of years of several soils formed in calcareous material (Robinson, 1950).



LEGEND

Soils

A1-Tama Silt loam and other prairie soils (Argiudolls) in deep loess over dolomite (& residuum)
A5-Seaton, Fayette, Dubuque and other forest soils (Hapludalfs), loess over residuum/dolomite
A6-Dubuque, Fayette silt loams and other forest soils (including Valton, a Paleudalf) in loess, etc.
A7-Seaton, Fayette silt loams: forest soils (Hapludalfs) in deep loess over residuum over dolomite
C5-Sparta, Plainfield and other loamy sands (Entic Hapludolls and Psamments) in sandy outwash
C9-Dakota sandy loam, Sparta sand (Mollisols), prairie soils on outwash terraces
D1-Steep rocky land with Hixton, Norden loams: forest soils (Hapludalfs) in sandstone & siltstone
J2-Wet alluvial soils, undifferentiated (with muck)
Bodies of Medary silt loam
Tour Route

FIGURE 1.--Generalized soil map of field trip area.

TABLE 1.--Soil types and locations sampled in Wisconsin and estimated and relative ages of certain Wisconsin soils based on the total amount of carbonates removed by leaching

Sample number	Soil Name	Location				Carbon-ates %	Depth of leaching inches	Est. Age	Rela-tive age**
		County	Section	Town-ship	Range				
1	Fayette si.l.	Grant	N. Center 28	2 N.	2 W.	19.75	64	20,200	4.73
2	Fayette si.l.	Grant	N. Center 20	1 N.	1 W.	17.29	96	25,900	6.07
3	Fayette si.l.	Grant	SE1/4 5	2 N.	1 W.	15.71	74	17,800	4.17
4	Fayette si.l.	Grant	Center W. Side 34	1 N.	2 W.	15.57	88	21,000	4.93
5	Tama si.l.	Grant	SW1/4 31	4 N.	3 W.	15.50	112	26,800	6.28
6	Waukesha si.l.	Rock	SW. Corner 34	3 N.	13 E.	25.61	52	23,200	5.45
7	Warsaw si.l.	Rock	SE1/4 25	4 N.	10 E.	26.15	32	15,700	3.68
8	Miami si.l.	Walworth	Center 28	1 N.	15 E.	32.42	35	20,600	4.83
9	Carrington Si.l.	Dodge	Center 7	11 N.	15 E.	25.59	67	28,000	6.57
10	Campia si.l	Barron	S. Center 26	36 N.	13 W.	7.18	88	8,800	2.06
11	Medary si.l	Grant	NE. Corner 18	2 N.	2 W.	12.00	40	7,700	1.81
12	Superior l.	Winnebago	SW. Corner 25	17 N.	16 E.	35.38	15	12,850	3.00
13	Kewaunee l.	Brown	SE. Corner 29	22 N.	20 E.	26.46	16	8,500	2.00
14	Kewaunee l.	Ontagamie	NE. Corner 34	21 N.	16 E.	24.71	16	8,230	1.94
15	Kewaunee l.	Fond du Lac	Center 28	17 N.	17 E.	27.38	16	9,400	2.20
16	Onaway l.	Shawano	W. Center 13	25 N.	17 E.	30.38	22	12,200	2.84
17	Onaway loam	Shawano	S. Center 8	25 N.	18 E.	30.47	21	11,650	2.73
18	Ontonagon l.	Ashland	SE1/4 27	47 N.	4 W.	10.86	26	4,940	1.16
19	Hibbing l.	Bayfield	NE1/4 25	48 N.	5 W.	9.50	27	4,380	1.02
20	Cushing l.	Burnett	Center 8	37 N.	18 W.	9.01	35	4,270	1.00
21	Cushing l.	Burnett	Center 26	38 N.	19 W.	7.06	48	4,480	1.04

**Calculated on the basis of Cushing loam = 1.00 and assuming equal rates of leaching for all soils.

mite. The assumed great age of the thick red clay layer has been used to justify classifying this as a paleosol (Mollic Paleudalf, fine-silty, mixed, mesic). The Rockbridge silt loam is a rare soil surviving on valley walls at a few sites. It is a forest soil formed in 0.4 to 0.8 m (15 to 30 in.) of loess over reddish clayey cherty gravelly colluvium. This is probably a paleosol, but is not so designated (Typic Hapludalf, fine-loamy, mixed, mesic). Medary silt loam is found on valley benches (terraces). It is a forest soil (Typic Hapludalf, fine, mixed, mesic), that has formed in 0.4 to 0.8 m (15 to 30 in.) of loess over slack-water-deposited reddish silt and clay that is calcareous at a depth of 0.9 m (3 ft). Robinson (1950) estimated the age of one Medary soil profile to be 7700 years (fig. 2, table 1). Madenford (1974) suggested 2460 B.P. as the date of deposition of the red silt and clay, which he proposed was washed from red till of the Superior Lobe into glacial Lake Aitkin at the headwaters of the Mississippi River, from which it was washed during dissection initiated by bedrock rebound. The Arenzville, Orion, Ettrick and alluvial land soil bodies are in alluvium at lowest valley positions, below the bench soils (Bertrand, Jackson, Curran, Richwood, Toddville, Rowley silt loams; Dakota sandy loam; Sparta and Plainfield loamy sands). All of these alluvial soils bury a sedimentary-pedological record in deep valley fill.

Soil landscape patterns displayed on the photomosaic sheets of the soil survey reports for Crawford, Grant and Vernon Counties (Slota and Garvey, 1961; Robinson and Klingelhoets, 1961; Slota, 1969) roughly parallel contour lines on uplands and follow drainage lines in valleys. Detailed as these modern soil survey maps are, they do not attempt to show the ultimate detail needed for research purposes, such as differentiation between phases of Fayette silt loam in nose, hollow, footslope and bench positions. Considerable acreages of steep stony land and of alluvial land contain complexes of soils that were understandably not mapped in detail. Studies and surveys to date have only begun to reveal the vast amount of information contained in the soils of this area.

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PRE-WISCONSINAN DEPOSITS IN THE BRIDGEPORT TERRACE OF THE LOWER WISCONSIN RIVER VALLEY

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INTRODUCTION

The lower Wisconsin River extends across the Driftless Area from the late Wisconsinan terminal moraine in south-central Wisconsin to its junction with the Mississippi River. A prominent feature in the lower valley is the Bridgeport terrace, a landform containing deposits that have been used to support various interpretations of the Quaternary history of the region (fig. 1, fig. 2). The Bridgeport terrace, named for a small community near the mouth of the Wisconsin River, has long been known to contain pre-Wisconsinan glacial deposits, but there has been disagreement on the source of the sediments and on whether they are till or outwash. The present study is an attempt to resolve these issues.

Prior to the present study, no detailed examination of Bridgeport ter-

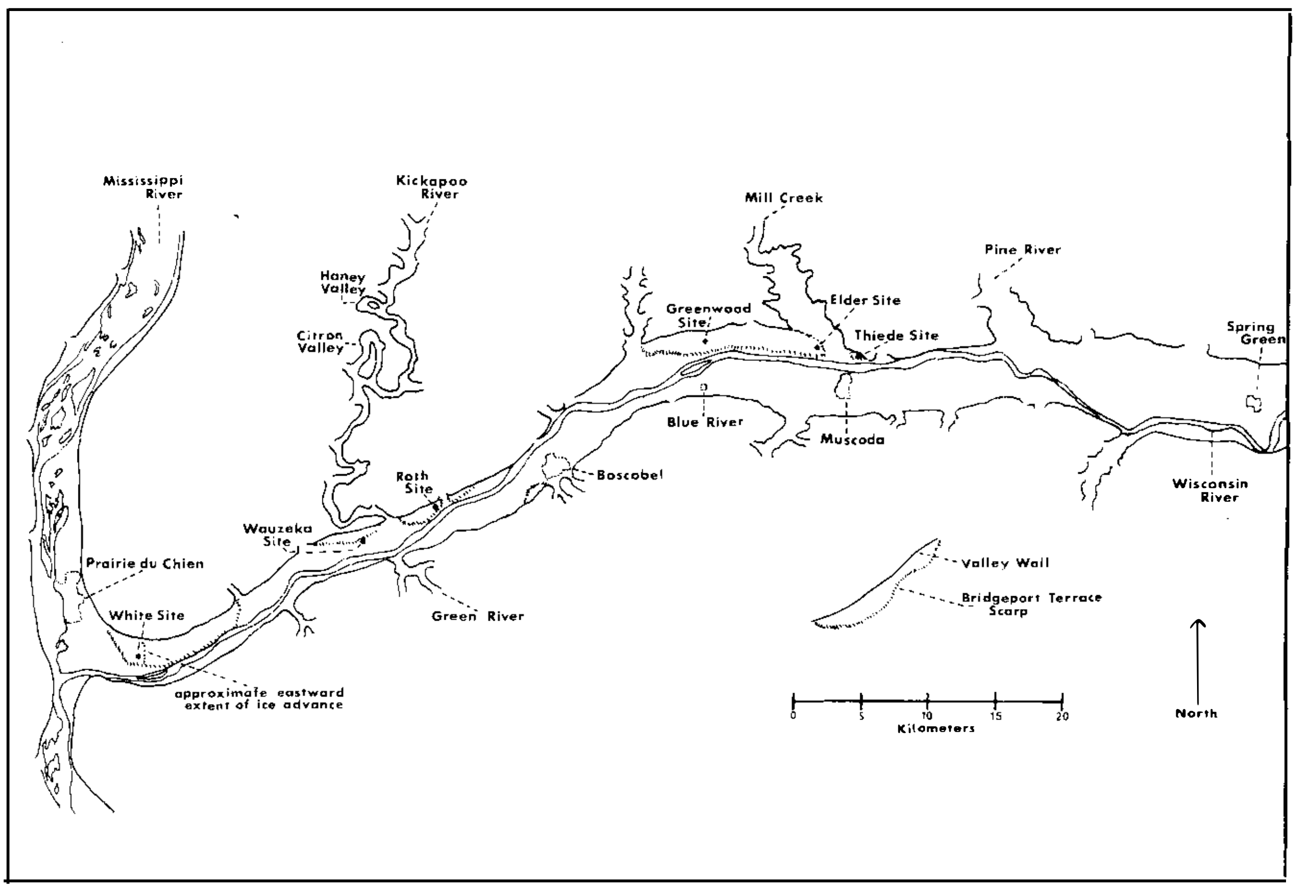

FIGURE 1.--Distribution of the Bridgeport terrace in the lower Wisconsin River valley.

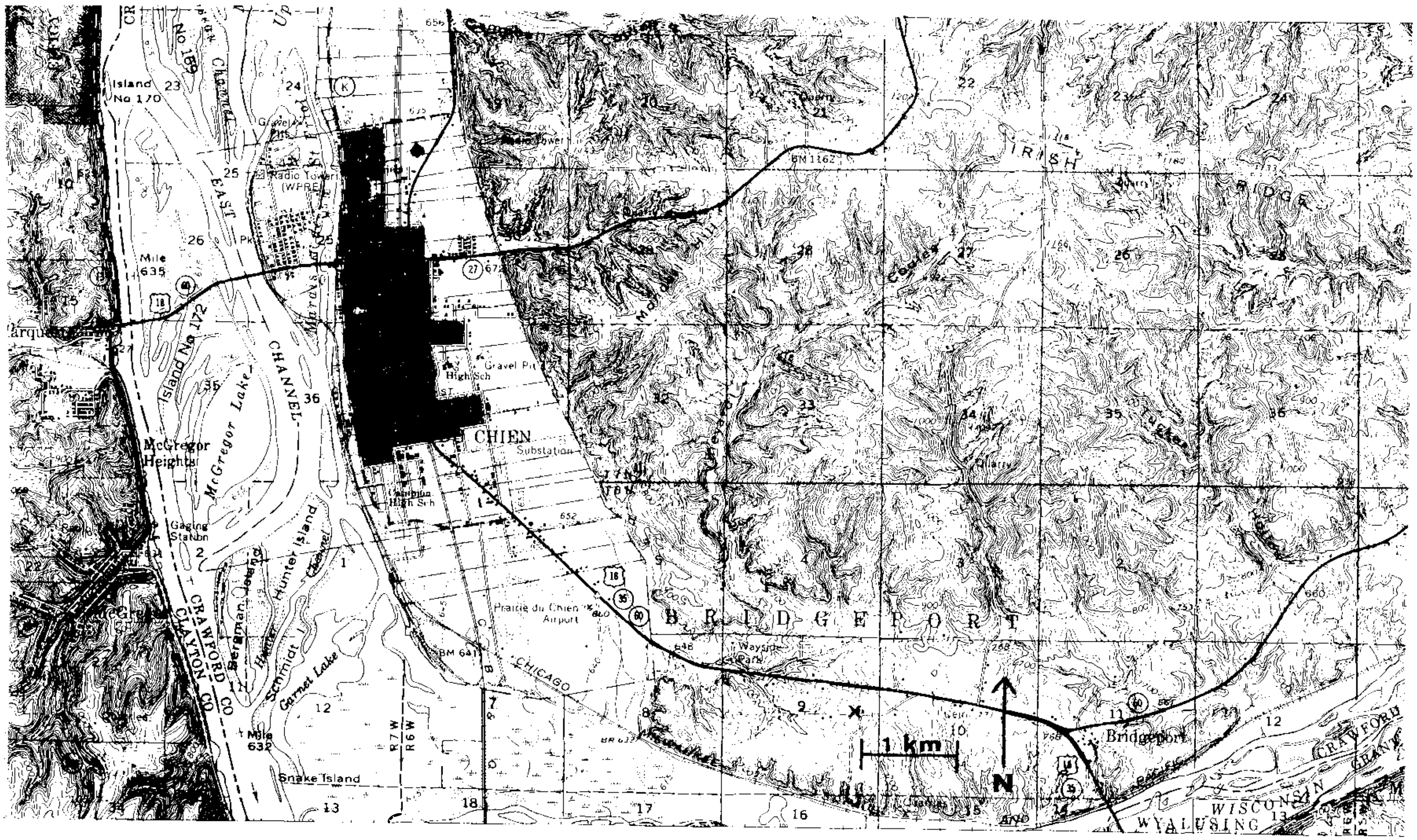

FIGURE 2.--The X near the center of section 9, Bridgeport Township, shows the location of the White Site drill hole where glacial till was encountered at a depth of about 2 m. We tentatively recognize the low ridge that extends from north to south across the west half of adjacent section 10 as the terminal moraine of an ice lobe that extended into the lower Wisconsin River valley. Clay mineralogy of the till indicates that it is very similar to the classical Kansan till of northeastern Iowa.

race deposits had been conducted since the early twentieth century when Alden (1918), MacClintock (1922), and Thwaites (1928) examined various sections of the valley. In comparison to these early investigations, this study had the advantage of a drill rig, aerial photography, and topographic maps. The interpretations and conclusions in the present study are based principally on stratigraphic sequences identified in five drill holes systematically spaced on the Bridgeport terrace in the lower-most 60 km of the valley. The sites of the drill holes are identified on figure 1 as White, Roth, Greenwood, Elder, and Thiede. The Wauzeka Site, also shown on figure 1, is a surface exposure.

THE BRIDGEPORT TERRACE

The longitudinal gradient of modern topography on the Bridgeport terrace slopes very gently eastward in sharp contrast to the westward and relatively steep gradient of the present Wisconsin River (fig. 3). The very irregular longitudinal profile of the modern surface is related to uneven loess distribution and to the highly dissected character of the Bridgeport surface. Local relief commonly reaches 30 m on the terrace remnants. Sediments encountered during drilling and in surface exposures indicate that the Bridgeport surface in most places overlies a clayey, sandy gravel that we

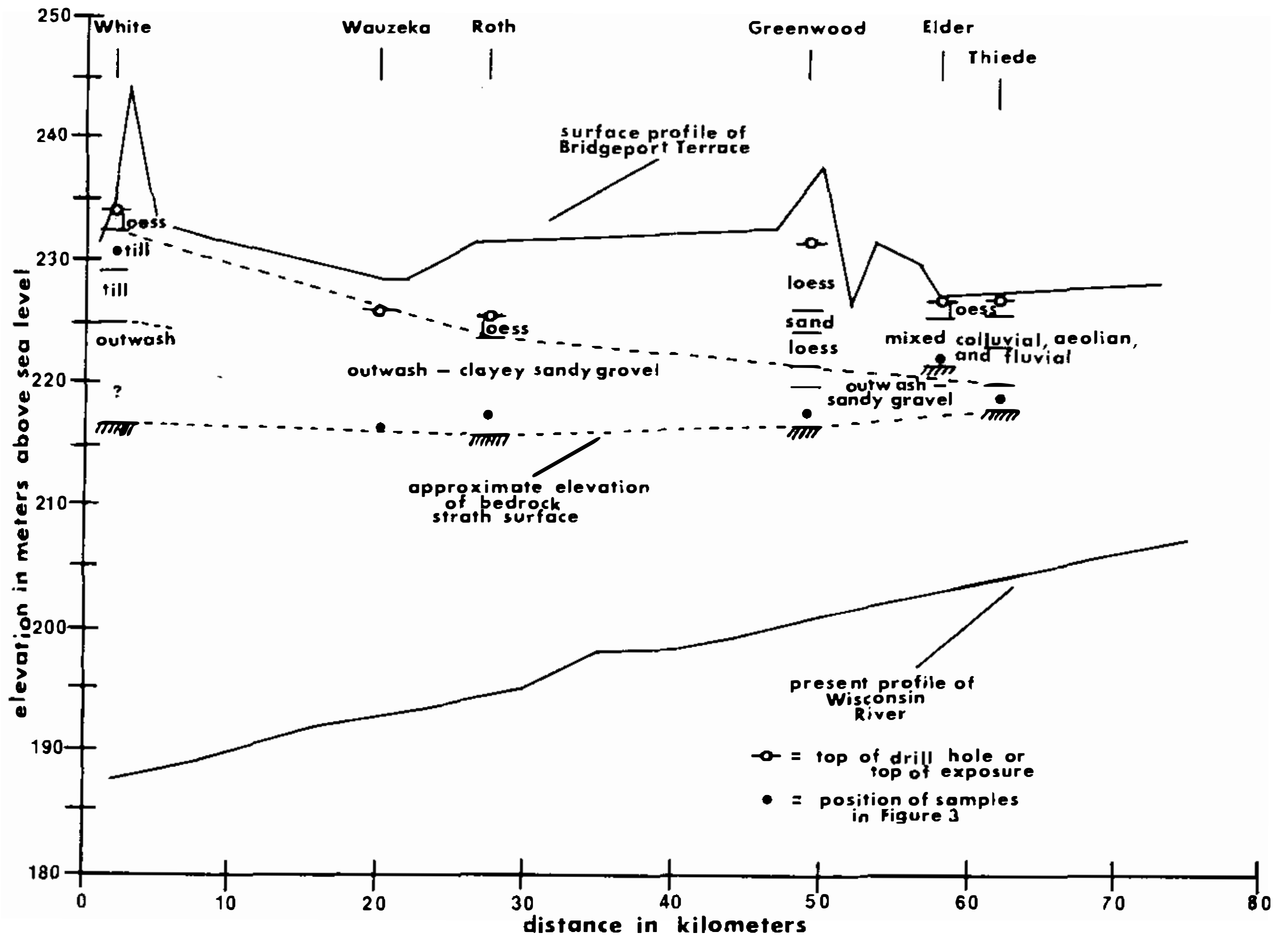

FIGURE 3.--Longitudinal topographic and stratigraphic profiles. Clayey sandy gravels that represent outwash from the ice lobe that blocked the mouth of the Wisconsin River indicate that river flow was temporarily reversed at that time. Sediments overlying the western-derived outwash between the Greenwood and Thiede Sites include eastern-derived outwash sediments that also appear to be of pre-Wisconsinan age.

interpret as outwash. The topographic gradient on the outwash is eastward and is much steeper than the eastward gradient on the modern Bridgeport surface (fig. 3). The outwash gravels rest directly on a strath surface cut into dolomite and sandstone. The longitudinal gradient of the strath has not been precisely determined, but elevations estimated from topographic maps indicate that it is close to horizontal (fig. 3).

PRESENT HYPOTHESES AND PRIOR INVESTIGATIONS

Deposits in the Bridgeport terrace have been a source of controversy regarding the Quaternary history of the lower Wisconsin River. We present evidence in this paper to support our hypotheses that a pre-Wisconsinan ice lobe advanced eastward approximately 3 to 4 km into the lower Wisconsin valley and caused temporary reversal in the flow of the Wisconsin River. These hypotheses are supported by our recognition of glacial till on the Bridgeport terrace near the mouth of the valley and on our interpretation that the sandy gravels resting directly on the dolomite and sandstone strath (fig. 3) are outwash. We also hypothesize that in the mid-section of the valley, near Blue River and Muscoda, the terrace contains outwash derived from the east. Since the outwash derived from the east occurs stratigraphically above the Bridgeport outwash derived from the west, it is younger than the Bridgeport deposits of the lower valley.

Alden (1918, p. 172) was one of the first to suggest the presence of till when he indicated that a large quartzite boulder (0.3 m x 0.6 m x 0.9 m) resting on the Bridgeport surface about 3 km upstream from the valley mouth was ...deposited by an ice front crossing the Mississippi Valley from the west and thrusting itself into the lower part of the Wisconsin Valley. A few years later, Thwaites (1928) argued that the quartzite pebbles and boulders in the Bridgeport terrace at the mouth of the valley were derived from upstream in the present Wisconsin valley near Baraboo. Thwaites (1928, p. 638-639) presented a series of arguments against recognition of till and against reversal of the Wisconsin River. Two key points were Thwaites' difficulty in finding an eastern outlet that would not involve ponding and the formation of delta deposits and his belief that the highest elevations of the Bridgeport terrace occurring between the mouth of the Wisconsin River valley and Wauzeka (fig. 3) could be explained by the ...superior resistance of the bedrock bench on which they rest. Our data suggest that Thwaites was incorrect in his assessments.

The most detailed of the early studies concerning Bridgeport terrace deposits was that by MacClintock (1922). MacClintock also suggested the existence of glacial till in the terrace at the mouth of the valley. He hypothesized that a tongue of ice may have projected into the lower Wisconsin valley a distance of about 6 km. However, he believed that the Bridgeport surface was underlain by two pre-Wisconsinan deposits which had different source regions. He agreed with Alden (1918) that the pre-Wisconsin deposits down-valley of Wauzeka were derived from the west, but he concluded that deposits in the Bridgeport terrace on the north side of the Wisconsin River opposite Blue River and Muscoda were derived from the east (fig. 1, fig. 4). In the following discussion we present evidence that supports MacClintock's interpretation of two outwash deposits, but our data suggest that MacClintock's relative age assignments were incorrect.

OBJECTIVES AND PROCEDURES

The prior studies of the Bridgeport deposits suggested three critical issues. The first is whether glacial till exists in the lower valley; the second concerns possible flow reversal of the Wisconsin River in response to ice blockage at the river mouth; and the third concerns the existence of one

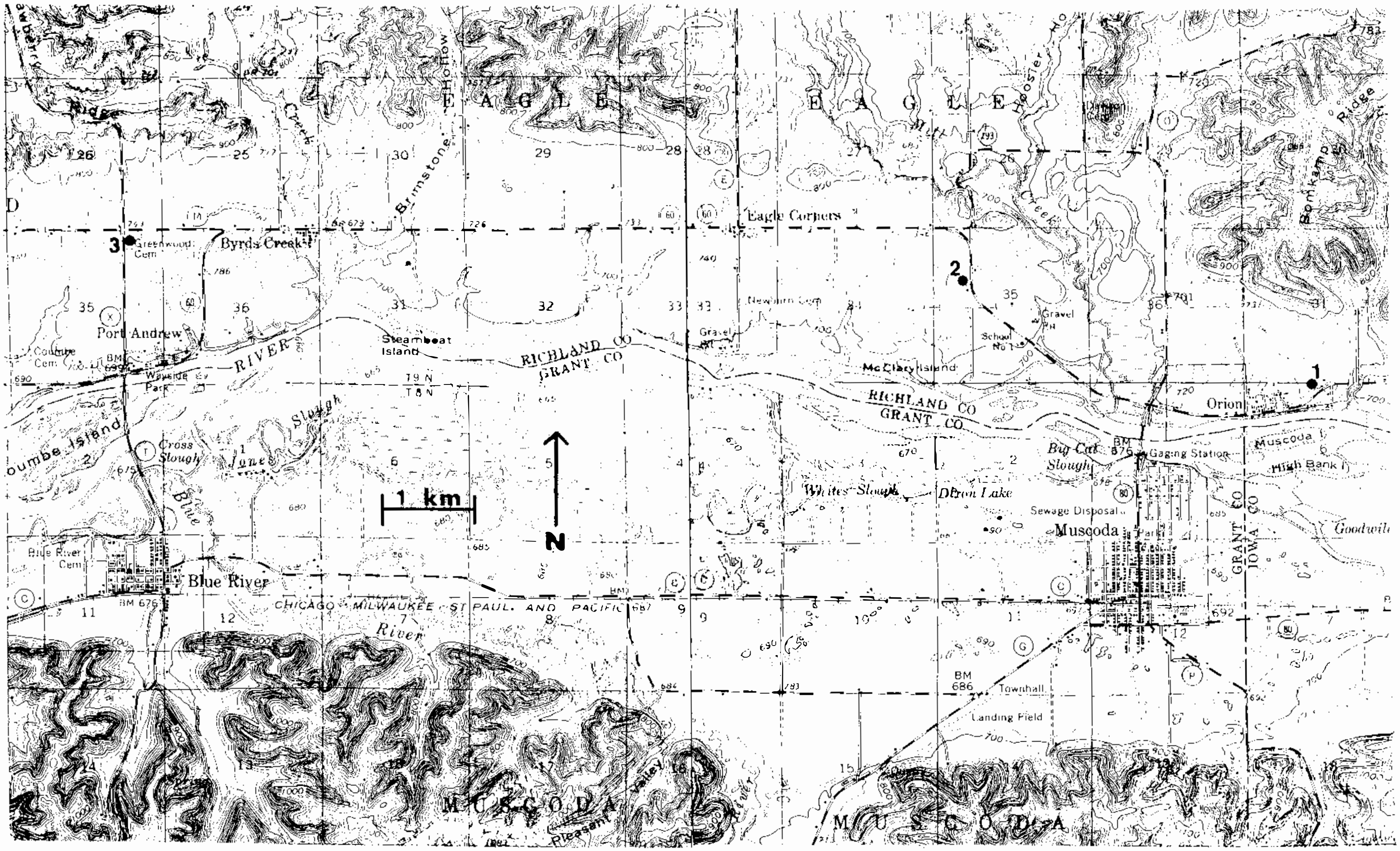

FIGURE 4.--The large dots numbered 1, 2, and 3 respectively represent the Thiede, Elder, and Greenwood drill sites. Note the difference in topography on the Bridgeport terrace north of Port Andrew in comparison to that east of Byrds Creek. The greater relief and dissection on the western end of the terrace remnant appear to be related to greater thicknesses of loess there compared to the eastern end where loess is relatively thin.

or two pre-Wisconsinan outwash deposits represented in the Bridgeport deposits. Five drilling sites were located to maximize the effects of down-valley differences in the sedimentological properties of the Bridgeport deposits in light of the above possibilities. Drilling was conducted with a truck-mounted Mobil-50 rig, using 10-cm diameter screw augers and sampling from the augers at 30 cm intervals. The procedure involved drilling in 1.5-m sequences, whereby each drilling of 1.5 m into new sediments would be followed by retrieval without rotation of the entire auger-string so that sampling could be conducted on the bottom auger-flight. After sampling, the augers were cleaned, reassembled, and drilling proceeded downward into the next 1.5 m of sediment.

Brief descriptions of the sedimentology were taken during drilling, but most analyses were deferred to the laboratory. At the time of this writing not all samples have been analyzed. Summary statistics of results to date are given in appendix I for samples that have been analyzed in detail. The grain size distribution of the less than 2 mm fraction was determined using standard sieve and hydrometer procedures. The lithology of the very coarse sand fraction (1 to 2 mm) was determined by identifying at least 150 grains from each sample. A quarter-phi interval particle size analysis of the sand fraction (0.06 to 2 mm) was conducted on several samples to determine the degree of sorting. Silt and clay were removed from the sand fraction by wet sieving. The sand fraction was dried and sieved on a sonic-sifter.

ICE-CONTACT DEPOSITS NEAR THE VALLEY MOUTH

The drill hole at the White Site about 3 km upstream from the mouth of the Wisconsin River was established to determine whether ice-contact deposits occur beneath the Wisconsin loess at the mouth of the valley. Our results indicate that a sandy loam till at least 6.4 m thick occurs at the site (appendix I). The characteristics of the till allow it to be separated into an upper and lower portion. The drilling was not to bedrock, which may be another 10 m below the base of the hole. The sandy character of the till may reflect the influence of outwash in the adjacent Mississippi valley that was presumably overriden by the ice that deposited the till. The average sand:silt:clay ratios in the upper and lower portions of the till are respectively 50:30:20 and 51:24:25. A clayey material (35 to 45 % clay) about 1.5 m thick separates the upper and lower portions. Quartz grains in the sand fraction are predominantly angular. The ratio of angular to rounded quartz grains at the White Site tends to be much greater than typical of those for other sites in the valley (appendix I).

The crystalline erratics observed during drilling were largely of pebble and granule size. The small sizes probably reflect the bias of sampling from augers. We have observed erratics in surface exposures that include small boulders of quartzite, and Alden (1918, p. 172) observed in the same area a quartzite boulder that was 60-cm diameter on the intermediate axis. The common occurrence of coarse clastics in a matrix that is about 50 % silt and clay indicates that fluvial processes are unlikely as a mechanism for deposition. Pebbles in the unit mapped as till contain striations (Alden, 1918; MacClintock, 1922).

We have tentatively assigned the low ridge on the terrace surface in the west one-half of section 10, Bridgeport Township as the terminal moraine for an ice lobe extending into the lower Wisconsin River valley a distance of about 3 km (fig. 2). A deeply incised gully just to the east of the ridge in section 10 shows fluvial sand and gravel that apparently represents outwash from the terminal moraine. Since we have not drilled on all of the high surfaces of the terrace near Bridgeport, it is possible that till may exist slightly further eastward. MacClintock suggested that ice may have extended about 6 km into the valley, but we find no evidence to support his conclusion. The high elevation of the surface we recognize as the terminal moraine has no counterpart elsewhere on the Bridgeport terrace in the lower valley (fig. 3).

The stratigraphic sequence at and near the White Site at the mouth of the Wisconsin River valley indicates that the Bridgeport till was associated with an ice lobe that advanced from the west. To test for correspondence between the Bridgeport till and tills of northeastern Iowa, five samples were sent to the Iowa Geological Survey clay-mineralogy laboratory. The semiquantitative calculations from the analyses conducted at the Iowa Survey lab showed that the clay mineralogy averaged 67 % expandables, 16 % illite, and 17 % kaolinite plus chlorite (table 1). These results agree closely with average values for clay mineralogy of the Iowa Wolf Creek Formation that averages 62 % expandables, 17 % illite, and 21 % kaolinite plus chlorite (Hallberg, 1980, p. 15).

Pre-Illinoian Pleistocene deposits in northeastern Iowa have classically been recognized as Kansan and Nebraskan. Hallberg (1980) redefined the Kansan and Nebraskan of east-central Iowa as undifferentiated pre-Illinoian stages for purposes of time-stratigraphic classification. On the basis of rock stratigraphic units, Hallberg subdivided the pre-Illinoian deposits into the Alburnett and Wolf Creek Formations. Each of the formations was recognized as including multiple tills, but the clay mineralogies within formations were relatively similar. Hallberg reported that the older Alburnett Formation averages about 43 % expandable clay minerals whereas the younger Wolf Creek Formation averages 62 % expandable clay minerals.

The clay mineralogy of the Wolf Creek Formation is broadly similar to the clay mineralogy of the Hersey till located in Wisconsin on the northwestern margin of the Driftless Area (Baker and Simpson, 1981). A comparison of average values for clay minerals in the two units is given in table 2. Baker (personal communication, 1982) has found that remnant magnetism in lower horizons of the Hersey till and associated lake deposits has reversed polarity, indicating that the reversal may represent the Brunhes/Matuyama boundary at about 730,000 B.P. If these correlations are correct, they imply that the Bridgeport till is of classical Kansan age as suggested by MacClintock (1922) and Trowbridge (1954).

REVERSED FLOWAGE OF THE WISCONSIN RIVER

Surface exposures and drill hole stratigraphy were used to map the gradient of outwash eastward from the till at the White Site. The upper boundary of the outwash of clayey sandy gravel that includes abundant crystalline erratics forms an abrupt contact with overlying silty and clayey younger deposits. The eastward gradient on the surface of the Bridgeport outwash between the Wauzeka and Thiede Sites is approximately 0.15 m/km, a magnitude that is approximately one-half of the westward gradient of the present Wisconsin River between the same two sites (fig. 3). We were unable to determine the significance of any former isostatic adjustments on the gradient of the outwash. Given the relationship of the eastward slope of the outwash to the relatively flat surface on the dolomite and sandstone strath below it, we conclude that the Wisconsin River experienced reversed flowage during the period when glacial ice blocked the mouth of the valley near Prairie du Chien.

Particle size characteristics of the Bridgeport outwash sands also support the interpretation of eastward flowage. Figure 5 presents graphic moment statistics (Inman, 1952) to quantitatively document change in particle size characteristics with distance. The sand sample from till at

TABLE 1.--*Clay mineralogy of the Bridgeport till* *

	EX	ILL	K+C
White Site, 3.2 m depth	63 %	20 %	17 %
White Site, 3.5 m depth	65 %	17 %	18 %
White Site, 5.0 m depth	72 %	12 %	16 %
White Site, 6.3 m depth	68 %	15 %	17 %
White Site, 9.9 m depth	64 %	18 %	18 %
mean	67 %	16 %	17 %
standard deviation	3.7 %	3.1 %	0.8 %

*These samples were processed by the Iowa Geological Survey through the courtesy of G.R. Hallberg, Chief, Geological Studies Division. A description of procedures is given in Hallberg (1978). Two additional samples were processed at the University of Wisconsin, Madison. We found somewhat higher percentages of illite and lower percentages of expandables and kaolinite plus chlorite, but laboratory procedures were different from those at the Iowa laboratory.

TABLE 2.--*Average clay mineralogy of tills to the west and northwest of the Driftless Area*

Hersey till*		Wolf Creek Formation **	
montmorillonite	50-60 %	expandables	62 %
kaolinite	25-30 %	illite	17 %
mica	15 %	kaolinite + chlorite	21 %
quartz	5-10 %		
chlorite	0-5 %		

* Data provided by R.W. Baker, University of Wisconsin, River Falls. The Hersey till percentages represent average values of samples from the C soil horizon.

** Data from Hallberg (1980, p. 15).

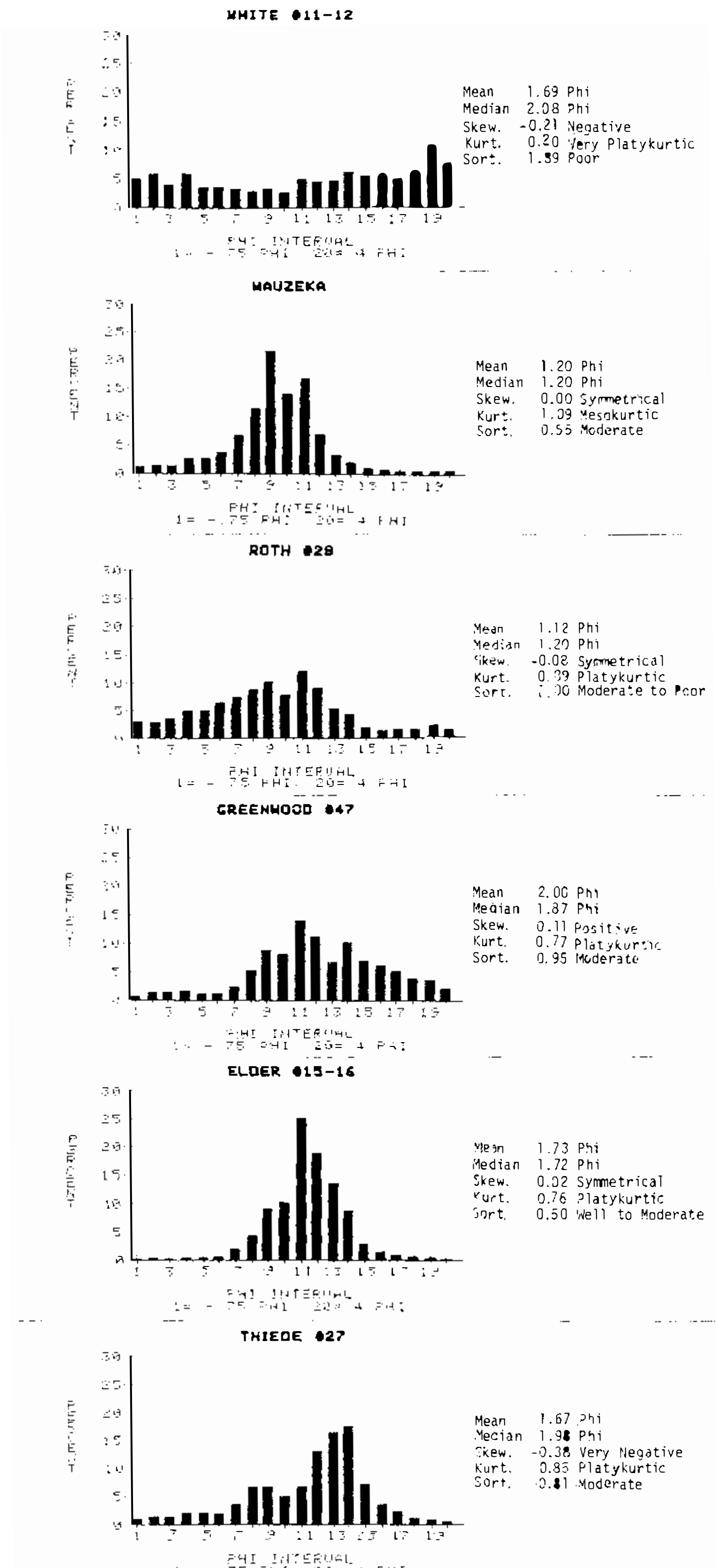

FIGURE 5.--Quarter-phi interval histograms of sand fractions from basal sediments underlying the Bridgeport surface. The White Site represents glacial till and the other five sites represent outwash. The Elder Site sands are from an eastern-derived outwash, but the other four histograms represent sands from the western-derived outwash. Note that the western-derived sands become finer in an eastward direction.

the White Site is poorly sorted in comparison to the fluvially deposited sands at other sites in the valley. The other histograms represented on figure 5 describe fluvial outwash sediments sampled from basal positions of stratigraphic sequences overlying bedrock. The samples from Wauzeka, Roth, Greenwood, and Thiede represent former bedload sediments from outwash that flowed eastward from the glacial till at the White Site. The Elder Site, which is located between the Greenwood and Thiede Sites (fig. 1), is underlain predominantly by outwash that was carried westward from central Wisconsin. Although the Elder sample also is stratigraphically close to the underlying bedrock surface, its relative elevation is higher than the adjacent outwash from the till near Bridgeport at the mouth of the Wisconsin River valley (fig. 3).

Comparison of the graphic moments statistics for the Wauzeka, Roth, Greenwood, and Thiede Sites shows that the median particle size tends to become finer and that fine sands tend to constitute a greater proportion of the total sand fraction with distance eastward from the White Site. The younger outwash from an eastern source at the Elder Site tends to be better sorted and slightly coarser than the underlying older western derived outwash at the Greenwood and Thiede Sites (fig. 5).

Further support for temporary reversal of the Wisconsin River when the valley mouth was blocked by glacial ice is suggested by bedding in the outwash sediments. MacClintock (1922, p. 676) observed cross-bedding in sandy layers that dipped eastward at Wauzeka. We also observed poorly-developed eastward dipping bedding in Bridgeport outwash deposits directly overlying bedrock in a gully wall located near the center of section 36, Eagle Township, about 3 km northeast of Blue River. The gully exposure is approximately 1.5 km southeast of the Greenwood Site (fig. 1).

POSSIBILITY OF TWO PRE-WISCONSINAN OUTWASH DEPOSITS

The stratigraphy of the Bridgeport terrace along the north side of the Wisconsin River near Blue River and Muscoda suggest the possibility of two pre-Wisconsinan outwash deposits. Sands and clayey cherty sands with crystalline erratics are sandwiched between overlying Wisconsinan loess and the underlying eastward-sloping outwash derived from the mouth of the Wisconsin River valley (fig. 3). The stratigraphy of the intermediate unit is complex and appears to include elements of colluvial, aeolian, and fluvial sediments. A prominent characteristic is abundance of angular chert fragments of various sizes. By comparison, the underlying western derived outwash has relatively little chert (appendix I).

The particle size properties of the sand fraction in the eastern-derived outwash has both similarities and differences with the sand fraction in the western-derived outwash. There is a tendency for the modal fraction to cluster around 1.50 to 1.75 phi (medium sand) in both, although the basal sand at the Thiede Site has a mode of 2.25 phi (fine sand). The most distinctive difference between the two units occurs in the coarse sand fraction. The western-derived outwash has greater quantities of coarse sand that is dominated by quartz. Coarse sand in the eastern-derived outwash has little quartz and it is dominated by chert fragments. The eastern sands are slightly better sorted than the western sands.

The deposits of eastern-derived Bridgeport sediments differed somewhat between drill holes at the Thiede, Elder, and Greenwood Sites. At the Thiede Site the sands contained considerable quantities of clay and chert and most chert pebbles and granules were very angular as previously noted. The abundance of chert and clay appears to be derived from local colluvial and fluvial contributions. The Thiede Site is only 0.7 km from the base of a high steep valley wall and tributary entrance (figs. 1 and 4). At the Elder Site, Wisconsinan loess overlies relatively clean outwash sand. The sand includes crystalline erratics and chert pebbles but is not characterized by the quantity of clay and chert typical of the Thiede Site. Further down valley at the Greenwood Site a very well-sorted and leached sand with chert granules occurs beneath the calcareous Wisconsinan loess and a paleosol (appendix I). We believe the Greenwood sand is correlative with the eastern-derived outwash sands at the Thiede and Elder Sites because the sedimentological characteristics are similar. The sandy loess that underlies the Greenwood sand may be contemporaneous with the eastern sediments, but it may also be related to a late phase of the basal Bridgeport outwash from the west.

MacClintock (1922), working primarily with surface exposures, also suggested that Bridgeport terrace deposits near Blue River and Muscoda were different in age from Bridgeport deposits in the lower valley. Because he found more carbonate clasts in deposits of the lower valley relative to the midcourse valley, MacClintock concluded that deposits near Blue River and Muscoda were oldest. However, because the lower valley Bridgeport deposits pass under the midcourse Bridgeport deposits thought to be oldest by MacClintock, his relative age assignments were incorrect. The differences in content of carbonate clasts between the two units probably can be explained by source regions of the sediments. The western source area for the basal eastward-dipping outwash includes southeastern Minnesota and northeastern Iowa where carbonate bedrock occurs commonly at the surface. We believe the source region for the pre-Wisconsinan Bridgeport sediments, that overlie the eastward-dipping outwash near Blue River and Muscoda, is central and northern Wisconsin where carbonate bedrock is less commonly exposed at the surface.

The sizes of erratics recovered during drilling in the upper outwash near Blue River and Muscoda were very small and indicate that the ice front probably was of considerable distance to the east. The largest erratics observed were very small pebbles (4 to 8 mm intermediate axis). It is very likely that larger particles exist but were not recovered on the drill augers. Erratics recovered by drilling in the basal western-derived outwash at the same sites were as large as small pebbles (8 to 16 mm intermediate axis), but small boulders up to 120 mm diameter on the intermediate axis were exposed in a gully that cut into the basal outwash. Another indicator that the ice front probably was of considerable distance to the east is the low frequency of erratics in the upper outwash by comparison to the lower outwash where they are abundant. These differences suggest that the ice front that was related to the upper outwash probably was further away than the down-valley distance represented between the White Site and the basal outwash in the midcourse valley near Blue River and Muscoda.

SUMMARY

The Bridgeport terrace is composed of isolated remnants of a once more extensive terrace that extended throughout the lower Wisconsin River system. Its preservation at selected localities relates to protection from lateral stream erosion provided by a buried dolomite and sandstone strath. Terrace deposits on the strath are dominantly pre-Wisconsinan age, and appear to represent at least two major pre-Wisconsinan depositional events. The first event is associated with a 3 to 4 km penetration of an ice lobe into the mouth of the valley. This penetration resulted in deposition of glacial till on the Bridgeport terrace at the valley mouth and in deposition of fluvial glacial outwash that grades eastward toward central Wisconsin. The eastward gradient, eastward-dipping bedding, and eastward decreasing particle size in Bridgeport outwash sediments indicate that the Wisconsin River was temporarily reversed while ice blocked the mouth of the valley. The discharge probably flowed to the Rock River paleodrainage via Black Earth Creek valley because deep fills occur there and in the Madison area. Alternatively, flow may have passed eastward into Lake Michigan drainage. Clay mineralogy of the Bridgeport till is similar to the clay mineralogy of tills in the Wolf Creek Formation in adjacent northeast Iowa and in the Hersey till on the northwestern margin of the Driftless Area in Wisconsin. These correlations suggest that the Bridgeport till is of classical Kansan age.

The second pre-Wisconsinan event represented in Bridgeport terrace sediments includes deposits of sands and clayey cherty sands with crystalline erratics that are found sandwiched between outwash of the Bridgeport till and overlying Wisconsinan loess. These deposits appear to represent fluvial glacial outwash from a source to the east in central and northern Wisconsin. Their age is unknown, except that they occur stratigraphically above outwash from the Bridgeport till and are therefore younger. They are best preserved in the midcourse valley along the north side of the Wisconsin River near Blue River and Muscoda.

The deposits near Muscoda, which are located near the center of the southern sector of the Driftless Area, have been cited by Black (1970) as evidence for glaciation of the Driftless Area. Our examination of surface exposures and of sediments from three drill holes that penetrated to bedrock in the immediate area indicate that the sediments suggested as till by Black are in fact fluvial glacial outwash. We found no evidence to indicate that glacial ice penetrated the lower Wisconsin River valley more than about 3 or 4 km at the mouth of the Wisconsin River near Prairie du Chien and Bridgeport.

ACKNOWLEDGMENTS

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APPENDIX I

WHITE SITE*

Field Description Unit	Depth ft	Depth m	Moist Munsell Color	Calcareous	S:Si:Cl	Gr	XL	Q/F	Chert	Ang/Rd
Silt and clayey silt	0.0-5.0	0.0-1.5	7.5YR4/4	No						
Silty sand w/ erratic pebbles	5.0-9.0	1.5-2.7	10YR6/4	No	59:23:18	7	8	21	1	3.4
Sandy silt-silty sand w/ erratic pebbles	9.0-15.3	2.7-4.7	7.5YR4/4-4/6	No						
	13.0	4.0		No	51:31:18	4	5	15	0	
	14.0	4.3		No	52:29:19	7	9	13	1	3.8
	15.0	4.6		No	48:31:21	5	15	17	4	8.6
Sandy clay	15.3-20.0	4.7-6.1	10YR5/4	No						
	17.0	5.2		No	36:19:45	0	6	9	17	1.5
	19.0	5.8		No	43:22:35	5	14	0	5	0.9
Clayey silty sand	20.0-29.0	6.1-8.8	10YR4/6-5/6	No						
	21.0	6.4		No	51:26:23	4	13	13	4	3.4
	22.0	6.7		No	50:23:27	6	5	15	5	6.5
	23.0	7.0		No	51:23:26	2	12	17	1	5.6
	24.0	7.3		No	52:23:25	2	6	13	2	0.2
	26.0	7.9		No	57:19:24	6	1	11	0	0.1
Sand	29.0-35.0	8.8-10.7	2.5Y6/4	No						
	31.0	9.5		No	68:14:17	8	20	12	0	-
	34.0	10.4		No	82:8:10	0	4	9	4	1.7

Roth Site*

Field Description Unit	Depth ft	Depth m	Moist Munsell Color	Calcareous	S:Si:Cl	Gr	XL	Q/F	Chert	Ang/Rd
Clayey silt	0.0-5.5	0.0-1.7	10YR5/6	No						
Clayey cherty sand w/ erratics	5.5-10.5	1.7-3.2	10YR4/6-5/6	No						
Clayey cherty sand	10.5-16.0	3.2-4.9	7.5YR4/4	No	72:10:17	0	1	1	49	1.15
Sand with minor silt and clay	16.0-19.0	4.9-5.8	10YR6/6	No	74:12:14	1	7	6	17	1.6
Clayey sand w/ erratics	19.0-23.5	5.8-7.2	7.5YR4/4	No	74:9:17	3	3	5	5	0.2
Clayey gravelly sand between beds of sand	23.5-28.0	7.2-8.5	7.5YR4/4	No	77:2:21	1	7	12	9	0.8
Clayey gravelly sand	28.0-31.0	8.5-9.5	10YR3/4	No						
Weathered bedrock (dol)	31.0-33.5	9.5-10.2	2.5Y7/6	Yes						

Greenwood Site*

Field Description Unit	Depth ft	Depth m	Moist Munsell Color	Calcareous	S:Si:Cl	Gr	XL	Q/F	Chert	Ang/rd
Silt and clayey silt	0.0-7.0	0.0-2.1	10YR5/6	No						
Coarse silt	7.0-9.0	2.1-2.7	10YR5/4	Yes	3:80:17	0	0	0	0	0
Coarse silt	9.0-18.0	2.7-5.5	10YR6/6	Yes						
Paleosol?-silt	18.0-18.3	5.5-5.6	7.5YR4/4	Yes						
Paleosol?-silt	18.3-18.7	5.6-5.7	10YR6/8	Yes						
Silt	18.7-19.5	5.7-5.9	2.5Y7/4	Yes						
Sand with chert granules	19.5-24.0	5.9-7.3	2.5Y8/4	No	91:1:8	0	10	0	37	1.73
Clayey silt	24.0-29.0	7.3-8.8	2.5Y6/4	No						
Sandy silt - silty sand	29.0-34.0	8.8-10.4	10YR6/4	No	47:30:13	0	0	0	0	0
Sand	34.0-39.5	10.4-12.0	10YR6/4	No	83:9:8	12	16	8	12	0.11
Rounded cherty gravel in silty sand	39.5-41.5	12.0-12.7	10YR4/4	Yes	72:17:11	3	17	7	6	1.71
Gravelly sand with silt	41.5-50.0	12.7-15.2	10YR6/4-5/4	Yes	82:8:10	5	5	7	4	0.16
Weathered bedrock (ss)	50.0-50.5	15.2-15.4								

Elder Site

Field Description Unit	Depth ft	Depth m	Moist Munsell color	Calcareous
Silt	0-3.0	0.0-0.9		No
Silty sand	3.0-6.5	0.9-2.0	7.5YR5/4-4/4	No
Silty sand	6.5-10.0	2.0-3.1	7.5YR4/6-5/6	No
Coarse sand and erratic pebbles and chert	10.0-16.0	3.1-4.9	7.5YR6/6	No
Sand with silt and erratic pebbles	16.0-18.0	4.9-5.5	7.5YR6/6 - 10YR6/6	No
Weathered bedrock (ss)	18.0-20.0	5.5-6.1		No

Thiede Site*

Field Description Unit	Depth ft	Depth m	Moist Munsell color	Calcareous	S:Si:Cl	Gr	XL	Q/F	Chert	Ang/rd
Sand, silty sand	0.0-5.0	0.0-1.5	7.5YR6/6 - 10YR6/6	No						
Silty cherty sand	5.0-7.5	1.5-2.3	7.5YR5/6	No						
Cherty clayey sand	7.5-8.0	2.3-2.4	10YR3/3-3/4	No						
Clayey sand and cherty sand	8.0-14.0	2.4-4.3	7.5YR6/8-5/8	No	75:9:16	0	4	0	37	3.2
Silty sand and cherty sand	14.0-17.0	4.3-5.2	7.5Y 5/6	No						
Clayey silty sand	17.0-19.0	5.2-5.8	7.5YR5/6	No						
Clayey cherty sand and silty sand	19.0-24.0	5.8-7.3	7.5YR5/6	No						
Sand with granules and pebbles of erratics	24.0-29.0	7.3-8.8	10YR6/4 - 6/6	No	90:3:7	1	9	9	2	1.4
Clayey silty sand with common erratics generally less than 5-10 mm diameter	29.0-31.0	8.8-9.5	10YR5/6 - 4/6	No						
Weathered bedrock (ss)	31.0-32.0	9.5-9.8	green	No						

*Gr = granite
XL = other crystalline rock
Q/F = quartz + fedlspar
Ang/rd = ratio of angular to round quartz grains

The record at Devils Lake can be tied to the regional glacial record. The lake basin was formed when the advancing Woodfordian ice reached the Baraboo Ranges and flowed westward along the exposed edges of the more massive Southern Range. The deposition of outwash in the basin dates back to this time. This event also marks the time when meltwater from the ice north of the Baraboo Range was cut off from its drainage route down the lower Wisconsin River. The discharge ponded against the highlands to the west, finally to spill into the valley of the Black River (Bretz, 1950).

The cessation of outwash deposition and the onset of organic sedimentation in Devils Lake mark the retreat of the Green Bay Lobe from the Johnstown Moraine. The timing of this event is usually based on bog-bottom dates--the radiocarbon age of the first organic material found on the unglaciated terrain. These dates are variously suspect. They may be far too young if the depositional site was occupied by a buried ice block (Florin and Wright, 1969), or far too old if the basin formed in drift that was calcareous or that contained particles of coal or older peat picked up by the glacier. Estimates for the recession of the Green Bay Lobe range from 12,000 to 13,000 B.P. (Black and Rubin, 1967-68) to 15,560 $\pm$ 150 B.P. (WIS-442; Bender and others, 1971) or even more.

Finally, Devils Lake sediments contain little or no carbonate. Although the blocking moraines may contain some calcareous material, its presence is not obvious, and none was noted in the lake. The water is soft, and the surrounding quartzite terrain contributes little soluble material to the lake (Twenhofel and McKelvey, 1939). The organic carbon in its sediments thus can provide better dates than most lakes in the Midwest.

The lake is 2.1 km long and 1 km wide; its surface area is 1.53 km^2. Its maximum depth is 12.5 m; its median depth and mean depth, respectively, are 9.2 m and 7.8 m. The elevation of the water surface is higher than the land surface outside the blocking moraines; the lake surface forms a saddle in the east-west groundwater divide that extends along and under the quartzite ridge. It is not known what proportion the water leaves the basin by evaporating or by percolating through the sediment fill. The age and thickness of unconsolidated sediment under the lake are not known. A water well at the south edge of the basin penetrated 117 m without striking quartzite (Black, 1967-68).

The size of Devils Lake makes it hard to study. Twenhofel and McKelvey (1939) described its surficial sediment, obtained by Ekman dredge and short cores from an ice cover. They reported the off-shore material generally was a normally-sorted black silty clay with few macrofossils. They found a few small clams in the shallow water, and a couple of fish vertebrae. They reported the presence of diatoms, sponge spicules, and bacteria, but no mention was made of pollen. The surface sediment offshore was described as a soupy black "sludge" up to 1 m thick. Their average sediment water content of 82.5%, and average loss on ignition of about 19% were similar to those I measured on the upper sediments. Twenhofel and McKelvey did not reach to the base of the organic sediments. They mentioned that the muds gave no hint of layering, and commented that, although this might normally be judged the result of complete mixing by organisms, small worms were the only bottom fauna they had noted.

I first visited Devils Lake in the early 1960's when I began teaching general geology and led field trips to the Baraboo area. As there was no coring equipment to deal with so large a lake, F. W. Bachhuber and I took a 2-inch Livingstone core from Hansen Marsh, which occupies one of the old proglacial lake basins (NW1/4 sec. 16, T. 11 N., R. 7 E.) 5 km northeast of Devils Lake. After penetrating a meter of peat and gyttja, the core entered brown and bluish laminated clays. Bedrock, probably Cambrian sandstone, was reached 8.5 m below the surface.

The pollen sequence on the final pollen diagram (Bachhuber, 1966) was typical for the Midwest. The unusual thing about it was that the late-glacial spruce zone extended for 7 m, while the postglacial record was restricted to the top meter! No funds were available for carbon dating at the time. The peat was later resampled (Davis, 1975) for material to date the end of the local spruce zone. The resulting date was 10,735 ± 105 B.P. (WIS-740; Bender and others, 1976), which agrees with other dates from the region.

It is never certain that samples from such thin peat units are not contaminated by roots. A site was needed with a higher postglacial sedimentation rate to improve the resolution for both dating and palynology. Devils Lake offered the best possibilities if its sediment was not too mixed and if it contained well preserved pollen.

SAMPLING

Samples of the flocculent sediments from the deeper part of the lake were obtained in 1977 for student projects. The water/sediment interface was sampled with a "frozen finger" sampler (Wright, 1980). Pollen was abundant and well preserved. The upper few centimeters contained a zone rich in ragweed that indicates the effect of land clearance accompanying settlement in southern Wisconsin.

To assess problems that might arise from coring the flocculent sediment, a SCUBA diver inserted open plastic tubes in the sediment and corked both ends before withdrawal. The cores were allowed to stand upright while they were frozen applying dry ice only to one side. The frozen cylinders were extracted and cut into samples for pollen analysis. The sediment in the core was never more than a fraction of the thickness the tube was inserted; the top of the frozen core was clear water. The open structure of the flocculent sediments apparently is disturbed by the core tube, and the solids then compact to occupy less space. The ragweed zone was still present in the upper part of the collapsed core. This information influenced later coring.

Heavy-duty coring equipment (Cushing and Wright, 1965) was assembled on the ice-covered lake during the late winter of 1978. The Livingstone core tubes were 10 cm in diameter so that samples for radiocarbon dating could be taken from a small sediment interval. The large cores would also provide material for macrofossil analysis as well as pollen. A special device was made for measuring the upper position of the soft bottom sediment to establish a coring datum at the surface. This consisted of an iron weight to which was attached a balanced spatulate metal vane. When the device was lowered from the surface, a mercury switch attached to the vane closed an electrical circuit when the vane tilted on reaching the flocculent sediment. This rang a bell at the surface. The wire holding the device was graduated in meters, and indicated the depth of water that must be traversed to start the first core. The device functioned well and gave readings reproducible to within 2 cm.

The position for the core was selected to be near the deeper part of the lake, well removed from sources of sediment in the small delta of Messenger Creek at the southwest shore and from the sandy bathing beaches. The core was taken 750 m from the north shore and 220 m from the talus blocks at the west side. The water depth at the coring site was 11.75 m. The first drive was taken through a depth of 1.5 m using a long core tube. The core tube was capped at both ends and allowed to stand upright at the surface until it froze. A tubular metal casing was lowered from the surface to help support and guide the drive pipe. Drives of a meter or less were used in the deeper sediments. The core yield was excellent, and the segments were wrapped in plastic and metal foil for storage. The fact that the casing did not quite reach to the sediment caused problems when penetration had reached a depth of 6 m. The pressure needed to push the Livingstone sampler deeper bowed the drive pipe which sheared off just below the casing. This last core and the sampler were still attached to the surface by the small-diameter pipe holding the piston. The pipe proved to have enough tensile strength to pull the corer from the hole. The sampler then jammed in the casing, and the entire casing had to be removed in one pull to retrieve the apparatus.

The core of 1978 did not extend to the mineral sediments that were expected to occur below the organic sediments, although later coring showed we had reached to within a centimeter of our goal. That additional coring was completed during the late winter of 1979. The station was reoccupied by triangulation. The casing was lengthened to 14.5 m so that it extended through the water column and well into the sediments. A 5-cm diameter Livingstone sampler was used, and the interface with the lower mineral sediments was reached on the first drive. The lower sediments were black, laminated silty clay. Coring continued 2.2 m below the interface with the mineral sediments. Time and available drive pipe limited the penetration.

LABORATORY PROCEDURES

The core segments were taken to my pollen laboratory at Madison for further study. The frozen core from the upper 1.5 m of sediments contained only 1.1 m of sediment under clear ice; the ice was discarded. The remaining sediment was allowed to melt in a foil-lined horizontal trough. Pollen samples were taken by packing sediment into a spoon of measured volume. The samples were also weighed and water content measured so that pollen concentration could be expressed in grains/cm^3 or grains per gram (dry or wet). The basic sampling interval was 5 cm; extra samples were taken near the top and base of the sediment.

Ten samples for radiocarbon analysis were chosen from various depths in the core. The base of the ragweed zone and the top of the spruce zone were selected by scanning raw sediment and glycerin smeared on microscope slides. The other samples for dating were taken at regular intervals. The samples were analyzed at the Radiocarbon Laboratory, Center for Climatic Research, University of Wisconsin, Madison (Bender and others, 1980).

Duplicate pollen samples were taken from two levels in each of the six original core segments. These twelve samples were used to estimate the number of marker grains needed for concentration measurements in the various parts of the core. The markers consisted of darkly stained *Lycopodium* spores embedded in tablets (Stockmarr, 1973; batch 212761 has a tablet mean and standard deviation of 12,489 $\pm$ 491). The darkly stained *Lycopodium* spores are used as an abundance standard in the sample, to which the indigenous pollen and spores are compared. Five tablets were placed in each of the twelve duplicate samples which were then processed in the usual fashion: successive 10-minute treatments in 10% HCl, 10% KOH, and 48% HF; glacial acetic acid washes before and after three minutes of acetolysis, washing with water; dehydrating with tertiary butyl alcohol, and mounting in 2000 centistokes silicone fluid for storage and microscope analysis (Maher, 1977). Rough counts suggested that five tablets would produce a reasonable number of markers in all the core segments except the fifth, from the interval between 430 and 525 cm in the sediment (fig. 1); ten tablets were added to pollen samples from that segment.

The cores obtained in 1979 from the lower lake sediments were treated in a similar manner, and two additional samples were submitted for radiocarbon analysis.

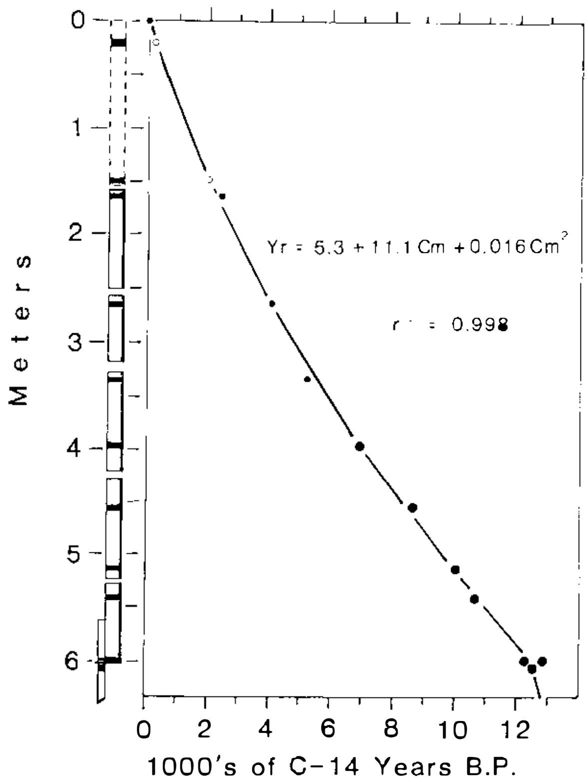

FIGURE 1.--Age versus depth relationships of the radiocarbon samples from Devils Lake. See also Table 1. Diameters of black dots include two standard deviations from mean value. Second order function fit to the data by least-squares criterion regressing radiocarbon age onto depth in cm; n = 11 (10 dates and surface). The two open circles and dashed upper segment are discussed in the text.

The pollen statistics were done using the methods described in Mosimann (1965) and in Maher (1981). In an attempt to avoid operator bias, all the pollen counting was completed, and the core was zoned by numerical methods before a pollen diagram was drawn.

LITHOLOGY

The sediment from above 603 cm in the core is an olive black (Munsell color wet: 5Y 3/1) silty clay. The sediment is massive and has little indication of stratigraphy except some occasional diffuse darker banding. The sediments have not yet been sieved for macrofossil studies. Aside from some fossil leaves noted in the field, little coarse material was found when the pollen samples were taken. The water content varies from 80% of the fresh sediment weight near the surface, to as little as 48% at 600 cm. Loss on ignition over this same depth interval ranges from 25% to 8%. Fragments of charcoal were noted at all levels; no attempt was made to quantify their abundance. As the pollen stratigraphy will clearly demonstrate, the homogenous appearance of this lake sediment results not from thorough mixing, but from a lack of contrasting sediments with which to define discrete layers.

The mineral sediment from depths greater than 603 cm differs markedly from that at shallower depths. Its color ranges from 5GY 2/1 to 5G 2/1 to 2.5GY 3/1--greenish black to dark olive gray. Its water content in fresh sediment is in the range of from 60% to 22%. When cores were newly extruded in the field, thin varve laminae were clearly visible on the surface. The material is in a highly reduced state. Vivianite (hydrated iron phosphate) and aggregates of tiny pyrite crystals were identified by electron-probe analysis. When the cores were opened in the laboratory, the laminations exposed on fresh surfaces rapidly darkened due to oxidation. Spot measurements of their thickness at a number of places within a meter of the contact with the overlying organic sediment suggest the sediment accumulated at an average rate of 0.128 cm/yr.

RADIOCARBON CHRONOLOGY

The radiocarbon dating of the Devils Lake sediments is important for three reasons. First, it provides a method of estimating the time since the ice of the Green Bay Lobe started to retreat. Second, it allows the pollen stratigraphy to be dated by an independent means. And third, it provides a way of measuring sedimentation rates which are needed for estimating pollen influx.

TABLE 1.--Devils Lake radiocarbon dates,

Bender and others (1980)

SAMPLE	DEPTH BELOW LAKE BOTTOM (Cm)	AGE BP	COMPRESSED DEPTH (cm) (See text)
WIS-993	18 - 25	245 ± 55	12 - 16
WIS-994	149 - 154	2055 ± 65	105 - 109
WIS-995	164 - 169	2430 ± 65	
WIS-996	263 - 267	4105 ± 65	
WIS-997	334 - 338	5245 ± 65	
WIS-998	395 - 399	6920 ± 75	
WIS-999	455 - 459	8640 ± 85	
WIS-1000	514 - 518	10,080 ± 100	
WIS-1001	541 - 547	10,620 ± 105	
WIS-1004	599 - 603	12,880 ± 125	
WIS-1073	599 - 603	12,260 ± 115	
WIS-1075	603 - 611	12,520 ± 160	

The twelve radiocarbon samples from the cores are listed in table 1 and shown in stratigraphic position in figure 1. The lowest sample from the 1978 coring operation (WIS-1004) was found essentially equal in spore content to the sample taken in 1979 (WIS-1073) from just above the interface with the lower mineral sediments. The average of their C-14 ages is 12,570 BP, which is in agreement with the date of 12,520 BP for WIS-1075 from the sediment just below the interface.

The depth versus time curve shown on figure 1 is defined using only the solid black dots that include the surface (0 cm = A.D. 1978 = -28 B.P.) and the ten lower carbon samples. (The two samples WIS-993 and WIS-994 from the top segment are used for a different purpose described below.) The curve is the best-fit second-order function defined by regressing radiocarbon years onto depth in centimeters below the water/sediment interface. The value of r^2 indicates that the function correlating radiocarbon age with depth accounts for almost all the variance in the data.

The two dates from the upper core segment were not used to derive the depth vs. time function. The drive through the uppermost 150 cm yielded only 110 cm of core. An almost perfect linear increase in age exists in the compressed upper core segment from the surface (-28 BP) through the lower two carbon dates (yr = [19.5] cm - 27.8; r^2 = 0.999+). It appears that the flocculent sediment is progressively less dense toward its upper portion, but when disturbed by the core tube, collapses to a substance of constant density. The samples' positions in the compressed core were translated to age B.P. Solving for depth in the quadratic equation in figure 1 allowed the compressed samples to be spaced reasonably within the upper 1.5 m. This reconstruction tends to confirm that little sediment is missing in the top segment. It should be noted that this exercise is necessary only when the reconstructed sediment column needs to be shown such as in figure 1. For all other purposes the core is taken as measured in the laboratory. For example, the sedimentation rate to be used for pollen samples from the upper segment is that which comes from the compressed core: 0.051 cm/yr.

The curve of figure 1 shows a progressive change of sedimentation rate through time. I interpret the rate change as resulting from compaction of the older sediment under the weight of younger. The line segment below 6 m shows the increased sedimentation rate in the underlying varved clays. The age at the base of the analyzed core is taken to be about 12,800 B.P., judged from the length of the varved segment and the average number of laminae per cm.

IMPLICATIONS FOR THE GREEN BAY LOBE

I am impressed by the orderly increase of radiocarbon age with depth, and that figure 1's curve so closely approaches the present age at the surface; it predicts +5 B.P. at 0 cm rather than the actual age of -28 B.P.

The Devils Lake dates suggest the lake received outwash until about 12,500 B.P. Normal postglacial sedimentation could not have occurred at Devils Lake until the ice of the Green Bay Lobe had receded enough to open a drainageway along the eastern side of the Baraboo Range near the present Wisconsin River; water from the proglacial lake to the north would no longer flood Devils Lake. Now if the Two Creeks Forest was drowned about 11,850 B.P. by rising Lake Michigan waters dammed by the Greatlakean advance, the time available for the Green Bay and Michigan Lobes to retreat some 350 to 400 km--exposing and then reblocking a drainage route to the east--is limited to about 700 yr. This amounts to an average retreat rate of 50 to 60 km per century.

PERCENTAGE POLLEN DIAGRAM

Wright (1971) and Bernabo and Webb (1977) have summarized the trends noted on Midwestern pollen diagrams, and discussed their significance in terms of vegetation and climate.

Pollen studies generate too much detail to be shown on a single page. Figure 2 is an abridged diagram of 15 pollen taxa in the Devils Lake sediment. The vertical dimension is scaled in 1000's of radiocarbon years; the depths of the samples were converted to

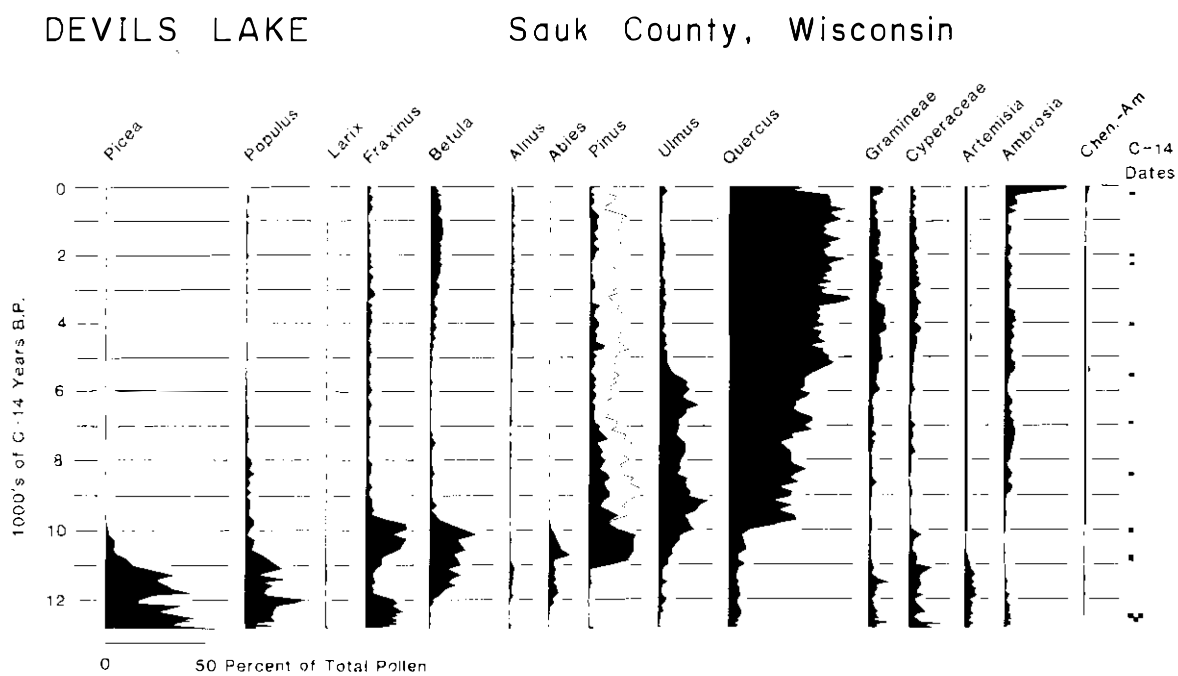

FIGURE 2.--Percentage pollen diagram. Vertical dimension transformed from depth to radiocarbon age by function in figure 1. Black part of pine curve represents Diploxylon or yellow pines; white part represents Haploxylon or white pines.

years using the information in figure 1. Scientific and common plant names used in this paper are listed in table 2.

TABLE 2.--Scientific and common names

Acer	maple
Alnus	alder
Ambrosia	ragweed
Artemisia	sage
Betula	birch
Cyperaceae	sedge family
Fraxinus	ash
Gramineae	grass family
Larix	tamarack
Ostrya	ironwood
Picea	spruce
Pinus	pine
Populus	aspen, cottonwood
Quercus	oak
Tilia	basswood
Ulmus	elm

Note: "Chen.-Am." refers to the Chenopodiaceae and Amaranthaceae families which are referred to in the text as the Chenopodiineae. This pollen taxon includes the goosefoot and pigweed families.

Picea is the dominant pollen type before 11,000 B.P. Spruce then rapidly declines, reaching 15% of total pollen by 10,800 and 1% by 9800. The diagram differs from many in the Midwest in that Populus is shown as a discrete curve. Populus pollen is easily destroyed, and its abundance in the older sediments may in part reflect the excellent pollen preservation in these sediments. Taking the diagram at face value, however, the combined abundance of Picea and Populus suggests the forest combined a good deal of aspen with spruce. Larix must have occurred in the region throughout the record. Larix pollen dissemination is low, and not much significance should be attached to the fact that its percentage curve is continuous in the early part of the diagram.

Fraxinus percentages exhibit an interesting double expansion. The first period of abundance ends about 12,000 B.P.; a second ash rise begins about 11,000 only to crash after 10,000 B.P. Fraxinus pollen can be subdivided on several morphological grounds. McAndrews and others (1973) suggest that Fraxinus nigra and F. quadrangulata tend to have three furrows whereas F. pennsylvanica and F. americana tend toward four furrows. There is a significant statistical difference in the proportion of 3-furrowed ash grains in the sediment: before 9600 B.P. they comprise 85% of all ash grains; after that date their proportion falls to 50%. I conclude the black ash (Fraxinus nigra) is partially responsible for the twin peaks of abundance; it became less important later in the postglacial.

Betula pollen percentages start to rise about 12,000 B.P., fall rapidly at 9800, and then maintain low values for several millennia until a gradual rise starts about 5000 B.P. Alnus pollen occurs at low percentages throughout the record.

Pinus percentages are exceedingly low at Devils Lake prior to 11,000 B.P. Leading authorities conclude the genus simply was not in the region (Wright, 1971). Pine pollen is distributed prodigally by wind, and the trace (<2%) in the older sediments at Devils Lake may well have come from trees growing in the Rockies. It is difficult to visualize how migrating pines could contribute almost no pollen to the lake and then burst on the scene. The pine percentage goes from 1.5 to 10.5 of total pollen within about a century. The pollen grains of these first pines have smooth furrow membranes belonging to the so-called yellow pines (subgenus Diploxylon), which in the Midwest probably included Pinus banksiana (jack pine) and P. resinosa (red pine).

By 9600 B.P. pine pollen with exine flecks on the furrow membranes--the white pines (subgenus Haploxylon)--became more common at Devils Lake. In the Midwest the only pine of this group is Pinus strobus (white pine). Jacobson (1979) summarizes the migration of white pine from a refugium in--or east of--the Appalachians. The white pine was in northern Indiana by 9800 B.P., in northern Wisconsin by 8500 B.P., and then spread westward, reaching northwestern Minnesota by 2700 B.P. Davis (1976, p. 21), implied that white pine spread to Wisconsin around the north side of Lake Michigan. White pine at Devils Lake at 9600 B.P. (see also King, 1981) indicates a migration path around the south side of Lake Michigan.

By the time white pine arrived at Devils Lake, yellow pine was already declining. The percentage of yellow pine continues to drop gradually until about 7000 B.P. when it stabilizes at a relatively low level, leaving white pine pollen the dominant type of pine.

Ulmus percentages begin an upward climb about 11,000 B.P. From 10,000 to 5500 B.P. elm forms a larger proportion of the terrestrial pollen than at any later time. Although not included on the abridged diagram, Acer, Ostrya, and Tilia also achieve their largest percentages during this time.

Quercus, which had represented 3 to 9% of the pollen sum during the spruce zone, experienced a marked increase about 10,000 B.P. With many fluctuations, it continued to dominate the terrestrial pollen until the time of settlement.

The pollen from anemophilous herbs never acquires the large percentages at Devils Lake that are found at sites farther west on the prairie (Wright and others, 1963; Van Zant, 1979). The nonarboreal pollen (NAP) is relatively high in the spruce zone, but decreases by 9000 B.P. to very low values. Ambrosia and Chenopodiineae obtain their highest pre-settlement values from 9000 to 3600 B.P., suggesting the mid-Holocene expansion of prairie. Gramineae and Cyperaceae reach their highest postglacial values after 5000 B.P.; this last interval also shows increases in birch and oak.

The effect of the land disturbance accompanying agriculture is most easily seen in the meteoric rise in ragweed pollen and the concomitant decrease in oak and pine. The southern part of Wisconsin experienced a marked population increase about 1850, some 125 yr before the core was obtained. The carbon date of 245 B.P. (WIS-993), Table 1, came from sediments deposited before the ragweed rise. But as radiocarbon dates go, it is not distinguishable from the present. Considering the flocculent nature of the upper levels of the lake sediment, it is amazing the ragweed rise is so prominent.

POLLEN ZONES

Pollen zones are useful simply to call attention to a part of the diagram; I have already made reference to the spruce zone. Zone boundaries may point to times of change. Given a series of curves like those on figure 2, one tries to see patterns or intervals where the pollen assemblage is similar.

I need to consider zones in the Devils Lake core for two reasons. First, I acquired the pollen data so that they could be used for percentage diagrams like figure 2, but also for calculating the numbers of pollen grains/cm^3 of sediment (pollen concentration) and the number of pollen grains/cm^2/yr (pollen influx). These measures have certain advantages over percentage data, because the various taxa can be considered in a frame of reference separate from the other types of pollen in the assemblage. Percentage curves tell the proportion made up by each pollen taxon. The changes in a taxon's proportion in slides from different levels can result from the interactions of a myriad of factors that have nothing to do with the abundances of that taxon in the landscape (Davis, 1963). When the percentage and influx diagrams look alike, the analyst is asked why he bothered with the influx measures. But when the diagrams differ during certain intervals of time, it becomes more interesting.

For reasons discussed in Maher (1981) and not elaborated here, pollen influx numbers are of practical interest, but they are subject to grievous errors due to uncertainties in the rate of sedimentation. It is simply not possible to know whether the average rate over an interval of core applies to a single sample taken from within the interval. The ambiguities of pollen influx data can be made less severe if the geometric mean of a taxon's concentration in several adjacent samples within a segment of core is multiplied by the geometric mean of the estimated rates of sedimentation in the same segment. If one is to average samples, it makes sense to average intervals of core that are more or less alike; that is, to average over zones.

The second reason for considering Devils Lake pollen zones also has to do with treating the data as concentrations and influxes as well as percentages. It is now fashionable to zone pollen diagrams by numerical methods which are sometimes assumed to be better--or at least less operator-dependent--than the older method based on "experience." It is not generally realized that a numerical method best suited for one form of data presentation is not necessarily suited to another, nor will it necessarily define the same zones. It would be a pity for the percentage and influx diagrams of Devils Lake to have different zones! With this in mind, I zoned Devils Lake using a nonparametric measure of similarity based on the rank order of abundance of the taxa in each sample. The same zones will be defined whether the pollen data are percentages or concentration or influx. The measure is the Spearman rank correlation coefficient (Siegel, 1956) which was used in palynology by Ogden (1977) but then was generally put aside for other numerical methods (Prentice, 1980).

Figure 3 is a Spearman rank matrix for the 122 levels of Devils Lake. Each level is compared to the other 121 levels, yielding 7381 pairs. When, as in this case, the correlation coefficients are based on the rank order of 31 pollen taxa in each level, the null hypothesis that the samples are uncorrelated can be rejected at the 0.01 level of confidence when the rank coefficient is greater than 0.42. Most of the coefficients in figure 3 are greater. In looking at a few thousand years' pollen record in a particular lake, surrounded by a constant topography in one geographic region that is characterized by westerly winds, it is not surprising the pollen samples are not a random assemblage! But high values of the rank coefficients can be used to indicate samples that are quite similar as judged by the order of abundance of their taxa.

The ages of the levels in figure 3 are plotted along the hypotenuse in 1000's of years B.P. At this simplified representation of the rank coefficients in the core, it is easy to see that the levels older than about 10,000 B.P. are more like themselves than like the younger levels. Similarly, the sediment levels younger than 10,000 B.P., although generally alike, seem capable of division at about 5000 B.P. Comparison of figure 3 with the percentage pollen diagram (fig. 2) shows the zone >10,000 B.P. tends to have high Picea, Populus, Fraxinus, Betula, Abies, and NAP. The zone <10,000 B.P., by contrast, contains Quercus, Ulmus, and white pine. The subdivision of this zone at 5000 B.P. corresponds, among other things, to the drop-off of Ulmus and the rise in Betula and Cyperaceae.

These few comments are suggestive of a method for dividing samples by starting with all combined and then breaking the whole into parts more similar within themselves than to each other. This has been called a "divisive" procedure by Gordon and Birks

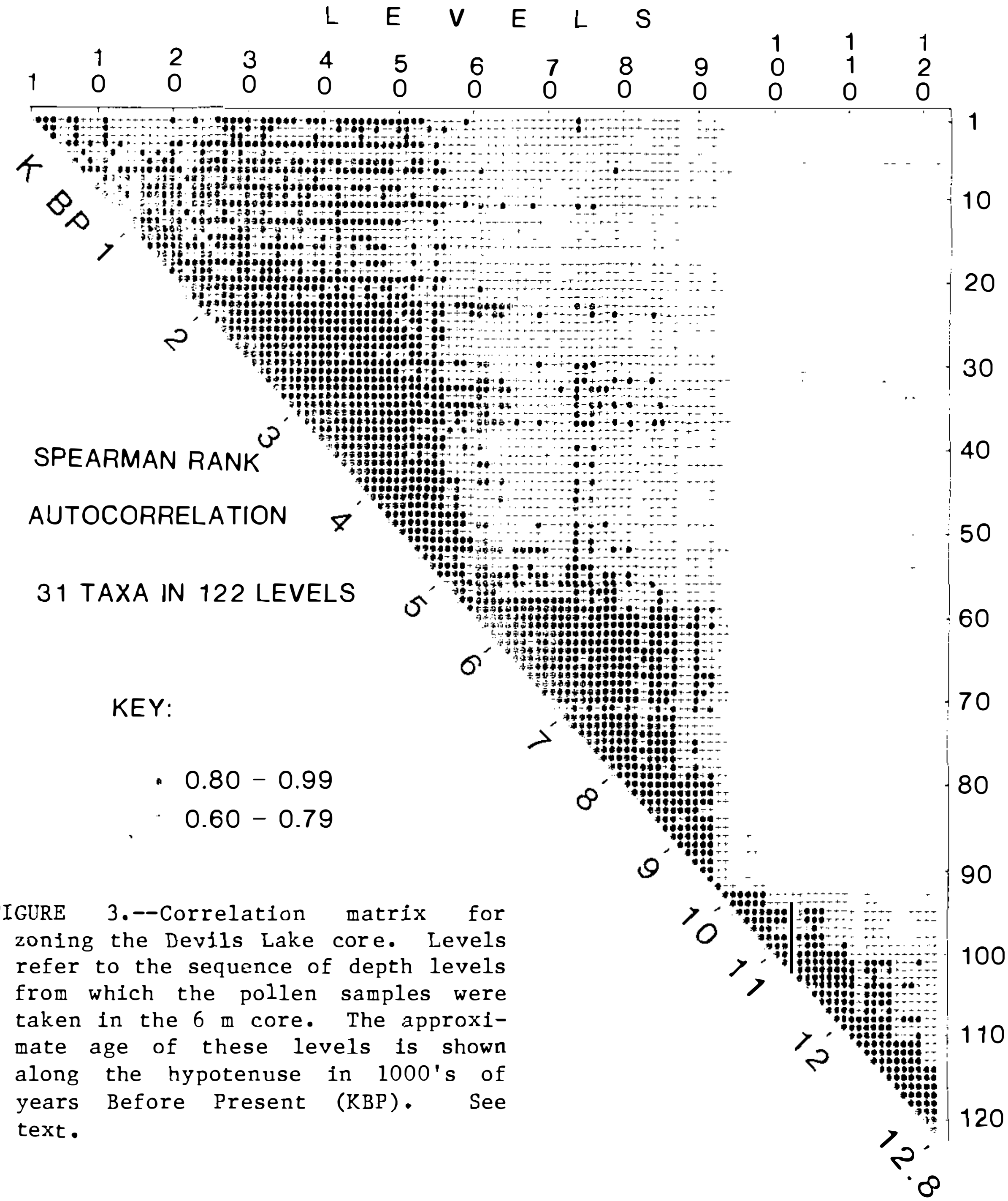

FIGURE 3.--Correlation matrix for zoning the Devils Lake core. Levels refer to the sequence of depth levels from which the pollen samples were taken in the 6 m core. The approximate age of these levels is shown along the hypotenuse in 1000's of years Before Present (KBP). See text.

(1972). The opposite "agglomerative" technique starts with the individual samples, which are combined through established hierarchical procedures with similar samples in stratigraphic contact, to form a dendrogram showing the relationships perceived within the data. It is good practice also to plot the data in a matrix like figure 3. A glance can then tell whether the agglomerative results make sense. The practice allows any computer with the same instructions to divide the data at the same places on any given day; this is difficult for humans to do. Users of dendrograms soon learn there can be as many zones as there are samples!

For my purposes, I needed an objective method for dividing the core into small segments containing samples with a related assemblage of pollen. An agglomerative technique was used with the Spearman rank coefficients to zone the Devils Lake core. I selected 13 zones, calling that at the base zone 1. All the zones had four or more samples except zone 11 with three. The least value of the coefficient between any of a zone's sample members was 0.75. Zones 1 to 4 comprise the lower grouping in figure 3--samples over 9800 B.P. The middle grouping (9800 to 5300 BP) contains zones 5 to 7, and the upper grouping contains zones 8 to 13. In discussing the influx of pollen to the Devils Lake sediments, the pollen influx stated will be a value averaged over an entire zone.

INFLUX POLLEN DIAGRAMS

Pollen analysts have long known that percentage data are easy to obtain, but difficult to interpret. Influx does not reveal the vegetation of the past or its associated climate; but it seems likely that a pollen taxon's influx variation through time will provide better environmental information than will a taxon's percentage in a pollen count. In the latter case its very measure of abundance has meaning only when it is grouped with things other than itself.

Not all sites are suited to pollen influx study; some have sedimentation rates so poorly known or sediment so unsuitable that any influx figure is meaningless. Still, if the extra time is spent at a few sites that are especially favorable, then some additional insights can be gained.

Figure 4 is a graph of the average influx of terrestrial pollen in the 13 zones. NAP reached its highest influx of 6600 grains/cm^2/yr, mostly ragweed, since the region was settled during the last century. The NAP influx is considerably less before European settlement, fluctuating within a narrow range between 1000 and 2400. The

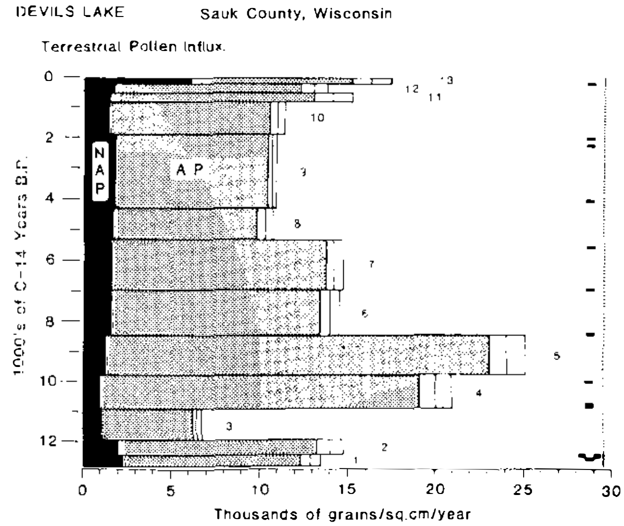

FIGURE 4.--Mean terrestrial pollen influx in the 13 Devils Lake zones. Zone numbers are to the right of each bar. Nonarboreal pollen (NAP) is in black, and arboreal pollen (AP) in gray. The white portion of the bars shows 0.95 confidence interval (Maher, 1981). The dotted line at the right indicates the position of the 122 levels from which pollen samples were taken. The radiocarbon samples are shown by the small black bars.

arboreal pollen influx is greater, and it oscillates more widely (5300 to 22,500 grains/cm^2/yr).

The arboreal species have many prolific pollen producers, and their height exposes their flowers to the stronger winds aloft. Trees growing in favorable habitats may contribute quantities of pollen to the sediments. Nonarboreal plants may produce quantities of pollen, but its chance for transport is limited because it is released close to the ground. A reduction in the number of trees might decrease the AP influx without producing a concomitant increase in the influx of NAP. It seems reasonable that the AP influx would range more widely than the NAP, and this appears to be the case in figure 4.

The influxes of zones 1 and 2 are about the same in spite of the fact that the mean sedimentation rate for zone 1 (0.081 cm/yr) is heavily weighted by estimates based on varve thickness, whereas the rate in overlying zone 2 (0.033 cm/yr) entirely depends on radiocarbon dating (fig. 1). Although this may be due to chance, it is comforting to find the sedimentation rates in the two zones combine with their different pollen concentrations in just such a manner as to produce similar estimates of influx.

The mean influx of terrestrial pollen in figure 4 is 12,600 grains/cm^2/yr. Most of the zone averages fall near the mean, except zones 4 and 5 which have influxes higher than average (20,000 to 25,000), and zone 3 which has the lowest influx (6000 to 7000). The boundaries of zone 3 are placed at 11,900 and 10,900 B.P. Zone 3 begins about the time the Two Creeks Forest was drowned (Broecker and Farrand, 1963). Palynologists (for example, West, 1961) have long sought evidence in their pollen diagrams for cooler weather thought to accompany the post-Two Creeks ice advance of the Michigan Lobe (Greatlakean Substage; the Valders of some authors). Whether this advance was caused by a change in climate or a surge occasioned by dynamic processes within the ice (Wright, 1971), one result would have been the same. Blocking the drainage from the Green Bay lowland created a vast proglacial lake that extended southwest toward the Wisconsin River to within only a few miles of Devils Lake. Such an expanse of water would affect the area's humidity. It is tempting to see the low influx of terrestrial pollen during zone 3 as a result of this moisture increase. Pollen is very efficiently removed from the atmosphere by rain. Although this may help bring pollen from local plants to Devils Lake, it effectively cuts off the regional pollen supply that is an important component in large lakes.

Figure 5 is an abbreviated pollen influx diagram showing the same taxa as figure 2, the percentage diagram. The influx diagram looks more schematic because each taxon's reported influx represents its average value over a whole pollen zone. The averaging process sacrifices some details in order to improve the estimates of sedimentation rate. At first glance there appears to be little difference between

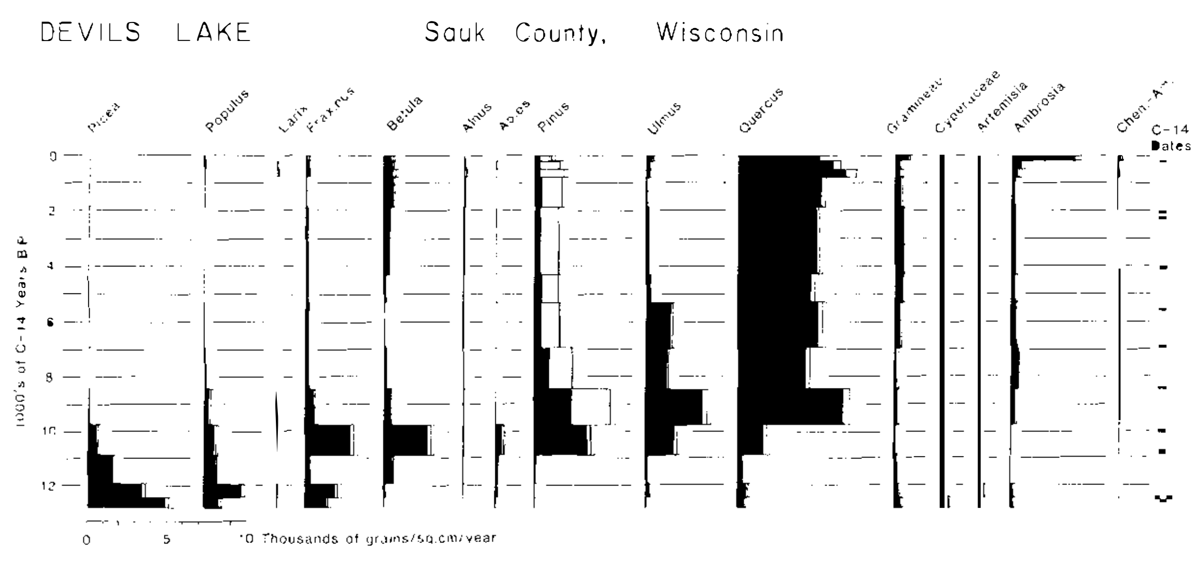

FIGURE 5.--Pollen influx diagram. Data plotted as a taxon's mean influx during the 13 pollen zones shown in figure 4. Black part of pine curve represents sents yellow pines; white part represents white pines. The narrow white strip visible on some longer bars represents 0.95 confidence interval (Maher, 1981).

the two diagrams. However, certain interesting differences do exist.

Late-Glacial Zones 1 to 4 (12,800 to 9800 BP)

The influx of Picea experiences a general decline during the interval without the complex fluctuations of figure 2. The laminated Late-Woodfordian clays give way to organic lake sediment during zone 1, signaling the beginning retreat of the Green Bay Lobe. Zone 2 should contain the pollen of the Driftless Area vegetation as well as that from plants spreading into the newly ice-free region. Today a record of this vegetation is found in the type Two Creeks site and its correlatives. Zone 3 is contemporary with the Greatlakean advance. The influx at Devils Lake is very low during this time. In the absence of competing pollen grains, those that did get to the sediment seem important in the percentage diagram. During zone 3 the percentages of Picea, Populus, Betula, Alnus, and Abies of the AP, and Gramineae, Cyperaceae, and Artemisia of the NAP, seem larger than their influx suggests they were.

The influx diagram shows that only Betula and--to a very minor extent--Pinus are higher in zone 3 than in zone 2. The decline of Picea influx continues in zone 4, but there are much larger influxes of Fraxinus, Betula, Ulmus, and Quercus, and the yellow pines arrived in the vicinity. Zone 4 is somewhat transitional to the Holocene. Some taxa from the older zones (Fraxinus, Betula, and Abies) reach their all-time influx highs. Ulmus and Quercus influxes begin rises that will culminate in later zones.

Early Holocene Zones 5 to 7 (9800 to 5300 B.P.)

Influx of the groups typical of the Late-Glacial (Picea, Populus, Fraxinus, Betula, and Abies) have much declined. The white pine arrived by 9600 B.P., early in zone 5. Ulmus influx is higher in zone 5 than before or since; this also applies to the unfigured taxa Acer, Tilia, and Ostrya. NAP influx is rather low in zone 5, but the gradual buildup of Ambrosia and Chenopodiineae probably represents an eastward expanse of prairie. Bernabo and Webb (1977, fig. 89) show prairie over more than a third of Wisconsin. I find this hard to believe, given the large pollen influx of mesic tree taxa mentioned above.

These Early Holocene zones provide another example of the differences between percentage and influx data. The large proportion of Ulmus and other mesic deciduous taxa depress the Quercus percentages which seem to rise gradually from zones 5 through 7. The Quercus influx, however, is at its highest rate in zone 5, suggesting that the oaks established a niche at the start of the Holocene that they have maintained ever since. Ulmus percentages form two equally-strong maxima, the first at about 9000 B.P. in zone 5, and the second in zone 7 at 6000 B.P. The elm influx at 9000 B.P. was decidedly the larger.

During zone 6 (8400 to 6900 B.P.), quite a number of taxa experience decreased influx compared to that which they had in zone 5. Examples include Pinus (both yellow and white), Ulmus, Quercus, and most of the other AP. The NAP, especially Ambrosia and Chenopodiineae, increased their influx, and this interval from 8400 to 6900 B.P. may have seen the largest development of prairie in southern Wisconsin. The whole interval in Devils Lake from 9800 to 5300 B.P. (zones 5 to 7) has Ambrosia and Chenopodiineae influxes that suggest the influence of prairie. It is somewhat strange, then, to see the contemporary high influxes of mesic elements like Ulmus, Acer, Tilia, and Ostrya.

Late Holocene Zones 8 to 13 (5300 B.P. to Present)

The influx and percentage diagrams seem much alike after 5300 BP. There is a markedly decreased influx of Ulmus (and also *Acer*, *Tilia*, and *Ostrya*), and there is a concomitant rise in the influx of *Betula* and perhaps Cyperaceae. It is tempting to relate the opposite behavior of these taxa to a common climatic factor. Perhaps the less droughty conditions experienced in the Midwest during the Late Holocene have favored rising water tables and peat growth. Large areas once supporting mesic woodland may have changed to fens, sedge meadows, and wet prairies.

CONCLUSIONS

The location, mineralogy, and geologic setting make the sediments of Devils Lake a unique resource. They provide a direct tie to the glacial record, and their composition makes them unusually well suited to radiocarbon dating. The pollen is well preserved, and the homogeneous sediment is nearly perfect for influx studies.

The information contained in the pollen record has hardly been tapped. Many very close similarities can be noted between the Devils Lake diagrams and other pollen diagrams in Wisconsin and neighboring states. These will be useful for estimating the age of some cores whose sediments are not suited for radiocarbon dating. When broad regional similarities are noted, the subtle variations across the region are easier to trace.

The sediments contain diatoms and other algae, and the charcoal record should provide insight into the fire history of southern Wisconsin. It is appropriate that such a lake is protected as part of the Wisconsin Ice Age National Scientific Reserve.

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SOIL DEVELOPMENT IN LATE WISCONSINAN AND HOLOCENE VALLEY DEPOSITS, BRUSH CREEK VALLEY, WISCONSIN

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River valleys in the Driftless Area of Wisconsin contain a variety of Late Wisconsinan and Holocene geomorphic surfaces on which soils have formed. There are significant differences in soil morphology from one surface to the next. An understanding of the stratigraphic relationships among these surfaces, coupled with radiocarbon dating of the sediments which form them, allows estimation of the age of the soil associated with each surface. The evidence discussed here indicates that differences in soil development are due primarily to age and climatic changes during the Holocene.

Brush Creek is a small, east-flowing tributary of the Kickapoo River. It drains an area of 70 km^2 in the upper Kickapoo River watershed, Vernon and Monroe Counties, Wisconsin. Geomorphic and pedologic investigations have been conducted in Brush Creek valley since 1972 as part of the region-wide study of modern and Holocene fluvial geomorphic activity by Knox and coworkers (Knox and Johnson, 1974; Johnson, 1976; McDowell, 1980; Knox, McDowell and Johnson, 1981). This valley, the site of fifteen radiocarbon dates, is one of the key sites for the Holocene alluvial chronology of the Driftless Area. Because the Late Wisconsinan and Holocene stratigraphy of Brush Creek valley is well known, it is a good site for studying the comparative development of soils.

As part of a stratigraphic and sedimentologic investigation of the Holocene alluvial deposits of Brush Creek, soil morphology was studied as a tool for the correlation of alluvial deposits (McDowell, 1980). Alluvial sediments and the soils formed in them were studied at over thirty sections, representing all of the stratigraphic units present in the valley. This paper is concerned with the sequence of soils in Brush Creek valley found on surfaces dating from the Late Wisconsinan to the historical period. First the stratigraphic setting of the soils and the evidence bearing on the age of each soil will be discussed. Then the textural variations among the parent materials of each soil will be described. The morphology of each soil is described in the fourth section. The latter part of this paper is devoted to some tentative conclusions on the influence of two factors of soil formation, time and climate, on the sequence of soils in Brush Creek valley.

STRATIGRAPHY AND AGE OF SOILS

Like most Driftless Area streams, Brush Creek flows in a valley deeply incised in nearly flat-lying Paleozoic sandstone and dolomite. The landscape of the watershed consists of an upland mantled by Late Wisconsinan loess (correlative to the Peorian Loess in Illinois) and a valley containing Late Wisconsinan and Holocene colluvial, alluvial and mass-wasting deposits. This study concentrates on the valley deposits. The surfaces of seven distinct stratigraphic units in the valley have been exposed to subaerial soil formation for periods ranging from more than 10,000 years to less than 100 years. The stratigraphic units, and the zones of soil formation associated with their surfaces, are shown in figure 1. The stratigraphic units are labelled with letters, in alphabetical

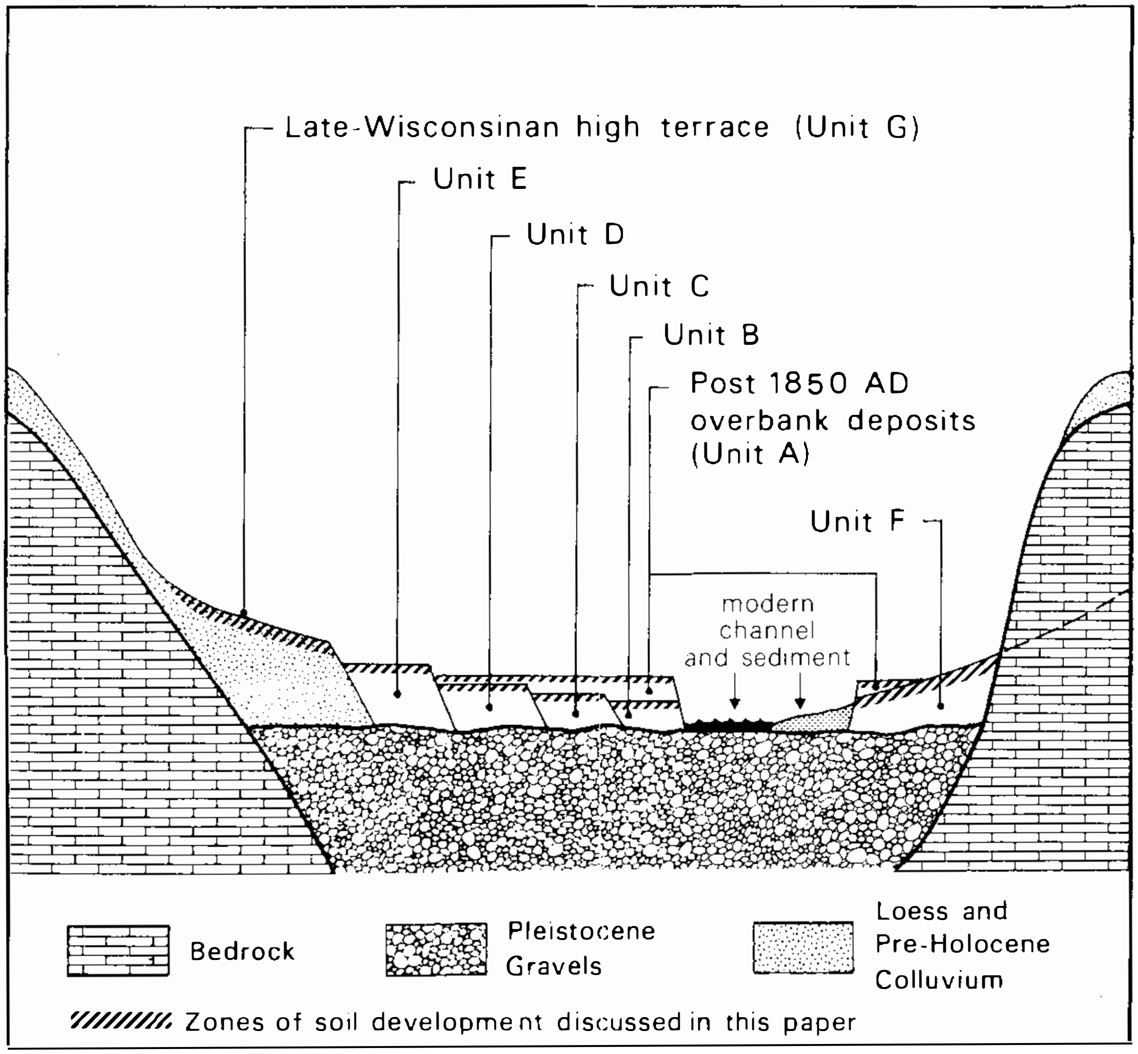


FIGURE 1.--Schematic cross-section showing the relationships among stratigraphic units and soils in Brush Creek valley. Not to scale.

order from the youngest unit (Unit A) to the oldest unit (Unit G). Each soil is designated by the same letter as the unit on which it formed. Radiocarbon dates associated with each unit are listed in table 1.

Geomorphic stratigraphic concepts can be used to estimate the length of time over which pedogenic processes have been operating in each soil. The age of the upper surface of each stratigraphic unit indicates the length of time that this surface has been available for soil formation. The ages of these surfaces can be established only indirectly, based on stratigraphic relationships among the surfaces and radiocarbon dates on the deposits which form the surfaces. The age of a geomorphic surface may, of course, be somewhat younger than the deposits which underlie it. The geomorphic surfaces in Brush Creek valley can generally be dated to within 1000 years, and in some cases less. The bases for dating these surfaces are discussed below, and the times of soil formation on each surface are summarized in figure 2.

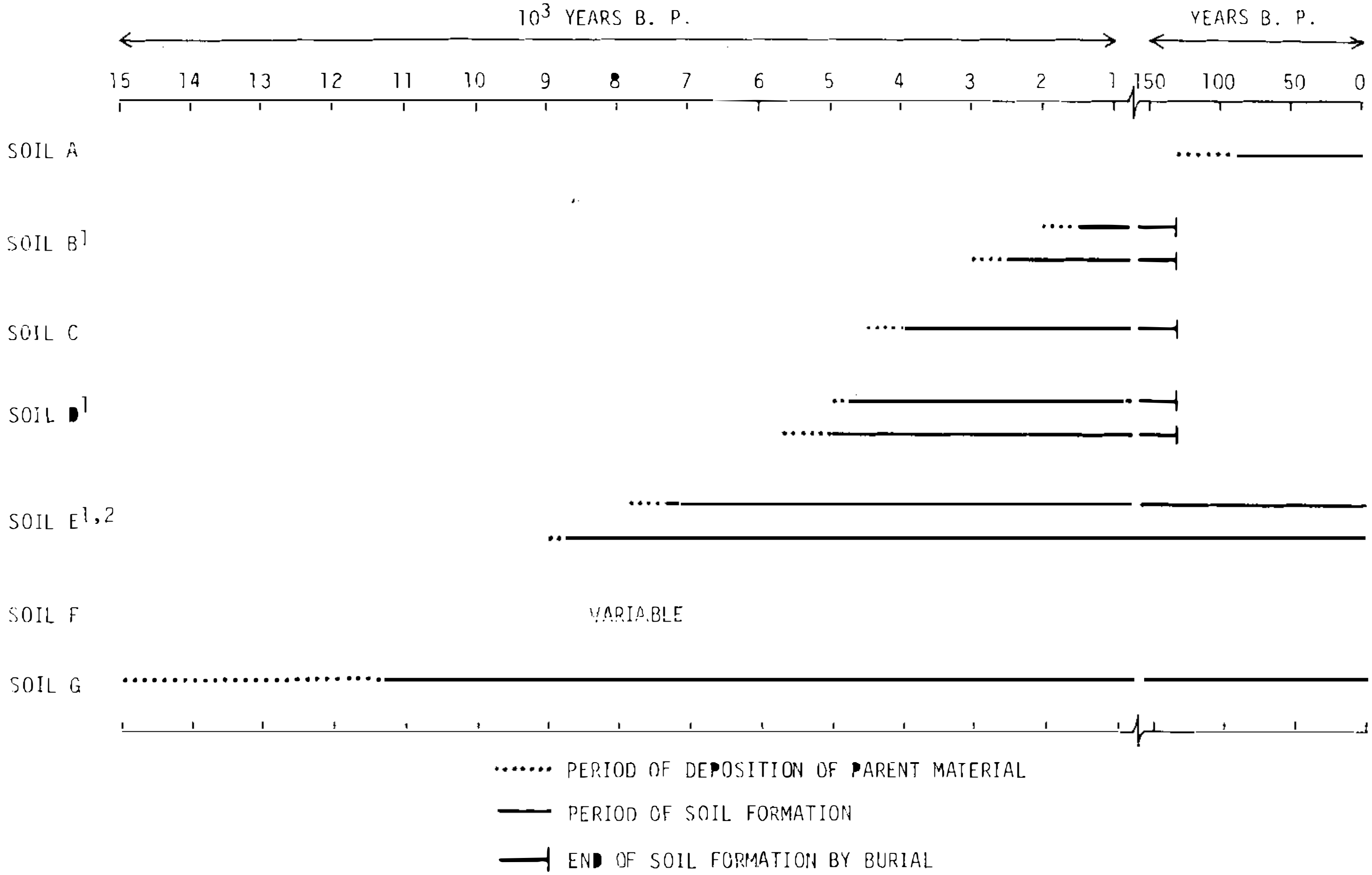
TABLE 1.--Stratigraphic units present in Brush Creek valley

Stratigraphic Unit	Associated ^{14}C dates or approximate age
Unit A (Historical Alluvium)	since A.D. 1850, based on historical artifacts and documents
Unit B	2065 ± 55 B.P. (WIS-808) 2445 ± 60 B.P. (WIS-767) 2715 ± 55 B.P. (WIS-678) 2940 ± 60 B.P. (WIS-757)
Unit C	4380 ± 65 B.P. (WIS-810) 4410 ± 75 B.P. (WIS-1022) 4440 ± 65 B.P. (WIS-666) 4540 ± 70 B.P. (WIS-1044)
Unit D	5045 ± 70 B.P. (WIS-813) 5055 ± 65 B.P. (WIS-674) 5145 ± 65 B.P. (WIS-1071) 5735 ± 70 B.P. (WIS-758) 5790 ± 85 B.P. (WIS-1224)
Unit E	7810 ± 95 B.P. (WIS-1046) 9060 ± 95 B.P. (WIS-1018)
Unit F	not dated; some fans deposited before 5700 B.P., others may date from later Holocene
Unit G	not older than 29,000 B.P.[a] older than 10,500 B.P.[b]

[a] Beginning of loess deposition (Hogan and Beatty, 1963).

[b] Oldest date on subsequent valley incision (Knox and Johnson, 1974).

The oldest stratigraphic unit exposed in Brush Creek valley is the Late Wisconsinan colluvium, which forms high terraces with gently sloping surfaces 3 m or more above the modern channel of Brush Creek. The geomorphic surface on these deposits dates from between 29,000 B.P. (maximum date of the loess from which the colluvium is derived; Hogan and Beatty, 1963) and 10,500 B.P. (oldest radiocarbon dates in the region on alluvial deposits set into the terraces; Knox and Johnson, 1974). It has been suggested that mass-wasting and colluviation in this region were related to the period of cold climate during the Late Wisconsinan (Schafer, 1962; Knox, 1980). If these colluvial deposits were formed under the cold glacial climate, then the surface may have been stabilized by 11,300 B.P., the approximate end of the glacial climate and beginning of the warmer post-glacial climate (Webb and Bryson, 1972).



[1]Two soil development lines are shown for Soils B, D and E, representing the range in ages of the corresponding stratigraphic units.

[2]At most sections, Soil E was not buried by Unit A, so soil development is shown as continuing up to the present. At a few sections, Soil E was buried by Unit A about 130 years ago.

FIGURE 2.--Periods of soil formation in Brush Creek valley.

Below the high terraces lie a series of six Holocene alluvial stratigraphic units. These units were created by episodes of adjustment in the fluvial system, resulting from changes in climate and vegetation cover (McDowell, 1980). During the Holocene, Brush Creek has not significantly incised the coarse gravel fill on which it flows, and most of the morphological adjustment in the stream has been lateral rather than vertical. Each episode of alluviation has resulted in construction of a new floodplain level, slightly lower than the previous floodplain. As a result, the surface of each unit was exposed for soil development from the time of construction until about 130 years ago. Beginning about A.D. 1850, most of these surfaces were buried by sediment eroded from hillslopes and uplands that were being cleared for agriculture.

The surfaces of the Holocene alluvial units can be dated more precisely than the high terrace surface because fifteen radiocarbon dates on these units are available from Brush Creek valley. These dates were obtained mainly from detrital wood samples taken from the base of the units. Therefore, the age of each surface is not known exactly. The maximum age of each surface is the date from the sediments which form it. Several dates from different sites were obtained from each unit, and the dates from a single unit span periods of 200 to 1200 years. The minimum age of each surface is the radiocarbon date of the next youngest stratigraphic unit which truncates it. This age range could be narrowed by knowing the length of time necessary for alluviation from the base of each section to the floodplain surface which caps it. There is no direct evidence for the length of time necessary for floodplain alluviation in this valley. However, delicate sedimentary structures are preserved in the middle and lower parts of many sections, and there is no evidence of soil formation below the surface zone of each section. This evidence suggests that alluviation of each section was fairly rapid, occurring perhaps within a few hundred years at each site. I therefore assume that the age of each floodplain surface is approximately 500 years younger than the radiocarbon date on the base of the stratigraphic unit which forms it. These alluvial surfaces present four major surfaces for soil development; the early Holocene surface (formed by 8500 to 7300 B.P., Unit E); the mid-Holocene surface (5200 to 4500 B.P., Unit D); and two late Holocene surfaces (4000 B.P., Unit C, and 2500 to 1500 B.P., Unit B). In addition, there is a stratigraphic unit (Unit F) consisting of alluvial fan deposits formed by small steep tributaries entering the valley floor of Brush Creek. The age of this unit is not well known. There are no radiocarbon dates on this unit, but stratigraphic relationships indicate that at least one fan may have been constructed before 5700 B.P. Unit F probably includes fan deposits of several different ages.

Unit A consists of overbank deposits of historical age. Deposition probably began shortly after settlement of the region by European-descended people, about A.D. 1850. The thickness of Unit A varies, depending on the underlying topography, but it is 50 to 100 cm thick at many sites. The surface of Unit A has not yet completely stabilized. In several Driftless Area valleys, discontinuous bodies of sediment were deposited on this surface during the large floods of July 1978. However, there is weak soil development in the upper 20 cm of Unit A, and this zone is underlain by undisturbed fine laminae of overbank deposits, indicating that most of the thickness of Unit A was deposited too quickly for soil development to destroy the structure. After the initial burst of sedimentation beginning A.D. 1850, the rate of sedimentation probably has been slow. I assume that the very weak soil formed in the top of Unit A may be about 75 years old.

SEDIMENTOLOGY OF THE STRATIGRAPHIC UNITS

There are two major sources of sediment in the Brush Creek watershed: (1) the Late Wisconsinan loess, which is a source of silt, and (2) the local sandstone, which provides well rounded, medium-to-fine sand. Gravels derived from local sandstone, dolomite and chert are a minor component of the stratigraphic units discussed in this paper. The colluvial deposits are derived primarily from loess, and they are silt loam in texture. The Holocene alluvial deposits are derived mainly from loess and sandstone. At a few sites, small amounts of the local gravels are included in the alluvial deposits. Only one of the units, Unit F (alluvial fan deposits), commonly contains a significant amount of gravel. The texture of the alluvial deposits varies from silty clay loam to sandy loam.

Each unit, and each section or exposure belonging to that unit, shows some variation in texture. The alluvial deposits consist of sandy point bar sediments at the base and silty overbank deposits in the upper part. They do not, however, show a progressive upward decrease in grain size, and the overbank deposits may include layers of sandy sediment.

In addition, there is some variation among the alluvial units. Unit F (alluvial fan deposits) consists of alternating layers of sandy gravels, sandy loam and silt loam, and it is the coarsest of the units. The next coarsest unit, Unit D, contains little gravel, but it has a larger proportion (about 50 %) of sandy point bar deposits than the other units. Unit E is dominated by silty overbank deposits, and it is the finest textured of the alluvial units. Units A, B and C are intermediate in texture, including silt loam and sandy loam.

SOIL MORPHOLOGY

Distinctly different soils are found on each of the geomorphic surfaces shown in figure 1. The range of soil characteristics found on each surface was established by systematic field and laboratory examination of thirty-four sections, including streambank exposures and Giddings probe cores. This sample of sections included: four profiles of Soil G; two profiles of Soil F; four profiles of Soil E, including two with radiocarbon dates; fifteen profiles of Soil D, including three with radiocarbon dates; five profiles of Soil C, including three with radiocarbon dates; and four profiles of Soil B including three with radiocarbon dates. Unit A with Soil A was present at twenty-one sections.

The following pedogenic features were regularly noted: Type and thickness of horizons; Munsell color (moist); shape, size and grade of structure; field texture; nature of horizon boundaries; presence or absence of cutans; consistence; biogenic features; color, size, frequency and distinctness of mottles; and organic inclusions. Several sections of different ages were tested for the presence of calcium carbonates with HCl, and all of the soils tested showed no reaction or a very weak reaction. In each section, samples were collected from each horizon, or at least every 40 cm. The grain size distribution and soil texture of each sample were determined by sieving and hydrometer analysis.

The high terrace soil, Soil G, is not mottled, but mottling is common among all of the Holocene soils Virtually all of the Holocene soil profiles have orange or red mottles in the B or C horizons, and many sections have an unmottled, gleyed zone below the mottled horizons. Frequency and size of mottles is controlled primarily by soil texture and elevation above the stream channel.

The soil characteristics which seem most representative of differences among the soils of different ages are presence or absence of the B horizon, B (or C) horizon color, grade of structure, and presence or absence of argillans. The characteristics of each soil type are summarized in table 2, and representative profiles are described in the appendix. Differences among the soils are discussed below.

Soil G shows the strongest soil development of all the soils studied. This soil, developed in silt loam, has a pronounced argillic horizon with visible argillans. The ratio of clay content between the Bt horizon and the A1 horizon ranges from 1.5 to 1.7 or more. The structure of the B horizon in this soil is moderate to strong, and the color of the B horizon is dark yellowish brown (10YR 4/4). Exposures of soil G in Brush Creek resemble the Tama series (typic argiudoll) and the Fayette series, valley phase (typic hapludalf).

Only two alluvial fan sections in Brush Creek valley were examined, and these two sections may be of different ages. Within an alluvial fan deposit, soil development is strongly affected by the number and position of gravel layers. It is not possible, therefore, to identify a typical example of Soil F. At one site (north center, SW1/4 sec. 35, T. 15 N., R. 3 W., Monroe County), a stratigraphic contact between Unit F and Unit C shows that the fan in question is older than Unit C (4600 to 4400 B.P.). The soil formed in this fan resembles Soils E and D in color and presence of argillans. Therefore, this fan may date from the middle or early Holocene

Soil E, developed in silt loam, has high chroma in the B horizon, moderate pedogenic structure, and visible argillans in the B horizon. In all four profiles studied, the B horizons have the increase in clay content necessary to qualify as argillic horizons. In comparison to Soil G, Soil E has slightly weaker structure and slightly higher value in the B horizon. Soil E resembles the Ettrick series (typic argiaquoll), except that Soil E in Brush Creek valley is usually better drained and has higher chroma in the B horizon than the Ettrick soil.

Unit D has relatively well-developed soils formed in silt loam and sandy loam. Soil D is similar to, but not as strongly developed as, Soil E. Soil D is easily distinguished from younger soils, Soils C and B, on the basis of stronger pedogenic structure, lower values and higher chromas in the subsoil, and the presence of a B horizon. The B horizon probably qualifies as a cambic horizon in most sections, based on pedogenic structure and slightly higher chroma in the B horizon than in the C horizon. The chroma of the B horizon and argillan development are slightly weaker in Soil D than in Soil E. Soil D generally resembles the Ceresco series (fluvaquentic hapludoll), although some exposures more closely resemble the Coffeen series (fluvaquantic hapludoll, with siltier texture than the Ceresco) or the Boaz series (mollic haplaquept).

Soils developed in Brush Creek alluvium after 4500 B.P. are very different from those which began developing before 4500 B.P. The younger soils (Soils C and B) do not have B horizons, and they have gleyed subsoil with low chroma. Additionally, the younger soils are typically rich in detrital organic material, exhibited in thin, black or nearly black strata of silty clay loam, and strata of sandy sediment rich in organic debris. Gleying and preservation of detrital organic material indicate soil genesis under relatively wet conditions. In contrast, the early and middle Holocene soils are oxidized and low in detrital organic material, indicating a drier soil environment during at least part of the soil's history.

The main features of pedogenic development in Soil C and Soil B are the A1 horizon and mottling. Generally

TABLE 2.

Summary of Pedogenic Characteristics

Soil Type	Age (yr B.P.)	Type of soil profile	B or C horizon color: Hue	Value	Chroma	Grade of pedogenic structure	Thickness of historical alluvium, cm	Other criteria
Soil A (historical alluvium)	75	A/C	10YR	2 - 4	1	very weak		horizontal laminae
Soil B	1500 to 2500	A/C	10YR	2 - 3.5	1	weak (massive)[a]	100 - 150	
Soil C	4000	A/C or O/A/C	10YR, 5YR	3.5 - 5	1 - 2.5	weak (moderate)	70 - 90	peaty layer
Soil D	4500 to 5200	A/B/C	10YR	4 - 6	2 - 3.5	moderate to weak	40 - 85	few argillans in B horizon

(TABLE 2 continued)

Soil E	7300 to 8500	A/Bt/C	10YR (7.5YR)	4 - 6	4 (3)	moderate (strong)	0 - 30	argillans in B horizon
Soil F	variable	variable		variable		variable	variable	gravelly layers
Soil G	about 11,000	A/Bt/C	10YR	4	4	strong to moderate	none	forms high terraces; argillans

[a]Parentheses indicate a condition which was noted in a few profiles but is not the dominant condition.

the late Holocene soils display little or no development of pedogenic structure, and alluvial stratification is preserved below the A1 horizon. At two of the five sections studied, Soil C has a peaty O horizon, about 15 cm thick, overlying the A1 horizon. At one of these sites (SE1/4 SW1/4 and SW1/4 SE1/4 sec. 35, T. 15 N., R. 3 W., Monroe County), Unit C grades laterally into an extensive body of peat which is up to 2 m thick. Accumulation of this peat apparently began at the same time as deposition of Unit C, about 4500 B.P. There is no evidence of peat accumulation associated with Soil B.

Soils C and B are both capped by 70 cm or more of historical alluvium (Unit A). This historical alluvium is considered part of the solum when these soils are classified and mapped in soil surveys. Soils C and B, with their overlying historical alluvium, are usually represented on soil maps as Kickapoo or Arenzville series (typic udifluvents), Orion series (aquic udifluvent), or Lawson series (cumulic hapludoll).

Unit A is by far the youngest stratigraphic unit in Brush Creek valley, and it exhibits no evidence of pedogenic development other than an incipient A1 horizon. Unit A is easily recognized in the field on the basis of its pale color, clearly preserved depositional structure, and abrupt lower boundary. The A1 or Ap horizon, in the upper 15 to 20 cm of Unit A, is characterized by weak granular or platy structure, and very dark grayish brown color (10YR 3/2). In this horizon, depositional structure has been largely destroyed. Below the uppermost horizon, Unit A displays the depositional structure typical of overbank deposits, horizontal laminae 1 mm to 1 cm thick. The color of the subsoil is about the same as the color of the surface horizon. A large part of the soil organic matter in Soil A probably was derived from the eroded hillslope soils which were the source of sediments forming Unit A. At the base of Unit A there is an abrupt, straight boundary with the underlying A1b horizons of soils B through D. The A1b horizons are darker (10YR 2/1 or darker) than the A1 horizon of Soil A.

THE INFLUENCE OF TIME AND CLIMATIC CHANGE ON SOIL DEVELOPMENT

The sequence of six major soil types studied in Brush Creek valley provides a useful illustration of the effect on soil morphology of the soil-forming factors--climate, organisms, relief, parent material, and time. Topographic differences among soils B, C, D and E are very slight, and the prehistorical vegetation cover on these soils was probably very similar. Before agricultural land clearance, the vegetation cover of Driftless Area valley bottoms included both lowland hardwood forests and wet prairies (Curtis, 1959; Finley, 1976), and Finley indicates that Brush Creek valley was occupied mainly by forest, with an area of brush at the eastern end of the valley. Vegetation cover probably was controlled by valley width and nearness to the valley walls rather than by the distribution of the stratigraphic units. Vegetation cover on the valley bottom may have varied in response to Holocene climatic changes. Soil G probably had somewhat different vegetation cover and soil drainage conditions due to its topographic position on the high terraces. Parent material is also relatively uniform among the six soil types, except for the slightly sandier texture of Soil D. Relief, vegetation cover and parent material do not account for the morphological differences among the soils in this sequence.

I believe that most of the differences in the soil sequence are due to the effects of time and climatic change. The soil sequence certainly shows the effects of time on soil morphology. The degree of soil profile development, particularly B horizon development, is strongest in the oldest soil (Soil G), and it decreases with decreasing age of soil. Additionally, the older soils in the sequence experienced a range of climatic conditions different from those experienced by the younger soils, as a result of Holocene climatic change in the region.

Ruhe (1974) pointed out the importance of past environmental conditions, particularly climate, in explaining the morphology and distribution of soils of Holocene age. According to Ruhe, correlation between soil type and modern climatic conditions may be largely coincidental. In Brush Creek valley, the same can be said for other environmental characteristics which are commonly thought to be responsible for differences among valley floor soils--elevation (first bottom vs. high bottom), and soil drainage classes. Many of the differences among soils mapped on valley floors in the Driftless Area may be due more to age and past climatic conditions than to modern environmental conditions.

Brush Creek valley, and the Driftless Area in general, have experienced moderate climatic changes during the Holcocene, as documented by pollen records from a number of lakes and bogs in the Upper Midwest (Bartlein and Webb, this volume). Rapid warming at about 11,300 B.P. marked the end of the glacial climate (Webb and Bryson, 1972). Warming at a moderate rate and decrease of precipitation continued until about 7000 B.P. The climate remained relatively warm and dry until about 4500 B.P. Since 4500 B.P. precipitation has increased. The period between 8000 and 4500 B.P. was the maximum in warm and dry conditions, as evidenced by the migration of prairie into the uplands of the Driftless Area (Bartlein and Webb, this volume).

These climatic variations probably caused changes in a number of factors influencing soil genesis: precipitation, evapotranspiration, water table height in the valley bottom, and vegetation. In particular, the warm/dry maximum at 8000 to 4500 B.P. left its mark on the older soils in the Brush Creek valley sequence. Soils G, E and D experienced all or part of the warm/dry maximum. These soils all show B horizon development, at least incipient clay translocation, and oxidized soil colors. Soils which started forming after about 4500 B.P. (Soils B and C) do not have B horizons.

It is particularly remarkable that Soil D more closely resembles Soil E (2000 to 3200 yr older) than it resembles Soil C (only 500 to 1200 yr younger than Soil D). The difference in age between Soil D and Soil C does not seem sufficient to account for the contrasts between these two soils. I believe the evidence indicates that soil environmental conditions which allowed clay translocation and oxidation existed during the early and middle Holocene, and these conditions ceased to exist in the valley bottom environment by about 4500 B.P. Clay translocation occurs in seasonally dry soils (Soil Survey Staff, 1975, p. 19). During the dry season, colloidal clay is translocated downward by movement of water into relatively dry subsoil, and withdrawal of this water into capillary-sized pores deposits the clay on ped surfaces. Conceivably the increase in precipitation at about 4500 B.P. could have raised the water table in the valley bottom sufficiently to stop clay translocation and prevent oxidation of the subsoil of the late Holocene soils. An increase in wetness in the valley bottom at about 4500 B.P. is also supported by the fact that peat accumulation in Brush Creek valley began at this time. If this explanation of environmental change is correct, then the argillic horizons of Soil E and argillans of Soil D must be considered relict features. In contrast to Soils E and D, Soil G, located on the high terraces, was probably not affected by the rise in water table. Presumably the conditions favoring clay translocation and oxidation have been present on the high terraces throughout the late Holocene.

LENGTH OF TIME NECESSARY FOR Bt HORIZON FORMATION

A number of earlier workers have estimated the length of time necessary for formation of a Bt horizon. Parsons, Scholtes and Riecken (1962) found that, on loess-derived Indian mounds in northeastern Iowa, Bt horizons formed in about 2500 yr. Dietz and Ruhe (1965) reported on Bt horizons in silty alluvial soils less than 2000 yr old in western Iowa. Fenton, Dietz and Riecken (1974) found that argillic horizons formed in less than 6000 yr but more than 2000 yr in loess in central Iowa. In Pennsylvania, Bilzi and Ciolkosz (1977) studied a 2000 yr old soil in alluvium with a well-developed cambic horizon grading to an argillic horizon.

The evidence from Brush Creek valley supports these results. Argillic horizons are not forming under present conditions in Brush Creek valley, but they did form prior to 4500 B.P. Soil D, which experienced clay translocation for 700 yr or less, does not have an argillic horizon. Soil E, which experienced clay translocation for periods ranging from 4000 to 2300 yr, does have an argillic horizon. Therefore, the length of time necessary for formation of an argillic horizon under mid-Holocene conditions was less than 2300 yr but more than 700 yr.

SUMMARY

Soils formed on colluvial and alluvial deposts of Late Wisconsinan and Holocene age were studied. Six major soil types, associated with six major stratigraphic units, were identified. The age of each soil type was determined based on radiocarbon dates and stratigraphic relationships. Four orders, mollisol, alfisol, inceptisol and entisol, and eight subgroups are represented in this soil sequence. Age and climatic history account for most of the differences among the six soil types. Soils dating from before 4500 B.P. retain relict features of a different soil-forming regime which was active during the mid-Holocene warm/dry period.

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APPENDIX

TYPICAL SOIL PROFILE DESCRIPTIONS

Soil B (with Unit A):
Section PWJ1: NE1/4 NW1/4 sec. 3, T. 14 N., R. 2 W., Vernon County; Powell farm; bank exposure on right (east) bank of Brush Creek located about 40 m downstream from State Highway 33 bridge.

depth cm	description
0-105	Historical overbank deposits (Unit A); 10YR 3/2 silt loam; weak, coarse and medium, subangular blocky structure; includes few layers of pale, coarse silt (0.5-1.0 cm thick); friable; abrupt boundary.
105	Top of Soil B and Unit B.
105-140	Alb horizon; 10YR 2/1 loam; weak to moderate, medium, subangular blocky structure; friable; clear boundary.
140-165	C horizon; 10YR 3/1 sandy loam; very weak, coarse, angular blocky structure; friable; few, medium, faint, orange mottles; gradual boundary.
165-210	2.5Y 3.5/1 loamy sand to sand; massive; faint strata of paler sand and darker silty clay loam, sandier layers becoming more common with depth; strata 0.2-0.5 cm thick; contains wood fragments near base.
210	Water level.
305	Approximate level of wood sample dated at 2065 ± 65 B.P. (WIS-808).

Horizon	Depth, cm	% Sand	% Silt	% Clay
Unit A	20 - 25	36.6	49.4	14.0
Unit A	40 - 45	65.2	26.2	8.6
Unit A	60 - 65	27.9	56.8	15.3
Unit A	80 - 85	24.5	58.1	17.4
Unit A	100 - 105	30.6	53.3	16.1
Alb	110 - 115	46.7	39.4	13.9
Alb	125 - 130	50.8	36.9	12.3
C	150 - 155	62.3	27.9	9.8
	170 - 175	75.6	17.9	6.5
	185 - 190	82.5	11.4	6.1
	205 - 210	89.6	6.7	3.7

Soil C (with Unit A):

Section KUA2: North center, SW1/4 sec. 35, T. 15 N., R. 3 W., Monroe County; Kuder farm; bank exposure on right (north) bank of Upper Brush Creek, about 100 m east of fence along property line.

depth cm	description
0-50	Historical overbank deposits (Unit A); 10YR 3.5/3 sandy loam; weak, fine, platy and angular blocky structure; clear, fine horizontal stratification, with some thicker (1 cm) layers; clear boundary.
50-82	Historical overbank deposits (Unit A); 10YR 4/2 sandy loam; weak, fine, platy and angular blocky structure; clear, fine horizontal stratification, with some sandy layers 1-3 cm thick; few, fine, orange (7.5YR 4/3) mottles, increasing with depth; abrupt boundary.
82-88	Alb horizon; 10YR 3/1 silty clay loam; fine gray and black stratification; common orange mottles; abrupt boundary.
88	Top of Soil C and Unit C.
88-104	Olb horizon; 10YR 1.7/1 (black) fibrous peat; clear boundary.
104-116	Alb horizon; 10 YR 1.7/1 silty clay loam; massive; common fibers and roots; clear boundary.
116-149	C horizon; 10YR 3.5/1 silt loam; massive; abrupt boundary.
149-178	Irregular strata (1-3 cm thick) of white sand and black silty clay loam; 10YR 3/1 loamy sand mixed; abrupt boundary.
178	Wood dated at 4410 ± 75 B.P. (WIS-1022).
178-210+	Gravels.
210	Water line.

Horizon	Depth, cm.	% Sand	% Silt	% Clay
Unit A	20 - 25	71.1	22.0	6.9
Unit A	35 - 40	56.2	34.1	9.7
Alb	108 - 112	19.3	51.7	29.0
C	120 - 125	27.9	55.2	16.9
C	135 - 140	36.6	45.7	17.7
	159 - 165	83.5	10.3	6.2

Soil D (with Unit A):

Section LEE1: SE1/4 NW1/4 sec. 35, T. 15 N., R. 3 W., Monroe County; Peter Leis pasture; bank exposure on left (north) bank of Upper Brush Creek on downstream end of meander bend, about 150 m west of road.

depth cm	description
0-40	Historical overbank deposits (Unit A) 10YR 3/2 silt loam; moderate, medium, angular blocky structure; faint, fine, horizontal stratification; very friable; few, medium, faint orange mottles; common inclusions of darker silt loam from below; clear boundary.
40	Top of Soil D and Unit D.
40-77	Alb horizon; 10YR 1.7/1 loam; moderate to weak, medium, subangular blocky structure; very friable; few, fine, faint orange mottles in lower part of horizon; clear boundary.
77-85	Bl horizon; 10YR 4/2 silt loam; moderate coarse and medium, subangular blocky structure; friable; common, fine, faint orange mottles; inclusions of darker loam from above; clear boundary.
85-111	B2 horizon; 10YR 4/3.5 silt loam; weak, medium subangular blocky structure; friable; common, fine, light orange, and few, fine, dark orange mottles; few, faint, patchy, brown cutans and void fillings; clear boundary.
111-130	B31 horizon; 2.5Y 5/3 sandy loam; moderate to weak, coarse and medium, subangular blocky structure; firm; mottles as above; clear boundary.
130-160	B32 horizon; 10YR 4.5/6 sandy loam; weak, coarse, subangular blocky structure; firm; common, coarse, medium orange mottles; few *in situ* roots and black organic inclusions; gradual boundary.
160-250	C horizon; 10YR 3/1 sandy loam; massive; firm; few, coarse, orange mottles; few roots and organic inclusions; abrupt boundary.
188	Water level.
230	Wood sample dated at 5145 ± 65 B.P. (WIS-1071).
250+	Gravels; top of gravels varies between 230 and 270 cm.

Horizon	Depth, cm.	% Sand	% Silt	% Clay
Unit A	20 - 25	13.6	70.4	16.0
Alb	45 - 50	44.7	41.1	14.2
B2	87 - 92	34.4	49.6	16.0
B31	115 - 120	64.2	24.9	10.9
B32	145 - 150	56.9	28.3	14.8
C	175 - 180	56.0	32.0	12.0

Soil E:

Section HEA1: NE1/4 SW1/4 sec. 36, T. 15 N., R. 3 W., Monroe County; Helmuth farm; bank exposure on right bank of Upper Brush Creek, about 200 m downstream from barn.

depth cm	description
0-34	A1 horizon; 10YR 1.7/1 silt loam, becoming paler with depth; moderate, medium and fine, subangular blocky and granular; friable; clear, irregular boundary.
34-46	AB horizon; 10YR 4/3 and 1.7/1 silt loam; moderate, medium and fine, subangular blocky structure; friable; clear, irregular boundary.
46-79	B1 horizon; 10YR 4/3 to 5/4 silt loam; strong, coarse and medium, subangular blocky; firm; pale, dusty ped faces; few, faint, patchy argillans; gradual boundary.
79-115	B21 horizon; 10YR 4/4 silt loam; strong, coarse and medium, subangular blocky; firm; few, faint orange mottles; ped faces not as pale and dusty as above; some distinct argillans; gradual boundary.
115-140	B22 horizon; 7.5YR 6/4 silt loam; moderate, coarse, subangular blocky; firm; few, coarse, orange (5YR 4/6) and fine, dark red mottles; few argillans; abrupt, smooth boundary.
140-146	C1 horizon; stratified 10YR 5/3 and 7.5YR 4/6 silt loam; fine, horizontal strata (0.5-0.3 mm thick), continuous for length of exposure; abrupt boundary.
146-163	C2 horizon; 7.5YR 6/3 and 5/4 silt loam; weak, coarse, subangular blocky structure; heavily mottled; irregular inclusions of pale strata from above; clear boundary.
163-199	C3 horizon; 2.5Y 6/2 silt loam; weak, coarse, subangular blocky; common, medium and coarse, orange (5YR 4/6) and red (2.5YR 3/5) mottles; abrupt boundary.
199-270	10YR 4.5/1 silt loam; massive; some irregular lenses of sand; common, coarse, orange (7.5YR 5/4) mottles; contains woody debris; abrupt boundary.
250	Sample dated at 9060 ± 95 B.P. (WIS-1018).
255	Water level.
270	Approximate level of top of gravels.

Horizon	Depth, cm.	% Sand	% Silt	% Clay
A1	9 - 15	33.1	53.9	13.0
A1	21 - 26	34.0	52.7	13.3
AB	38 - 42	37.8	51.5	10.7
B1	50 - 55	36.5	51.2	12.3
B1	65 - 70	22.9	60.6	16.5
B21	85 - 90	29.2	51.9	18.9
B21	105 - 110	18.0	63.0	19.0
B22	120 - 125	12.2	66.4	21.4
C1	140 - 146	8.1	69.9	22.0
C2	153 - 158	17.4	64.3	18.3
C3	170 - 175	7.0	66.4	26.6
C3	185 - 190	5.5	69.9	24.6
	205 - 210	21.5	61.1	17.4

Soil G:

Section VRLL: NW1/4 SE1/4 sec 35, T. 15 N., R. 3 W., Monroe County; exposure at top of bluff (about 10 m high) undercut by Upper Brush Creek. Bluff is located on right (south) side of Upper Brush Creek, downstream from Von Ruden farmhouse.

depth cm	description
0-40	A1 horizon; 10YR 2/2 silt loam; clear boundary.
40-115	B2t horizon; 10YR 4/4 silt loam to silty clay loam; strong, medium, subangular blocky structure; well-developed brown argillans on ped faces.
115-190+	B3t horizon; 10YR 5/4 to 4/4 s)lt loam to silty clay loam; moderate, medium, platy structure; argillans less prominent than above.

Horizon	Depth, cm.	% Sand	% Silt	% Clay
A1	20 - 30	24.0	60.6	15.4
B2t	60 - 70	12.9	62.9	24.2
B2t	100 - 110	6.9	64.8	28.3
B3t	140 - 150	6.8	64.4	28.8
B3t	180 - 190	9.6	64.4	26.0

GLACIATION OF THE DRIFTLESS AREA: AN EVALUATION OF THE EVIDENCE

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INTRODUCTION

For more than 150 years people have recognized that the characteristic topography and lack of erratic clasts in southwest Wisconsin demonstrate a significantly different Quaternary history from surrounding areas. Most writers have concluded that this area was never glaciated, and the term "Driftless Area" has been used for more than 100 years. Two writers (Sardeson, 1897; Squier, 1897, 1898) questioned the validity of the Driftless Area concept during the last century but most other authors have accepted the concept that evidence of glaciation of the Driftless Area is lacking. Not until 1960, when Black (1960, p. 1827) published an abstract stating that evidence "confirm(s) glaciation of much, if not all, the Driftless Area during the early Wisconsinan stage or all of it during a pre-Wisconsinan stage", was the concept of an unglaciated area in southwest Wisconsin seriously questioned. Since that time evidence has been discussed by Akers (1964), Frye, Willman, and Black (1965), Palmquist (1965), Black and Rubin (1968), and Black (1970a,b).

The concept of a glaciated Driftless Area appears to have been accepted by many people, and Black's interpretations have never been critically evaluated in print. Although we cannot prove that the Driftless Area has never been glaciated, we do feel that evidence of glaciation is weak.

In this paper we will examine published evidence and arguments for glaciation of the Driftless Area. In particular, we will try to answer three questions:

1. Is there evidence demonstrating that the Driftless Area was glaciated?
2. If it was glaciated, when did the glaciation take place?
3. What are the boundaries of the driftless area?

EVIDENCE FOR GLACIATION

Early evidence of Squier and Sardeson

Sardeson (1897) examined railroad cuts in the vicinity of Dodgeville and the Pecatonica River valley south of the Wisconsin River (fig. 1, location 1). He found unstratified deposits of chert, limestone, and sandstone blocks in a matrix of silt and clay that he interpreted to be deposits of a local valley glacier. Apparently the deposits contain no erratic material or other features typically associated with glacial deposits (for example, striated clasts). Smith (1949) reexamined the deposits and concluded that they appear to be products of mass wasting from the valley side. Squier's papers provide little information on location, so it is impossible to relocate his exposures. The deposits he describes also appear to be mass-wasting deposits and possibly fans similar to those described by Knox (1980, 1981). He attributed the rounded, bowl-shaped appearance of valley heads to a possibly glacial origin but admits that evidence of glaciation is weak. In the summary of his 1897 paper, he states "while these deposits must be regarded as essentially non-glacial, there does not appear to be anything inconsistent with the assumption that occasionally during periods of exceptional activity, glaciers may have advanced on to them for short periods" (Squier, 1897, p. 192).

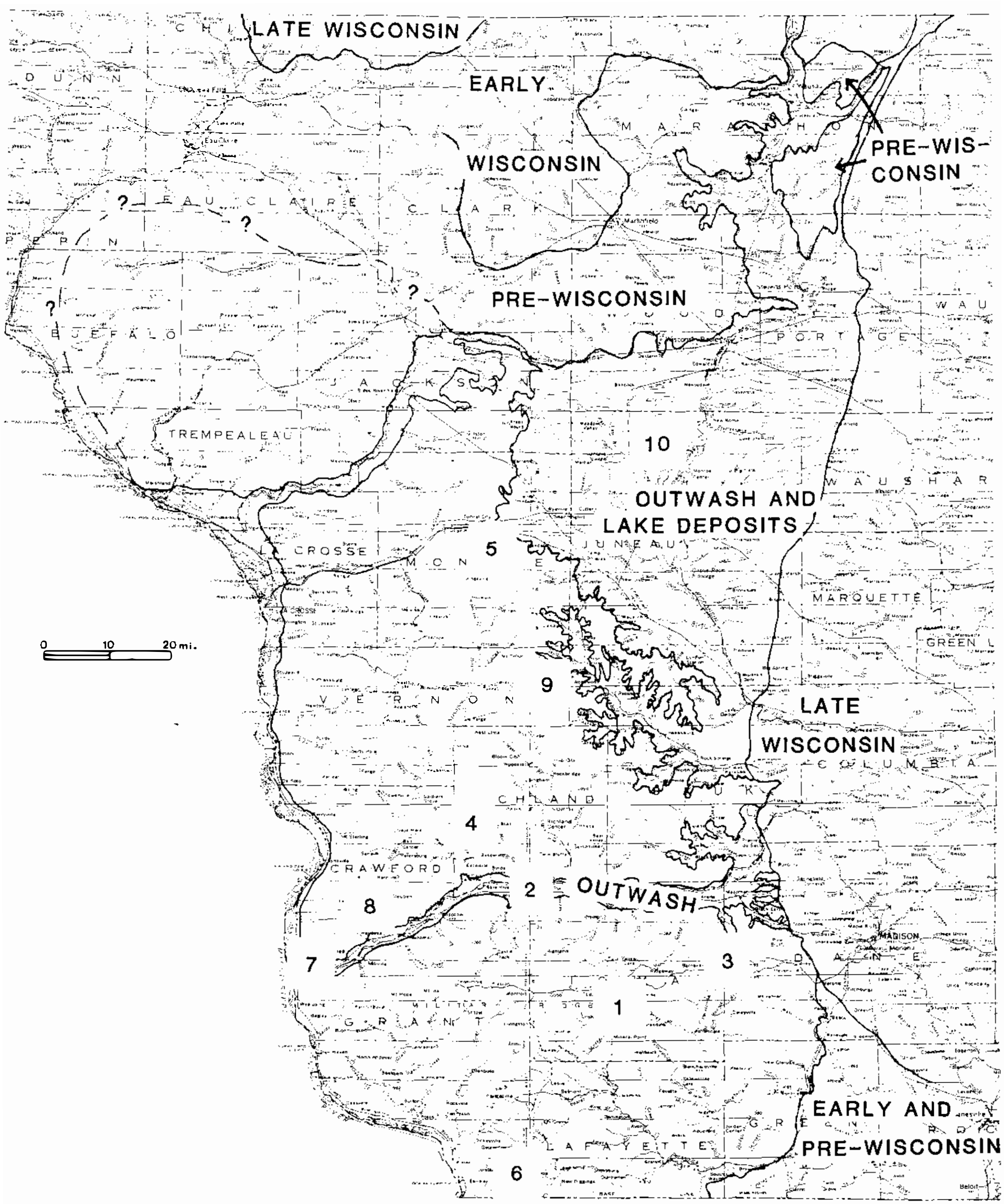

FIGURE 1.--Map of Driftless Area of Wisconsin showing age of boundaries and locations mentioned in text. Scale 1:1,000,000.

The evidence provided previous to 1960 cannot be taken to document glaciation in the Driftless Area. Both Sardeson and Squier suggested the presence of local glaciers, and it is indeed possible that perennial snowfields did occur in some places in the Driftless Area but better evidence is needed.

Evidence published since 1960

Black (1960), in a short abstract, stated that the Driftless Area was glaciated, much of it during "Rockian" (late Altonian) time. Unfortunately, the evidence provided is very general and has been only briefly expanded upon since (Frye, Willman, and Black, 1965; Black and Rubin, 1968; Black, 1970a, b). Several of Black's students, however, pursued this theme, and some details of the evidence are provided by theses of Akers (1961, 1964) and Palmquist (1965).

We will now evaluate each of the arguments in as much detail as possible.

Erratics near Muscoda

The Quaternary history of the Wisconsin River Valley is discussed in another section of the guidebook, and we will not repeat this discussion here. It is important to note, however, that one evidence of glaciation used by Black (Frye, Willman and Black, 1965; Black 1970a) is the presence of erratic boulders and cobbles near Muscoda in the center of the Driftless Area (fig. 1, location 2). These were described by MacClintock (1922), and we have examined fresh exposures and drilled in the deposits. All of the Precambrian erratic materials are associated with fluvial-glacial outwash, including the oldest sediments in the Bridgeport terrace (Knox, Attig, and Johnson, this volume). We see no evidence that these deposits indicate glaciation of the Driftless Area.

Distribution of erratics

Black (1965, p. 54) stated that "positive evidence of glaciation in the "Driftless Area" of Wisconsin comes from abundant fragments of Precambrian igneous rocks and Paleozoic chert and sandstone that rest on younger formations". Unfortunately, with the exception of one Precambrian pebble found below loess on West Blue Mound (Black, 1970a), no other sites (away from the boundary of the Driftless Area) where Precambrian erratics exist have been given except in sites described by Akers (1964). We (especially Knox) have spent a large amount of time doing detailed field work in the Driftless Area and have never seen an erratic that was clearly of ice-contact glacial origin. All of the counties in the central part of the Driftless Area have been mapped recently by field parties of the Soil Conservation Service; although scattered upland erratics associated with early glaciations are present along the edges of the Driftless Area, there have been no reports of erratics beneath loess found in the Driftless Area (A. J. Klingehoets, F. Anderson, R. Cheetham, Soil Conservation Service, Madison, personal communication).

We feel that the use of the distribution of supposed erratics to document glaciation of the Driftless Area is unsound. There is no documentation of erratics being present beneath loess except in outwash derived from outside the Driftless Area and in areas along its northeast and southern boundaries. Knox has observed crystalline erratics of cobble size at an upland archeological site in north-central Grant County. Scars and markings on these erratics indicate they were being used as tools by the Indians. Black (1965) reports upland gravel with erratic material near Hazel Green (fig. 1, location 6) in southern Grant County. Although we have not examined this site, it is in an area of upland outwash near the southern boundary of the Driftless Area. Similar occurrences are reported in Illinois (J. C. Frye and H. B. Willman, 1981, personal communication to Knox).

The development of East Blue Mound

Along the drainage divide south of the Wisconsin River, East and West Blue Mound stand over 100 m above the surrounding landscape (fig. 1, location 3). The geomorphic development of the area has been discussed by numerous authors, and a brief history of ideas is given by Black (1970a). Our discussion here is confined to the supposedly glacial origin of the flat top of East Blue Mound.

West Blue Mound, the higher of the two, is capped by silicified Silurian dolomite. East Blue Mound is capped by the Maquoketa Formation, a unit containing shale and limestone (fig. 2). Black (1970) has pointed out the broad, nearly flat top of East Blue Mound (fig. 3) and concluded "thus, only one agency--that of glacial ice--seems capable of beveling the top of East Blue Mound to so flat a surface" (Black, 1970a, p. H-10).

As noted by Black (1970) and Smith (1949) the siliceous Silurian dolomite is very resistant to erosion, weathering to large blocks that are now scattered around the slopes of West Blue Mound. However, the degree of silicification over east Blue Mound or the surrounding area is unknown. It is possible that a considerable difference in the amount of silicification led to the preservation of the cap on West Blue Mound, but other factors such as joint density and the location of headwater streams could have been equally responsible.

It seems more likely that the present landscape was produced by slope processes and fluvial erosion which were perhaps more severe during glacial time. The peculiar flat top on East Blue Mound is probably caused by removal of shale down to dolomite units in the Maquoketa Formation.

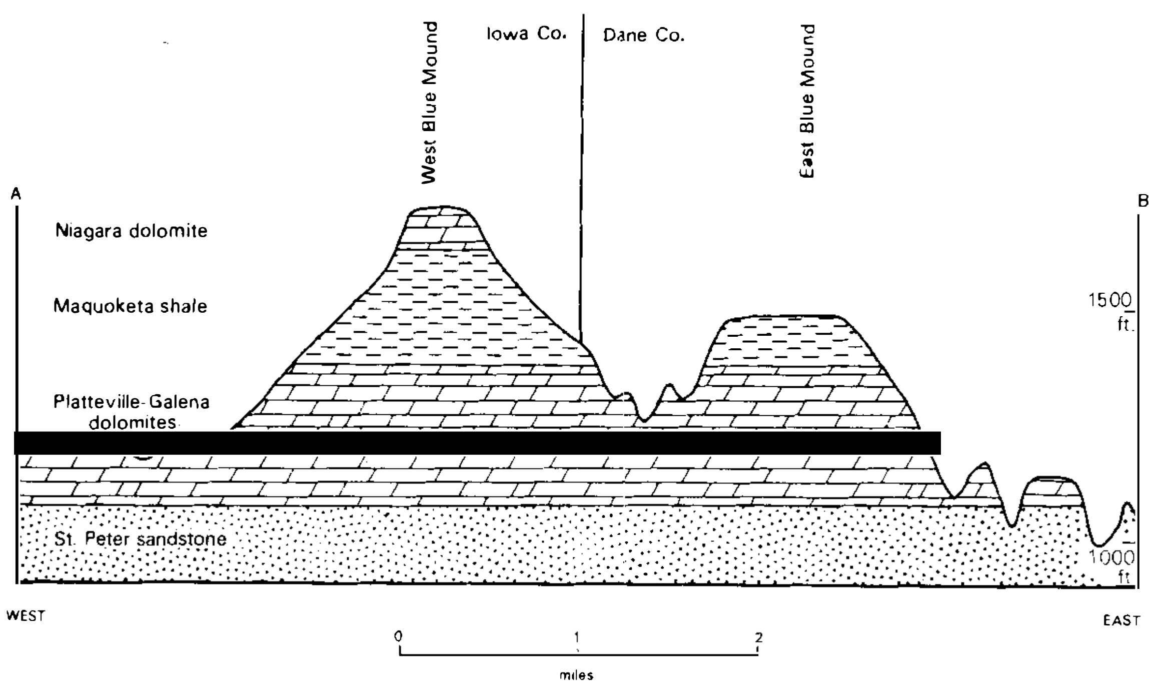

FIGURE 2.--Cross-section from east to west across East and West Blue Mounds. From Black (1970a).

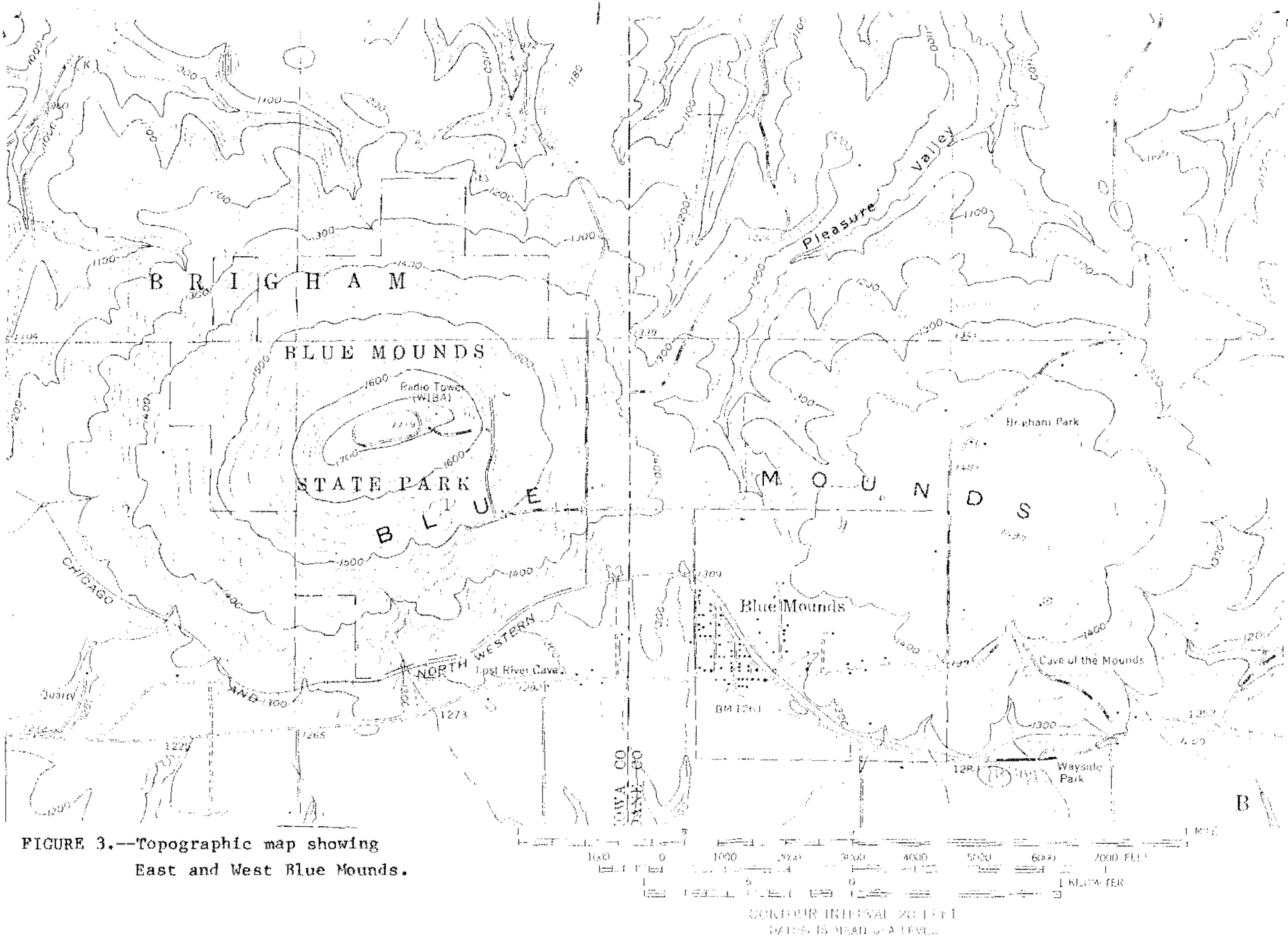

FIGURE 3.--Topographic map showing
East and West Blue Mounds.

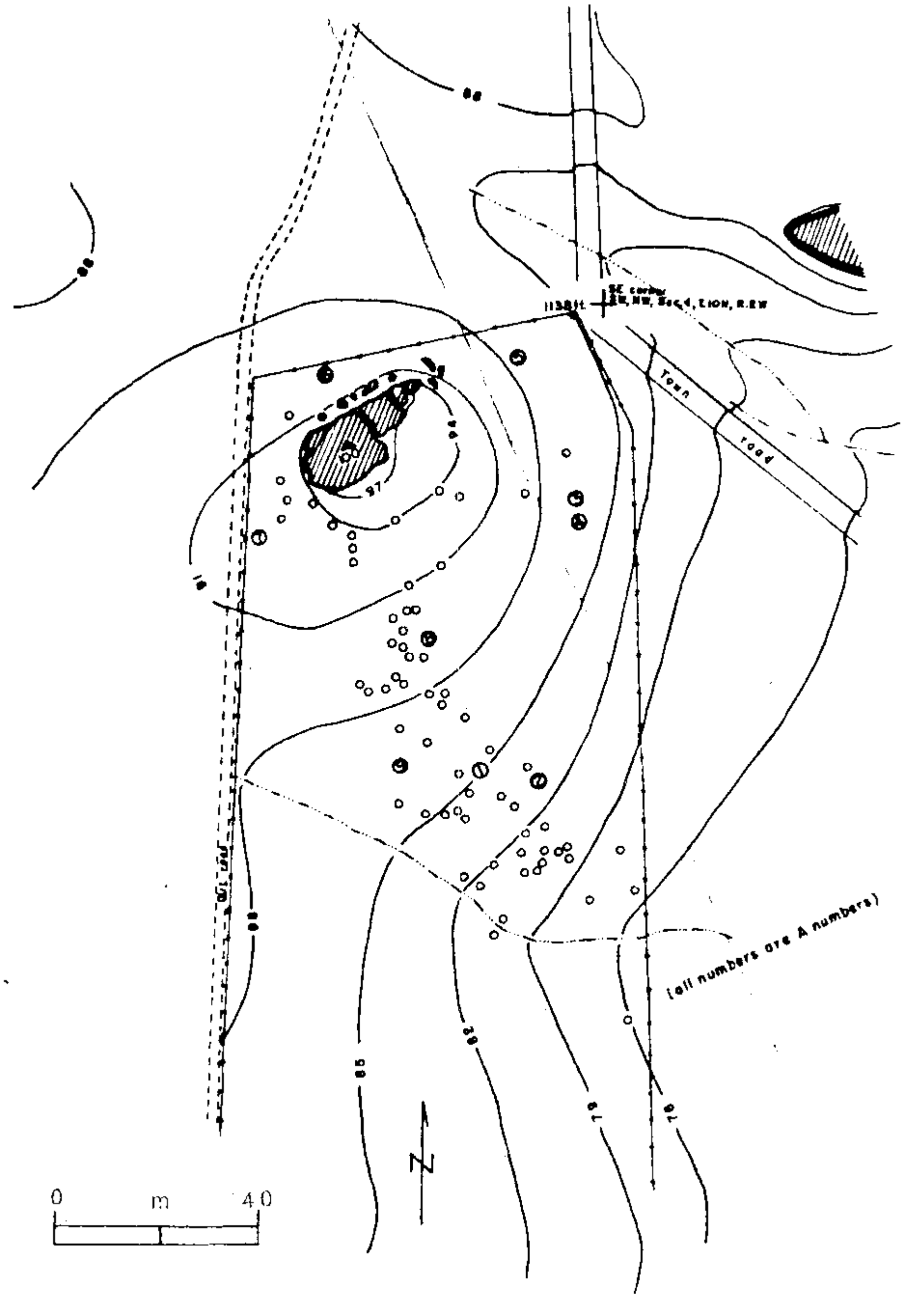

FIGURE 4.--Contour Map showing boulder distribution circles around low sandstone pinnacle (cross hatched) near Bosstown. From Akers (1965).

Boulder train near Bosstown

Black (Frye, Willman and Black, 1965) has inferred the direction of ice flow in the Driftless Area from a "boulder train" described in detail by Akers (fig. 1, location 4). Black concluded that the distribution of blocks around a low sandstone pinnacle "is so striking that it suggests glaciation" (Frye, Willman and Black, 1965, p. 54). A map of the "boulder train" is shown in figure 4. Note that the blocks are distributed nearly perpendicular to the contours, not "obliquely across the slope" as described by Akers (1964, p. 40). It seems likely that the arrangement of boulders simply indicates the toppling of a high sandstone pinnacle toward the southeast, followed by downslope movement that carried the blocks further downslope. We see no reason to invoke glaciation to explain the distribution of boulders.

Erratics and other evidence of Akers (1964)

In his Ph.D. thesis, Akers described not only the "boulder train" near Bosstown but several other features that he considers indicative of glaciation. Between Boaz and Tomah (fig. 1) he described 17 sites with cobbles and boulders above bedrock. At some locations the boulders are below or mixed with loess, but evidently at many the cobbles and boulders were on the surface.

At twelve of these locations he concluded that the lithologies present were what would be expected from the underlying or overlying lithologies. At four locations (DD, I, H, and G, Akers, 1964) he found what he interpreted to be chert of the Oneota Formation (Prairie du Chien) on St. Peter Sandstone or younger stratigraphic units. If his interpretation of the age of the chert at these locations is correct, a process such as glaciation is necessary to explain the fragments of chert atop younger stratigraphic units.

Because these are localized occurrences they are difficult to explain. It may be that the younger Ordovician dolomite units do contain oolitic chert at these localities and that the chert is not out of place. The scattered blocks, up to 30 cm in diameter, could have been brought to ridge tops by Indians. Concentrations of artifacts of chert and often quartzite have been found on ridge-tops at some places in the Driftless Area (Stoltman, 1982, personal communication).

Akers (1964) found several sites near Tomah (fig. 1, location 5) where gravel and a presumed till indicated glaciation. At one location Akers reported a deposit of red, pebbly, stoney clay that he interprets as till. All of the included clasts have a local origin (Akers, p. 114), and the interpretation of the material as till is based on its topographic position atop a ridge. Because of the possibility of changes in valley-wall positions, we do not feel that this is a strong argument. In fact, the site is close to the base of a long, high hillslope and is only separated from it by a low saddle. We interpret this to be an ancient mass-movement deposit.

At another location (EE, Akers, 1964) a supposed kame is reported. The material at this site was considered by Andrews (1958) to be reworked Windrow Formation. The site contains several meters of sand and gravel, which reportedly contains a small proportion of igneous pebbles. Although the cut is covered with a scatter of igneous pebbles from the adjacent roadbed, we searched with about 10 graduate students for over 1 hour and were unable to find any igneous pebbles within the unit itself. This does not prove that igneous rocks are not present, but the deposit is clearly not typical of ice-contact stratified deposits in other parts of the state. The Windrow Formation, which has been considered to be Cretaceous, contains igneous pebbles in places, carried by streams flowing from areas of Precambrian outcrop to the north.

Thus, the evidence of Akers raises some questions concerning the position of the northeast boundary of the Driftless Area, but it does not prove that the Driftless Area was glaciated. We believe that Black's statement that "isolated deposits explainable only by glaciation are on the crests of ridges in all but La Crosse County in the classical "Driftless Area" of southwest Wisconsin" (Black, 1960, p. 1827) overstates the case. In fact, there are no deposits described by Black or others that suggest more than a slight change in the position of the boundary, and even this evidence is not convincing.

Residuum and Paleosols on Uplands

Upland interfluves in the Driftless Area often contain relatively shallow thicknesses of red clay residuum in relation to an amount that might be expected from a long period of weathering. Black (1970b) concluded that the small thicknesses of residuum resulted from stripping of uplands by glacial ice. We find this conclusion unacceptable because there are no striated rocks or erratics to indicate glaciation. We conclude that most of the upland residuum has been removed by accelerated mass wasting and slope wash during episodes of periglacial climate (Knox and Johnson, 1974; Knox, 1980, 1981). Evidence of severe mass wasting during Woodfordian time is indicated by colluvial terraces, often up to 10 meters thick at the bases of hillslopes. These colluvial deposits include large very angular blocks of dolomite embedded in reworked Woodfordian loess.

Whittecar (1979) has shown that mass wasting was also rather intense during Altonian time in the southeastern Driftless Area. He identified solifluction deposits under peat that has a bottom radiocarbon date of 40,500 ± 1700 B.P. (ISGS-562). Our interpretation of severe upland stripping during the Wisconsinan is consistent with the conclusion of Frye and others (1969, p. 6) who in describing paleoenvironments for adjacent northwestern Illinois, concluded:

> "...of even greater significance are the unusually intense episodes of erosion to which the region has been subjected. This region was virtually surrounded by glacial ice during early Woodfordian time when the Green River Lobe was at its maximum extent. This glacial configuration contributed to a climate that accentuated solifluction, sheet wash, and eolian scour and deposition. Similar, but probably less intense conditions existed during late Altonian time..."

Black (1970b) suggested that pre-Woodfordian paleosols are very rare and that rareness might be attributed to past glaciations in the Driftless Area. We agree that Pleistocene paleosols of pre-Woodfordian age are not abundant, but we believe that fluvial erosion and mass wasting account for their absence in most places. However, because of the limited number of exposures on upland sites it is possible that pre-Woodfordian paleosols may be more common than believed. For example, a new roadcut through a narrow upland interfluve in southeastern Monroe County exposes a well-developed buried soil that separates Woodfordian loess from an older loess (Knox, Road Log, Stop 5, this volume). The Monroe county paleosol is located in a landscape position that would have been very vulnerable to erosion by glacial ice as postulated by Black. Its preservation argues against glaciation.

Frolking (1978, and this volume) found that the distribution of upland red clay residuum is nonrandom. He observed that the quantity of upland red clay varies closely with the distribution of chert-rich bedrock formations. Frolking indicated that red clays are characteristically absent on ridgetops underlain by the non-cherty upper Galena Dolomite as in northern parts of Grant and Iowa Counties, but where the cherty lower Galena Dolomite forms upland surfaces as in Green County, red clays are relatively thick. He also found abundant red clays overlying the chert-rich lower Prairie du Chien Formation along the north-central part of the Driftless Area in Monroe and Juneau Counties. The association of red clay with underlying chert-rich bedrock suggests that stone lines of chert lag deposits are important for the preservation of upland red clay. Where chert lag is not available most of the red clay has been removed by fluvial erosion.

Chert Rubble in the Valleys

Black (1970a; 1970b) and his students (Palmquist, 1965; Akers, 1964) have argued that valley fills in the Driftless Area are too small in volume and contain too little chert rubble relative to that which might be expected from the solution of many meters or tens of meters of rock. They suggested that the small quantities of fill and chert rubble could be best explained by removal through Pleistocene glaciation. We disagree with the glaciation hypothesis because (1) major valley incision by fluvial processes has occurred in the Driftless Area during the Pleistocene, (2) valley fills typically contain pre-Late Wisconsin alluvium of local origin, and (3) many valleys of the Driftless Area contain relatively large quantities of chert rubble.

It seems likely that most of the older valley fills have been removed by fluvial erosion during the Pleistocene. Frye (1974, p. 279) suggested that major valleys throughout much of the midcontinent region experienced maximum entrenchment during the early Kansan. Trowbridge (1966) concluded that major valley incision in adjacent northeastern Iowa occurred since the Nebraskan because glacial deposits of that age were restricted to upland sites but Kansan glacial deposits occurred at lower elevations on hillslopes and terraces. Glacial till that appears to be correlative with the northeastern Iowa Kansan till mapped by Trowbridge occurs on the Bridgeport strath (fig. 1, location 7) at the mouth of the Wisconsin River Valley (Knox, Attig, and Johnson, this volume). The strath surface underlying the till is approximately 85 m higher than the elevation of the adjacent bedrock valley floor (Trowbridge, 1954, p. 803), indicating major incision since the strath surface represented the valley floor. If the surface of the Bridgeport strath was the valley bottom elevation during the Kansan as suggested by Trowbridge (1954, p. 803), major entrenchment of valleys has occurred since then.

Other deposits in the region indicate that episodic aggradation and degradation of valley bottoms has occurred in post-Kansan time. The best evidence of post-Kansan erosion and deposition is represented by valley fill sediments in the Citron Valley (fig. 1, location 8) area of the lower Kickapoo River (Knox, Road Log, Stop 1, this volume). Here the buried bedrock valley floor in the cutoff bedrock valley meander of Citron Valley is approximately 6 m higher than the buried bedrock valley floor underlying the alluvium of the present course of the Kickapoo River adjacent to Citron Valley. Tests of remnant magnetism preserved in silts of the Citron Valley cutoff suggest that the fill is post-Kansan. This interpretation is supported by the observation that the bedrock floor of Citron Valley is incised approximately 30 m below the elevation of the bedrock surface on the Bridgeport strath of supposed Kansan age.

Drilling by Knox in many Driftless Area valleys has shown that pre-Late Wisconsinan valley fill deposits typically are restricted to protected sites such as the insides of valley meanders and to where bedrock straths have prevented erosion by lateral channel migration. The occurrence of the pre-Late Wisconsinan fluvial sediments at these protected sites is testimony to the important role that fluvial erosion has played in removing pre-Late Wisconsinan alluvium from elsewhere in the valleys of the Driftless Area.

Black (1970a) emphasized the paucity of pre-Woodfordian alluvium and colluvium in valley fills to support his argument for coverage of much if not all of the Driftless Area by ice as recently as about 30,000 years ago. Actually, pre-Woodfordian deposits are relatively common in most valleys of the Driftless Area as noted above. The pre-Woodfordian sediments appear to be as old as Kansan, as in the fluvial-glacial outwash on the Bridgeport strath and as young as Altonian and Farmdalian in most other valleys. None of these deposits, with the exception of the Bridgeport sediments at the mouth of the Wisconsin River Valley, suggests ice-contact deposition.

Naturally occurring glacially-derived erratics are restricted to outwash deposits in the major valleys that originate outside of and pass across the Driftless Area. Many of the stream valleys that originate within the Driftless Area contain a highly weathered sandy gravel underlying the Woodfordian sandy gravels (Knox, 1980). There are no radiocarbon dates to document whether the basal sandy gravel is older than 30,000 years, but its highly weathered condition suggests that it probably is at least as old as early Wisconsinan.

A radiocarbon date of 20,270 ± 650 B.P. (ISGS-558; Knox, 1980) was obtained for basal Woodfordian sediments 0.1 to 0.5 m above the surface of the weathered basal gravels in the Platte River valley of Grant County. In the bedrock gorge of the lower Kickapoo River adjacent to Citron Valley, the basal gravels are separated from overlying Woodfordian and Holocene sediments by an oxidized and leached silt and paleosol (Knox, Road Log, Stop 1, this volume). The widespread occurrence of pre-Woodfordian sediments in valley fills, especially those of apparent early Wisconsinan age, contradict Black's claim that glaciation has scoured the valleys relatively free of sediments older than about 30,000 years.

Black (1970a, p. H-10) argued that only one stream valley in the Driftless Area -- the headward reaches of the Baraboo River at Hillsboro (fig. 1, location 9) -- contained the quantity of chert rubble that might be anticipated from the weathering and erosion of upland bedrock. Black (1970a) used the valley fill stratigraphy of the East Branch Pecatonic River to illustrate his claim for limited chert rubble in stream valleys. However, Knox and Johnson (1974) showed that the anomalously small quantity of chert rubble in the Pecatonica system was a result of dominance by sandstone bedrock in the valley walls such that accessibility to chert rich limestones and dolomites by eroding streams has been very restricted in the late Pleistocene.

Whittecar (1979) conducted a detailed drilling program in the East Branch Pecatonic River valley and found that considerable quantities of silty solifluction lobes flowed from the hillslopes onto the floodplains during periglacial conditions of the Altonian. A peat bed overlying soliflucted sediments had top and bottom dates of 26,820 ± 200 B.P. (ISGS-561) and 40,500 ± 1700 B.P. (ISGS-562). The soliflucted sediments with the overlying dated peat are presently exposed as a terrace along the Pecatonica valley on the southeastern margin of the Driftless Area. Because Woodfordian sediments are inset below the Altonian sediments (Whittecar, 1979), it is apparent that mid-Wisconsinan valley incision occurred and may have removed some of the cherty rubble from the system.

We therefore conclude that where chert rubble is scarce the paucity usually is a result of either noncherty bedrock or Pleistocene fluvial erosion. Where chert-rich bedrock is more common, as in the Platte River system of Grant County, considerable quantities of pre-Woodfordian cherty-rich sandy gravels are present (Knox, 1980; Knox and Johnson, 1974).

Boundaries of the Driftless Area

The term "Driftless Area" has been used primarily in reference to the physiographic area in southwestern Wisconsin and northwestern Illinois. It is obviously a misnomer because glacially derived sand and gravel (outwash) and loess are present. In some cases the term has been used to indicate the unglaciated area, but in this case definition is a problem because the glacial boundary is poorly defined in places.

In this paper we do not attempt to define the Driftless Area, nor are we arguing for minor revisions in its boundaries. We discuss the boundaries here only to provide further insight on the problem of the glaciation of the Driftless Area as a whole.

It appears that south of Trempealeau County the west boundary of the Driftless Area is the Mississippi River (Trowbridge, 1966) except for a small area near Bridgeport (fig. 1, location 7) at the mouth of the Wisconsin River. Here till is evidently present on a high terrace (Knox, Attig, and Johnson, this volume). The distribution of the till and associated deposits in the Wisconsin River valley suggest that ice advanced toward the east but that only a small tongue of ice entered the Wisconsin River Valley a distance of about 3 to 4 km. To our knowledge there is no evidence that this ice advance came onto the uplands of southwest Wisconsin.

The southern boundary of the Driftless Area in Illinois is not discussed here. Black (Frye, Willman, and Black, 1965) reports gravel with erratics under loess just north of the state line at Hazel Green (fig. 1, location 6). The Hazel Green gravels appear to be correlative with the high-level outwash gravels described in Illinois by Willman and Frye (1969). The southeast border of the Driftless Area was mapped in detail by Alden (1918). Till in this area was subsequently examined by Bleuer (1970). Illinoian and supposed Kansan till is present to the boundary and the boundary is fairly abrupt. No erratics are reported outside the glaciated area shown in figure 1. To the north, in Dane and Sauk Counties, late Wisconsin till, sand and gravel, and lake sediment form an abrupt boundary marked by an end moraine in most places (Alden, 1918; Mickelson and McCartney, 1979). Along a small portion of the boundary in northern Dane County the moraine is absent, and late Wisconsinan age till thins to a scatter of erratics. Even in this area, however, the boundary cannot be more than 1 or 2 km from the position mapped by Mickelson and McCartney (1979).

In northern Sauk County and northward (fig. 1) sand and gravel and fine-grained lake sediment lie adjacent to Driftless Area uplands. The southwesternmost extent of till beneath lake sediment has not been mapped. Black (1974) reports drilling into till at two nearby locations (fig. 1, location 10). Ice-rafted material is present throughout the area covered by these lakes (Martin, 1932). The deposits discussed by Akers lie near the western edge of deposits of glacial Lake Wisconsin (fig. 1, location 5). On the basis of the till under glacial Lake Wisconsin, Knox, McDowell, and Johnson (1981) mapped the northeastern boundary of the Driftless Area at the base of the escarpment along the western edge of the former lake.

The glacial deposits of north-central Wisconsin were mapped by Weidman (1907), Hole (1943), and more recently by Stewart (1973) and Mode (1976). The area adjoining the Wisconsin River north of Stevens Point, included as part of the Driftless Area by Weidman (1907) has clearly been glaciated (Thwaites, 1956; Stewart, 1973).

The pre-Wisconsinan Wausau till and the early Wisconsinan Merrill till in the northeast portion of figure 1 are distinguished by texture and clay mineralogy (Stewart and Mickelson, 1975). Dates of greater than 36,000 B.P. (IGS-262) and 40,800 $\pm$ 2000 B.P. (IGS-256) provide a minimum date for deglaciation of the area covered by Merrill till. The Wausau till is considerably more weathered than the Merrill, suggesting that it is pre-Wisconsinan age. Both of these units were traced westward to Clark County by Mode (1976), and there is no indication of glaciation after these units were deposited. If the Driftless Area was glaciated during very late early Wisconsin time ("Rockian"), as proposed by Black, it seems unlikely that the ice source could have been to the northeast.

The margin of the Driftless Area in the north is poorly defined. Robert Baker, (personal communication, 1981) is mapping glacial deposits to the north and west but has not attempted to redefine the southernmost extent of till. There is no evidence, however, that Wisconsinan age ice came from the northwest into the Driftless Area. Surficial units directly north of the Driftless Area are almost certainly pre-Wisconsin. Chamberlin and Salisbury (1885), Leverett (1899, 1929), Weidman (1907, 1913), and Thwaites (1955) all show slightly different northern boundaries of the Driftless Area, and we have no way of choosing between them.

The "clean-ice" concept

Black suggests that few erratics or other signs of glaciation appear in the Driftless Area because "clean, slow-moving ice did little work and left little obvious evidence of its former presence" (Black, 1964, p. 176). It is hard to imagine how the Laurentide Ice Sheet, which obviously carried abundant Precambrian erratics into other parts of the state, could have carried none into the Driftless Area if, in fact, the ice sheet covered southwest Wisconsin. We know of no other large area being covered by an ice sheet that left no trace. If Black and others are correct, then a local ice cap seems to be required. Most models of the Laurentide Ice sheet (for example, Prest, 1970; Sugden, 1977, Hooke, 1977; Denton and Hughes, 1981) assume that the terminal zone was a wastage area and that accumulation occurred far back from the ice margin, with no local ice caps in the Midwest. However, Hooke and others (1976) have suggested that significant near-margin accumulation occurred on the late Wisconsin Des Moines Lobe, and perhaps accumulation during earlier glaciations was great enough to sustain an ice cap in the Driftless Area.

In order to cover the areas discussed by Black and to "plane off" the top of East Blue Mound, it would have to be a large ice cap, covering all of the Driftless Area and areas beyond. It seems unreasonable to hypothesize such an ice cap without some substantial evidence such as till, morainic topography, boulder trains, striations, ice-marginal channels, or proglacial lakes.

CONCLUSIONS

To our knowledge, we have examined all of the documented "evidence" of glaciation in the Driftless Area. We see nothing in this so-called evidence that convinces us that the Driftless Area was glaciated. In fact, the lack of evidence for glaciation is very impressive! Lack of evidence does not prove that it was not glaciated but we cannot accept the concept of Black that it was glaciated during Wisconsin time.

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